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Cretaceous Pacific Superplume Activity:

**Aspects from Geology, Petrology and Geochemistry of the Ontong Java Plateau Basalts
from Malaita and Santa Isabel, Solomon Islands**

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Doctor Thesis

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Abstract

Mid-Cretaceous (from 125 Ma to 80 Ma) is the anomalous period that global environment has drastically changed. The cue to understand the most recent "pulse" era in the earth history is veiled in Western Pacific oceanic plateaus in the period. In this study, the Ontong Java Plateau (OJP) was target on because the OJP is the most voluminous oceanic plateau and has the area exposed on land in Solomon Islands.

In the Chapter I, on the basis of geological evidences of Solomon Islands, the emplacement of the Ontong Java Plateau were discussed and subducted OJP were reconstructed. The Malaita and Santa Isabel of the Solomon islands, represent the fragment of southern margin of the OJP. The basement of these two islands formed as a huge oceanic plateau during the first Cretaceous pulse event at the position of the Southern Pacific Superplume at 122 ± 3 Ma. The OJP has then migrated horizontally passing through two hotspots Samoan and Rarotonga, and finally to the north Solomon trench, off Malaita and Santa Isabel islands at 20-25 Ma, and subducted to form a series of accretionary complex on the leading edge of Australian plate. By 4 Ma, the accretionary package has grown enough to appear above sea-level to transport the eroded materials of OJP basalts to the north Solomon trench.

Based on the paleogeographic reconstruction, the subducted portion of OJP can be estimated to be ca. 1000km long at least, presumably 2000-3000km by the calculation of relative plate motion 12.6 cm/yr since the first contact at trench. The huge buoyant oceanic plateau must have easily subducted into deep mantle remaining a minor piece of fragments as an accretionary complex on the surface.

In the Chapter II, to clarify the petrogenesis of the Ontong Java plateau (OJP), we have carried out the petrological and geochemical study on the systematically enormous collected samples in central Malaita and Santa Isabel. We determined the major, trace and rare earth elements (REE) compositions of whole-rock and major element compositions of igneous minerals from lava, and major compositions of daughter minerals in olivine-hosted melt inclusions from olivine gabbro. On the basis of major element compositions, the OJP lavas from central Malaita are divided into three types; High-TiO₂ type, Low-TiO₂ type, and High-MgO type. High-TiO₂ type lavas are higher

in TiO_2 , Fe_2O_3^* , P_2O_5 , and REE and lower in Al_2O_3 and CaO contents at given MgO than Low- TiO_2 type. Both have flat REE patterns, although high- TiO_2 type lavas have slightly LREE enriched patterns. High- TiO_2 type and Low- TiO_2 type lavas correspond in major and REE compositions to Singgalo- and Kwaimbaita-type basalts distinguished by previous study, respectively. High-MgO type lavas have high MgO and Fe_2O_3^* , slightly high TiO_2 , P_2O_5 and REE and low SiO_2 , Al_2O_3 and compatible trace element contents. They have similar REE patterns to high- TiO_2 type. The feature is compared with those of no other OJP basalts. The stratigraphic relation of these type lavas are ordered as High-MgO, High- TiO_2 , and High- TiO_2 type lavas beginning at the top although the all the sequence is not conformably exposed because they are disturbed and repeated by the faults. Daughter minerals in olivine-hosted melt inclusions have anomalous features compared with host olivine gabbro. Amphibole-bearing melt inclusions are dominant unless it is absent in host olivine gabbro. Daughter amphiboles and clinopyroxenes are higher in Fe, Al and Ti. They are reverse zoned. A few globules have carbonate. Although the petrological features of melt inclusions are modified by complicated process of closed-system after trapped, the core compositions of mafic minerals suggest that they are generated from SiO_2 -deficient melt like High-MgO type lavas. High-MgO type lavas are comparable with no melting experiment of peridotite, eclogite and peridotite and basalt mixture. The high-pressure melting of garnet clinopyroxenites is roughly explainable for major, trace element and REE features of High-MgO type lavas. The most reasonable source of High-MgO type lava is refractory of eclogitic recycled oceanic crust extracted SiO_2 -rich melt. High-MgO type lavas infer that the plume generating OJP include the ancient oceanic crust. Other type lavas are expected to be mixture of totally molten basalt and a partial melt of peridotite.

In the Chapter III, to understand the pacific super plume history, Paleogeography of continents, oceans and oceanic large igneous provinces (LIPs) since 1000 Ma were reconstructed in order to decode history of the Pacific superplume activities, on the basis of the most confident recent data for accretional geology from orogens associated with ophiolite and greenstone belts, and geochronologic and paleomagnetic data from each continent in literature. Several conclusions were led through our reconstruction; (1) Pacific superplume was born at 750 Ma to break-up supercontinent Rodinia. (2) African superplume was born at 250-200 Ma to beak-up Gondwanaland. (3) Episodic activity of

Pacific superplume was present with culminations at 400-500 Ma, 300-250 Ma and 150-75 Ma. (4) Presence of collage of immature oceanic island arcs at 900-600Ma in Sahara-Saudi Arabia, at 600-500 Ma in Kazakhstan Block, at 200-100 Ma in Alaska-Kamchatska region, and since Miocene in the western Pacific. These brought juvenile arc crust through convergent plate boundary process of microplates. (5) Role of Tethyan-Indian Ocean in Wilson cycle. Break-up, dispersion and amalgamation of continents was not simple reversal as suggested by Hoffman (1991). Instead, the Tethyan-Indian Ocean played an important role, which is to form double-sided convergent boundary around continental lithosphere or microplate. Opening and closing of the Iapetus Ocean were contemporaneous to break-up of Rodinia and to formation of Gondwana, respectively. At c.1000Ma, the Grenville orogeny, which formed the Rodinia, prevailed to form the Pacific-type orogens along active region of the Pacific superplume. These suggest that the numerous volumes of subducted lithospheres along the Pacific-type orogens triggered the birth of the Pacific superplume

Chapter I. Geology of Solomon Islands

Abstract

The Malaita and Santa Isabel of the Solomon islands, represent the fragment of southern margin of the Ontong Java plateau (OJP). The basement of these two islands formed as a huge oceanic plateau during the first Cretaceous pulse event at the position of the Southern Pacific Superplume at 122 ± 3 Ma. The OJP has then migrated horizontally passing through two hotspots Samoan and Rarotonga, and finally to the north Solomon trench, off Malaita and Santa Isabel islands at 20-25 Ma, and subducted to form a series of accretionary complex on the leading edge of Australian plate. By 4 Ma, the accretionary package has grown enough to appear above sea-level to transport the eroded materials of OJP basalts to the north Solomon trench.

Based on the paleogeographic reconstruction, the subducted portion of OJP can be estimated to be ca. 1000km long at least, presumably 2000-3000km by the calculation of relative plate motion 12.6 cm/yr since the first contact at trench. The huge buoyant oceanic plateau must have easily subducted into deep mantle remaining a minor piece of fragments as an accretionary complex on the surface.

1-1. Introduction

To understand the dynamic behavior of superplume, the Cretaceous hotspot volcanic rocks and the related geologic events such as formation of oil, gas and coals, global warming, frequency change of magnetic reversals, and X_{CO_2} flux in atmosphere (Larson, 1991a), are the best example on the present-day Earth, because the oceanic lithosphere subducts within 200 m.y. after formation at mid-oceanic ridge. The Cretaceous pulse time defined from 125-81 Ma, has been an unusual period, when extensive hotspot volcanism occurred as well as 150% higher production rate of MORB than the normal period, but it happened only in the Pacific domain (Larson, 1991b).

The Pacific superplume has been extensively active in this period, and formed a series of microplates such as Magellan, huge oceanic plateaus like Ontong Java plateau (OJP), Caribbean plateau, and seamount chain, e.g., mid-Pacific seamount chain. The reconstructed size of huge plateau corresponds to the continent as large as South America, and presumably major part of this composite plateau and seamounts, must be partly above sea-level.

Origin and source material of those LIPs on the Pacific ocean-floor are the key to solve the global material circulation in space and time through the Earth history. By this reason, lots of trials of drilling on the LIPs on the ocean floor have been performed (e.g., DSDP289; ODP803, 807, 1183-1187). Particularly the Ontong-Java plateau has attracted many geologists and geochemists, because of the largest size of oceanic plateau on the Earth (Fig. 1-1).

1-2. Tectonic setting

The Solomon Islands form a linear double chain of islands, oriented NW-SE between latitudes 5-10 South, and longitudes 156-163 East in the southwest Pacific (Fig. 1-2A). Solomon Islands are a main component of the Greater Melanesian Arc system which marks the collision zone between the Australian and the Pacific plates which includes from NW to SE, the islands of New Britain and Bougainville in Papua New Guinea, Solomon Islands and Vanuatu (New Hebrides). The present-day relative plate motion between the Pacific and Australian is to WNW direction with a speed of 12.8

cm/yr (Fig. 1-2B).

The Solomon Islands are bounded on both sides by trench, one located to the north (North Solomon trench) and another to the south (South Solomon trench). Active seismicity is present underneath the narrow oceanic lithosphere from Bougainville, Choiseul, Isabel, Malaita, and San Cristobel to Guadalcanal islands over ca. 150 × 1000 km (Fig. 1-2B). These earthquakes occur along the double-sided subduction zones predominantly along the south Solomon trench, rather than the north Solomon trench (Fig. 1-3).

The active volcanoes form a chain along the south Solomon trench, however, the volcanic front is very close to trench axis, less than 80km, compared to the normal volcanic front such as Japan ca. 200km inward from trench. 110 km above the Benioff-Wadati plane, the frontal arc volcano stands in NE Japan, but the volcanic arc in New Georgia stands 30-40 km above the Benioff-Wadati plane, clearly demonstrating the source magma of Calc-alkaline andesite and strongly alkaline magma must have come by slab melting of MORB crust and from sub-lithospheric deep mantle, respectively (Fig. 1-3).

Underneath the New Georgia Group, the spreading ridge axis may be present (Fig. 1-2B). The Woodlark basin was opened at 5 Ma, presumably related to the southward subduction of the Pacific plate, by which a back-arc basin was formed like the similar way as opening of Japan Sea. The propagating ridge axis is close to the southern tip of the Papuan island (Fig. 1-2B).

The geology of the Solomon Islands is not simple, as shown below. Both islands of Malaita and Isabel, and partly San Cristobel, are composed of pieces of the OJP as an accretionary complex, but the rest of islands such as Choiseul, Guadalcanal and New Georgia volcanic islands do not contain basalts with chemically similar to LIPs. Instead, the MORB-like altered basalts with Cretaceous age appear at the structural bottoms in some islands. The relationship between the basement MORB and the OJP basalts is seen in the Isabel Island where the sliced OJP basalts underthrust the Cretaceous oceanic crust and mantle as shown later.

1-3. Geology of Solomon Islands

The islands of Malaita and Santa Isabel are major two targets for our study

where are present the extensive exposure of OJP rocks (e.g. Petterson et al., 1997; Tejada et al., 1996).

Off the north Solomon trench, the great bulk of OJP is located, much of it standing at a general elevation of ca. 2000m above surrounding the ocean floor (e.g. Kroenke, 1972). Ar-Ar dating of drilled basalts by ODP and DSDP boreholes into the OJP, and of basalts from Santa Isabel and Malaita suggest a bimodal distribution of ages around 122 ± 3 Ma and a second population of clustering age around 90 ± 4 Ma (Mahoney et al., 1993a,b; Tejada et al., 1996).

Since the pioneering works of arc-trench seismic profile off island of Choiseul by Bruns et al. (1989), presence of subduction-accretion structure of the OJP rocks to the hanging wall of oceanic microplate including Santa Isabel and Malaita has been widely recognized. Recently more extensive seismic reflection works have been completed off the OJP (Mann and Taira, 2004). The large-scale velocity structure was obtained along the NW-SW traverse line from OJP, through the north Solomon trench, Malaita and south Solomon trench, to the Woodlark basin (Fig. 1-4). Thickness of the OJP crust reaches to ca. 35 km, and the pelagic to terrigenous sedimentary cover sequence over a few km rests on the OJP outwards the trench axis. Under the compression regime due to the double subduction on both sides, a series of low-angle thrust develops in both fore-arc on the north and the arc-trench gap on the south (Fig. 1-4A). The tight folding with seaward vergency well-develops on both sides of accretionary complex.

As the Solomon islands are the collage of crustal terranes, Coleman (1965, 1966, 1970) and Hackman (1973) divided this region into the following three geologic provinces; (1) The Pacific Province, (2) The Central Province, and (3) The Volcanic Province.

(1) The Pacific Province

This province appears to the northeast on the Pacific side, seems to be uplifted and largely unmetamorphosed portion of the OJP, and forms the basement of Malaita, Ramos, Ulawa, and east of KKFS fault (Kaipito-Korighole Fault) of Santa Isabel islands (Hawkins and Barron, 1991; Tejada et al., 1996).

(2) The Central Province

This is adjacent to the Pacific Province in the SW and contains variably metamorphosed Cretaceous and early Tertiary seafloors and remnants of NE-facing arc

sequence that grew during the early to middle Tertiary above the Pacific plate, before the arrival time of OJP.

The widespread occurrence of altered N-MORB consists largely of the basement. However, the origin of those rocks is not yet clearly understood. These may have been formed during the opening time of a series of basins such as Coral Sea, Tasman Sea, and other ocean basins in the late Cretaceous to early Tertiary. Alternative interpretation, following its N-type MORB signature, is that it may have been a trapped Pacific ocean floor sometime before the Eocene CA volcanism to penetrate the crust.

(3) The Volcanic Province

This Province extends along the SW flank of the Central Province, and is an island arc sequence composed of volcanic and intrusive rocks with active volcanoes. The age of this province appears to be younger than 4 Ma (Petterson, 1995). This zone may overlap the Central Province, because CA volcanism overlapped the Central Province. The volcanism is due to the northward subduction of old Solomon Sea Plate on the west, young (<5Ma) oceanic plate in the Central, and the old Louisiade oceanic plateau on the east (Fig.1-2B). By this reason, the distance between volcanic front and trench is variable, 110km on the west, 0-70km in the central, and 110km to the east.

1-3-1. Malaita

The Malaita island, 150km x 35km in size, is an unique rugged rocky island covered by the thick rainforest in an equatorial region. The NE-SW trending elongated shape is slightly oblique against the general trend of oceanic microplate of Bougainville-Guadarucanal (Fig.1-2). Malaita belongs to the Pacific Province together with major part of Santa Isabel Island.

The first reconnaissance geological map of Malaita (Rickwood, 1957) only covered the northern portion of the island. Allum (1967) first published the map of the entire island. Since then, Hughers and Turner (1976, 1977), Hine (1991) and Barron (1993) investigated in detail in several small regions carefully. The sedimentological investigations from the Cretaceous to Tertiary were also performed (McTavish, 1966; Coleman, 1966; van Deventer and Postuma, 1973).

The petrographic and chemical studies have been made on Oligocene alnöites by numbers of petrologists and geochemists by their unique characters of oceanic

kimberlite (Rickwood, 1957; Nixon and Coleman, 1978; Neal, 1985; 1988; Nixon and Neal, 1987; Neal and Davidson, 1989; Ishikawa et al., 2004).

It was Petterson et al. (1997) that have extensively studied the structure of Malaita particularly the northern half. Fig. 1-5, 1-6 and 1-7 show the geological map, profile and stratigraphy of Malaita, respectively (Petterson et al., 1997). They have established the basic stratigraphy of the Malaita volcanic-sedimentary sequences. The basement rock is the Cretaceous OJP forming at least 3-4 km thick volcanic rocks dominantly aphyric pillowed basalts. Those are usually unaltered and weakly deformed pillow lavas with minor sheet flows and feeder dikes. As minor, the fractionated gabbroic rocks are present. These rocks are named as the Malaita Volcanic Group (MVG) by them. The detailed descriptions will be given in Chapter II together with petrography, mineralogy and chemistry. The mode of field occurrences of the lavas, e.g. size, shape, flow unit, flow boundary, and other field evidences are shown in photo and plate 1-1 – 1-17.

On the top of the accreted OJP, the Kwaraae Mudstone Formation composed of bedded radiolarian chert with Lower Cretaceous ages is present as thick as 100-270 m (Petterson et al., 1997). This formation is composed of a parallel laminated, very fine-grained, radiolarian, foraminiferal siliceous mudstone. Mudstone becomes more siliceous in the upper parts. The Kwaraae Mudstone Formation is conformably overlain by the 300-1000 m thick upper Aptian/lower Albian to Eocene Alite Limestone Formation. This limestone is composed of an alternating sequence of hard, porcellaneous, very fine-grained calcilutites with regular inter-bedded cherts. Cherts are red, white, green and vary in thickness between 10 to 30 cm. The viscosity contrast between host limestone and chert layers produces a very characteristic folding style dominated by schvron to polyclinal folds, thrusts, boudinage, and pinch-and-swell structures well-documented in Japanese accretionary complex.

The Kwaraae and Alite formations formed as a result of relatively slow pelagic sedimentation at deep ocean environment far from the continental margin.

Depending on the place, highly vesiculated alkali basalts 44.2 ± 0.2 Ma (Tejada et al., 1996) called the Maramaskite Formation are present. This is highly brecciated and vesicular, amygdaroidal alkaline basalts reaching a total thickness of 100-500 m.

The 300-850 m thick, Upper Miocene to Lower mid-Pliocene Suafa

Formation conformably overlies the Haruta Formation. The Suafa Formation contains a spectrum of litho-facies with one end member being typical deep-water pelagic limestone, and the other end-member of current bedded, relatively poorly sorted, calcareous siltstones containing benthonic fauna. Some facies contain up to 10% allochthonous grains of plagioclase, pyroxene, hornblende, biotite and lithic fragments, suggesting the input of terrigenous volcanic arc materials, as close proximity of landmass (Fig.1-7).

The 34 Ma ultramafic alnöites breccias and fine-grained pipes and sills intrude the Malaitan cover sequences as high as Lower Haruta Formation (Pettersen et al., 1997). These contain complex mineralogy of olivine, clinopyroxene, melilite, phlogopite, ilmenite, spinel, nepheline, melanite, apatite as well as discrete megacrysts and lherzolitic autholiths (Allen and Deans, 1968; Nixon et al., 1980, Neal, 1985, Ishikawa et al., 2004). Megacrysts of garnetite, ilmenite, clinopyroxenite may reach to the size up to 50 cm across.

The major unconformity develops locally on the Suafa Formation which consists of the Upper Pliocene to Pleistocene limestone and conglomerate.

Tectonic divisions of Malaita

The fundamental structural characteristics of Malaita are the well-developed NW-SE structural elements as shown well by the folding axis orientation. Four major periclinal anticlines are documented by Pettersen et al (1997) who named the Tombaita, Fateleka, Kwaraae, and Kwaio anticlines. The folding axis is parallel to the trench axis to the north. The extensive folding seems to have developed sometime during the early Miocene by Pettersen et al. (1997). Most of them display vergence of folding to NE suggesting the subduction of Pacific plate. As shown in the eight EW cross sections of Malaita Island (Fig.1-6), the NE-directed vergence of folding and associated low-angle thrusts are well-developed, indicating a typical structure of subduction-accretionary complex. Locally as shown in the section 2, back-thrust with opposite sense of vergence is seen. A series of west-dipping low-angle thrust suggests that this island is composed of typical accretionary complex, and that if correct, the structural top as seen on the west must be the oldest. Timing of formation of the oldest accretionary complex, as suggested by the systematically different oceanic plate stratigraphy from top to the bottom structural levels, would be sometime in middle Miocene, ca. 20-25 Ma, as

described by Petterson et al. (1997). Since then, the trench turbidite start to mix mineral and volcanic fragments from the south.

The 4 km deep sub-horizontal detachment master fault underneath the central range (<1500m high) is estimated by Peterson et al. (1997) and a series of branched thrusts are delineated in their NE-SW profiles (Fig. 1-6).

Fault systems

Two types of fault, normal and thrust faults are present (Fig. 1-5). Although the investigated areas are extensively weathered by equatorial lateritization, Petterson et al. (1997) investigated the Malaita in great detail, by which the distribution of those two types of faults, became clear in conjunction with the folding system. We Titech Group have spent three seasons in this region, however, we have not yet completed our own geologic maps. During our mapping, we have confirmed their faults systems.

The low-angle thrusts 10-40 degrees NW-dipping act as boundary faults of horsts, and are cut by the high-angle normal faults at later stages, presumably twice or more. The major secondary extension tectonics to collapse the thickly piled up accretionary complex, by four NW-SE trending major high-angle normal fault zones (Fig.1-5) occurred sometime in Pleistocene, presumably by the stop of Pacific subduction. Most of active fault systems identified by the recent IODP cruise are the products of Quaternary incipient subduction of Pacific plate again (Mann and Taira, 2004).

We titech group also recognized some shear zone along Singgalo River (Fig. 1-8).

Accretionary complex

By the lack of knowledge of accretionary complex, Petterson et al. (1997), in spite of their detailed works, not only the definition of each accretionary horse, but also the oceanic plate stratigraphy (OPS) has not been clarified. Nevertheless, the rough estimate of megascopic structure of the island may be possible, on their maps and cross sections (Fig. 1-6).

Considering their description of sedimentary rocks and their stratigraphy summary (Fig.1-7), the OPS for the westernmost portion of accretionary complex, i.e., structural top can be estimated as in Fig. 1-7b. Although need the critical fossil evidence

and confirmation in the field, the formation of accretionary complex began on the western margin at 20-25 Ma on the basis of the age of calc-alkaline volcanic activity on Solomon Islands (Petterson et al., 1997). The recent subduction of the Pacific plate, and related major re-organization of the fault system in the trench-fore-arc region, have not yet become steady-state (Fig.1-2).

The OPS-forming rock types are, predominantly pillow lava, with subordinate amounts of sheet flow, dolerite and minor gabbro, pelagic limestone, and chert (red, green, white). Rarely, hyaloclastite, reworked hyaloclastite, volcanic breccia, particularly in the case of Eocene alkali basalts are present. The total thickness reaches maximum to 8-9 km for the OJP. The early Cretaceous cessation of OJP flood basalt volcanism promoted the pelagic sedimentation ranging in thickness 100-270m, followed continuously by the Alite pelagic limestone interbedded with chert, ranging 300-1000m in thickness. These thick pelagic sediments were covered by 44.2 Ma highly vesicular alkali basalts less than 500m thick. Foraminifella-dominated pelagic limestone interbedded with chert deposited on the alkaline volcanics until sometime in mid-Miocene.

Fore-arc basin deposits

As well-documented by the several seismic reflection profiles on the Solomon Islands, most of accretionary complex are discontinuously overlain by the in-situ deposits particularly in the well-developed fore-arc basins. A typical fore-arc basin is observed on the south of the island named as the Central Solomon intra-arc basin (Fig.1-4). To the south of this extensional basin, the basement high is not yet named, but the normal faulting is extensively developed. Further to the south, the stress regime changes suddenly to strongly compressional to form a typical accretionary complex by the young hence buoyant oceanic plate.

On the contrary, the northern flank of island of Malaita does not show a simple geometry of fore-arc basin, instead, more complex topography appears. However, the arc-trench gap may be subdivided into two regions, extensional inner and compressional outer zones (Fig.1-4).

The age of the recent tectonic reorganization can be estimated by the bottom of the fore-arc basin deposit on the south, and the bottom sediments which cover the low-angle thrusts. According to Mann and Taira (2004), it began sometime in

Pleistocene.

1-3-2. Santa Isabel

A small elongated island, 220km × 20km, next to the west of Malaita, is called the Santa Isabel. Rock types in this island are similar to those on the Malaita except the basement of the Cretaceous MORB on the Santa Isabel. Along the boundary between the southern basement rocks and the northern major part of fragments of OJP, blueschist and serpentinite are exposed locally. The boundary fault zone is called as the KKFS (Fig.1-9). The Tokyo Tech group carried out the survey in area A-C (Fig. 1-10 – 1-12).

KKFS is a boundary fault dividing the Pacific Province from the Central Province where the early Cretaceous MORB-like volcano-plutonic sequence (ophiolite) is extensively exposed. The KKFS dips ca.60 degrees southward on which the early Cretaceous oceanic basement rocks are present, suggesting the KKFS was once the plate boundary under which the OJP has subducted. The presence of strongly serpentinitized peridotite, and the sodic-amphibole bearing schistose rocks also support the KKFS as a consuming plate boundary.

The origin of peridotite, gabbro and Cretaceous MORB-like pillowed basalts, overlapping in the ascending order to the south, is debated, whether or not some of these are the Eocene subduction zone magmatic products, whole or partly.

The northern part is composed of a series of accretionary complex same as the Malaita. The predominant rock type is the aphyric basalt called as Signa Basalt (Geol. Surv. Solomon Is.). A thick pile of feebly altered pillowed basalts, massive sheet dolerite flows are repeatedly overlain, at least a few km thick, although the exposure is poor. The general strike is NW-SE, and dip gently to southwestwards. Inter-layered pelagic chert and limestone appear more predominantly northwestwards and northeastwards in the south. The possible occurrence of west-dipping thrust is suggested, but not yet fully confirmed.

Arc volcanic rocks are present particularly to the south of KKFS. Those are believed to have been formed in the Eocene by the Pacific subduction from the north.

Significance of blueschist

In spite of minor occurrence, the rock blueschist is critically important, because this metamorphic rock marks the ancient consuming plate boundary. By this reason, geochronology and petrography is the key to decipher the tectonic history of subduction, although the data are not yet available. The blueschist have been found by 2000 expedition, Drs. Kaneko, Ohta, Parkinson, and Mr. Suzuki (Fig. 1-12). If considering this belt to the eastern possible extension, New Caledonia and the western extension of Papua New Guinea, this belt may be formed in the Cretaceous or Eocene.

1-3-3. Choiseul and San Cristobel

The KKFS extends westward to over the Santa Isabel Island, to off the northern coast of island of Choiseul (Fig.1-2). The Choiseul belongs entirely to the Central Province. The structurally lowest of the Choiseul is the Barrovian-type regional metamorphic belt widely exposed on the eastern end of the island, which is a western extension of blueschist belt in island of Santa Isabel.

The basement rock is underlain by the Cretaceous altered MORB, penetrated later by subduction zone magmas since Eocene time. The Quaternary calc-alkaline volcanic piles remain the shape of volcanoes, presumably by the northward subduction of the Woodlark basin.

Along the southern margin of the San Cristobel Island, the basement rocks of the Central Province are exposed (Geol. Surv. Solomon Is.). Although the mapping program is still on-going, if this is correct, the KKFS may pass the southern part of San Cristobel Island. Major part of San Cristobel, is hence belongs to the Pacific Province. This means the OJP is exposed extensively in San Cristobel Island, too. To estimate the size, particularly the lateral extension of the OJP, the geology of San Cristobel Island would be a key.

1-4. Tectonic history of the Solomon Islands

Using the constraints described above, the following tectonic history of the Solomon Islands are estimated.

(1) Early Cretaceous time

The basement of Solomon Islands is back to the early Cretaceous, when N-

MORB was formed, although the geochemical signature has not yet fully understood. If it is correct, the original tectonic setting must be in the Pacific domain, and later migrated and accreted to the continental margin of Australia by the Eocene time.

(2) Eocene time

The trapped Pacific oceanic crust was subducted by the Pacific from the north, to form the extensive calc-alkaline volcano-plutonism. Moreover, along the consuming plate boundary, deep-seated accretionary complex was subjected blueschist facies metamorphism and returned back to the surface. The trench line was the present KKFS, dipping southwestward, migrated northward with time.

The Coral Sea was opened sometime in the Eocene, by a series of hydrous mantle plumes originated from the underlying wet mantle transition zone in the western Pacific.

(3) 25-20 Ma to 8.5Ma

The OJP migrated to the North Solomon trench, and initiated subduction at sometime 25-20Ma, the collision of buoyant OJP presumably modified extensively the shape of subduction zones in this region. The earlier subduction made the westernmost end of accretionary complex on the Malaita Island, and successively continued to pile up the thick accretionary complex until 8.5 Ma.

(4) Opening of Woodlark basin at 5Ma

The Woodlark basin opened at 5Ma, probably by the rising wet mantle plume underneath. The ridge has propagated westwards with time to break-up the Paleozoic orogenic belt. The young oceanic lithosphere started subduction northeastwards to form the subduction zone volcanoes since 4Ma from Bougainville, Choiseul, New Georgia islands, to Guadalucanal islands, as well as formation of accretionary complex in front.

(5) Growth of accretionary complex and emergence above sea-level at 2Ma

The continuous accretion of OJP made the accretionary complex grow effectively, and it appeared above sea-level sometime around 2Ma. However, the Pacific subduction stopped since then, until very recently.

(6) 1Ma-present

The OJP initiated subduction again reorganizing the geometry of subduction zones to develop numbers of high-angle normal faults, flip of subduction active faults ocean wards in general, but not always. By this reorganization, the North Solomon trench is not yet a single trench line, instead forming an array of overlapping active fault

lines (Fig. 1-2).

1-5. Reconstruction of the subducted OJP

The present-day distribution of OJP (Fig.1-2) suggest more than half of original huge plateau must have subducted into deep mantle from the north Solomon trench. To understand the original volume of plateau, an application of principle of accretionary geology, would be critical.

According to the oldest age of terrigenous materials, the arrival time of OJP at Malaita was 20-25 Ma (Pettersen et al., 1997). If so, the already subducted SW margin of OJP must have reached to the margin of North Solomon trench at 20-25Ma. Using the relative plate motion back to the arrival time, we can reconstruct the site of northern Solomon trench, and the OJP at 20-25Ma. The relative plate motion between Australia against the Pacific plate was 12.8cm/yr during the past 36Ma (Engebretson et al., 1985).

Although the opening of Woodlark basin since 5Ma changed the relative plate motion between Malaita-Bougainville microplate and Australia, but this change do not alter the paleogeographic relation before 5Ma.

Using the value of relative plate motion and the direction between both plates, the original sites of OJP and the north Solomon trench was reconstructed 20 Ma, following the paleogeographic maps after Mann and Taira (2004). The result (Fig.1-13) at 20 Ma shows the western extension of OJP was ca.2500km and at least the western margin of OJP must have reached to the position of Malaita Island. It indicates that the extra-portion of OJP must have been much larger ca. 2500km, as large as the present OJP size (Fig. 1-13).

The latter conclusion may be more plausible, because the shape of the present OJP as shown by contours of -4000m and -2000 m suggests the central thickest portion is cut by the northern Solomon trench (Fig. 1-2). This implies the half of the original OJP may have subducted in deep mantle. The OJP is in contact with the trench, over 1500km, from the north of Bismarck sea, through off Bougainville, Choiseul, Santa Isabel, Malaita, to San Cristobel islands. First tectonic contact to Malaita, although not yet confirmed, if so, the original shape of the OJP may be triangular.

The Ontong Java plateau, the largest oceanic plateau on the present-day ocean-floor reaches to the crustal thickness ca.35km. However, Gladzenko et al. (1997)

calculated ca. 26km thick, instead of an-usually thick 35km. The recent estimate by the velocity structure through IODP program support again anomalous 33km thick oceanic crust, assuming the lower crust with 7.0-7.6 km/sec P-wave velocity (Miura et al., 2004).

The size of oceanic plateau is of essential importance on origin of plume, because the size of plume to form the OJP magma must be bigger than 670km across sphere under the assumption of lherzolitic homogeneous mantle and 30% partial melting (Coffin and Eldholm., 1994, Fig.1-14). The recent observation that the OJP was formed by the twice of flood basalt volcanism at 120 and 90 Ma, suggests the presence of two independent plumes at different places under the Pacific ocean-floor. Therefore, the argument of the size larger than 670 km across may not be essential. The observation in the Malaita, however, is crucial to discuss the original volume of OJP and age of flood basalts. The thick basaltic piles with a close proximity to OJP is as old as 122Ma without 90 Ma fragments, although minor amounts of 44Ma alkaline basalts and 34 Ma input of alnöite magma were added. If this is correct, the ca. 90Ma flood basalt may have covered only the eastern portion of OJP and never covered the major and western portion of OJP. Moreover, the size of ca.120 Ma mantle plume, if assuming the homogeneous lherzolitic mantle, would be undoubtedly bigger than the radius of upper mantle. This is one of the critical constraints on the origin of OJP source mantle as discussed in Chapter II.

1-6. History of OJP from S. central Pacific to the N. Australia

In this section, we will discuss the history of OJP before subduction. The OJP was born in the southern Pacific domain ca.120Ma, migrated westwards, then arrived at the north Solomon trench in Miocene to accrete minor fragments at trench. Using the paleomagnetic constraints, the original site of OJP and its travel history are summarized in this section.

1-6-1. Birth at central south Pacific

The OJP was formed above a major Cretaceous plume head in the southern central Pacific, presumably on the S. Pacific superplume, between 35 and 50 south, with

major plateau-building magmatic events at 122Ma and 90 Ma (Mahoney et al., 1993; Tejada et al., 1996; Neal et al., 1997). The portion of the OJP now exposed on Malaita probably formed at the southwestern edge of plateau, thousands of kilometers away from the position of the present-day Solomon Islands.

The extensive volcanism at RRR junction at this period, has promoted the formation of oceanic microplate called the Magellan (Nakanishi et al., 1998).

1-6-2. Migration and overprinting by the Samoan and Raratongan hotspots

From the Aptian to the mid-Eocene the cover sequence of Malaita records the deep water pelagic sedimentation within an intra-oceanic environment. During the Eocene, the Pacific plate began to subduct beneath the Australian plate at the north Solomon trench which resulted the Eocene calc-alkaline volcano-plutonism along the NE-facing Solomon arc (Kroenke, 1984).

The Eocene was also a period of change within the cover sequence of Malaita. This change commenced with the eruption of highly vesicular alkaline Maramaskite Formation basalts (44Ma) which may have been erupted as the OJP passed over the Samoan, or possibly Raratongan hotspot (Tejada et al., 1996). On the OJP, several newly grown seamounts may have grown by this volcanic period. This event may have provided the extension deformation of the OJP to form the development of Faufaumela basin. The alnöite volcanism (34Ma) followed this event, which may have been erupted along the rifted OJP basin.

1-6-3. Arrival at trench and subduction into deep mantle

The first attach of the OJP to the North Solomon trench was sometime between 25 and 20Ma (Pettersen et al., 1997), and may have caused the stop of arc volcanism (Coleman and Kroenke, 1981; Yan and Kroenke, 1993). The Eocene Calc-alkaline magmatism in the Solomon arc stopped at 24Ma, as seen in the youngest Calc-alkaline rocks (24Ma) in the Guadarucanal (Chivas, 1981). In spite of arrival of OJP at trench, Peterson et al. (1997) considered that nothing happened on the hanging wall, therefore, they claimed this as a soft touch.

Cessation of Calc-alkaline volcanism began at 24Ma, and continued until

15Ma. Since 15Ma, the Calc-alkaline volcanism started again well-exemplified in Choiseul, until Pleistocene, suggesting active subduction again since then. Within the Solomon Island, there is no definitive evidence for subduction-related volcanism until the Upper Miocene. The oldest radiometric dates for this volcanic phase is a K-Ar date is 6.4 ± 1.9 Ma in the Guadalucanal island (Hackman, 1980). The Calc-alkaline volcanism may have begun from the east, and then gradually moved westwards from the Fiji islands to Guadalucanal (Pettersen et al., 1997).

Most workers consider that nothing happened during the first touch of the OJP

From 24Ma to 15Ma, by the buoyancy of OJP. Instead, anticlinorium within the OJP formed because of collision of the OJP and the westward movement of the Pacific plate (Kroenke, 1972). Yet, this anticlinorium was formed by normal faulting near the surface.

1-6-4. Accretion to form the Solomon Islands

Most workers consider the formation of accretionary complex began very recently from the north, called as an incipient subduction (Mann and Taira, 2004), and since 5-4Ma from the south as exemplified well along the seismic profile.

However, we would like to emphasize, based on geology on Malaita, that the accretion of the OJP began at sometime between 20-25Ma to form the horse on the structural top of the Malaita, then successively continued until the easternmost coastal area, by underthrusting of the OJP, until 8.5Ma at least.

At 5 Ma, the Woodlark basin began to open and continued up to now. At the same time, the young and hence buoyant ca. 8km thick oceanic lithosphere began to subduct northwards under the Solomon arc, by which the already accreted thickest portion of the OJP about 35km lifted up to form the Malaita anticlinorium.

1.7. Discussion & Conclusions

1-7-1. Tectonic history of the OJP

There is a consensus that the thick pile of basaltic lava flow is a fragment of

the OJP in terms of petrochemical character, ages, and geophysical continuation to the OJP off north Solomon trench. Yet, on the tectonic history of the OJP specifically the timing and the tectonic transport process, there are three different opinions, (1) Yan and Kroenke (1993), and Petterson et al. (1997, 1999) interpreted the arrival time of the OJP was 25-20Ma, but no accretion of the OJP until subduction polarity reversal occurred at 6-5Ma, when young Woodlark basin began to subduct to increase the compression on the Pacific side, to incorporate the thickest portion of the OJP as an accretional fragment. (2) Phinney et al. (2004), and Cowley et al. (2004) speculated the normal intra-oceanic subduction has long continued until 6 Ma at KKFS as a subduction zone. At 5 Ma, arrival of 33km thick OJP at trench marked as KKFS fault zone. Subduction-accretion occurred then; the top 20% of the OJP, remaining 80% is having been subducted since then until now. Subduction initiation occurs at San Cristobel trench and oblique-slip motion along the KKFS line. (3) We, based on the field geology on Malaita and Santa Isabel, the successive accretion of the OJP since sometime during 25-20 Ma, and the biggest accretion of the thickest portion of OJP crust occurred at around 5Ma (Fig.1-15).

1-7-2. Significance of easy subduction of the largest oceanic plateau

The tectonic mode of the Ontong-Java plateau, traditionally, affected to the model of dynamics on the Precambrian tectonics and continental growth of the Earth through time. For these important subjects, the OJP has a critical clue to judge the models.

First, the buoyancy of huge oceanic plateaus mainly due to the role of plagioclase, and concluded that the OJP cannot subduct into mantle, instead, a series of thrusting may develop at the boundary. Therefore, plate tectonics has not prevailed in the Achaean, until the mantle potential temperature cooled enough down to operate the modern-type plate tectonics. In this model, arrival of huge oceanic plateau has a strong effect to stop plate movement.

Second, the entire plateau can be easily trapped on the hanging wall on their margins, by its buoyancy (Nur and Ben-Avraham, 1982; Schubert and Sandwell, 1989; Abbott, 1996, Albarede, 1996). If so, the plateau would play a critical component of continental growth. Nevertheless, the Phanerozoic orogenic belts do not show such

component as major one. Presumably the answer is intermediate; some portion can be accreted, remaining major portion can be easily subducted into deeper mantle. The bottom lower than 30km would form garnet even before subduction. If it begins to subduct, the phase transformation to eclogite would yield much bigger slab-pull force than the normal mode of subduction.

The geology of Malaita strongly suggests that the huge oceanic plateau can easily subducts into mantle with time, although when arrives at trench, it stops subduction at once, and deforms the topology of subduction zones triggering the flip of subduction polarity like the case of OJP.

The Archean granite-greenstone terranes certainly contain numbers of greenstones with MORB and OIB origins (e.g. Condie, 1997). However, the size of each accreted slice of greenstone never support the idea of whole piece of accretion.

1-7-3. Estimate of lava volume of Cretaceous pulse

Based on the calculation of the volume of the OJP, as double of the previously thought value, $36-101.3 \times 10^6 \text{ km}^3$. The upwelling mantle flow under the Pacific superplume must contain huge amounts of basaltic component. In the case of OJP, the size of plume diameter must be larger than the thickness of upper mantle if the source mantle is homogeneous lherzolite (Fig. 1-14). Otherwise, the mantle must be enriched in recycled MORB crusts, as discussed in Chapter II.

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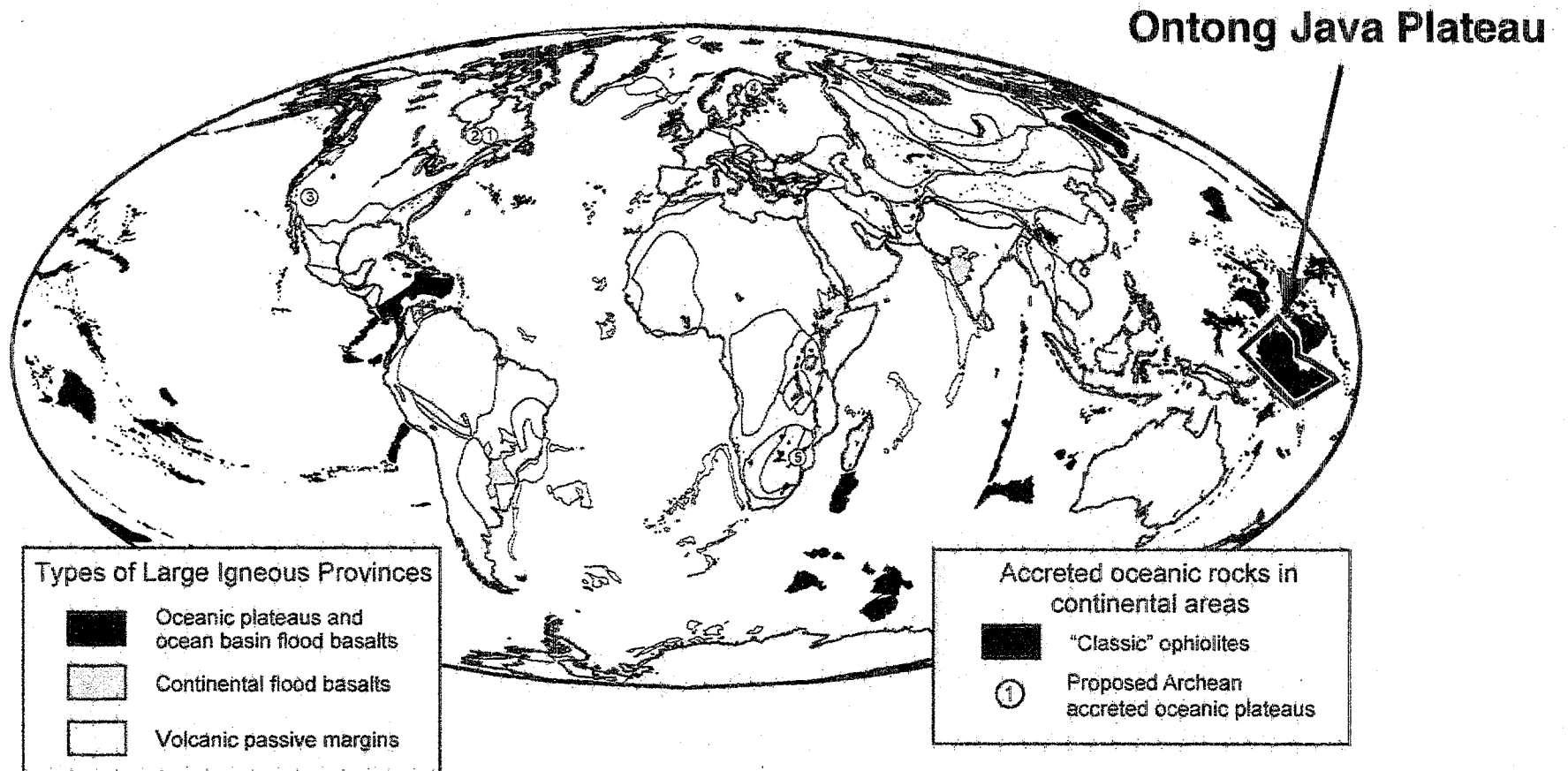


Fig.1-1. World distribution of LIPs after Mann and Taira (2004). Three different types of large igneous provinces, oceanic plateaus and ocean basin flood basalts, continental flood basalts, and volcanic passive margins. Moreover, on-land classic ophiolites are also shown. Some of the possible oceanic plateaus with Archean origin are also shown in the continents. The Ontong-Java plateau is the largest on the present-day ocean-floor. Also note the distribution of numbers of plateau basalts or seamount chains close to the Ontong Java plateau in the Western Pacific domain.

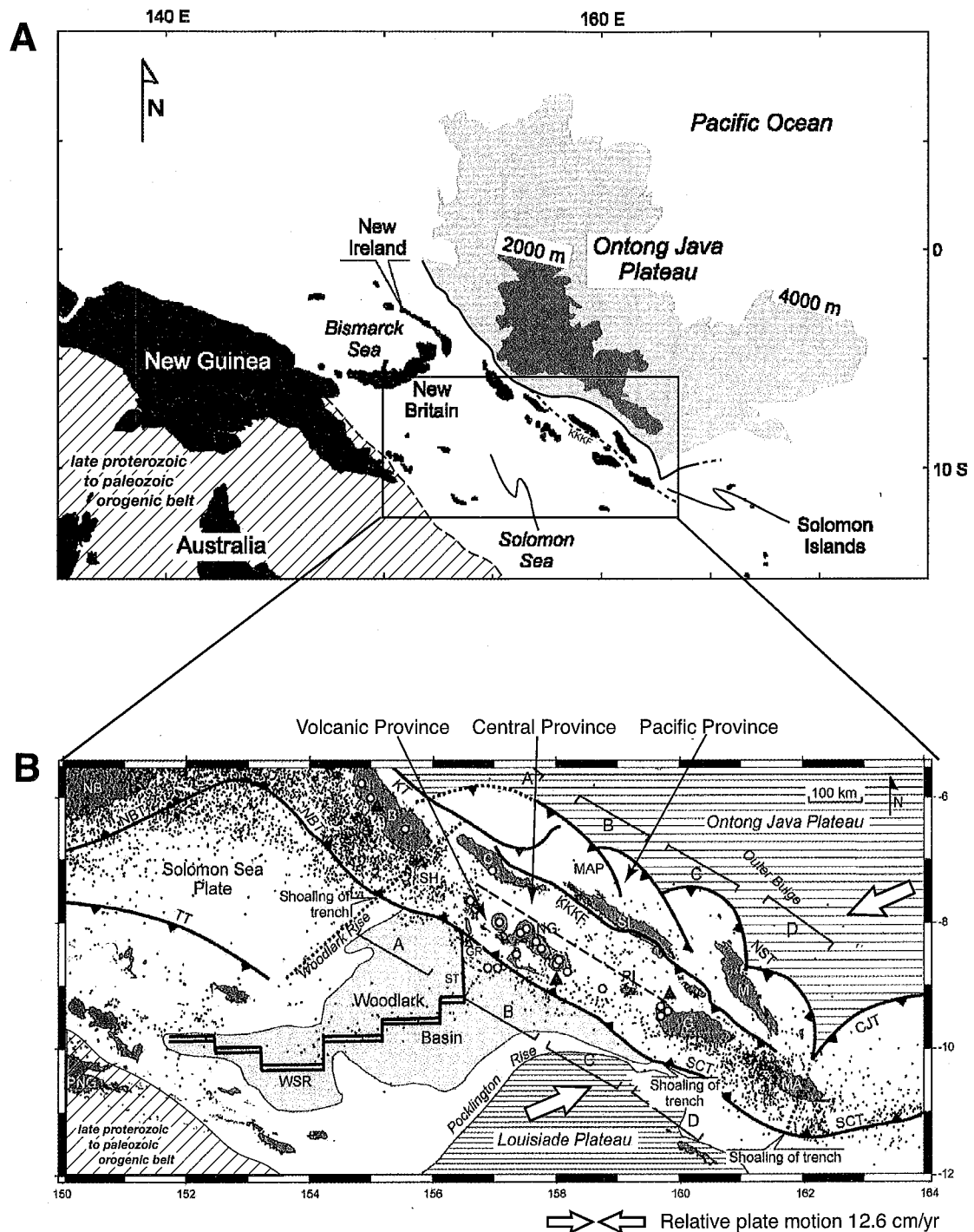


Fig.1-2. Above. The Solomon Islands and the Ontong-Java plateau. The Ontong Java plateau with 35km thick oceanic crust stands about 4-5km from the surrounding ocean-floor. The thickest portion occurs in SW part and cut by North Solomon trench, suggesting the half of the plateau has already subducted into deep mantle. Below. Tectonic outline of the Solomon Islands modified after Mann and Taira (2004). Hypocenters of earthquake are also shown, indicating subduction of Solomon Sea plate underneath the oceanic microplate (Solomon Island microplate, SIM hereafter) on which the Solomon Islands exist. The Indo-Australian plate is also subducting under the SIM. The ridge-transform system in the Woodlark basin formed at 5Ma, and is actively subducting under New Georgia Group. The Ontong Java plateau also subducts from north under the SIM. The north Solomon trench is not linearly arranged, a complex manner of incipient subduction. Active and Quaternary volcanoes are shown as red triangle and yellow circle. Note the position of cross sections in Fig.1-3.

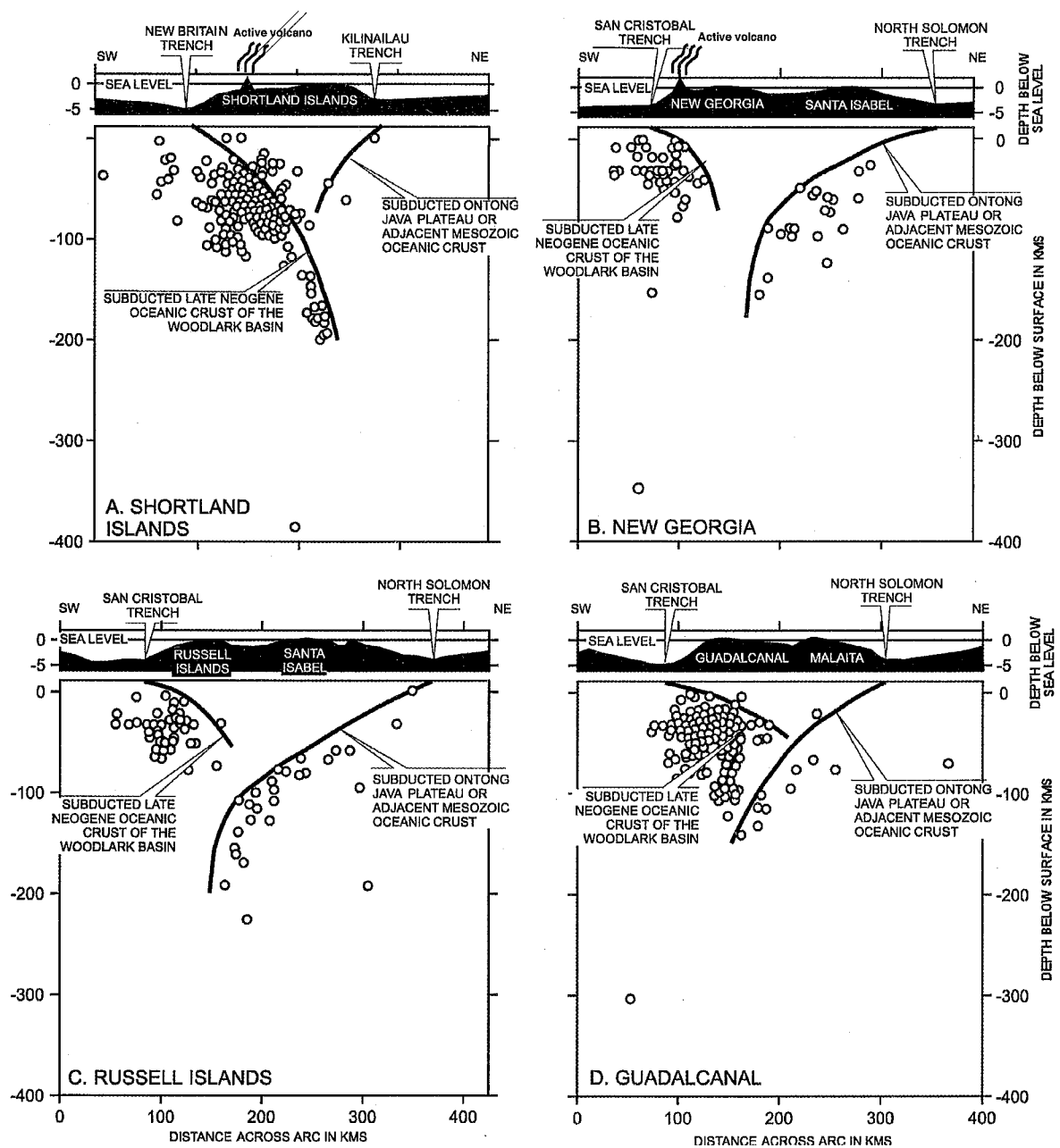


Fig.1-3. N-S cross sections of subduction zones, from A to D in the Solomon Islands slightly modified after Mann and Taira (2004). Note the near-trench volcanoes by subduction of young (<5Ma) oceanic lithosphere. The earthquakes in the young lithosphere occur shallower than 200 km. The Pacific subduction, in spite of old lithosphere, triggers, quite limited amounts of earthquakes, commonly shallower than 200km. rarely, deep earthquakes occur deeper than 300 km.

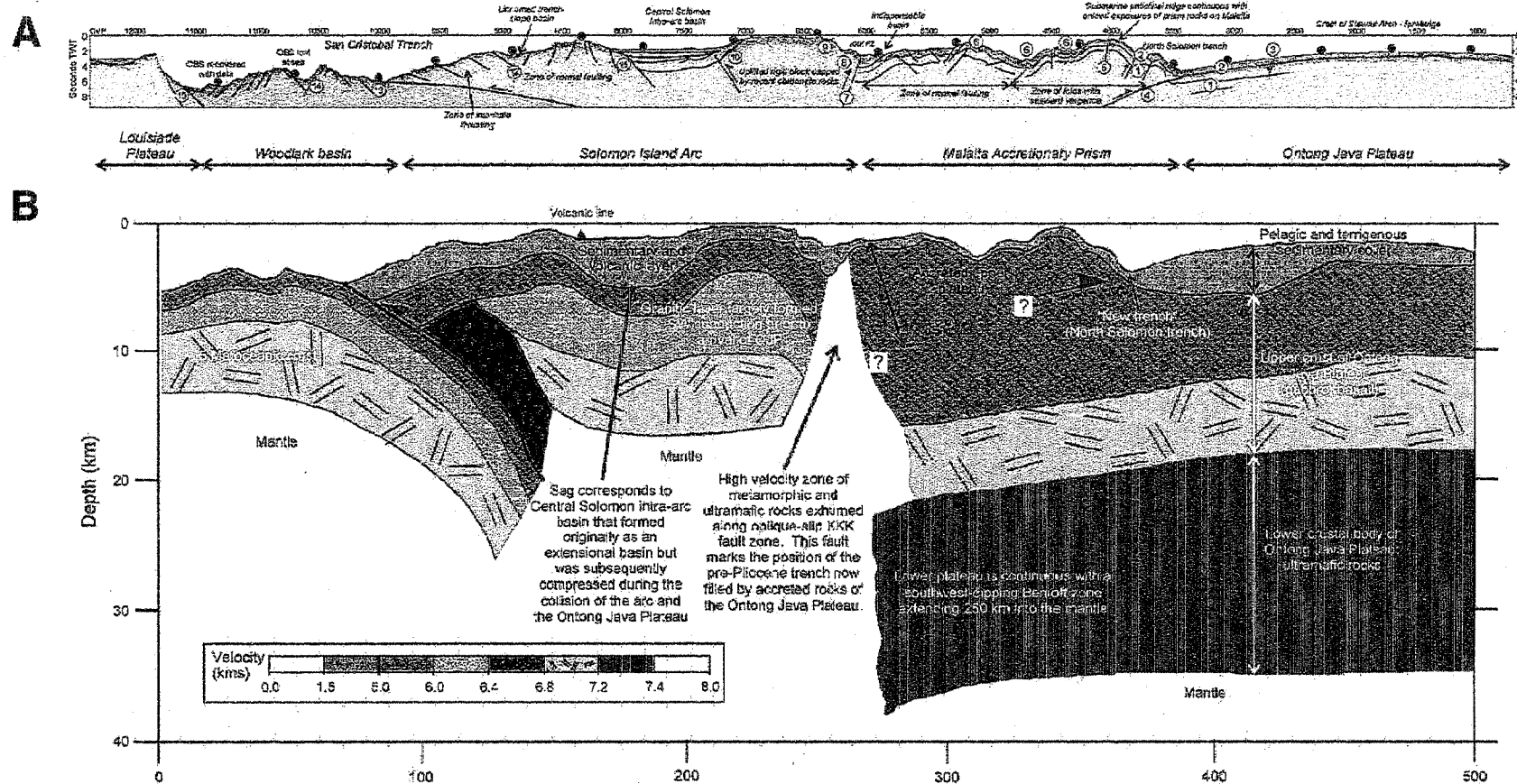


Fig.1-4. N-S profiles of Ontong-Java, through the Solomon Islands to Woodlark basin after Mann and Taira (2004). Above: Shallow structure <10 km based on the seismic reflection profile. Note the symmetric structure by subduction of oceanic lithosphere on the both north and south sides. SIM has a double arc fore-arc trench system. Below: Crustal velocity structures of OJP, SIM and Woodlark basin. Note the Calc-alkaline volcano stands 10-30km above the Benioff-Wadati plane on the left, whereas subduction zone volcanoes by the Pacific plate are absent. Thickness of crust of OJP, SIM and Woodlark basin are, 35 km, 15 km, and 6-7 km, respectively. A thick unit, named accreted upper part of plateau of OJP may be a thrust package of the OJP since 20-25 Ma or 8.5Ma (see the text).

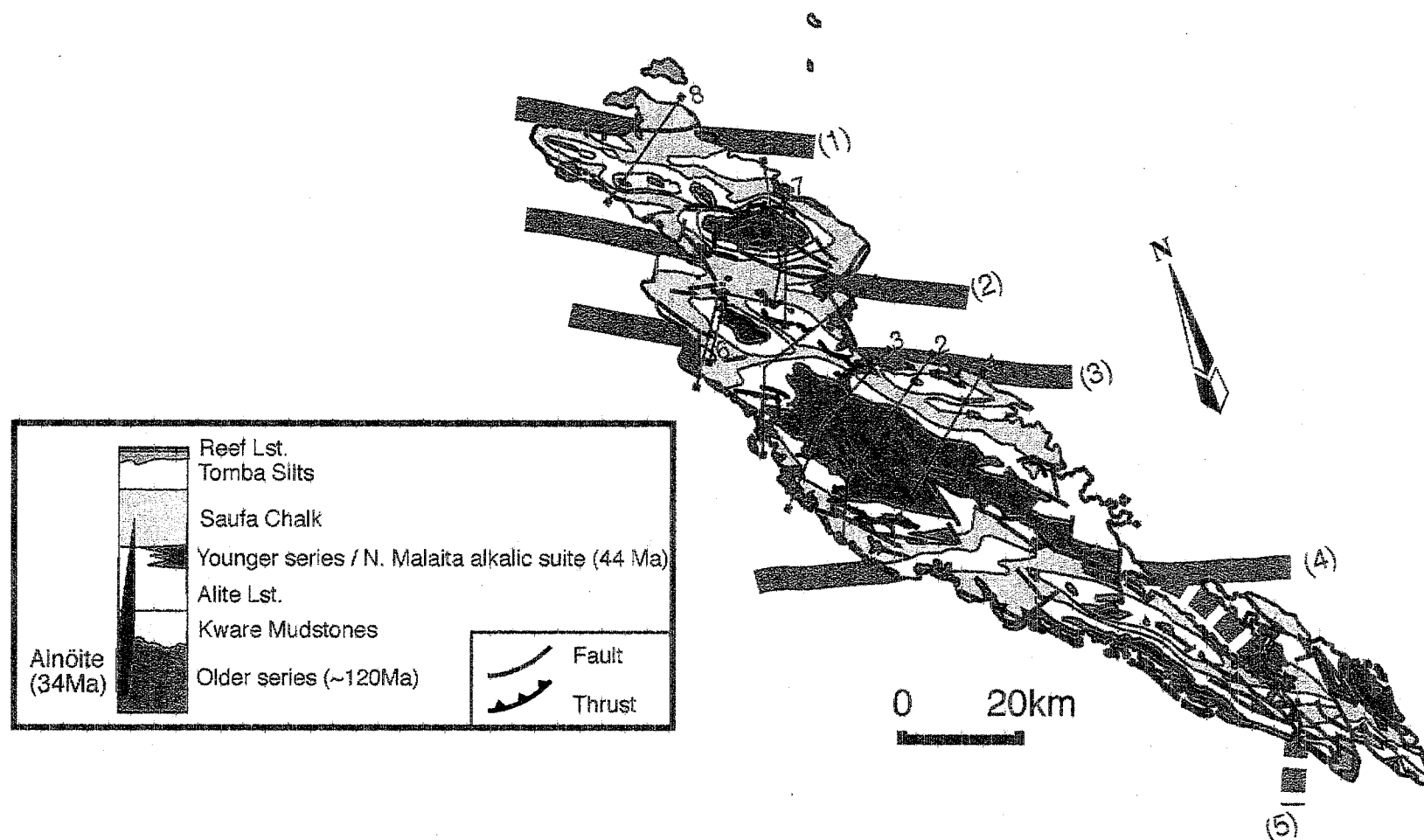


Fig.1-5. Geologic map of Malaita Island (after Petterson et al., 1997). Note the distribution of thrust (red solid teeth), and high-angle normal fault (black solid line). At least four tectonic domains are recognized by the late-stage extensional collapse of thick accretionary complex composed mainly of OJP. Distribution of orogen-parallel thrusts, cut by the late-stage normal faults. Ca. 44Ma alkaline suites appear on the southern and northern tips of the island. The 34 Ma alnöite occurs along the rifted margin between the two basement blocks to the north. The central mountain range as high as 1500m above sea-level, is composed mainly of ca.120 Ma fragments of OJP. Those are thick piles of pillow lava, massive sheet flow with minor reworked hyaloclastite, pelagic limestone, chert and hemi-pelagic mudstone, and trench-fill turbidite.

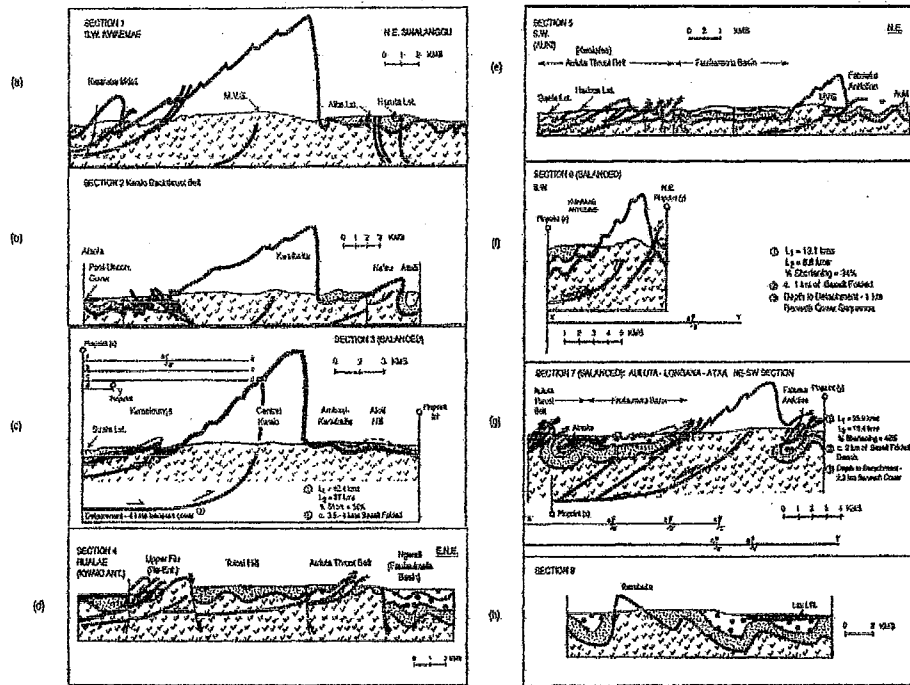
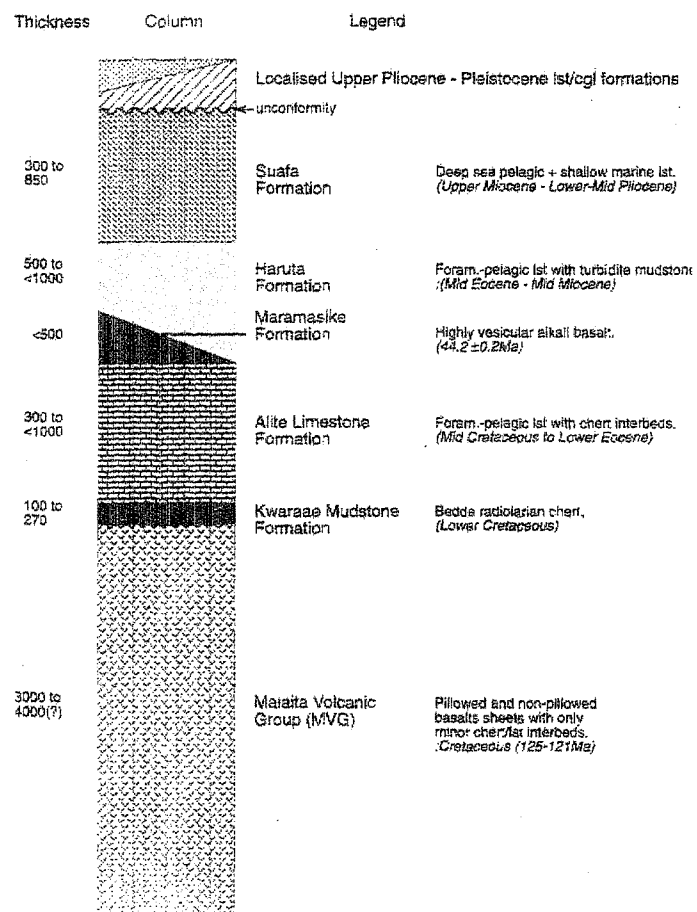
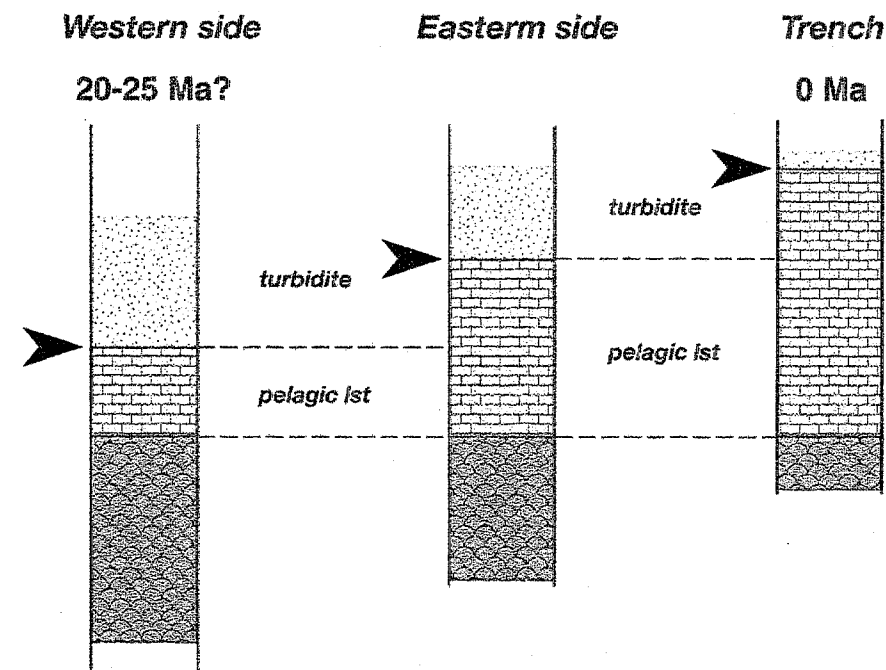


Fig. 1-6. Cross sections along the NS traverse in Fig. 1-5. After Petterson et al. (1997). Note the presence of thrusts and tight folding with east vergency. Locally a back-thrust is present (line 2). Late-stage extension collapsed the early thrust piles of the OJT. Note the most extensively elevated portion in the central Malaita by thrusting according to the speculation by Petterson et al. (1997).

(a)



(b)



Oceanic Plate Stratigraphy

Fig.1-7. An oceanic plate stratigraphy capped by in-situ sedimentary rocks after accretion, modified after Petterson et al. (1997). The oceanic flood basalts formed by the Pacific superplume at 125-121 Ma capped by the bedded chert and pelagic limestone, and by 44.2 Ma off-ridge volcanism of alkali basalts which in turn overlain by deep-sea pelagic sediments during the lateral transportation by plate movement. At sometime around 25-20 Ma trench sediments covered the OPS finally. All of these were covered by the in-situ sediments after late Pliocene.

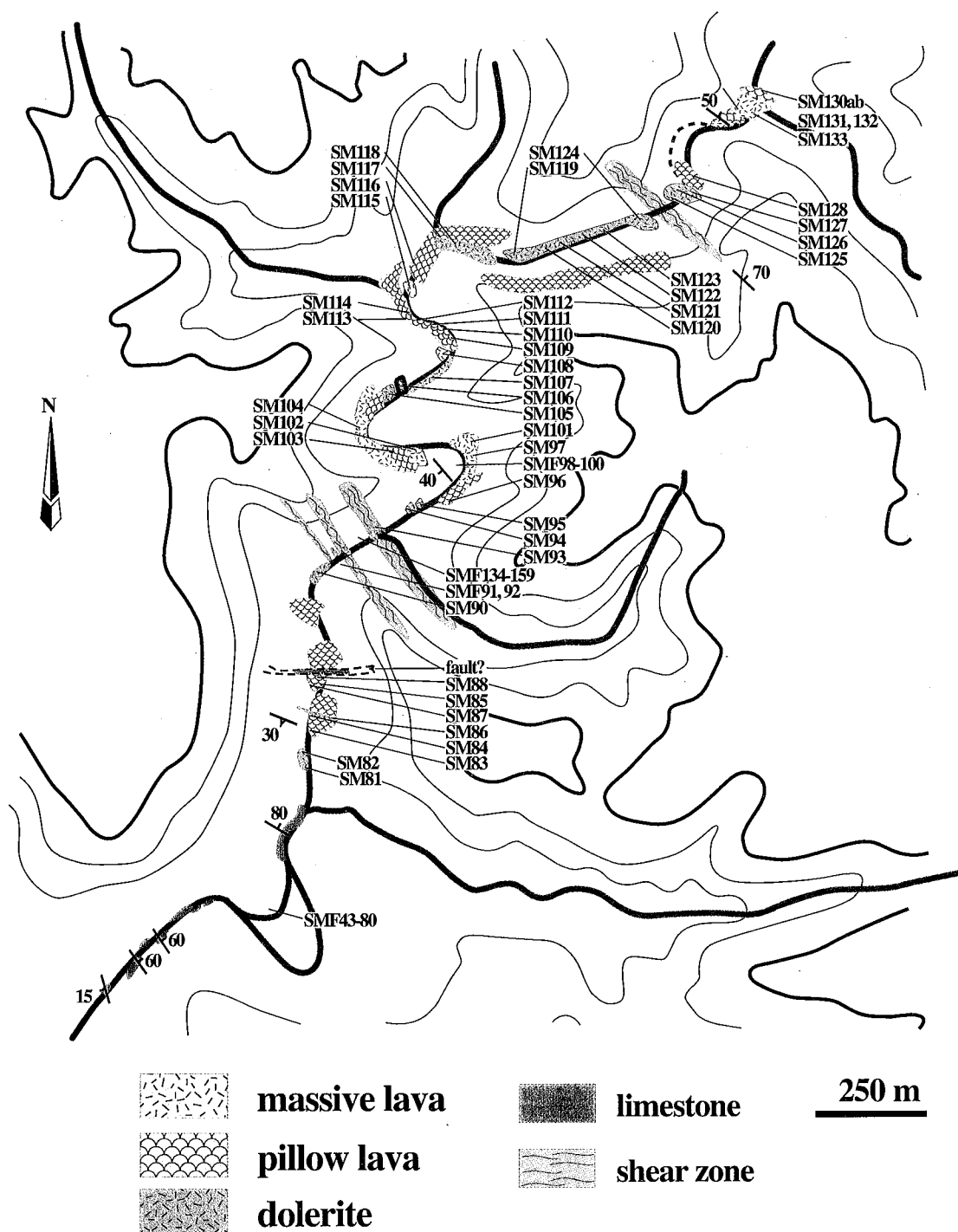


Fig.1-8. A NE-SW traverse of Malaita island and sampling map along the Singgalo River. Note the NW-SE trending shear zones dipping gently SW side, indicating thrust imbricated pile-up structure of the OJP accretionary complex.

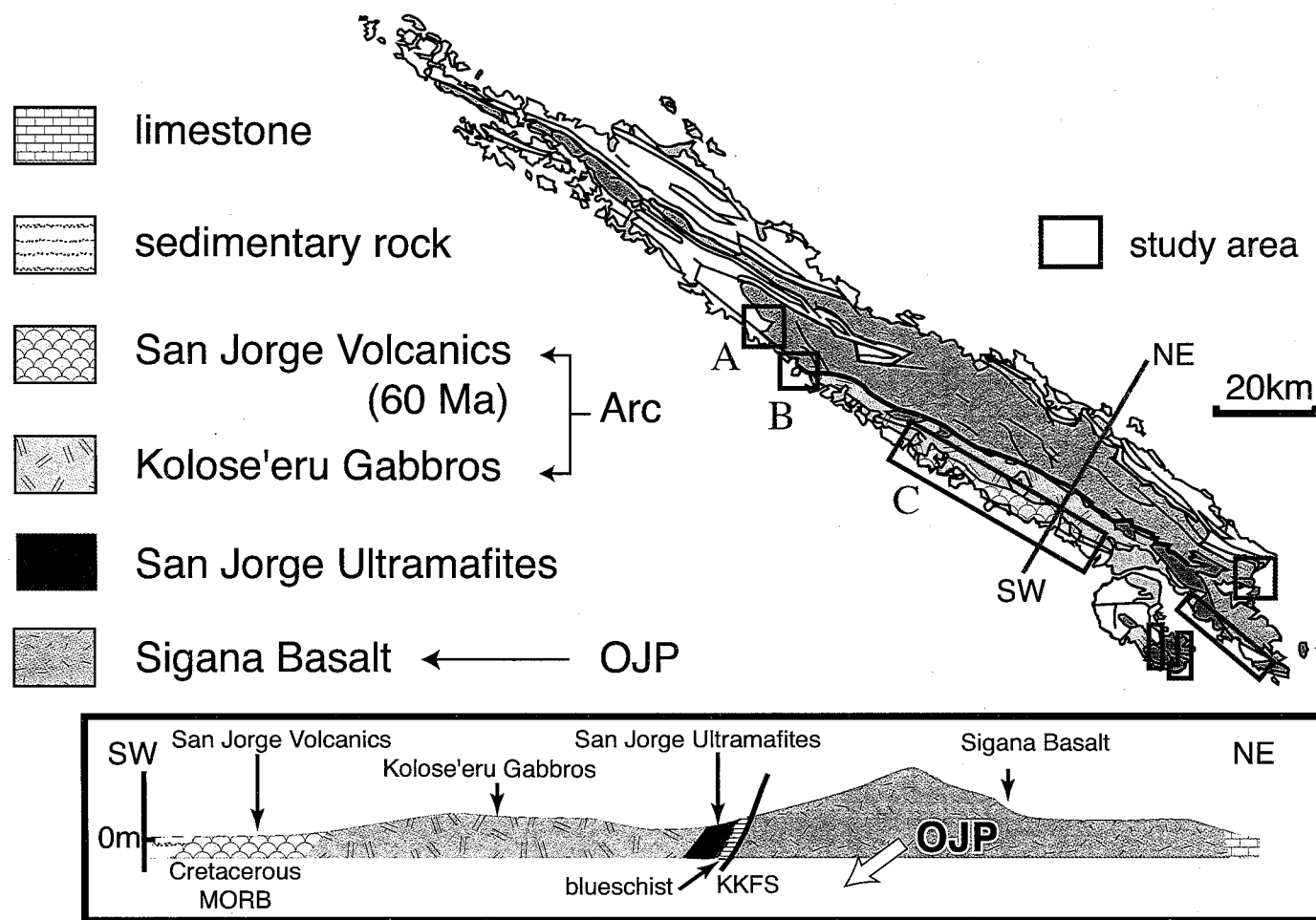


Fig.1-9. A geologic map of Santa Isabel. After Geol. Surv. Solomon islands (1996). The study area by Titech Group in 2000, 2001 and 2002 are shown. Major part of the island is occupied by accreted portion of OJP (Sigana Basalt). Separated by S-dipping KKFS thrust, the southern ophiolitic unit composed of peridotite, gabbro, and basalt is unconformably overlain by the Eocene arc volcanics (San Jorge Volcanics & Kolose'eru Gabbro). Along the KKFS thrust, minor blueschist blocks were found by Titech group in 2000.

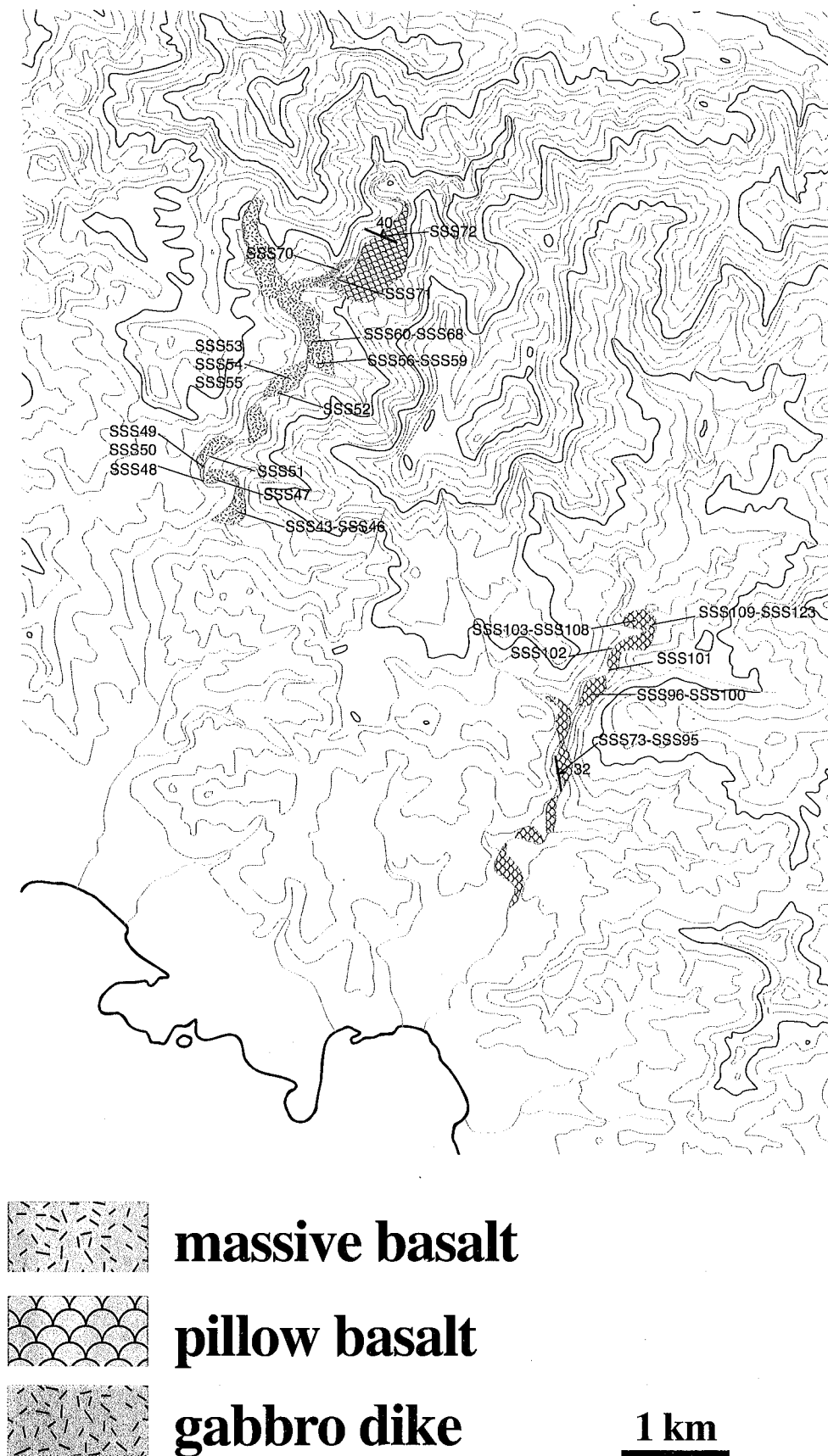


Fig.1-10. Sampling map of the Santa Isabel Island. Its area is A in Fig. 1-9.

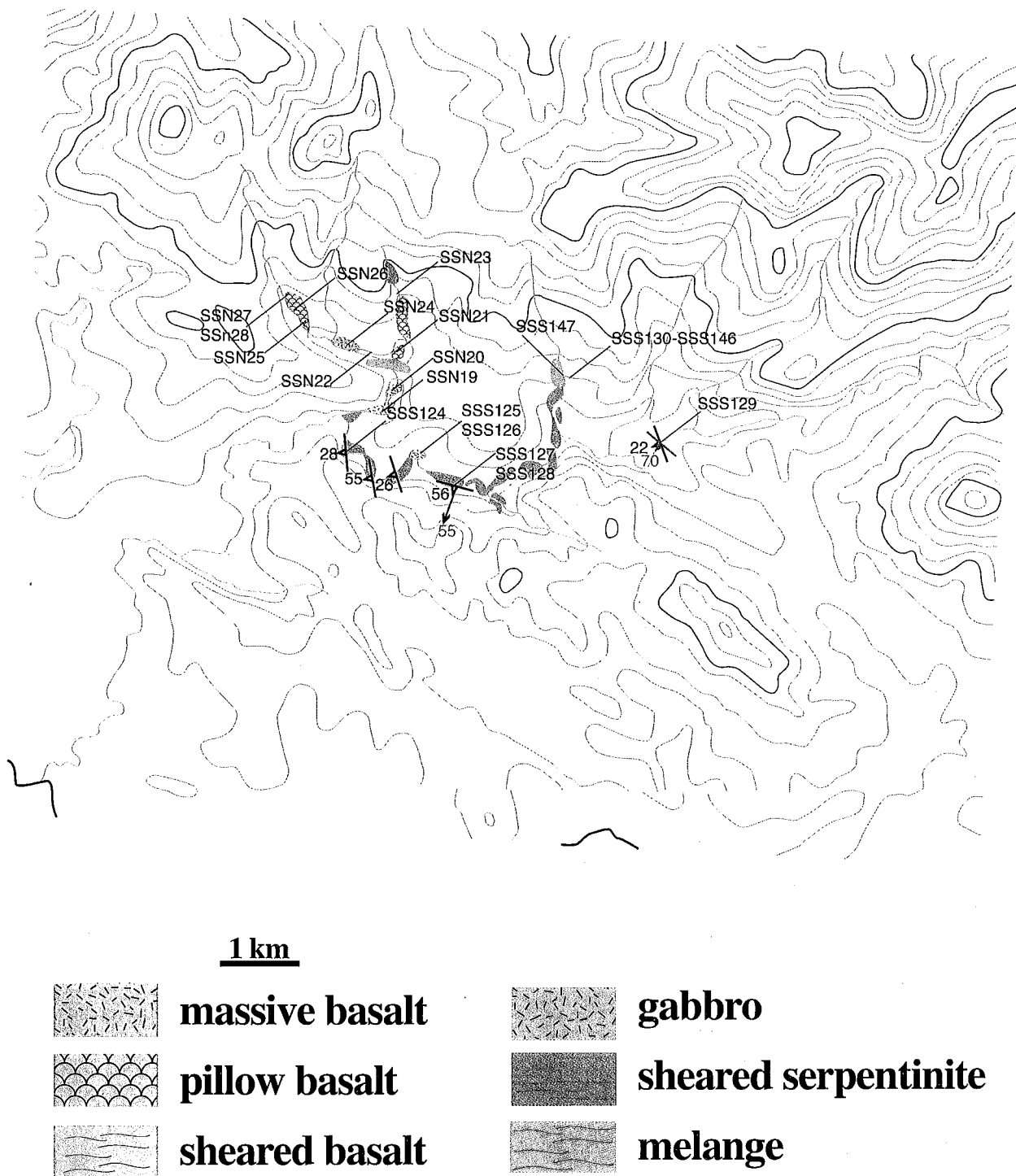


Fig.1-11. Sampling map of the Santa Isabel Island. Its area is B in Fig. 1-9.

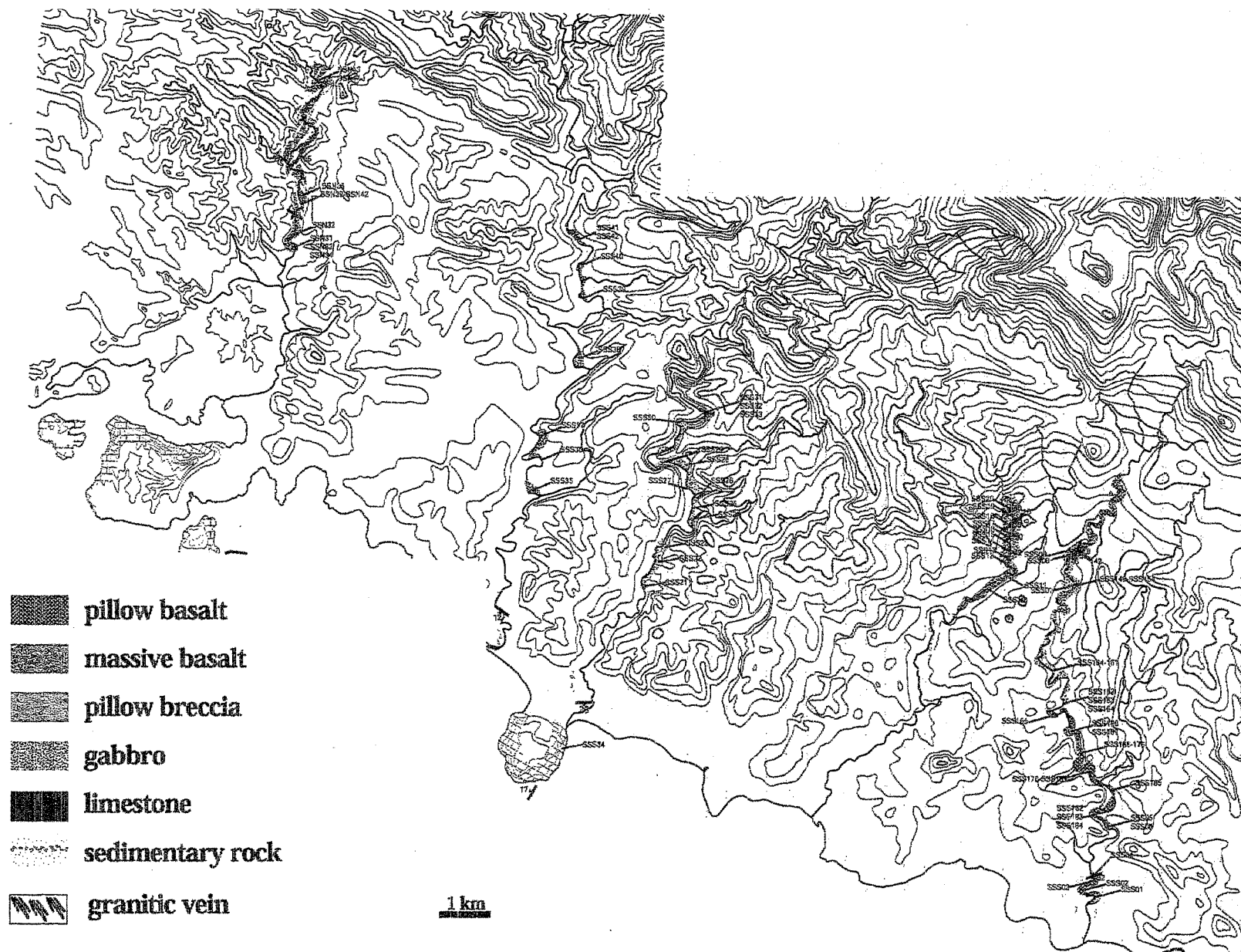


Fig.1-12. Sampling map of the Santa Isabel Island. Its area is C in Fig. 1-9.

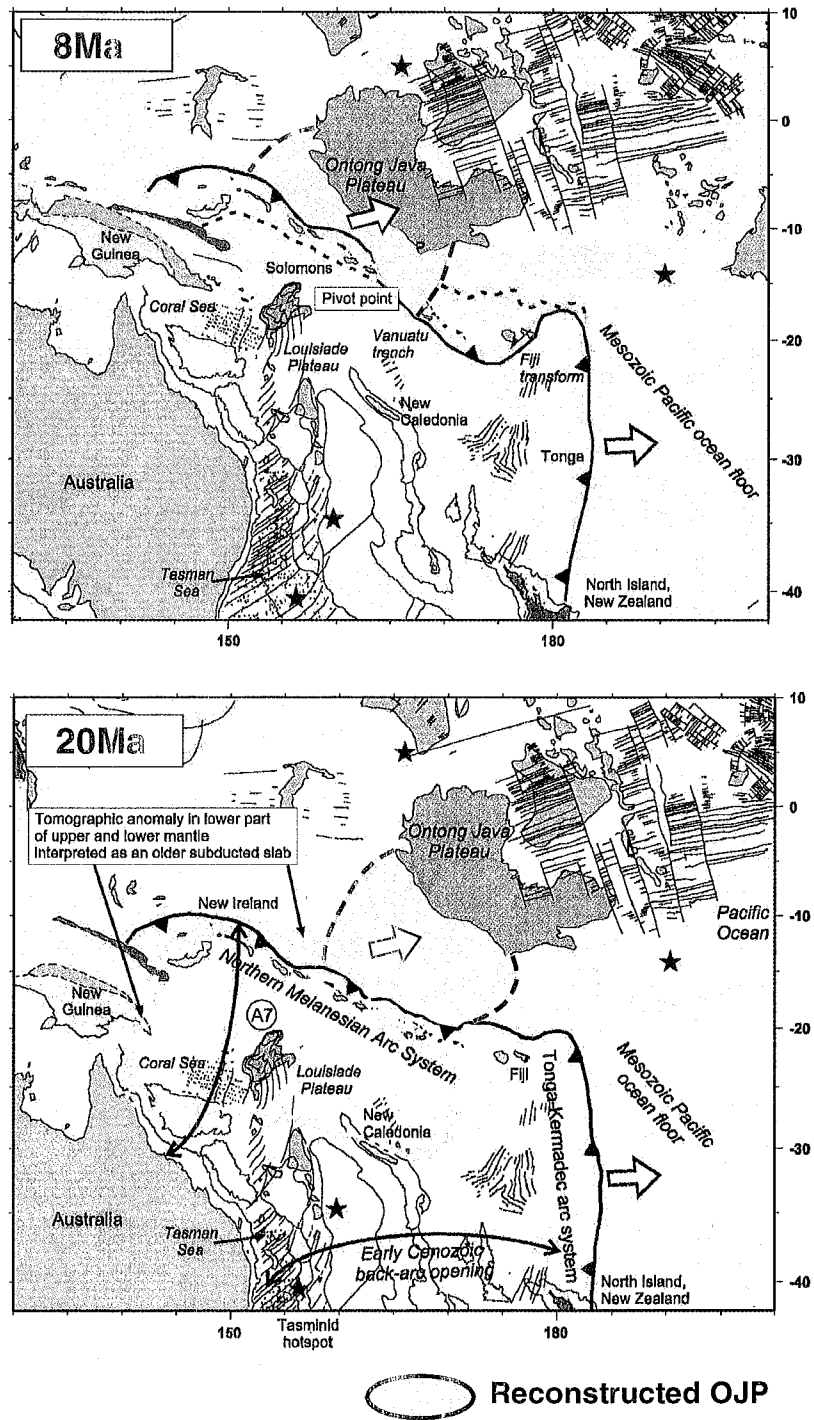


Fig.1-13. Paleogeographic reconstruction of OJP at 8Ma (above) and at 20Ma (below). Base maps are Mann and Taira (2004). Note the size of subducted portion of OJP. About 1000km wide OJP may have been present at least before subduction (above) or 2000km wide as large as the present-size of OJP may have been present (below).

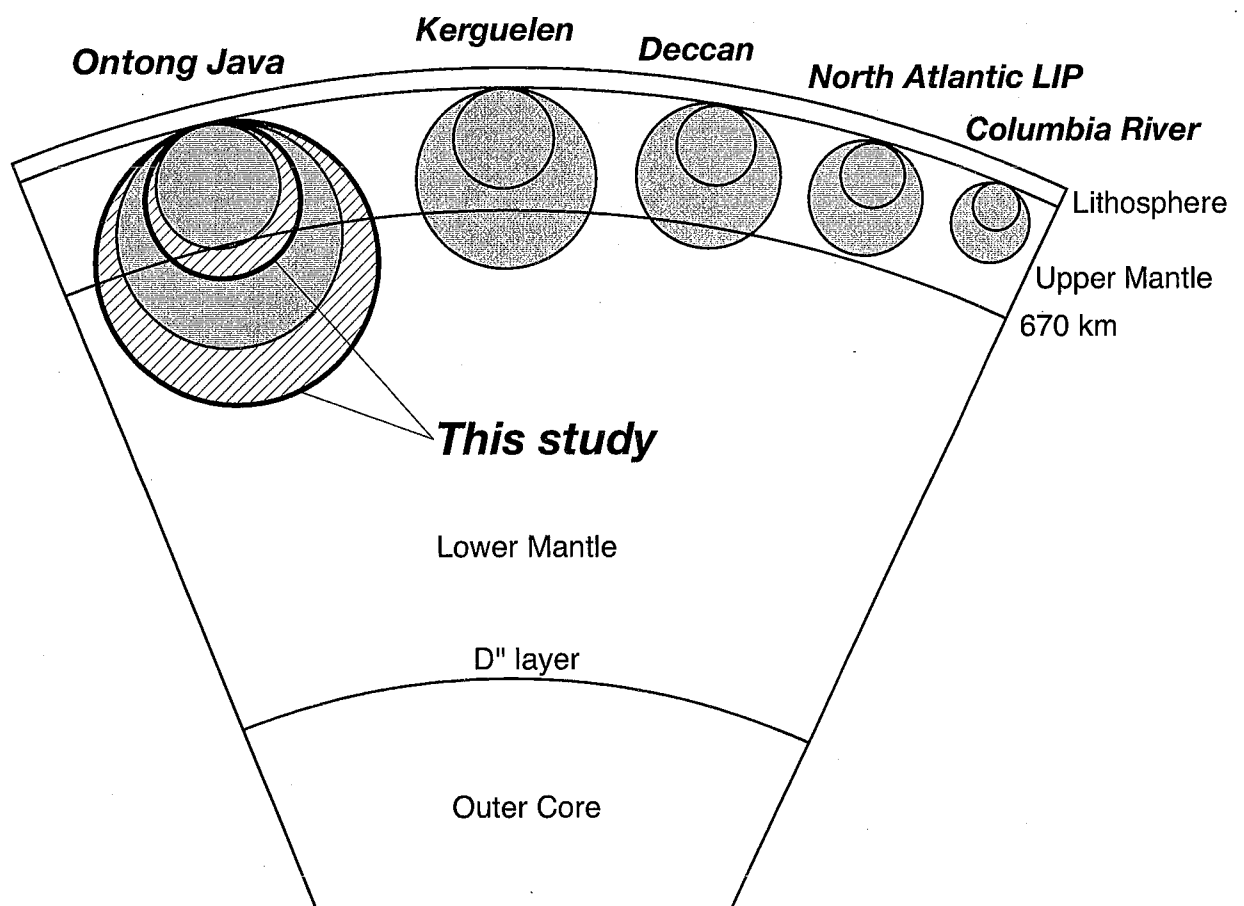


Fig. 1-14. Plume size of the OJP assuming the lherzolitic source mantle (after Coffin and Eldholm, 1994).

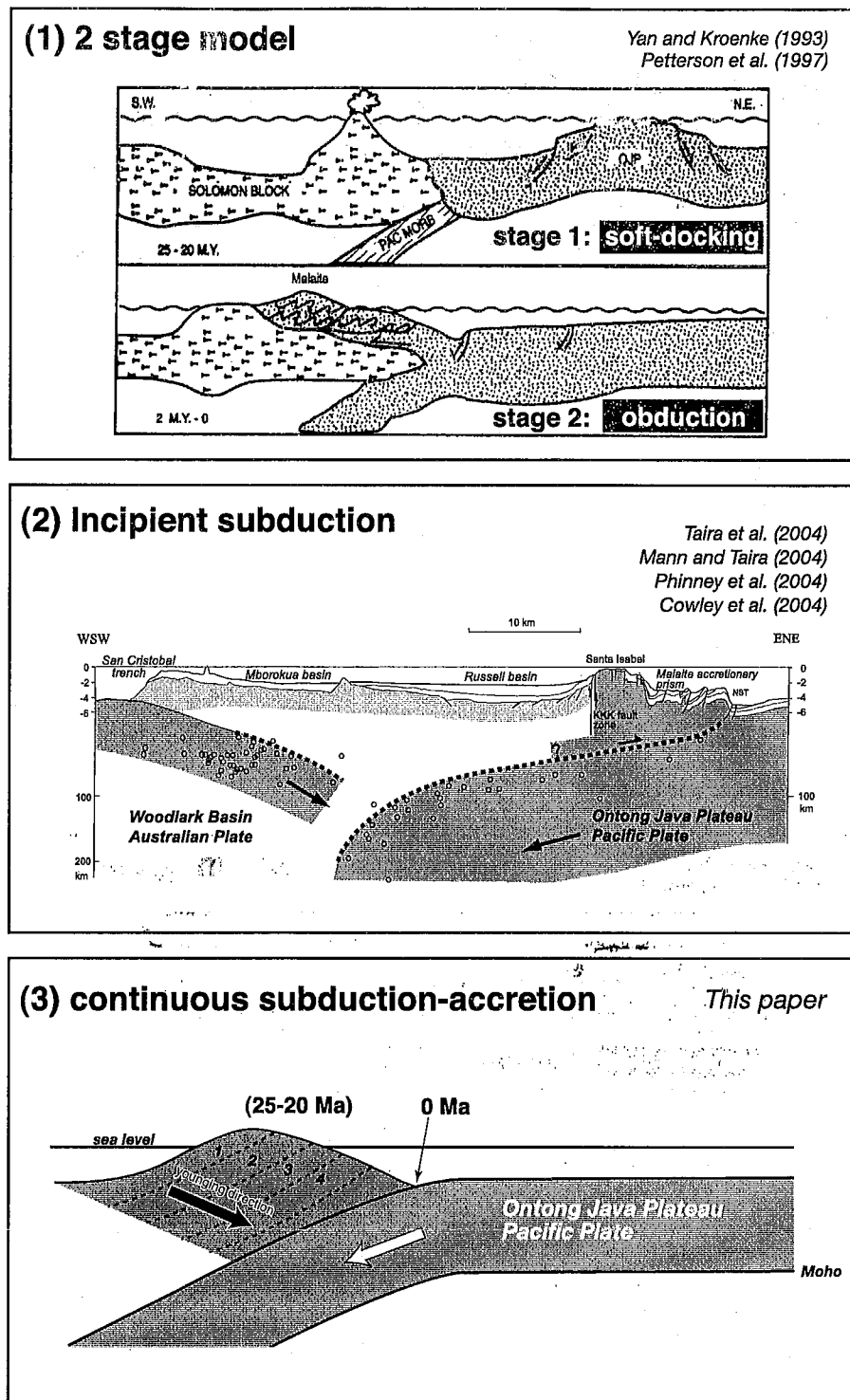


Fig.1-15. Three different models of the OJP accretion to the Solomon Island. (1) The soft-docking of the OJP at 20-25 Ma, and obduction at 2-4Ma to Malaita (Petterson et al., 1997), (2) Incipient subduction and accretion at 1 Ma by Taira et al. (2004), and (3) Continuous subductipn and accretion since 25-20Ma (this paper). Subduction and accretion of the OJP after the arrival time at North Solomon trench at 25-20Ma .



Plate 1-1. Surface skin of glassy pillow lava flows. Note the similarly looking surface of pillow to the Hawaiian modern pillowed lava flows. Singgalo River.



Plate 1-2. Same as Plate 1-1.



Plate 1-3. A typical pillow lava pile. Un-deformed. Note the thin skinned glassy margin, radial cracks, and open lava tubes in the core. Along Singgalo River.



Plate 1-4. A giant pillow with a lava tube in the center which was later filled by lava. A flat hollow was filled by late-stage segregated quartz (white portion). Along the Singgalo River.



Plate 1-5. A flat sheet of lava flow transforms to a pillow on the left and on the right. In the center, a flat vacant space about 10 cm thick is present. Along the Singgalo River.



Plate1-6. An un-deformed pillow piles. Along Singgalo River.



Plate 1-7. A close-up view of pillow lobe of Plate 1-4.



Plate 1-8. Surface skin of glassy pillow lava flows.



Plate 1-9. The pillow lava flows on the left is covered conformably by the sheet flows on the right. Along the Singgalo River.



Plate 1-10. The bottom 3 m of a massive sheet flow about 10m thick. Note the developed columnar joints vertical to the bottom boundary. Also note that layering parallel to the bottom boundary. Olivine crystals are accumulated on the bottom. Along the Singgalo River.



Plate 1-11. The close-up view of the bottom of a sheet flow. Note the presence of glassy zone on the bottom and the underlying pillow lava flow.



Plate 1-12. A fractionated silicic portion of massive sheet flow of Plate 1-10. Along the Singgalo River.



Plate 1-13. An outcrop pf hyaloclastite. Along the Singgalo River.

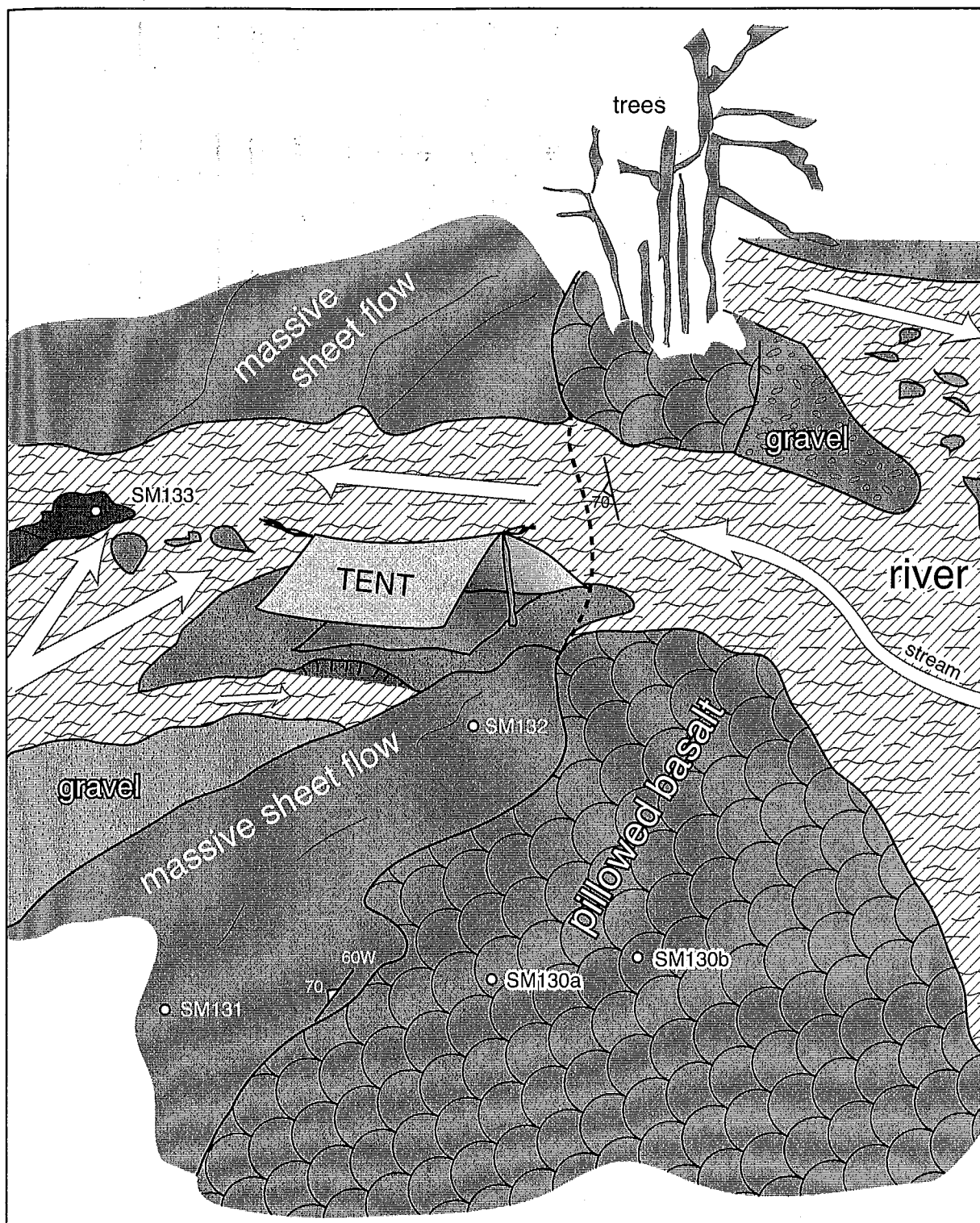


Plate 1-14. A sketch of the relation between pillowed lava flow and a massive sheet flow.



Plate 1-15. Our tent in the river Singgalo early morning after flooded in the night.

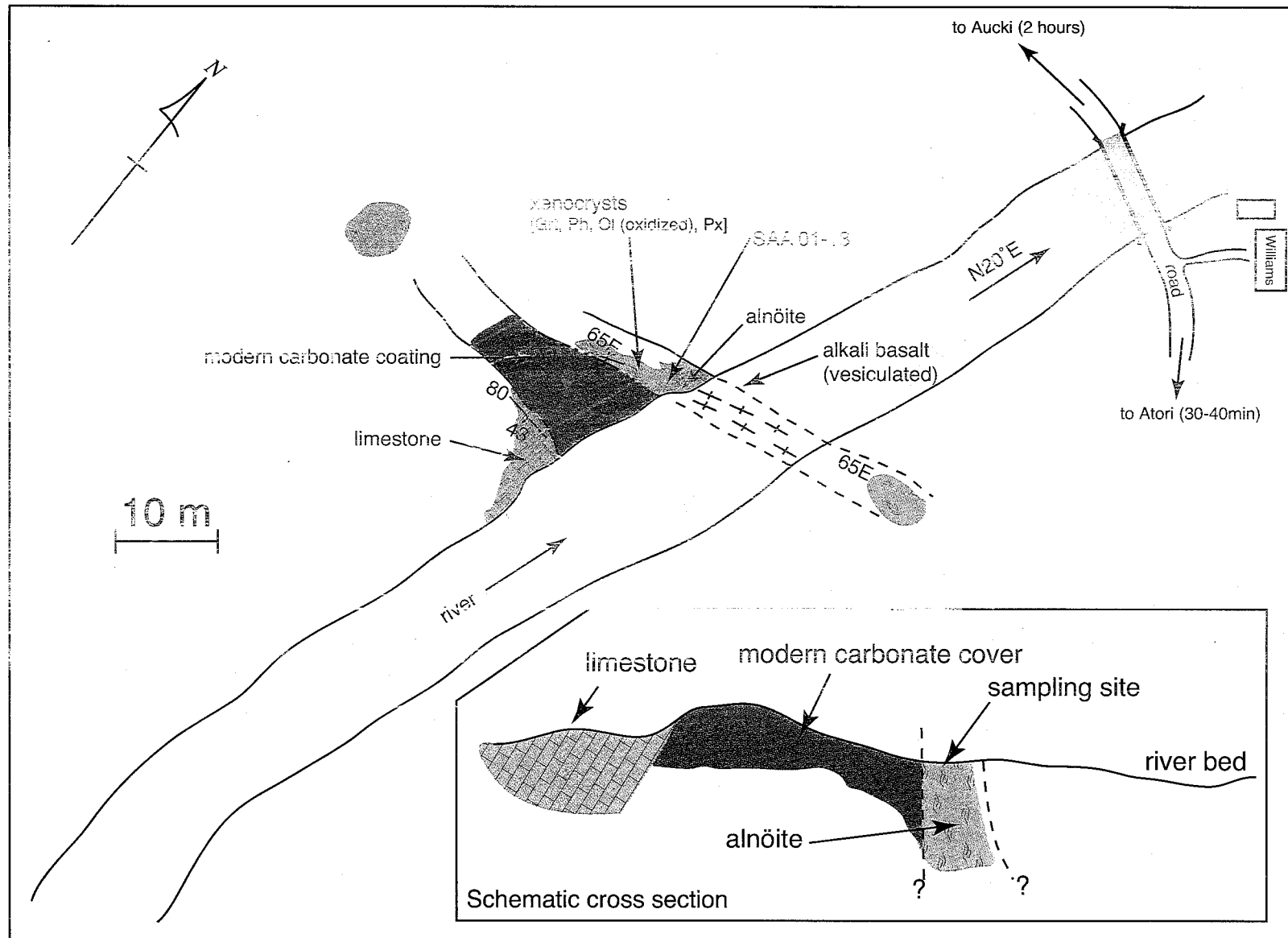


Plate 1-16. Alnöite outcrop in the northern Malaita. Alnöite occurs as a pipe trending 65 E intruding stratified limestone bed. Highly brecciated lava fragments are mixed with megacrysts of garnet, phlogopite, olivine, and clinopyroxene.

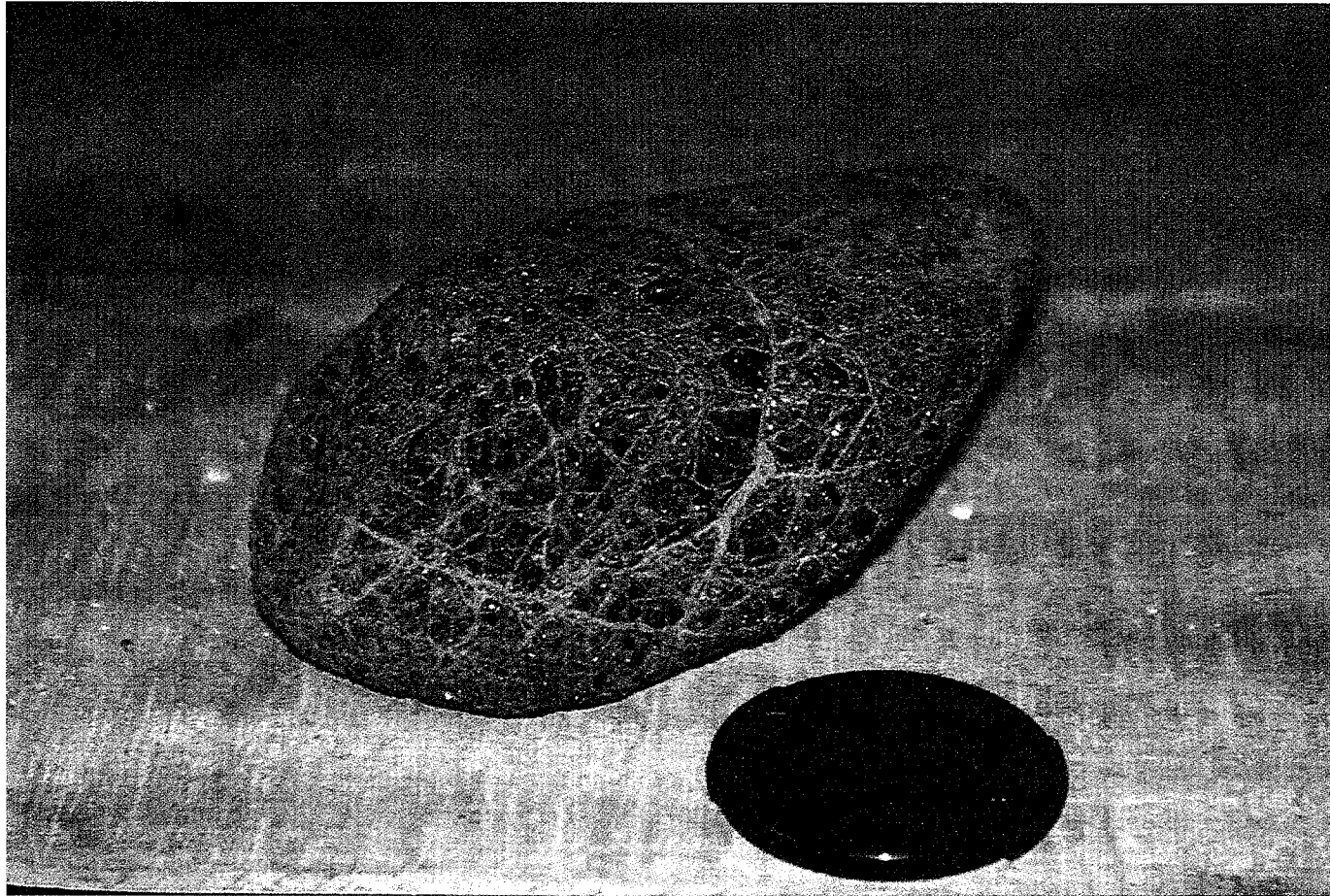


Plate 1-17. A garnet megacryst presumably derived from Alnöite pipe. The largest reaches to man-head size.

Chapter II. Petrology and Geochemistry of the Ontong Java Plateau basalts from islands of Malaita and Santa Isabel, Solomon Islands

Abstract

To clarify the petrogenesis of the Ontong Java plateau (OJP), we have carried out the petrological and geochemical study on the systematically enormous collected samples in central Malaita and Santa Isabel. We determined the major, trace and Rare Earth Elements (REE) compositions of whole-rock and major element compositions of igneous minerals from lava, and major compositions of daughter minerals in olivine-hosted melt inclusions from olivine gabbro.

On the basis of major element compositions, the OJP lavas from central Malaita are divided into three types; High-TiO₂ type, Low-TiO₂ type, and High-MgO type. High-TiO₂ type lavas are higher in TiO₂, Fe₂O₃*, P₂O₅, and REE and lower in Al₂O₃ and CaO contents at given MgO than Low-TiO₂ type. Both have flat REE patterns, although high-TiO₂ type lavas have slightly LREE enriched patterns. High-TiO₂ type and Low-TiO₂ type lavas correspond in major and REE compositions to Singgalo- and Kwaimbaita-type basalts distinguished by previous study, respectively. High-MgO type lavas have high MgO and Fe₂O₃*, slightly high TiO₂, P₂O₅ and REE and low SiO₂, Al₂O₃ and compatible trace element contents. They have similar REE patterns to high-TiO₂ type. The feature is compared with those of no other OJP basalts. The stratigraphic relation of these type lavas are ordered as High-MgO, High-TiO₂, and High-TiO₂ type lavas beginning at the top although the all the sequence is not conformably exposed because they are disturbed and repeated by the faults.

Daughter minerals in olivine-hosted melt inclusions have anomalous features compared with host olivine gabbro. Amphibole-bearing melt inclusions are dominant unless it is absent in host olivine gabbro. Daughter amphiboles and clinopyroxenes are higher in Fe, Al and Ti. They are reverse zoned. A few globules have carbonate. Although the petrological features of melt inclusions are modified by complicated process of closed-system after trapped, the core compositions of mafic minerals suggest that they are generated from SiO₂ deficient melt like High-MgO type lavas.

High-MgO type lavas are comparable with no melting experiment of peridotite, eclogite and peridotite and basalt mixture at high- and low-pressure. The high-pressure melting

of garnet clinopyroxenites is roughly explainable for major, trace element and REE features of High-MgO type lavas. The most reasonable source of High-MgO type lava is refractory of eclogitic recycled oceanic crust extracted SiO_2 -rich melt. High-MgO type lavas infer that the plume generating OJP include the ancient oceanic crust. Other type lavas are expected to be mixture of totally molten basalt and a partial melt of peridotite.

2-1. Introduction

Mid-Cretaceous (from 125 Ma to 80 Ma) is the anomalous period that global environment has drastically changed. Larson (1991b) summarized the magnetic reversal stratigraphy, world oceanic crust production (Larson, 1991a), high-latitude sea-surface paleotemperatures (Savin, 1977; Arthur et al., 1985), long-term eustatic sea level (Haq et al., 1988), times of black shale deposition (Jenkyns, 1980), and world oil resources (Irving et al., 1974; Tissot, 1979) from 150 Ma to present and discussed the relations between these observations. Secular variation of sea-surface temperature at high latitude estimated by $\delta^{18}\text{O}$ of microfossils indicates the paleotemperature for the mid-Cretaceous has the highest level of that in recent 150 m.y. (Savin, 1977; Arthur et al., 1985). World-widely spread black shale notably deposited in the period (Jenkyns, 1980) and world oil resource production has the peaks in the period (Irving et al., 1974; Tissot, 1979). Eustatic sea level rises from late Aptian through Turonian (Haq et al., 1988).

Larson (1991b) insisted that these geological consequences have been yielded by pulse in oceanic crust production. World oceanic crust production including oceanic plateaus in the period is 1.5 times to twice as voluminous as that of other periods (Larson, 1991a). Especially, Pacific regions are remarkable. The oceanic plateaus distributed in southwestern Pacific are representative. The outgassing of mantle volatile, especially CO_2 , and continental rearrangement accompanied by enormous amount of oceanic crust production increased in world temperature. Notable black shale deposition and the anomaly in oil generation were resulted from increases in excess nutrients and carbon supplied by pulse in oceanic crust production. The raised sea level is resulted from high spreading rate of mid-ocean ridge, elongation of ridge crest, and production of oceanic plateaus. However oceanic crust production alone cannot explain all observations, for example Eocene and Jurassic eustatic sea level rise, this is important source function.

The period corresponds to the Cretaceous normal superchron that is long interval of normal magnetic polarity without reversal (Harland et al., 1990; Larson, 1991b). However mechanism and process of magnetic field reversal stoppage is not well understood, the removal of heat and material from deep mantle by enormous amount of oceanic crust has some influences on it (Larson, 1991b).

It is the latest pulse in history of the earth (Larson, 1991a). To clarify the

petrogenesis of the Cretaceous oceanic plateau, especially distributed Pacific region, it is very notable to understand the pulse period in the earth history. But unfortunately, the almost pacific plateaus are subsided below the sea level. The study of the plateau basalts is dominantly undergone to treat the core samples by Deep Sea Drilling Project (DSDP) and Ocean Drilling Project (ODP).

In the Solomon Islands of Malaita, Ramos, Ulawa, San Cristbal and the northeastern part of Santa Isabel, the part of world's most voluminous oceanic plateau, the 5×10^7 Km³ Ontong Java Plateau (OJP) are exposed (e.g. Petterson, et al., 1997). OJP basalts exposed Solomon Islands are well studied by geochemical method (e.g. Tejada, et al., 1996; 2002). Their petrological and geochemical features are comparable with those of drilled samples from ODP/DSDP site on the OJP.

The petrogenesis of the OJP basalts is arguable. Firstly, on the basis of the geochemistry of the OJP basalts, they were generated by primitive mantle partial melting at 30% degree in homogeneous mantle peridotite plume (Tejada et al., 2002; Fitton and Godard, 2004). Secondary, based on the petrology of xenoliths in alnöite from Malaita, the mantle plume generating the OJP basalt include the recycled basaltic component and it has contributed to the generation of them (Ishikawa et al., 2004). Thirdly, by mentioned the evidences hardly to explain by the plume model, the OJP basalts were generated by impacting bolide (Ingle and Coffin, 2004).

Previous studies of the OJP basalt were restricted the sporadic locality. In this study, we have carried out the petrological and geochemical study on the systematically enormous collected samples in central Malaita and Santa Isabel. We determined the major, trace element and REE compositions of whole-rock, major element compositions of igneous minerals, and major compositions of daughter minerals in olivine-hosted melt inclusions. We discovered MgO-rich basalts and more chemical heterogeneous preserved in melt inclusions. On the basis of these, we discuss the petrogenesis of OJP and plausible mantle plume model.

2-2. Geological outline

Northern and central Malaita has several asymmetrical anticline structures whose axes are generally oriented northwest-southeast and verge to the northeast and

eastern limbs shallowly dip and western limbs steeply dip to overturn (Petterson et al., 1997). The cores of anticlines consist of Cretaceous basaltic basement. The basement rocks recently have been termed the Malaita Volcanic Group (MVG; Petterson et al., 1997). MVG is conformably overlain by Lower Cretaceous mudstone (Kwaraae Mudstone Formation), Mid Cretaceous - Lower Eocene pelagic limestone with chert interbeds (Alite Limestone Formation), Mid Eocene - Mid Miocene pelagic limestone with turbiditic mudstone (Haruta Limestone Formation) and Upper Miocene - Lower-Mid Pliocene pelagic and marine limestone (Suafa Limestone Formation) in order (Petterson et al., 1997; Petterson, 2004). During the Eocene (44 Ma) and Oligocene (34Ma), alkaline basaltic volcanism and alnöitic intrusion occurred, respectively (Davis, 1977; Tejada et al., 1996). Alkaline basalts were recently named the Maramaskite Formation (Petterson et al., 1997). The Youngest Pliocene – Recent shallow water clastics and reef limestone formations unconformably overlie the MVG and the pelagic sedimentary cover sequences. This unconformity was produced by the uplift and transpressive deformation related to the obduction of the OJP against the Solomon arc, mainly between 4 and 2Ma (Petterson et al., 1997; Peterson, 2004).

MVG consists of piled sheets of pillowed and non-pillowed tholeiitic basalts and the maximum exposed thickness have been estimated at ~4000m (Petterson et al., 1997; Petterson, 2004). The MVG was dated at 122 Ma by ^{40}Ar - ^{39}Ar method (Tejada et al., 1996, 2002). The age is almost indistinguishable from that of OJP basalts (Mahoney et al., 1993; Parkinson et al., 2001; Chambers et al., 2002). MVG is closely similar in geochemical and isotopic features to OJP basalts (Mahoney et al., 1993; Tejada et al., 1996, 2002, 2004; Fitton and Godard, 2004). MVG is divided into two groups that are Singgalo- and Kwaimbaita-types on the basis of geochemical and isotopic compositions. These were named after the two rivers across the island of Malaita where thickest exposures of each magma type have been found (Tejada et al., 2002). Singgalo-type has lower $\epsilon_{\text{Nd}}(t)$ and $(^{206}\text{Pb}/^{204}\text{Pb})_i$, and higher $(^{87}\text{Sr}/^{86}\text{Sr})_i$ than Kwaimbaita-type. Although slightly unobvious, Singgalo-type are higher in incompatible element contents than Kwaimbaita-type. Each type lava is remarkably homogeneous geochemical and isotopic compositions (Tejada et al., 2002). They are stratigraphically separated into upper and lower groups and these are Singgalo- and Kwaimbaita-types, respectively. Singgalo- and Kwaimbaita-type lava flows correspond to those of Unit A and Unit C-G at ODP site 807 in geochemical feature, respectively, and have similar stratigraphical relation

between Unit A and Unit C-G (Mahoney et al., 1993; Tejada et al., 2002). MVG in southern Malaita has Singgalo-type geochemical and isotopic features (Tejada et al., 1996).

Santa Isabel is divided into two tectonic provinces by Kaipito-Korighole Fault System (KKFS) oriented northwest-southeast. In northern region from the KKFS, basement lavas are known as the Sigana Basalts and overlain by Late Cretaceous – Late Paleocene pelagic limestones and volcanoclastic sediments. They are intruded by rare alkalic dikes termed the Sigana Alkalic Suite (Tejada et al., 1996). Sigana Basalts were dated at 121 Ma, 90 Ma and 62 Ma by ^{40}Ar - ^{39}Ar method (Tejada et al., 1996). Former two ages are indistinguishable from OJP basalts. Structural relation of ages is unclear. Many of them are similar to two types of MVG in geochemical and isotopic features but a few samples have high TiO_2 contents. Latter are termed High Ti Sigana Basalts (Tejada et al., 1996). Sigana Alkalic Suite has HIMU signature in isotopic compositions (Tejada et al., 1996). They are dated at 92 Ma and inferred that these have intruded concurrently with 90-Ma group of Sigana basalt (Tejada et al., 1996). In southern region from KKFS, ophiolitic assemblage referred to Jajao Igneous Suite by Tejada et al. (1996) is distributed. It is comprised of basalts (San Jorge Volcanics), gabbros (Kolose'eru Gabbros), minor leucocratic rocks, and ultramafites (San Jorge Ultramafites), overlain by clastic sediments (Tejada et al., 1996).

2-3. Methods

We have carried out the geological survey in Solomon Islands in 2000 and 2001. Samples were systematically collected with sketching route maps on enlarged 1:5000 scale topographic maps along Singgalo and Ambasi rivers for Kwaio Anticline in Malaita (Fig. 2-1 and 2-2). As same way, samples were collected along Beamalalo in Santa Isabel Island (Fig. 2-8).

248 and 64 samples are selected for analyses of the major elements excluding strongly altered ones for Malaita and Santa Isabel Island, respectively. Each sample was crushed into small fragments using a hammer after cutting into slabs about 1 cm thick and grinding to exclude iron spilt of the blade. The fragments were washed in purified water and high quality ethanol using a ultrasonic bath, followed by crushing into powder using a tungsten carbide vessel. Glass beads for major element and REE

analyses and pellets for trace element analysis are prepared from these powders. Each sample powder was mixed with the lithium tetraborate ($\text{Li}_2\text{B}_4\text{O}_7$) flux in the proportion of 1 to 10, and fused at 1200 °C in a Pt crucible and cast into a glass bead. Each pellet was formed with a oil hydraulic press. Major elements were analyzed using RIGAKU Symaltics 3550 Xray fluorescence (XRF) spectrometers at the Tokyo Institute of Technology (Tokyo Tech) by method described by Goto and Tatsumi (1994). 35 samples were selected for REE and trace element analyses. REE data were obtained by using ThermoElemental VG PlasmaQuad 2, quadrupole-based ICP-MS with a S-option interface at Tokyo Tech following the procedures described by Orihashi and Hirata (2003). Trace elements were determined using RIGAKU RX2100 Xray fluorescence (XRF) spectrometers at Tokyo Tech.

Major element compositions of the minerals were observed by scanning electron microscope (SEM) and major compositions of them were determined by EPMA (JEOL JXA-8800) at the Tokyo Tech. The analyses were operated at 15 kV accelerating voltage and 12 nA beam current. Counting time was 10 - 20 s. Beam diameter was focused into a spot of 1 to 3 μm in diameter. Data were reduced by oxide ZAF correction.

2-4. The mode of field occurrence of the Ontong Java Plateau basalts

2-4-1. Central Malaita

Figs. 2-1 and 2-2 show route map along Singgalo and Ambasi rivers in Central Malaita, respectively. The route is comprised of dominant piled pillow and sheet lavas, minor intrusive gabbros, and rare interbedded sediments (Fig. 2-3).

Pillow lavas comprise 20 % of lavas of MVG along Singgalo and Ambasi rivers. They preserve primary structures (Fig. 2-4a). They are mostly close-packed. All pillow lobes affect normal position. They vary in size from 20 cm to 10 and several meters in major axis. They are also diverse in form from elliptical cross section close to circle to just about thin sheet lavas. But their majority is 0.5 – 2 m in major axis and 0.3 – 1 meter in minor axis. They have massive core and radial joints in mantle. Some have drain cavity or hollow. In outer structure, they have contraction cracks and some parts of rinds have flaked. Some pillow lobes preserve ~1 cm thick chilled margin but the

quench glasses are displaced by clay minerals.

Sheet lavas comprise 80 % lavas of MVG along Singgalo and Ambasi Rivers. They vary in thickness from 20 cm to tens of meters thick. The majority is several meters thick. Many of relatively thick ones have well-developed columnar joints (Fig 2-4b). Rare sheet lavas preserve ~1cm chilled margin but their quench glasses are displaced by clay minerals. Ropy wrinkle structures were observed at several points.

Pillow lavas and sheet lavas are closely related. They are alternated one flow unit to another in places. Some of sheet lavas transit to pillow lava in flow foot parts. Certain lavas have flow foot breccia with plunging pillow lobes (Fig. 2-5).

MVG and submarine OJP basalts has been said to be non-vesicular (e.g. Tejada et al., 1996). But the lavas bearing vesicles were often observed in our traverse (Fig. 2-1 and 2-2). The pillow lobes have occasionally spherical vesicles in size from 2 mm to 8 mm in diameter (Fig. 2-6). They are concentrated in inner surface of cracked rind. Some sheet lavas have those of the same sort in the faces under several centimeters from the flow boundary. These lavas are not dominated but unable to be ignored. These lavas are distributed all over the route.

Although lavas of MVG commonly flowed on former lavas directly, lava flow units are conformably overlain by thin sedimentary rock layers in places (Fig. 2-7). These are green volcanoclastics or reddish brown mudstones and in thick from 2 cm to several meters. Interpillowed sediments are rarely observed. These are commonly greenish hyalocrastites and occasionally carbonates.

Intrusive gabbros are exposed in Ambasi river route. The largest one is exposed 250 m long along the river. It is stratigraphically positioned lowermost part of apparent MVG in the route. It intruded in the former volcanic stratigraphy and underplating. The other outcrops are exposed tens of meters thick as dyke or sill.

2-4-2. Northwestern Santa Isabel

The basement basalts along Beamalalo river of pillow and sheet lavas and intrusive gabbros (Fig. 2-8). These have pale green or reddish saufaces and more weathered than the MVG. These lavas are in morphology similar to the MVG along the Singgalo and Ambasi rivers. In the route, pillow lavas are dominant. Interpillowed sediments are mostly limestones. Limestone lenses are occasionally included in sheet

lavas (Fig. 2-9). Gabbro dykes intrude into basement lavas.

2-5. Petrography

Most lavas of MVG are aphyric. Even so phenocrysts occur in the thin sections, these are rare (less than about 5 vol% in mode). Groundmass assemblage is plagioclase + clinopyroxene + Fe-Ti oxide $\pm$ glass. Sulfide often occurs. The groundmass texture of chilled margin is hyaline with skeletal crystals. Toward to core, they generally transit from quench variolitic (Fig. 2-10a) or subvariolitic texture through hyalocrystalline intersertal (Fig. 2-10b) or intergranular texture to subophitic or holocrystalline ophitic texture (Fig. 2-10d). It often appears that subophitic or ophitic parts are bonded by interstitial glasses or intersertal parts in each thin section (Fig. 2-10c). Ophitic clinopyroxenes are pale brown. Phenocryst assemblages are olivine, olivine + plagioclase, olivine + plagioclase + clinopyroxene, plagioclase + clinopyroxene, and plagioclase. These indicate that olivines have been firstly crystallized, followed by plagioclase crystallization and lastly clinopyroxene have been formed. Lavas of MVG vary in the granularity of groundmass. Olivine phenocrysts are almost euhedral or subhedral (Fig. 2-11a). They are in size from 300 μ m to 800 μ m in diameter. But they were totally replaced by the clay minerals. Plagioclase phenocrysts are angular or orthophiric (Fig. 2-11b). They are almost in size from 300 μ m to 800 μ m in long side. The maximum size is up to 2 mm in long side. Clinopyroxene phenocrysts are euhedral or subhedral (Fig. 2-11b). They are almost in size from 300 μ m to 500 μ m in diameter. The maximum size is 800 μ m in diameter. They are colorless. These phenocrysts occasionally form glomeroporphyritic texture (Fig. 2-11a). Glasses are totally replaced by clay mineral. Occasionally, quartz, calcite or prehnite veins appear. Amigdules are filled with Na-zeolite, calcite or clay minerals. The grade of alteration or metamorphism is not related with stratigraphy.

Intrusive gabbros in the central Malaita have generally equigranular or ophitic texture (Fig. 2-12a and b). Their mineral assemblage is plagioclase + clinopyroxene + Fe-Ti oxide $\pm$ orthopyroxene $\pm$ sulfide. In only one sample, poikilitic olivines are occurred (Fig. 2-12c). Their rims were replaced by clay minerals but cores are preserved well. These olivines have often melt inclusions and no Cr-spinel.

Basement lavas in northwestern Santa Isabel are in petrography similar to

those of MVG. But they are highly altered than those of MVG. On the other hand, intrusive gabbros in the area are different from those in central Malaita (Fig. 2-12d). Their mineral assemblage is plagioclase + amphibole + clinopyroxene + Fe-Ti oxide. Amphiboles and clinopyroxenes are in form angular. Amphiboles are reddish brown and clinopyroxenes are pale pink. These petrographical features correspond to those of Sigana Alkalic Suits termed by Tejada et al., 1996.

2-6. Bulk Chemistry

All the samples are carefully collected to winnow the fresh parts in the field. Furthermore, analyzed samples were selected after detailed observation on rock slabs and thin sections in order to exclude highly altered samples. Most samples from central Malaita are over 98 wt % in total of major elements composition (Table 2-1) and only a few samples have comparatively scattering high K_2O and Na_2O contents (Fig. 2-13). The grade of alteration corresponds to those of samples measured by Tejada et al., 2002. The samples from northwestern Santa Isabel are rather altered than those of central Malaita (Fig. 2-13). Unless selected carefully, some of samples are less than 95 wt % in total of major elements compositions. These are more scattered in K_2O and Na_2O contents. These are consistent with the petrographical observation. The samples that are over 97 wt% in total of major elements composition and have no scattering signatures in Na_2O contents are available to followed discussion. These are also mainly based on the immobile elements (e.g. Bienvenu et al., 1990).

2-6-1. Central Malaita

Distribution of plots of the MVG from central Malaita on MgO vs TiO_2 contents diagram is not one narrow trend and unexplainable the one kind of mineral fractionation (Fig. 2-14). It suggests that they are evolved from more than two different magmas. Most of samples are plotted on the fields of Singgalo- and Kwaimbaita-type basalts. The plots of aphyric and fine-grained ones on and near the fields are clearly divided into two trend. One has relative low TiO_2 and low trend, the other has relative high TiO_2 and slightly steep trend. Each corresponds to Kwaimbaita- and Singgalo-type. Although the coarse-grained samples have more scattered, they also clustered in two

fields. Here, each is named Low-TiO₂ and High-TiO₂ type lava because Kwaimbaita- and Singgalo-type should be strictly labeled by obviously distinguishable isotopic compositions. Plotted on the transitional zone, these types are determined by the comparison with the samples from same or near stratigraphic layer. Plots departed from the fields are labeled by correcting presumable mineral accumulation. Some of the plots disengaged from the field and formed the flat trend at high MgO contents are not explainable the mineral accumulations. The trend is not to be explainable by mineral accumulation because aphyric one occurs on the trend and no presumable accumulation exists. The samples on the trend are different type magma from clustered ones similar to Singgalo- and Kwaimbaita-type lavas. Here, they are termed High-MgO type.

Other major elements variations against MgO are shown in Fig. 2-15. Although ambiguously, the plots are divided to three type groups that are High-TiO₂ type, Low-TiO₂ type, and High-MgO type. High-TiO₂ type is higher in P₂O₅ and Fe₂O₃* (total Fe represented as Fe₂O₃) contents and lower in Al₂O₃, CaO contents at given MgO contents than Low-TiO₂ type. These correspond to Singgalo-type like, Kwaimbaita-type basalts as same as those of TiO₂ contents, respectively. As exception of the two groups, the High-MgO type has picrobasaltic compositions that are characterized by low-SiO₂ (43 – 47 wt %). These have relative high-TiO₂, Fe₂O₃*, P₂O₅ and low-Al₂O₃ contents to other two groups at given MgO contents. At Site 1185 and 1187 of ODP LEG192, high-Mg basalts were found and named Kroenke-type basalt after the flanks of the submarine Kroenke Canyon where these sites are positioned (Fitton and Godard, 2004). They have Kwaimbaita-type basalt like isotopic feature and are inferred to be parental magma of Kwaimbaita-type basalt. High-MgO type lavas from central Malaita are lower in SiO₂ and Al₂O₃ contents and higher in TiO₂, Fe₂O₃*, and P₂O₅ contents than Kroenke-type basalt. Tejada et al., (1996) reported one sample that is high-Mg basalt with singgalo-type isotopic feature from Southern Malaita. It has higher SiO₂ and Al₂O₃ contents and lower TiO₂, Fe₂O₃*, and P₂O₅ contents than those of High-MgO type lavas from central Malaita. These High-MgO type lavas are distinguished from all other type OJP basalts that have been discovered both on land and submarine. High-MgO lavas are similar to other type in petrography, although they are slightly olivine-phyric and higher in intersertal Fe-Ti oxide.

REE and trace element compositions from the MVG show in Table 2-2 and 2-3, respectively. Lavas from the MVG have the REE patterns from flat to slightly

enriched (Fig. 2-16a). High-TiO₂ type lavas have relative high REE contents to those of Low-TiO₂ type (Fig. 2-17). These also correspond in REE characteristics to Singgalo- and Kwaimbaita-type basalts, respectively. High-TiO₂ type lavas have slightly more LREE enriched pattern and higher (La/Yb)_N ratios than those of Low-TiO₂ type. High-MgO type lavas have similar REE patterns to the High-TiO₂ type. The abundances are also similar to those of High-TiO₂ type and distinguished from Kroenke-type basalt (Fig. 2-17). Trace element relative abundance patterns to primitive mantle values (Sun and McDonough, 1989) of lavas from the MVG are shown in Fig. 2-16b. Rb and Th are more scattered than other elements and Sr disarray the patterns. These elements are mobile during the hydrothermal alteration (e.g. Bienvenu et al., 1990). Excepting these elements, the patterns for all types are roughly flat. High-TiO₂ type lavas have relative high trace element contents to those of Low-TiO₂ type (Fig. 2-18a). The variations of Nb/Zr between the types are more similar values than those of Ce/Yb that is the ratio of relatively low incompatible and influenced by whether present or absent of garnet (Fig. 2-18b). These also correspond in trace element features to Singgalo- and Kwaimbaita-type basalts, respectively, although Low-TiO₂ type lavas are slightly lower in Nb contents than Kwaimbaita-type. High-MgO type lavas have similar trace element patterns and abundance to the High-TiO₂ type. MgO variation diagrams of compatible trace elements are shown in Fig. 2-19. High-MgO type lavas are distinguished in incompatible trace element compositions from Kroenke-type basalt. High-TiO₂ type lavas have lower Ni and Cr contents than Low-TiO₂ type at given MgO contents (Fig. 2-19). These features are compared with the relations of Singgalo- and Kwaimbaita type basalts. Although their high MgO contents, High-MgO type lavas have lower Ni and Cr contents than Low-TiO₂ type lavas and similar to High-TiO₂ type. The feature is distinguished from those of Kroenke-type basalt.

Major element compositions of intrusive gabbros from are rather scattered than those of lavas (Fig. 2-20). The characteristics are explained by petrographically appropriate mineral accumulations on the lava compositions. They have flat REE patterns similar to lavas of MVG (Fig 2-16).

2-6-2. Geochemical stratigraphy of Central Malaita

The sampling points of each type lava are plotted on the columnar section (Fig.

2-21). Previous works simply suggested that the upper several 100-meter portion comprises Singgalo type basalt, whereas the lower portion is composed of Kwaimbaita type basalt based on the geochemistry of the upper 4000-meter portion of Malaita (Tejada et al., 2002). Our geochemical stratigraphy suggests that this section is not comfortable and simple across the island. The top unit has High-MgO type lavas overlain by cap carbonate. High-TiO₂ type lavas are exposed under the High-MgO type lavas section. Between High-MgO and High-TiO₂ type lavas, not fault or thick sedimentary rock, but thin volcanoclastite layer is interbedded. It indicates that High-MgO type lavas are not generated other hotspot. Below units have both High-TiO₂ type and Low-TiO₂ type lavas. High-TiO₂ type lavas are always superposed on Low-TiO₂ type lavas at any unit. In the many cases, the thin sedimentary rock layers are exposed between different type lavas as same as site 803 of ODP Leg130 (Mahoney et al., 1993). As unit is lower, occupation of High-TiO₂ type lavas decrease and Low-TiO₂ type lavas increase. The lowermost unit is only Low-TiO₂ type. The repeating signature suggests that shortened by the compression stress when the OJP accreted.

2-6-3. Northwestern Santa Isabel

Igneous rocks of Santa Isabel relative highly altered to the MVG (Fig.2-13), they are not available to detailed discussion. Although some basement lavas from northwestern Santa Isabel are similar to the Singgalo- or Kwaimbaita- type basalts in TiO₂ contents, Other have rather high TiO₂ contents (Fig. 2-22). These correspond to the High Ti Sigana Basalt termed by Tejada et al., 1996.

Intrusive gabbros in northwestern Santa Isabel are obviously different in major element and REE compositions from those lavas and lavas and intrusive gabbros of the MVG. They have high TiO₂ and P₂O₅ contents and highly LREE enriched and HREE depleted patterns (Fig. 2-16). These correspond to the Sigana Alkalic Suites termed by Tejada et al., 1996. It is consistent with the petrographical features that they include igneous amphibole and pale pink clinopyroxene (Fig. 2-12d). Sigana Alkalic Suites intruded in 90 Ma and have HIMU signature in Pb isotope compositions.

2-7. Mineral Chemistry

2-7-1. Minerals in lavas from the MVG

Clinopyroxene phenocrysts from High-TiO₂ and Low-TiO₂ type basalts are Wo₃₇₋₄₀En₅₀₋₅₂Fs₁₀₋₁₁ and Wo₃₆₋₄₁En₄₇₋₅₃Fs₉₋₁₂ augite, respectively. These from High-TiO₂ and Low-TiO₂ lavas are overlapped these components. But these from Kwaimbaita-type lavas have more diversity.

Al and Ti versus Mg# diagrams of clinopyroxenes are shown on Fig. 2-23. Mg# of clinopyroxene phenocrysts cores are similar that around from 80 to 84. Mg# and Al contents of groundmass clinopyroxenes from High-TiO₂ type lavas have steep negative correlation at higher Mg# (>75) and scattering positive correlation at lower Mg# (<75). Ti contents of groundmass clinopyroxenes from High-TiO₂ lavas have steep negative correlation at higher Mg# (>75) and scattering moderate positive correlation at lower Mg# (<75). On the other hand, Al contents of groundmass clinopyroxenes from Low-TiO₂ type lavas have steep negative correlation at higher Mg# (>78) and moderating negative correlation at lower Mg# (<78). Ti contents of groundmass clinopyroxenes from Low-TiO₂ type lavas have steep negative correlation at higher Mg# (>78) and moderate positive to flat correlation at lower Mg# (<78). Those correlation is reflected the coexisting melt compositions. These general correlations of Low-TiO₂ type lavas are similar to those of High-TiO₂ type lavas but more depleted in Al at crest on the trend and in Ti at lower Mg#. Differences of those are mineral fractionation. But more depleted in Ti contents at lower Mg#. Groundmass clinopyroxenes of Low-TiO₂ type lavas have also minor flat trend. This trend is explainable by the magma mixing and refreshing evolved magma.

The plagioclase compositions from both of High-TiO₂ and Low-TiO₂ type lavas are similar (Fig. 2-24). Plagioclase phenocrysts have slightly higher An contents (75-83%) in core. Most of the plagioclase phenocrysts have normal zoning. The groundmass plagioclase are lower in An contents than the phenocrysts cores.

2-7-2. Minerals in intrusive rocks from the MVG

Olivine from olivine gabblos have moderate Forsterite value (Fo#). Fo vs NiO and CaO variations are shown in Fig. 2-25. These have negative trend. These indicate that olivine crystallised from magma.

The composition of the pyroxenes from intrusive rocks were plotted on Di-En-Fs-Hd quadrilateral diagram (Fig. 2-26). Pyroxenes from poikilitic olivine gabbro and equigranular gabbros have slightly higher (~1200 °C) than doleritic gabbro (~1100°C).

Correlations of Mg# of clinopyroxenes and An contents of plagioclase coexisted with these is shown in Fig. 2-27. The correlation is concerned with the water contents of magma. Hydrous magma prefers to crystallize the high An contents plagioclase. Consequently, the An contents of plagioclase keep high, while Mg# of clinopyroxene decrease. Pairs of clinopyroxene and plagioclase from olivine gabbro has steep positive trend in the diagram and those from doleritic gabbro moderate. These trend similar to Hess deep and MAR, respectively (Bloomer et al., 1995; Dick and Natland, 1996). Some plots that departed from main trend have higher An content in plagioclase. They suggest that water contents are heterogeneous in magma chamber. It is possible that the enrichments of water occur in the mushy boundary layer along the wall (Sano and Yamashita, 2004).

2.8. Olivine-hosted melt inclusions from the olivine gabbro

2-8-1. Petrography

All the olivine-hosted melt inclusions are overall shape (Fig. 2-28). They vary in size from 20 to 100 μm . They have large void or high porous. They vary in assemblage of daughter mineral phases. These are divided to two groups that are amphibole-bearing and amphibole-free. Amphibole-bearing melt inclusions are dominant. Amphibole-bearing inclusions vary in daughter mineral assemblage. Its assemblage is amphibole + clinopyroxene + orthopyroxene $\pm$ sulfide $\pm$ Fe-Ti oxide $\pm$ apatite $\pm$ titanite. Amphiboles and Clinopyroxenes are euhedral or subhedral. Melt inclusions are fringed with orthopyroxene like reaction rim.

In contradiction to it, Amphibole-free inclusions are plagioclase + clinopyroxene $\pm$ orthopyroxene $\pm$ Fe-Ti oxide $\pm$ sulfide. The minerals are same as found in host rock.

2-8-2. Daughter mineral chemistry

Clinopyroxenes in melt inclusions are lower in Mg# than those from host rock. The clinopyroxene of amphibole-bearing melt inclusions have higher Al and Ti contents than those from host gabbro at given Mg# (Fig. 2-29). All clinopyroxenes are reverse zoning (Fig. 2-30).

The cores of amphiboles are enriched in Ti contents than rims. These features are similar to those of clinopyroxene. Amphibole is also reverse zoned (Fig. 2-31a). Unless Ti contents are enriched like alkali basalts, potassium contents related sodium is very low (Fig. 2-31b). Because no other daughter minerals are compatible, the K/Na of amphibole is directly reflected the magma composition.

Aluminous phases in melt inclusions are plagioclase and spinel. Plagioclase in amphibole-bearing melt inclusions are enriched in An contents than those of amphibole-free (Fig. 2-32a). Cr# of spinels in the melt inclusions are close to 0% (Fig. 2-32b).

Sulfides in melt inclusions are monosulfide solid solution (MSS) contrast to groundmass pyrite (Fig. 2-33). Carbonates in melt inclusions are Fe-rich.

Melt inclusions are crystallized in close-system. Volatiles are concentrated to melt, as host olivines overgrow on the wall of melt inclusions. It is not uncommon that melt inclusions have unusual crystal fractionation; e.g. amphiboles occur in melt inclusions from amphibole-free host rock (Schiano, 2003). But some of these features are not explained by the closed-system. Although the uncertainty of modal percentage are seemed to be large because the isotropy of daughter phase distribution is not assured, carbonate occurrences up to 8 mode% in melt inclusions (Fig. 2-26) require 2-3 wt% CO₂. Plume-related tholeiitic basalts (Kilauea, Reunion) are contained as much as 7000 ppm CO₂ (Bureau et al., 1999; Gerlach et al., 2002). It is unrealistic that wall olivine overgrowth accepts responsibility of over twice concentration of CO₂. The olivine caught the primary CO₂-rich melt. The petrographical feature that they have large void or high porous is consistent with it. Daughter amphiboles and clinopyroxene have higher Al and Ti contents in especially these cores. Although their reverse zoned compositional patterns are hardly explainable, their high-Ti and high-Al features are comparable with those of amphibole and clinopyroxene generated from SiO₂ deficient alkalic melt. It is inconsistent with the existence of orthopyroxene. But the concentrations of water and SiO₂ contents in melt inclusions accelerated by wall olivine overgrowth make orthopyroxene easier to be formed. Because host gabbro has tholeiitic signature relative Ti-poor pyroxenes (Fig. 2-29) and flat REE patterns (Fig. 2-16), the

melt in the magma chamber is heterogeneous in major element compositions. The magma that have alkaline feature may be minor and diluted by homogenizations in magma chamber. Although, the trace element and REE compositions were not determined, major elements compositions of trapped melts inferred by daughter mineral compositions is similar to High-MgO type lavas.

2-9. Discussion

2-9-1. Petrogenesis of the High-MgO type lavas

High-MgO type lavas that have firstly reported here have unusual geochemical features. Major element compositions indicate SiO_2 -deficient alkali signatures, although REE and trace elements are not strongly enriched. Despite the their MgO-rich signature, they are depleted in Ni and Cr that are compatible elements. High-MgO type lavas are geologically related with High- TiO_2 type lavas. High- TiO_2 type lavas are exposed under the High-MgO type lavas section. Boundary between High-MgO and High- TiO_2 type lavas are not disturbed by fault or thick sedimentary rock. It indicates that High-MgO type lavas are not generated at other hotspot but related with the formation of the OJP. Major element compositions of High-MgO type lavas have a trend that one end member is High- TiO_2 type lavas. The simulation of mineral fractionation from the relative primary compositions of High-MgO type lavas by MELTS program (Ghiorso and Sack, 1995) is shown in Fig. 2-34. The trend of the High-MgO type lavas is not explainable by the fractionation in crust (1kbar and 10kbar). It indicates that High-MgO type lavas are not parent magma of High- TiO_2 type. The trend is mixing with High- TiO_2 type.

Major element compositions of High-MgO type lavas are compared with high-pressure melting experiments. The partial melt from anhydrous peridotite is higher in SiO_2 and lower in FeO^* (total Fe as FeO) than High-MgO type lavas (e. g. Walter, 1998; Hirose and Kushiro, 1993; Kushiro, 1996). Eclogite (MORB pyroxenite) melting generates the andesitic melt and it is distinguished from High-MgO lavas (Fig. 2-35, e. g. Pertermann et al., 1998). Carbonated peridotite partial melt is higher in MgO contents than High-MgO lavas (Fig. 2-35, e. g. Hirose, 1997). Partial melts from peridotite + basalt mixture and SiO_2 -deficient pyroxenite (MIX1G) are roughly explainable to major

element compositions of High-MgO lavas (Fig. 2-35, Kogiso et al., 1998; Kogiso et al., 2003; Hirshmann et al., 2003). The Ni- and Cr- deficient signature of High-MgO type lavas indicates that olivine in source regions are very poor or absent. These suggest that MIX1G clinopyroxenite is the most plausible source. MIX1G partial melting at relative lower pressures (2.0 and 2.5 GPa) is more similar in major element compositions to High-MgO type lava than at higher pressure (5.0 GPa).

The REE patterns of partial melt from SiO₂-deficient clinopyroxenite are calculated as mentioned below. Garnet pyroxenite xenoliths in alnöite from Malaita are selected to using as source. The REE and mineral modal compositions data are referred to Ishikawa et al. (2004). Selected ones include no other mineral and their modal compositions 35-50 % garnet and 50-65 % clinopyroxene. The bulk REE compositions of source garnet pyroxenite are calculated by these modal compositions and REE contents in each mineral. Partition coefficients of REE for garnet and clinopyroxene are from Pertermann et al. (2004). The REE compositions of partial melts from these sources are calculated by the melt fraction and modal compositions of solid phase of partial melting experiments of MIX1G at moderately high-pressure (2.5 GPa, Hirschmann et al., 2003), assuming the batch melting. Although these have slightly heavy REE depletions or convex shape, these are roughly flat and similar to those of High-Mg type lavas (Fig. 2-36).

20-60 % partial melting of SiO₂-deficient clinopyroxenite at 1450-1500°C and moderately high-pressure (2.5GPa) are more explainable of the feature of major element, especially SiO₂-deficient and FeO*-rich, compatible trace elements and REE compositions of High-MgO type lavas than other candidate.

2-9-2. The mantle plume generating the OJP basalts

The petrogenesis of High-MgO type lavas indicates that mafic lithology is involved in source region. The garnet clinopyroxenite xenoliths that are assumed source for REE examinations are refractory of eclogitic pyroxenite after extraction of SiO₂-rich melt (Ishikawa et al., 2004). It is consistent with the “heterogeneous plume model” that mantle plume contains the ancient oceanic crust (Hauri, 1996; Takahashi and Nakajima, 2002). On the other hand, they were generated by 30% partial melting of primitive mantle in homogeneous plume at low pressure (Tejada et al., 2002; Fitton and Godard,

2004). This model has advantage to explain the rather homogeneous signatures and flat trace pattern. Isotopic differences between Kwaimbaita- and Singgalo-type are depended on a little difference of source and differences of trace and REE abundance are on different degree of partial melting. But High-MgO type lavas that have distinguishable major element compositions from other type lavas are not explained by homogeneous plume model. Also impact model that the OJP is formed by melting homogeneous lithosphere is not able to explain High-MgO type lavas. If ancient oceanic crust with basaltic compositions is involved in the mantle plume ascending along an adiabatic decompression path, then it begins to melt at greater depths than surrounding peridotite (Fig. 2-37, Takahashi et al., 1993; Yasuda et al., 1994). Even if the P-T conditions of adiabatic path are below the peridotite solidus, the basaltic materials will be almost completely molten. However, the eruption of the magma may not be possible until the P-T conditions exceed the peridotite solidus, because the totally molten basalt cannot diffuse effectively through matrix of subsolidus peridotite (Takahashi and Nakajima, 2002). In the case that the mantle plume reaches the peridotite progresses, it is possible that the imprint of entrained basaltic material will not be retained by the residual phases of the ambient peridotite. In this respect, the composition of the erupted magma is expected to be close to a mixture of totally molten basalt and a partial melt of peridotite. Geochemical differences between Kwaimbaita- and Singgalo- type is depended on the degree of contribution of peridotite component. These ancient crust component should have flat REE patterns. But the melt from basalt/peridotite sandwich melting experiments at 1400-1450°C and 2.7 GPa is alkaline picritic compositions (Takahashi and Nakajima, 2002). It is depended on the melt fraction and garnet existence. If achievement of equilibrium with peridotite reaches the relative small area to the amount of melt, the melt fraction make large and melt keep low incompatible element contents. A large amount of the OJP basalts indicated that the assumption is not worse. On the other hand, Based on the basalt/peridotite sandwich melting experiment at 1300°C and 3.5 GPa, the SiO₂-rich partial melt is ephemeral and reacted with peridotite (Yaxley and Green, 1998). In the result, eclogite + peridotite assemblage are changed to garnet clinopyroxenite + garnet orthopyroxenite. Garnet clinopyroxenite are not more melted at MORB solidus because it has higher solidus than MORB and slightly lower solidus than peridotite. Garnet clinopyroxenite in alnöite from Malaita are decompressed along the PMT = 1350-1450°C path and start remelting at 2-2.5 GPa.

These melts correspond to High-MgO lavas. The geochemical stratigraphy suggests the later erupted magma generated at lower PMT. But Ar dating of the Kwaimbaita- and Singgalo- type basalts are not distinguishable (Tejada et al., 2002).

2-10. Conclusions

On the basis of major element compositions, the OJP lavas from central Malaita are divided into three types; High-TiO₂ type, Low-TiO₂ type, and High-MgO type based on the previous classification that is Singgalo- and Kwaimbaita- type basalts. High-TiO₂ type lavas are higher in TiO₂, Fe₂O₃*, P₂O₅, and REE and lower in Al₂O₃ and CaO contents at given MgO than Low-TiO₂ type. Both have flat REE patterns, although high-TiO₂ type lavas have slightly LREE enriched patterns. High-TiO₂ type and Low-TiO₂ type lavas correspond in trace element and REE compositions to Singgalo- and Kwaimbaita-type basalts distinguished by previous study, respectively. High-MgO type lavas have high MgO, Fe₂O₃*, TiO₂, P₂O₅ and REE and low SiO₂ and Al₂O₃ contents. They have similar trace and REE compositions to High-TiO₂ type. The feature is compared with those of no other OJP basalts. The stratigraphic relation of these type lavas are ordered as High-MgO, High-TiO₂, and Low-TiO₂ type lavas beginning at the top although all the sequence is not conformably exposed because they are disturbed and repeated by the faults. High-MgO lavas are comparable with no melting experiment of peridotite and peridotite and basalt mixture at high- and low-pressure.

Daughter minerals in olivine-hosted melt inclusions have anomalous features compared with host olivine gabbro. Amphibole-bearing melt inclusions are dominant unless it is absent in host olivine gabbro. Daughter amphiboles and clinopyroxenes are higher in Fe, Al and Ti. They are reverse zoned. A few globules have carbonate. Although the petrological features of melt inclusions are modified by complicated process of closed-system after trapped, the core compositions of mafic minerals suggest that they are generated from SiO₂ deficient melt like High-MgO type lavas.

High-MgO type lavas are comparable with no melting experiment of peridotite and peridotite and basalt mixture at high- and low-pressure. The high-pressure melting of garnet clinopyroxenites is roughly explainable for major, trace element and REE features of High-MgO type lavas. The most reasonable source of High-MgO type lava is refractory of eclogitic recycled oceanic crust extracted SiO₂-rich melt. High-MgO type

lavas infer that the plume generating OJP include the ancient oceanic crust.

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Table 2-1. Major element compositions of igneous rocks from MVG

sample	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	total
SN001	43.75	1.86	12.68	16.91	0.23	11.01	12.94	1.72	0.09	0.16	101.35
SN002	44.55	1.8	13.3	15.81	0.19	11.32	12.75	1.65	0.09	0.16	101.62
SN003	44.29	1.62	13.48	14.76	0.2	10.08	14.26	1.6	0.1	0.15	100.56
SN004	44.45	1.72	13.5	15.29	0.19	10.54	12.35	1.75	0.09	0.16	100.03
SN005	44.53	1.73	13.46	15.23	0.22	9.92	13.49	1.72	0.02	0.16	100.48
SN009	45.56	1.67	13.37	15.41	0.19	9.24	12.92	1.82	0.11	0.15	100.43
SN010	45.69	1.66	13.34	15.08	0.18	9.31	12.23	1.8	0.12	0.15	99.57
SN011	45.82	1.61	13.57	14.69	0.18	9.17	12.5	1.8	0.08	0.14	99.58
SN012	46.12	1.7	13.56	14.59	0.2	8.66	12.3	1.85	0.06	0.16	99.2
SN015	46.84	1.59	13.93	13.91	0.18	8.23	12.03	1.9	0.11	0.15	98.86
SN016	46.83	1.75	14.44	13.02	0.22	7.99	12.23	2.1	0.08	0.16	98.82
SN018	48.15	1.78	14.61	12.71	0.19	6.46	12	2.17	0.14	0.16	98.36
SN019	48.33	1.71	14.17	13.24	0.21	6.22	12.12	2.18	0.14	0.16	98.46
SN022	48.52	1.7	14.32	13.8	0.22	7.12	10.82	2.18	0.14	0.15	98.98
SN023	48.66	1.62	13.94	13.46	0.18	7.15	11.36	2.14	0.14	0.15	98.78
SN024	48.16	1.63	14.37	11.71	0.19	6.86	12.74	2.19	0.15	0.15	98.15
SN025	49.37	2.61	12.4	19.43	0.26	4.56	8.95	2.6	0.26	0.24	100.67
SN027	49.02	1.68	14	14	0.2	7.19	11.25	2.05	0.09	0.16	99.63
SN029	48.56	1.62	14	14.28	0.21	6.95	11.73	2.05	0.04	0.15	99.59
SN030	48.75	1.65	13.83	14.31	0.17	7.24	10.91	2.06	0.12	0.14	99.18
SN031	48.29	1.59	13.81	13.57	0.18	7.19	11.24	2.08	0.11	0.14	98.2
SN032	48.04	1.65	14.18	14.43	0.22	6.41	12.11	2	0.03	0.15	99.22
SN033	48.67	1.64	13.89	13.8	0.17	7.21	11.04	2	0.14	0.14	98.71
SN034	48.89	1.65	14.04	13.87	0.21	6.65	11.61	2.06	0.1	0.15	99.22
SN035	48.42	1.62	13.91	13.79	0.2	6.87	11.48	2.05	0.13	0.14	98.61
SN037	48.96	1.64	13.85	14.12	0.2	6.7	11.53	2.14	0.14	0.15	99.43
SN040	49.34	1.54	14.37	13.08	0.21	6.61	12.23	2.15	0.12	0.14	99.79
SN042	48.48	1.69	14.54	13.04	0.23	7.08	11.65	2.25	0.14	0.15	99.27
SN043	48.41	1.45	14.28	13.38	0.2	7.28	12.07	1.95	0.11	0.12	99.25
SN044	49.2	1.55	14.19	13.7	0.24	5.96	12.2	2.02	0.08	0.14	99.27
SN046	48.6	1.51	13.96	13.48	0.21	6.9	11.91	2.01	0.11	0.13	98.83
SN047	47.98	1.17	13.86	11.81	0.17	7.92	11.08	3.04	0.13	0.11	97.28
SN049	48.11	1.2	14.31	11.84	0.22	8.85	12.18	1.88	0.06	0.11	98.77
SN051	47.91	1.22	14.42	12.5	0.18	7.52	13.08	1.71	0.02	0.11	98.67
SN052	49.07	1.19	13.81	12.13	0.17	7.81	10.88	3.47	0.14	0.1	98.78
SN054	47.67	1.2	13.47	12.29	0.18	8.06	11.3	3.07	0.1	0.11	97.44
SN055	48.49	1.18	14	11.53	0.22	8.28	12.35	2.11	0.38	0.12	98.67
SN058	49.53	1.63	14.22	13.68	0.22	6.36	11.16	2.22	0.17	0.17	99.36
SN059	48.52	1.19	13.93	11.89	0.21	8.09	11.57	2.78	0.13	0.11	98.43
SN060	48.72	1.19	13.83	11.87	0.21	8.35	11.6	2.41	0.41	0.12	98.69
SN061	48.6	1.63	14.02	13.84	0.21	6.52	11.22	2.1	0.14	0.15	98.44
SN062	48.47	1.68	14.39	13.82	0.21	7.3	11.66	2.19	0.03	0.15	99.91
SN064	49.25	1.66	14.12	13.89	0.22	6.78	11.37	2.13	0.13	0.15	99.69
SN065	49.29	1.6	14.4	13.52	0.2	6.68	11.55	2.13	0.12	0.15	99.64
SN066	49.07	1.66	13.99	14.21	0.21	6.56	11.6	2.12	0.11	0.15	99.68
SN067	48.82	1.64	13.97	14.16	0.22	6.31	11.56	2.16	0.14	0.15	99.13
SN068	48.59	1.52	14.12	13.5	0.2	7.17	11.89	1.98	0.11	0.12	99.2
SN069	48.81	1.5	14.14	13.71	0.21	6.32	12.05	2.05	0.07	0.14	98.98

Table 2-1 (continued).

sample	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	total
SN071	48.86	1.47	13.87	13.57	0.2	7.17	11.59	2.07	0.13	0.14	99.06
SN072	48.43	1.62	15.06	11.98	0.26	7.52	11.27	2.22	0.13	0.15	98.64
SN073	48.58	1.52	13.97	13.63	0.2	6.78	11.74	2.03	0.12	0.14	98.72
SN075	47.95	1.51	14	13.42	0.21	6.88	11.76	1.96	0.12	0.14	97.94
SN077	48.12	1.46	14	13.25	0.22	7.02	12.03	1.95	0.1	0.13	98.27
SN078	48.08	1.48	13.85	13.46	0.21	6.69	11.86	1.97	0.11	0.13	97.85
SN079	48.13	1.47	13.94	13.11	0.19	7.27	11.52	2.02	0.11	0.12	97.9
SN080	47.39	1.59	14.93	12.73	0.25	7.13	11.08	2.16	0.13	0.14	97.54
SN082	48.34	1.47	13.99	13.45	0.18	7.2	11.77	1.97	0.11	0.13	98.61
SN083	49	1.57	14.99	12.41	0.26	7.46	12.17	2.15	0.12	0.14	100.28
SN085	48.55	1.52	14.28	13.33	0.21	7.13	11.92	2.12	0.08	0.14	99.27
SN086	48.31	1.48	14.07	13.44	0.2	6.96	12.03	2.04	0.11	0.13	98.79
SN088	49.34	1.58	14.63	13.07	0.19	7.08	12.27	2.06	0.09	0.14	100.46
SN087	49.15	1.54	14.62	13.42	0.21	6.81	12.19	1.99	0.11	0.14	100.18
SN090	48.91	1.21	14.37	11.84	0.2	7.88	11.79	2.54	0.47	0.11	99.33
SN092	48.57	1.53	14.29	13.71	0.23	6.84	12.1	2.1	0.14	0.14	99.64
SN093	48.74	1.5	14.42	13.69	0.21	6.88	12.16	2.01	0.12	0.14	99.87
SN094	48.65	1.53	14.31	13	0.21	6.91	12.25	2.08	0.07	0.14	99.13
SN095	48.31	1.51	14.31	13.7	0.24	6.61	12.32	2.12	0.12	0.14	99.37
SN096	48.5	1.44	14.29	13.17	0.18	7.61	12.13	1.9	0.02	0.12	99.36
SN098	48.25	1.25	15.41	11.83	0.27	6.03	13.81	1.77	0.16	0.12	98.9
SN100	48.88	1.25	15.33	10.34	0.18	7.92	13.21	1.99	0.03	0.12	99.26
SN101	47.96	1.19	14.41	11.61	0.18	8.06	12.55	2.22	0.05	0.1	98.32
SN102	48.37	1.21	14.63	12.45	0.19	7.35	13.16	1.81	0.06	0.11	99.34
SN103	48.67	1.34	14.39	12.7	0.27	7.99	11.95	1.84	0.08	0.12	99.36
SN107	48.46	1.21	14.67	12.32	0.19	7.14	13.46	1.78	0.02	0.11	99.37
SN109	48.12	1.47	14.09	13.46	0.21	7.21	12.02	2.01	0.1	0.13	98.82
SN110	48.53	1.57	13.99	13.64	0.19	7.26	11.62	2.01	0.13	0.13	99.07
SN112	48.48	1.57	13.89	13.75	0.19	7.64	11.38	1.97	0.13	0.13	99.14
SN113	48.67	1.56	13.88	13.71	0.18	7.83	11.3	1.98	0.13	0.13	99.38
SN115	48.27	1.51	13.82	13.56	0.19	7.41	11.58	1.95	0.11	0.13	98.54
SN118	48.45	1.58	13.95	13.63	0.21	6.88	11.55	2.12	0.14	0.14	98.65
SN120	48.48	1.53	14.63	13.16	0.29	6.72	12.26	2.07	0.1	0.14	99.4
SN122	48.91	1.51	14.1	13.62	0.2	7.14	11.91	1.96	0.11	0.13	99.6
SN123	48.93	1.47	14.34	13.23	0.19	7.25	11.93	2.08	0.1	0.14	99.66
SN124	48.13	1.69	15.48	12.54	0.33	7.62	9.87	2.4	0.15	0.15	98.36
SN127	49.02	1.39	14.49	12.85	0.2	7.01	12.47	2.03	0.1	0.14	99.71
SN129	48.24	1.12	14.31	11.74	0.18	7.86	11.32	3.28	0.41	0.12	98.58
SN130	48.31	1.22	15.01	12.45	0.21	6.93	13.64	1.75	0	0.12	99.65
SN132	48.88	1.19	14.47	12.08	0.18	8.05	13.06	1.79	0.05	0.11	99.85
SN133	48.05	1.13	14.61	12.01	0.18	8.71	12.64	1.66	0.05	0.1	99.14
SN134	47.91	1.19	14.46	11.82	0.2	8.33	12.48	1.73	0.08	0.12	98.32
SN135	48.63	1.21	14.66	12.53	0.2	7.43	13.42	1.66	0.01	0.12	99.87
SN136	48.67	1.23	14.74	11.97	0.18	7.84	13.2	1.68	0.02	0.12	99.65
SN139	48.53	1.22	14.64	12.24	0.2	7.95	13.09	1.71	0.02	0.12	99.72
SN140	48.6	1.24	14.89	12.03	0.23	7.29	13.53	1.76	0.05	0.11	99.74
SN141	48.52	1.22	14.72	11.82	0.18	8.25	13.18	1.71	0.02	0.12	99.73
SN142	48.77	1.45	15.05	12.41	0.26	7.96	12.29	2	0.02	0.13	100.35

Table 2-1 (continued).

sample	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	total
SN144	48.55	1.23	14.76	12.43	0.21	7.12	13.47	1.74	0.04	0.12	99.67
SN143	49.12	1.29	15.35	10.98	0.31	7.11	13.22	2.38	0.02	0.12	99.89
SN146	47.52	1.2	14.63	12.34	0.17	8.02	12.25	2.5	0.1	0.11	98.84
SN147	48.51	1.19	14.56	12.01	0.17	8.51	12.67	1.68	0.03	0.11	99.43
SN149	48.52	1.24	14.89	11.48	0.23	8.28	12.65	1.79	0.05	0.12	99.25
SN150	48.8	1.16	14.4	12.14	0.17	8.14	12.86	1.72	0.07	0.11	99.57
SN151	48.25	1.19	14.38	12.32	0.18	7.71	13.11	1.71	0.04	0.11	99
SN152	48.76	1.2	14.79	12.29	0.22	7.31	13.41	1.8	0.04	0.11	99.93
SN153	48.74	1.31	14.31	12.89	0.2	8.01	12.15	1.78	0.08	0.12	99.6
SN155	48.64	1.32	14.22	12.89	0.2	8.12	11.92	1.8	0.08	0.12	99.32
SN157	48.56	1.4	14.27	12.98	0.21	7.33	12.12	1.85	0.08	0.13	98.92
SN159	48.04	1.37	14.27	12.81	0.2	7.78	12.18	1.74	0.01	0.13	98.52
SN161	48.3	1.5	13.99	13.75	0.21	6.96	11.89	2	0.12	0.14	98.85
SN162	44.99	1.16	14.9	9.88	0.16	6.75	13.71	2.78	0.31	0.11	94.75
SN164	48.67	1.49	13.94	13.53	0.2	7	11.96	1.96	0.11	0.14	99
SN165	48.87	1.44	14.36	12.99	0.2	7.01	12.22	2.14	0.11	0.14	99.47
SN166	48.7	1.59	14.98	12.81	0.32	7.13	12.17	2.1	0.12	0.14	100.07
SN168	48.57	1.6	15.1	12.49	0.28	7.14	12.42	2.14	0.06	0.15	99.95
SN170	48.49	1.61	15.1	12.25	0.34	7.5	11.65	2.23	0.12	0.14	99.43
SN172	48.61	1.6	15.26	12.15	0.3	7.49	11.79	2.21	0.12	0.14	99.67
SN174	48.6	1.58	15.27	12.52	0.33	7.09	12.31	2.13	0.09	0.14	100.06
SN175	48.91	1.39	14.68	12.47	0.22	7.61	12.3	1.94	0.08	0.13	99.74
SN177	49.22	1.51	14.6	12.85	0.22	6.89	12.15	2.14	0.12	0.14	99.83
SN178	49.47	1.56	14.84	12.98	0.26	7.25	12.23	2.24	0.1	0.14	101.08
SN179	48.43	1.67	15.81	12.95	0.37	8.1	10.34	2.24	0.06	0.15	100.12
SN180	49.13	1.52	14.46	13.4	0.3	7.35	12.18	2.07	0.1	0.13	100.64
SN182	49	1.47	14.33	13.36	0.21	7.13	11.88	2.11	0.12	0.13	99.74
SN183	48.74	1.48	14.38	13.04	0.2	7.25	12.03	2.01	0.1	0.13	99.35
SN186	48.59	1.54	14.47	13.49	0.23	6.75	12.12	2.09	0.11	0.14	99.53
SN189	48.89	1.54	14.16	13.38	0.2	7.15	11.92	2.06	0.12	0.14	99.57
SN190	49.37	2.09	12.61	16.59	0.26	6.36	10.24	2.19	0.16	0.16	100.03
SN191	49.2	1.54	14.27	13.31	0.23	7.19	11.86	2.09	0.11	0.14	99.94
SN192	48.72	1.52	14.09	13.54	0.21	7.11	11.96	2.05	0.1	0.13	99.44
SN193	48.65	1.35	14.28	12.84	0.18	7.77	11.84	1.85	0.08	0.13	98.96
SN194	48.53	1.29	14.32	12.77	0.19	8.02	12.01	1.83	0.07	0.12	99.15
SN195	47.88	1.45	15.01	12.39	0.27	7.84	11.94	1.87	0.02	0.14	98.8
SN196	48.25	1.36	14.27	12.76	0.2	7.66	12.08	1.88	0.08	0.12	98.66
SN198	48.31	1.2	14.48	12.27	0.21	7.44	13.11	1.7	0.05	0.11	98.88
SN199	48.42	1.19	14.53	12.25	0.21	7.47	13.1	1.76	0.05	0.11	99.1
SN200	48.36	1.2	14.38	12.29	0.19	7.33	13.15	1.75	0.06	0.11	98.83
SN201	48.14	1.19	14.4	12.28	0.18	7.51	13.03	1.75	0.07	0.11	98.66
SN202	48.29	1.22	14.81	12.06	0.18	7.37	12.96	1.81	0.08	0.12	98.89
SN203	47.85	1.27	13.6	12.56	0.19	7.87	10.76	3.18	0.72	0.11	98.08
SN204	48.08	1.17	14.37	12.25	0.18	7.99	12.64	1.63	0.21	0.11	98.62
SN205	48.05	1.2	14.46	12.32	0.21	7.32	13.16	1.73	0.04	0.11	98.6
SN206	48.54	1.22	14.9	12.39	0.23	7.07	13.42	1.73	0.01	0.11	99.6
SN208	48.3	1.32	14.42	12.54	0.22	7.8	11.99	1.82	0.08	0.12	98.62
SN210	48.07	1.2	14.62	12.04	0.24	7.5	13.26	1.72	0.01	0.12	98.77

Table 2-1 (cotinued).

sample	SiO2	TiO2	Al2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	total
SN211	48.12	1.21	14.83	11.83	0.27	7.28	13.4	1.78	0	0.11	98.85
SN212	48.02	1.21	14.62	12	0.22	7.47	13.17	1.82	0.01	0.11	98.65
SN213	48.06	1.22	14.65	11.66	0.24	7.88	13.07	1.71	0.01	0.11	98.61
SN214	48.02	1.2	14.56	12.42	0.21	7.27	13.19	1.79	0.01	0.11	98.8
SN215	48.04	1.2	14.52	12.1	0.2	7.38	13.23	1.7	0.01	0.11	98.5
SN218	48.07	1.66	13.86	14.46	0.22	6.74	11.79	2.08	0.04	0.15	99.06
SN219	49.5	1.71	14.42	13.2	0.2	6.73	11.22	2.18	0.15	0.15	99.46
SN220	48.71	1.64	13.84	14.17	0.17	7	11.28	2.03	0.07	0.14	99.06
SN221	48.8	1.66	13.88	14.38	0.2	6.78	11.63	2.08	0.04	0.15	99.59
SN222	49.59	1.61	14.08	13.39	0.17	7.22	11.18	2.1	0.14	0.14	99.6
SN224	49.09	1.56	14.13	13.51	0.2	6.72	11.72	2.04	0.13	0.14	99.26
SN225	48.94	1.69	13.66	14.11	0.18	7.13	10.71	2.06	0.12	0.15	98.74
SN228	49.54	1.71	14.08	13.67	0.19	6.73	11.3	2.17	0.13	0.16	99.68
SN229	48.65	1.7	14.5	13.74	0.19	7.34	11.33	2.05	0.06	0.15	99.72
SN230	49.39	1.67	13.97	13.62	0.18	7.32	10.85	2.08	0.12	0.14	99.33
SN231	48.78	1.56	14.1	13.02	0.19	7.02	12.04	2.01	0.13	0.14	98.99
SN233	48.96	1.63	13.74	13.98	0.16	7.41	10.34	2.06	0.14	0.14	98.55
SN234	49.07	1.58	13.75	13.68	0.17	7.45	10.85	2.03	0.14	0.14	98.85
SN235	49.11	1.58	14	13.43	0.2	7.75	11.07	2.11	0.13	0.13	99.52
SN236	49.16	1.61	13.95	13.29	0.18	7.44	11.07	2.1	0.12	0.14	99.06
SN237	48.94	1.63	13.76	13.93	0.18	7.22	10.75	2.07	0.15	0.14	98.77
SN278	48.73	1.61	13.95	13.73	0.18	7.15	11.38	2.05	0.1	0.14	99
SN279	48.57	2.04	12.77	16.12	0.25	6.14	10.23	2.22	0.17	0.17	98.67
SN281	48.17	1.21	14.43	12.41	0.21	7.53	13.13	1.75	0.01	0.12	98.96
SN282	48.06	1.17	14.64	12.14	0.18	7.79	13.01	1.73	0.07	0.11	98.91
SN284	47.81	1.16	14.94	10.91	0.17	8.83	12.77	1.57	0.05	0.11	98.32
SN285	48.08	1.24	14.91	11.51	0.2	7.56	13.55	1.71	0.03	0.12	98.91
SN286	48.17	1.18	15.22	10.84	0.18	8.38	12.84	1.63	0.09	0.11	98.64
SN287	48.5	1.23	15.01	11.65	0.24	7.89	13.07	1.75	0.04	0.11	99.5
SN288	47.96	1.24	14.44	12.08	0.18	8.26	11.77	2.29	0.06	0.12	98.41
SN289	48.28	1.24	14.61	12.21	0.16	8.48	12.36	1.78	0.06	0.12	99.3
SN292	48.97	1.21	14.45	12.16	0.18	8.09	11.31	2.59	0.07	0.11	99.14
SN293	48.27	1.21	14.6	11.95	0.17	8.03	12.48	1.59	0.04	0.12	98.47
SN295	47.36	1.25	14.49	12.11	0.18	8.03	11.56	2.62	0.08	0.12	97.8
SN297	47.84	1.29	14.95	12.44	0.22	7.02	13.33	1.59	0	0.12	98.82
SN299	48.8	1.21	14.4	12.2	0.22	8.23	11.9	2.38	0.02	0.11	99.48
SN300	48.06	1.26	14.08	12.14	0.2	7.93	12.01	2.37	0.07	0.12	98.24
SN301	48.09	1.28	14.96	12.18	0.19	7.37	13.31	1.6	0.01	0.12	99.11
SN303	48.88	1.25	14.16	12.23	0.24	7.75	12.13	1.74	0.04	0.12	98.54
SN306	48.85	1.25	13.95	12.07	0.18	8.13	10.89	2.83	0.14	0.12	98.41
SN307	47.59	1.31	15.51	12.21	0.18	6.18	13.74	1.64	0	0.12	98.48
SN309	44.97	0.72	12.57	12.77	0.19	16.3	9.33	0.94	0.04	0.08	97.92
SN310	48.57	0.73	17.12	8.22	0.14	8.06	14.52	1.4	0.09	0.08	98.92
SN311	48.85	0.79	16.2	8.05	0.14	8.2	14.39	1.88	0.05	0.08	98.63
SN312	48.49	0.8	15.66	9.07	0.15	8.66	14.19	1.41	0.06	0.08	98.56
SN313	48.57	1.87	12.45	15.9	0.23	5.82	9.88	3.02	0.2	0.17	98.1
SN314	50.09	1.81	12.65	14.7	0.21	5.72	9.2	3.73	0.16	0.18	98.45
SN315	49.91	1.04	14.06	9.93	0.16	7.81	12.74	3.04	0.13	0.09	98.92

Table 2-1 (cotinued).

sample	SiO2	TiO2	Al2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	total
SN318	50.84	1.97	12.19	16.35	0.26	5.41	8.14	3.89	0.19	0.21	99.43
SN319	50.51	1.33	13.83	13.64	0.2	6.4	10.47	2.96	0.14	0.11	99.59
SN317	51.2	1.1	13.33	10.85	0.18	7.46	11.46	3.13	0.36	0.1	99.18
SN320	47.69	2.29	11.95	18.55	0.25	4.51	9.72	2.4	0.08	0.21	97.66
SN321	49.24	1.25	14.07	12.37	0.19	7.42	11.43	2.65	0.14	0.12	98.88
SN322	48.86	1.09	14.68	11.48	0.19	8.46	10.75	3.01	0.05	0.1	98.66
SN323	47.4	1.22	14.26	12.68	0.19	8.54	11.02	2.59	0.03	0.12	98.06
SN325	48.95	1.18	14.73	12.11	0.17	8.09	12.4	1.72	0.06	0.11	99.52
SN326	48.83	1.28	14.92	12.32	0.21	7.32	12.91	1.82	0.05	0.12	99.76
SN327	48.34	1.25	14.69	12.24	0.23	7.95	13.06	1.75	0.03	0.12	99.66
SN329	48.71	1.21	13.74	11.37	0.17	8.03	10.56	3.5	0.11	0.11	97.49
SN331	48.81	1.26	14.88	12.31	0.18	7.52	13.1	1.72	0.04	0.12	99.93
SN332	48.07	1.28	14.84	12.46	0.21	7.11	13.13	1.72	0.02	0.12	98.95
SN334	48.13	1.24	14.15	11.88	0.19	8.38	12.19	1.84	0.08	0.12	98.18
SN335	48.52	1.16	14.61	11.8	0.18	8.72	12.31	1.66	0.09	0.11	99.15
SN337	48.73	1.19	14.87	11.4	0.2	8.66	12.55	1.76	0.04	0.11	99.51
SN338	48.37	1.19	14.53	11.68	0.18	8.04	12.73	1.69	0.1	0.12	98.62
SN339	47.64	1.19	14.96	11.74	0.17	7.81	12.48	2.18	0.09	0.11	98.37
SN341	48.68	1.27	14.88	12.13	0.18	7.56	13	1.78	0.08	0.12	99.66
SN342	48.68	1.16	14.52	11.66	0.16	8.8	11.88	1.83	0.09	0.11	98.9
SN345	48.63	1.19	14.63	11.7	0.17	8.46	12.5	1.79	0.15	0.11	99.32
SN346	48.68	1.21	14.8	12.49	0.22	7.64	13.44	1.72	0	0.12	100.32
SN348	48.03	0.91	15.52	10.5	0.16	9.28	13.45	1.51	0.03	0.09	99.47
SN349	48.29	0.98	15.51	10.57	0.17	8.72	13.26	1.9	0.04	0.09	99.53
SN350	48.72	1.31	12.28	13.29	0.21	8.14	13.04	1.51	0.05	0.11	98.64
SN351	48.85	1.3	12.35	12.98	0.21	8.09	13.04	1.49	0.06	0.1	98.47
SN352	49.15	1.5	13.5	13.33	0.2	7.02	11.88	2.93	0.17	0.11	99.79
SN353	47.57	1.73	13.14	15.96	0.23	6.28	10.25	2.35	0.05	0.13	97.68
SN354	48.7	1.24	14.59	12.47	0.19	7.41	13.24	1.78	0.06	0.12	99.8
SN355	48.64	1.18	14.38	11.98	0.17	8.04	12.94	1.71	0.04	0.11	99.19
SN356	48.31	1.22	14.6	12.41	0.18	7.47	13.16	1.75	0.05	0.12	99.27
SN362	48.47	1.55	14.23	13.65	0.24	7.16	11.84	2.03	0.11	0.13	99.39
SN363	48.35	1.48	14.19	13.49	0.2	7.16	11.98	1.95	0.09	0.12	99.02
SN365	47.92	1.82	13.35	15.55	0.21	7.62	10.23	2.03	0.12	0.15	99.01
SN366	48.39	1.64	14.43	13.01	0.19	7.23	11.47	2.14	0.14	0.14	98.76
SN367	48.85	1.77	13.77	14.19	0.2	6.4	11.31	2.17	0.14	0.15	98.95
SN368	48.72	1.65	13.9	13.81	0.18	7.02	11.11	2.07	0.14	0.14	98.74
SN369	48.73	1.56	13.91	13.6	0.21	6.9	11.6	2.12	0.14	0.14	98.91
SN370	48.6	1.61	13.96	13.71	0.2	6.83	11.5	2.05	0.12	0.14	98.72
SN371	49	1.68	14.03	13.72	0.21	6.82	11.43	2.12	0.12	0.15	99.29
SN372	48.94	1.62	14.02	13.74	0.18	6.89	11.55	2.07	0.12	0.15	99.28
SN373	49.27	1.68	14.37	13.15	0.24	6.54	12.01	2.12	0.06	0.15	99.58
SN374	49.08	1.63	13.98	13.93	0.2	6.57	11.8	2.09	0.1	0.15	99.53
SN376	48.72	1.61	13.92	13.83	0.19	6.62	11.65	2.03	0.14	0.15	98.86
SN377	48.75	1.4	14.55	12.78	0.18	7.44	12.02	1.97	0.1	0.12	99.33
SN378	49.39	1.37	14.9	12.12	0.18	7.21	12.2	1.99	0.11	0.14	99.62
SN379	48.98	1.46	14.37	12.54	0.2	7.47	12.09	2.01	0.11	0.12	99.36
SN380	49.24	1.47	14.18	12.86	0.19	7.22	11.92	2	0.09	0.14	99.31

Table 2-1 (cotinued).

sample	SiO2	TiO2	Al2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	total
SN382	48.84	1.62	14	13.9	0.2	7.1	11.28	2.02	0.13	0.14	99.24
SN385	49.87	1.66	13.9	13.5	0.18	7.1	10.74	2.02	0.13	0.14	99.23
SN386	49.62	1.7	14.06	13.41	0.2	6.77	11.29	2.27	0.12	0.15	99.6
SN387	49.18	1.44	14.04	12.31	0.18	7.49	12.08	1.97	0.08	0.12	98.89
SN388	48.93	1.66	13.82	13.65	0.16	7.34	10.6	2.07	0.14	0.14	98.51
SN389	48.65	1.06	14.03	9.95	0.17	8.47	14.13	1.64	0.03	0.09	98.21
SN390	48.28	1.2	14.45	11.53	0.18	7.8	11.37	2.92	0.19	0.1	98.02

Table. 2-2. Major element compositions of igneous rocks from western Santra Isabel

sample	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	total
SI001	49.95	1.19	14.91	10.14	0.31	6.32	7.96	3.54	0.55	0.12	94.98
SI003	49.30	1.30	14.79	10.80	0.16	5.78	10.14	1.99	0.12	0.13	94.51
SI004	49.27	1.20	14.37	10.46	0.16	6.64	10.27	2.13	0.14	0.12	94.76
SI005	50.03	1.98	15.71	10.09	0.17	7.13	10.29	3.04	0.51	0.22	99.19
SI007	51.68	2.39	13.00	10.73	0.22	6.06	10.23	4.47	0.31	0.22	99.31
SI010	47.41	2.18	13.51	12.90	0.22	7.08	10.27	2.44	0.08	0.21	96.28
SI011	47.44	2.21	13.71	13.29	0.23	6.99	10.32	2.28	0.21	0.21	96.89
SI012	47.31	2.19	13.76	13.26	0.24	6.92	9.85	2.72	0.10	0.21	96.55
SI013	46.92	2.20	13.65	13.16	0.21	6.80	10.63	2.07	0.31	0.21	96.15
SI014	51.16	1.70	14.69	12.90	0.21	4.36	7.92	2.28	1.19	0.17	96.59
SI015	48.22	1.37	14.83	10.80	0.18	7.26	10.45	2.37	0.13	0.13	95.74
SI017	48.48	1.53	14.72	10.95	0.33	8.59	9.74	3.59	0.02	0.13	98.08
SI018	47.86	1.52	14.66	11.20	0.38	8.56	9.40	3.85	0.03	0.13	97.59
SI019	48.58	1.52	14.39	11.01	0.25	8.24	9.03	4.31	0.01	0.13	97.48
SI020	46.99	1.47	14.45	10.98	0.34	8.50	9.30	3.89	0.03	0.13	96.09
SI021	47.85	1.50	14.68	11.03	0.36	8.50	10.17	3.21	0.02	0.13	97.46
SI022	47.77	1.59	14.70	10.72	0.55	8.35	10.06	3.75	0.02	0.14	97.65
SI023	50.10	1.56	15.03	12.62	0.15	5.71	8.82	3.00	0.19	0.14	97.33
SI024	49.52	1.57	15.17	12.39	0.17	5.93	8.94	2.97	0.23	0.15	97.04
SI025	49.80	1.53	14.99	12.23	0.15	5.81	9.04	2.70	0.21	0.14	96.60
SI026	49.76	2.61	16.18	10.39	0.26	3.62	6.49	4.27	3.85	1.08	98.50
SI027	48.86	2.56	16.19	10.65	0.19	3.50	7.22	3.75	4.36	1.05	98.33
SI028	50.01	2.61	15.95	10.42	0.29	3.66	6.58	4.14	3.85	1.07	98.57
SI029	48.17	2.54	16.30	10.75	0.19	3.53	7.96	3.87	3.58	1.02	97.90
SI030	49.25	2.58	16.40	10.83	0.19	3.52	6.27	4.30	4.06	1.06	98.46
SI032	49.80	2.62	17.07	10.67	0.21	3.53	6.45	4.25	3.54	1.02	99.16
SI033	46.89	2.57	16.86	10.26	0.17	3.79	8.24	4.06	3.48	1.03	97.35
SI034	52.19	2.01	17.47	9.88	0.18	2.78	5.08	5.16	4.02	0.79	99.57
SI035	50.51	2.35	17.05	10.10	0.19	3.28	5.84	5.22	3.51	0.91	98.95
SI036	51.11	2.22	17.22	9.97	0.18	2.98	5.58	5.00	3.92	0.86	99.05
SI041	45.44	2.69	16.70	10.59	0.17	3.79	8.73	5.13	3.10	1.18	97.53
SI042	44.86	2.70	16.16	10.57	0.17	3.79	9.40	4.73	3.27	1.17	96.82
SI048	47.76	2.76	16.42	10.88	0.18	3.83	7.11	3.72	4.24	1.12	98.02
SI053	50.16	1.88	14.89	9.11	0.15	6.02	10.77	3.43	0.53	0.19	97.14
SI063	45.58	2.78	15.66	11.13	0.22	3.74	9.33	3.76	3.32	1.14	96.67
SI065	47.82	2.21	13.81	13.22	0.18	6.80	11.23	2.48	0.02	0.20	97.96
SI066	48.06	2.19	13.87	13.25	0.20	6.82	11.25	2.59	0.01	0.20	98.44
SI067	47.87	2.20	13.74	13.37	0.22	6.97	11.11	2.62	0.01	0.20	98.30
SI068	47.37	2.58	13.25	14.19	0.22	6.31	11.11	2.52	0.02	0.23	97.80
SI070	51.16	2.23	14.79	8.64	0.18	6.91	10.74	3.56	0.14	0.20	98.54
SI072	49.30	2.51	17.23	10.28	0.19	3.34	6.50	4.89	3.33	0.95	98.52
SI073	49.19	2.53	17.26	10.30	0.19	3.34	6.52	4.90	3.21	0.96	98.39
SI078F	46.92	2.02	14.54	12.18	0.19	7.84	10.09	3.00	0.31	0.18	97.25
SI080F	47.28	4.10	13.61	13.44	0.11	6.17	11.54	3.52	0.04	0.15	99.95
SI082	49.78	1.11	15.14	10.03	0.15	8.20	9.26	2.88	0.25	0.12	96.94
SI085	49.47	1.37	14.26	11.91	0.21	8.03	9.56	2.56	0.29	0.13	97.80
SI086	50.00	1.09	15.08	9.90	0.18	7.73	10.21	2.48	0.22	0.11	97.02
SI087	49.95	1.10	15.03	10.01	0.18	7.81	10.26	2.42	0.15	0.12	97.02

Table. 2-2 (continued).

sample	SiO2	TiO2	Al2O3	Fe2O3	MnO	MgO	CaO	Na2O	K2O	P2O5	total
SI088	49.78	1.13	15.08	10.04	0.17	7.46	9.88	2.39	0.27	0.12	96.31
SI090	46.96	2.19	13.75	13.37	0.21	6.27	10.93	2.83	0.98	0.22	97.70
SI092F	49.34	1.11	14.93	8.96	0.11	8.78	12.17	2.89	0.04	0.12	98.44
SI095	47.45	2.32	13.79	13.05	0.24	6.08	11.23	3.14	0.69	0.23	98.21
SI096	47.71	2.49	12.77	12.74	0.23	5.94	11.02	3.05	0.68	0.27	96.89
SI097	48.20	2.51	12.79	14.50	0.21	6.32	10.35	2.41	0.02	0.22	97.54
SI098	49.22	2.43	12.95	13.23	0.32	6.74	10.33	3.32	0.02	0.23	98.80
SI099	46.06	3.60	11.98	17.12	0.25	5.76	9.36	2.24	0.04	0.33	96.73
SI100	49.21	1.89	13.98	11.11	0.22	5.82	10.49	3.62	0.66	0.22	97.22
SI101	47.60	2.17	13.71	13.22	0.22	7.13	10.75	2.70	0.01	0.20	97.70
SI102	48.69	2.58	13.17	13.67	0.26	6.24	10.05	3.48	0.10	0.24	98.47
SI105	48.24	2.29	14.06	12.41	0.19	6.38	12.10	2.49	0.04	0.24	98.45
SI106	47.95	1.87	14.14	12.12	0.21	7.30	12.21	2.58	0.01	0.18	98.57
SI107	50.02	2.00	15.24	10.01	0.18	6.72	11.40	3.09	0.56	0.19	99.41
SI108	46.57	1.87	13.97	12.40	0.19	7.48	11.97	2.16	0.02	0.18	96.82
SI109	47.26	2.48	14.16	13.06	0.24	6.41	11.78	2.80	0.06	0.23	98.46

Table 2-3. REE contents of lavas and intrusive rocks from MVG

	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu
SN001	6.53	14.63	2.03	12.69	4.00	1.38	4.69	0.75	5.37	1.12	3.12	0.43	3.26	0.47
SN004	6.36	15.19	2.11	11.23	3.94	1.33	4.43	0.75	5.36	0.96	3.12	0.39	2.72	0.41
SN010	6.01	13.12	2.01	11.28	3.26	1.30	4.35	0.70	4.91	0.93	2.94	0.40	2.83	0.41
SN023	5.57	12.77	1.86	10.92	3.50	1.14	4.31	0.62	4.41	0.92	2.52	0.36	2.76	0.41
SN025	11.11	25.45	3.35	20.17	5.83	1.83	6.91	1.08	7.89	1.53	5.21	0.56	4.74	0.72
SN035	6.13	13.85	1.85	12.25	3.33	1.28	4.35	0.62	4.18	0.95	2.90	0.35	2.72	0.38
SN044	5.92	13.61	1.86	11.17	3.60	1.25	4.60	0.67	4.83	0.94	3.11	0.43	3.23	0.39
SN052	3.13	7.34	1.11	7.06	2.31	0.87	2.90	0.45	3.12	0.65	1.95	0.27	2.24	0.32
SN059	3.48	8.89	1.32	7.56	2.87	0.84	3.20	0.55	3.76	0.69	2.22	0.32	2.41	0.32
SN062	5.90	13.57	1.92	11.62	3.45	1.34	4.54	0.74	4.86	0.97	2.86	0.42	2.91	0.37
SN071	5.25	12.04	1.75	10.65	3.06	0.98	3.82	0.61	4.07	0.88	2.64	0.34	2.80	0.38
SN085	5.32	12.99	1.80	11.02	2.94	1.15	4.44	0.58	4.77	0.91	2.55	0.39	2.41	0.41
SN094	5.96	13.36	2.06	11.56	2.87	1.20	4.39	0.60	5.18	0.94	2.74	0.40	2.92	0.42
SN098	3.68	8.61	1.31	7.93	2.35	0.99	3.40	0.51	3.52	0.76	2.28	0.31	2.49	0.33
SN107	3.68	9.16	1.34	7.95	2.67	0.93	3.58	0.53	3.54	0.87	2.07	0.34	2.46	0.31
SN113	4.95	11.51	1.69	10.94	2.91	1.11	3.70	0.61	4.45	0.87	2.37	0.36	2.70	0.38
SN124	7.44	15.00	2.20	12.94	3.37	1.23	4.19	0.66	4.33	0.81	2.59	0.34	2.60	0.37
SN136	3.31	8.21	1.24	7.63	2.58	0.97	3.40	0.55	3.87	0.75	2.34	0.34	2.40	0.29
SN157	4.03	9.70	1.36	8.45	2.75	1.09	3.18	0.61	4.09	0.84	2.48	0.35	2.32	0.33
SN170	5.66	13.87	1.95	10.49	3.17	1.16	3.96	0.68	4.72	0.97	2.65	0.34	2.90	0.34
SN180	5.12	13.04	1.67	10.56	3.10	1.06	3.55	0.60	4.35	0.83	2.69	0.31	2.66	0.32
SN190	6.79	15.77	2.15	12.72	4.07	1.32	4.42	0.77	5.40	1.09	3.40	0.46	3.65	0.42
SN196	4.57	10.49	1.42	8.88	2.87	0.91	3.39	0.57	3.92	0.81	2.56	0.32	2.42	0.24
SN206	3.49	7.98	1.24	7.83	2.40	0.91	3.29	0.52	3.90	0.72	2.17	0.31	2.19	0.31
SN215	3.69	8.81	1.33	6.78	2.53	0.95	3.10	0.52	3.60	0.72	1.91	0.29	2.41	0.30
SN221	5.84	13.46	1.90	11.06	3.50	1.25	4.19	0.62	4.78	0.90	2.80	0.40	2.91	0.40
SN279	7.62	17.73	2.51	14.02	4.35	1.39	4.59	0.78	5.83	1.12	3.37	0.41	3.28	0.53
SN307	3.17	7.91	1.09	7.22	2.28	0.87	2.52	0.44	3.09	0.67	1.99	0.26	2.04	0.30
SN309	2.16	5.37	0.74	4.90	1.55	0.59	1.71	0.30	2.10	0.46	1.35	0.18	1.39	0.20
SN335	3.43	7.21	1.17	7.95	2.49	0.91	2.91	0.45	3.75	0.71	2.16	0.31	2.48	0.31
SN348	2.58	5.86	0.92	5.23	1.69	0.60	2.18	0.42	2.54	0.57	1.65	0.18	1.46	0.24
SN372	6.02	13.04	1.96	11.14	3.81	1.23	4.03	0.66	4.74	0.98	2.96	0.37	2.80	0.39
SN378	5.64	12.84	1.91	10.22	2.98	1.13	3.90	0.59	4.44	0.92	2.80	0.36	2.85	0.32
SN390	2.73	6.38	1.02	6.72	2.12	0.82	2.71	0.41	3.02	0.64	1.65	0.24	1.96	0.24

Table 2-4. Trace elements contents of lavas and intrusive rocks from MVG

	Y	Nb	Rb	Zr	Sr	Pb	V	Th	Ni	Cr
SN001	27.67	4.26	2.40	81.70	127.80	3.48	365.82	0.57	56.43	59.46
SN004	26.31	4.53	3.23	83.40	140.64	4.01	358.79	0.99	68.77	133.48
SN010	26.91	4.60	2.76	84.76	136.60	3.47	332.49	0.80	70.04	127.18
SN023	27.52	4.76	2.72	83.64	142.89	3.00	38.25	0.57	78.97	123.58
SN025	37.83	7.35	3.93	120.68	131.20	3.07	369.10	1.10	24.53	25.00
SN035	27.37	4.42	2.42	84.97	137.14	3.38	331.49	0.72	78.12	132.58
SN044	26.33	4.61	1.00	80.09	132.50	3.15	314.34	0.26	76.70	117.28
SN052	21.44	2.47	0.60	59.85	96.94	2.77	308.32	0.80	90.13	137.98
SN059	22.66	2.98	1.81	68.73	212.77	3.07	298.71	0.49	103.68	184.65
SN062	27.67	4.70	0.80	87.77	143.52	3.81	340.28	0.64	79.25	128.98
SN071	27.06	4.10	2.48	83.58	134.36	3.35	299.90	1.04	73.02	117.28
SN085	26.27	4.18	2.02	78.64	135.28	3.07	329.64	0.87	82.50	120.88
SN094	27.48	4.51	1.26	81.49	135.97	3.32	317.37	0.64	83.63	120.88
SN098	23.17	2.90	5.80	69.44	115.16	2.77	309.77	n.d.	101.71	204.34
SN107	22.51	3.01	0.56	66.98	118.85	2.92	295.10	0.26	101.85	183.76
SN113	25.08	3.95	3.03	75.79	133.68	3.00	333.57	1.31	89.29	156.85
SN124	24.93	5.07	1.29	89.80	158.59	3.47	382.66	0.80	89.85	143.38
SN136	22.52	2.92	1.00	64.80	121.40	2.92	294.24	0.72	108.33	177.49
SN157	25.67	3.30	1.98	74.70	122.24	2.92	315.90	0.80	83.63	128.98
SN170	28.74	4.62	2.03	88.87	150.16	3.00	350.29	0.87	90.42	129.88
SN180	25.91	3.86	2.40	79.05	138.59	2.92	319.66	0.8	79.39	119.08
SN190	30.84	5.18	2.30	89.18	128.31	3.29	407.66	0.64	38.93	45.87
SN196	24.53	3.25	1.60	70.86	125.47	2.85	303.15	0.64	92.39	139.78
SN206	21.88	3.04	0.64	63.91	119.41	3.00	291.72	0.64	100.72	181.97
SN215	22.12	2.87	0.65	67.11	121.06	3.22	291.95	0.11	104.67	182.86
SN221	27.62	4.93	1.17	84.20	136.36	3.55	330.94	0.57	73.16	132.58
SN279	30.67	5.67	2.64	93.45	129.08	3.48	387.20	0.92	42.78	41.34
SN307	23.24	3.87	0.53	69.56	112.95	3.54	311.75	0.57	115.37	227.57
SN309	14.74	1.37	2.52	46.30	85.95	2.92	191.38	0.72	449.62	154.16
SN335	21.19	2.62	2.52	65.05	116.20	2.40	286.67	0.49	133.20	207.92
SN348	17.86	1.94	1.84	54.59	114.98	2.85	239.07	0.34	153.23	338.55
SN372	27.50	4.48	2.68	86.95	138.61	4.02	331.91	0.72	78.82	142.48
SN378	27.43	4.39	2.13	81.66	142.13	3.29	286.39	0.49	99.59	248.07
SN390	19.70	2.81	2.82	59.54	178.59	3.15	296.27	0.42	114.10	223.10

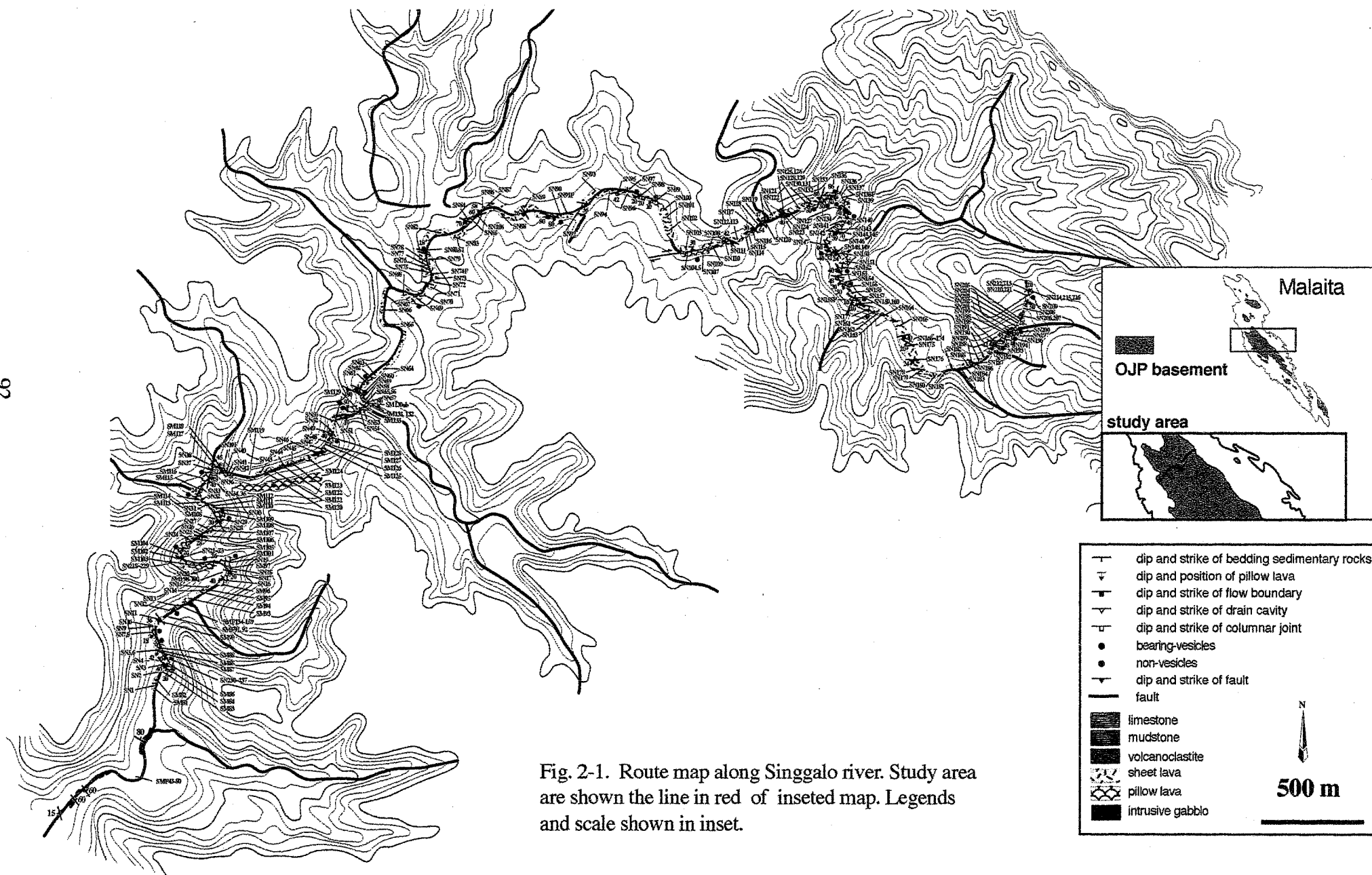


Fig. 2-1. Route map along Singgalo river. Study area are shown the line in red of inseted map. Legends and scale shown in inset.

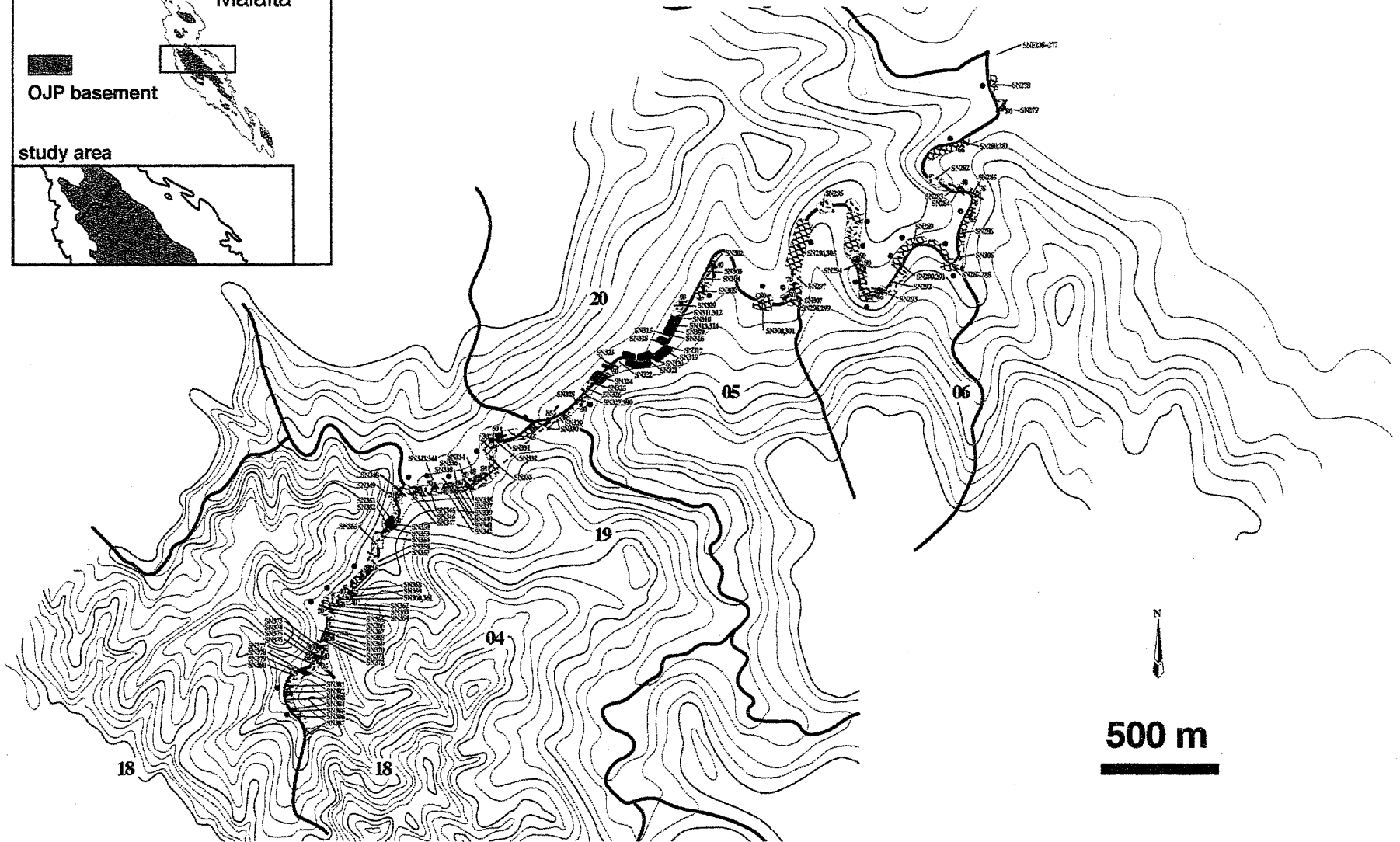
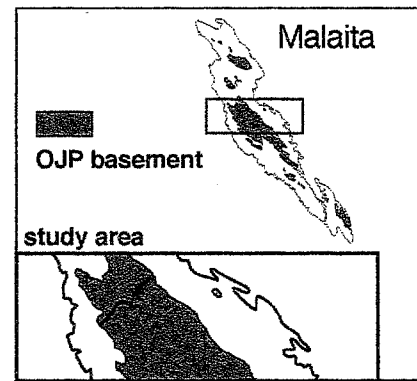


Fig. 2-2. Route map along Ambasi River. Study area is shown in inset map. Legends are same as Fig. 2-1.

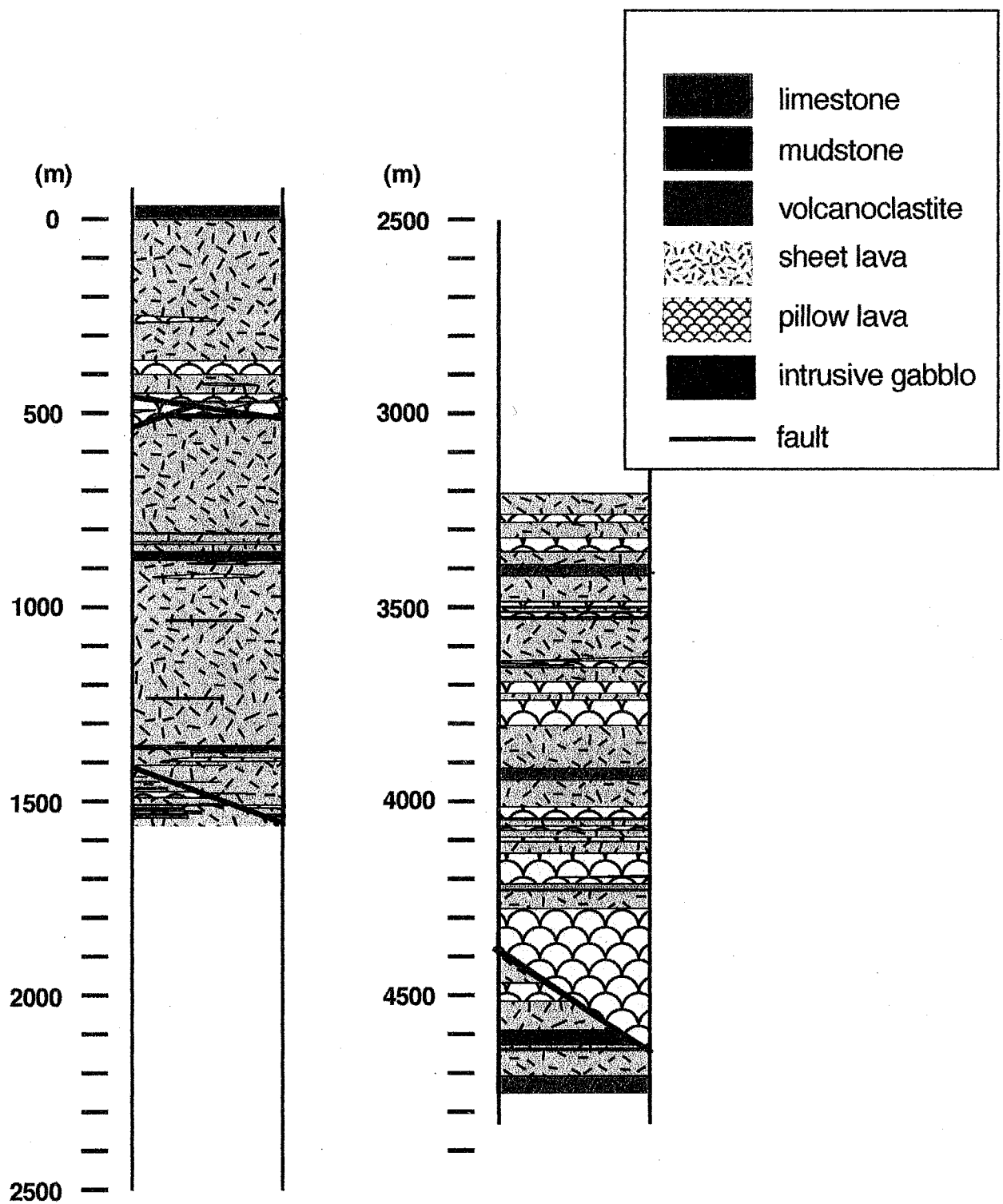


Fig. 2-3. Columnar section of MVG along Singgalo and Ambasi Rivers. Legends are shown in inset. Vertical axis expresses meters of below the capped sedimentary rocks. The MVG is discontinuous by several faults and maximum thickness is 1200m.

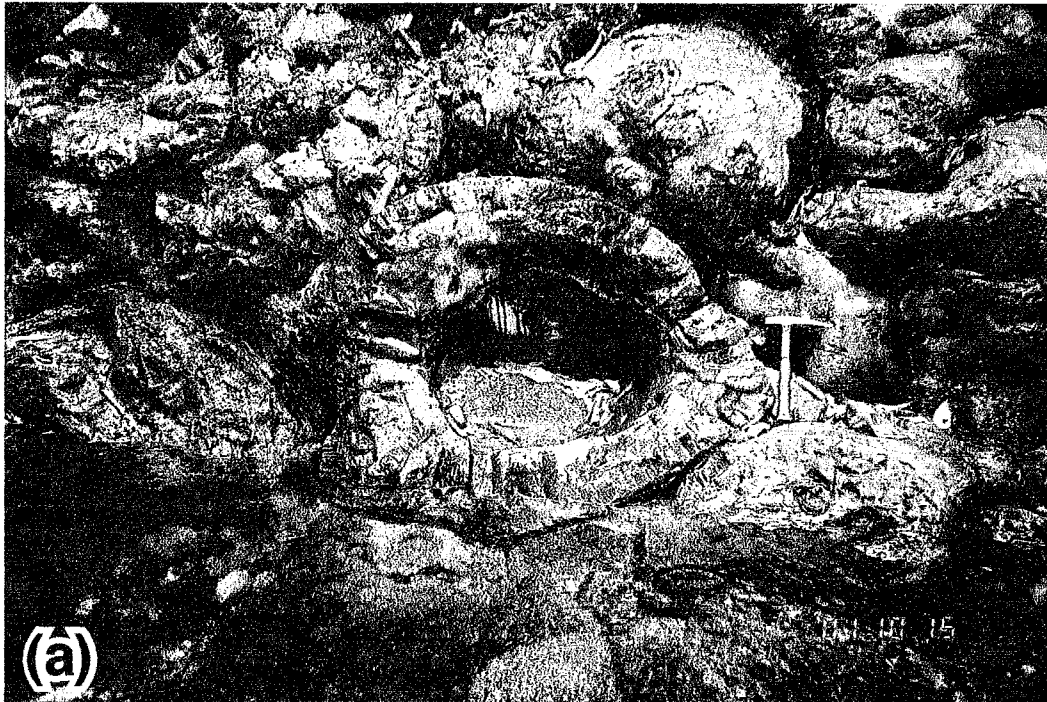


Fig. 2-4. The mode of field occurrences of lavas from the MVG.
 (a) Pillow lavas. Each pillow lobe has circular shape. The pillow lobes centered in photo has hollow. (b) Sheet lavas with developed columnar joints.

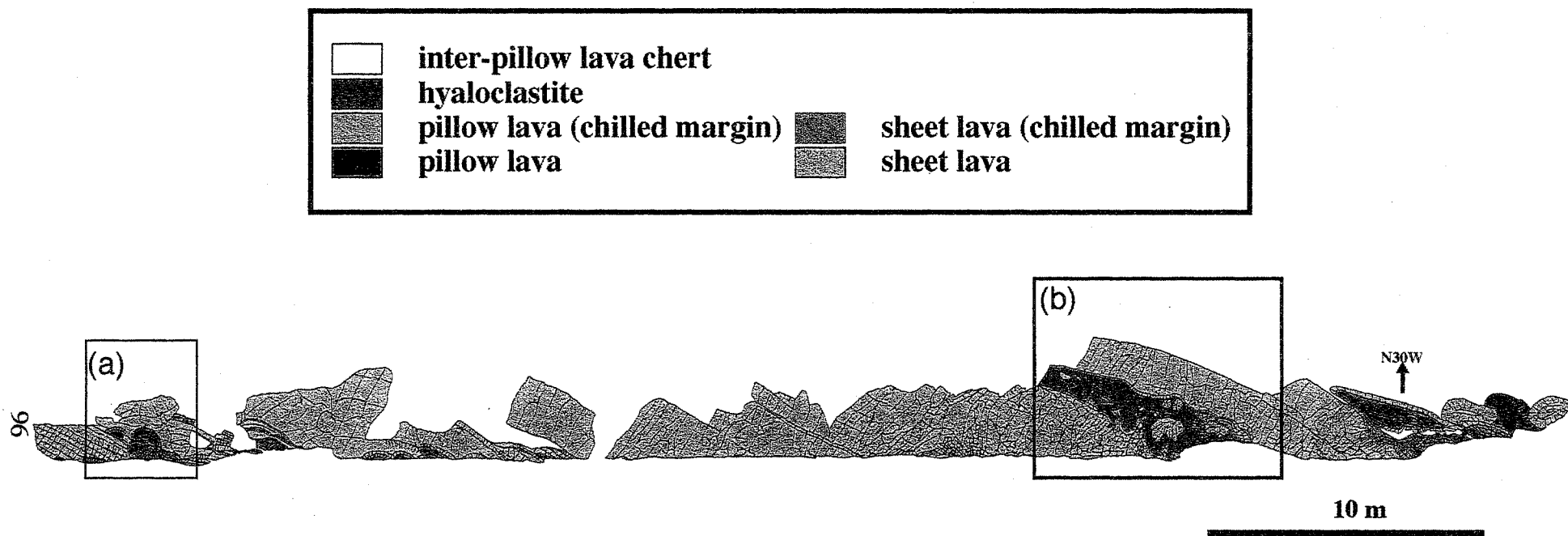


Fig. 2-5. The geological sketch of sheet lavas. Legends are shown in inset. (a) The sheet lavas have pillowed flow foot. (b) The sheet lavas have flow foot breccia. The lavas columnar joints are developed the upperpart of lavas.



Fig. 2-6. Pillow lobes with vesicles. scale is 10 cm. Vesicles with 2 - 10 mm in diameter exposed by peeled rind.



Fig. 2-7. Pillow lavas are overlain by greenish volcanoclastite. Empty space between pillow lobes are filled with volcanocrastite.

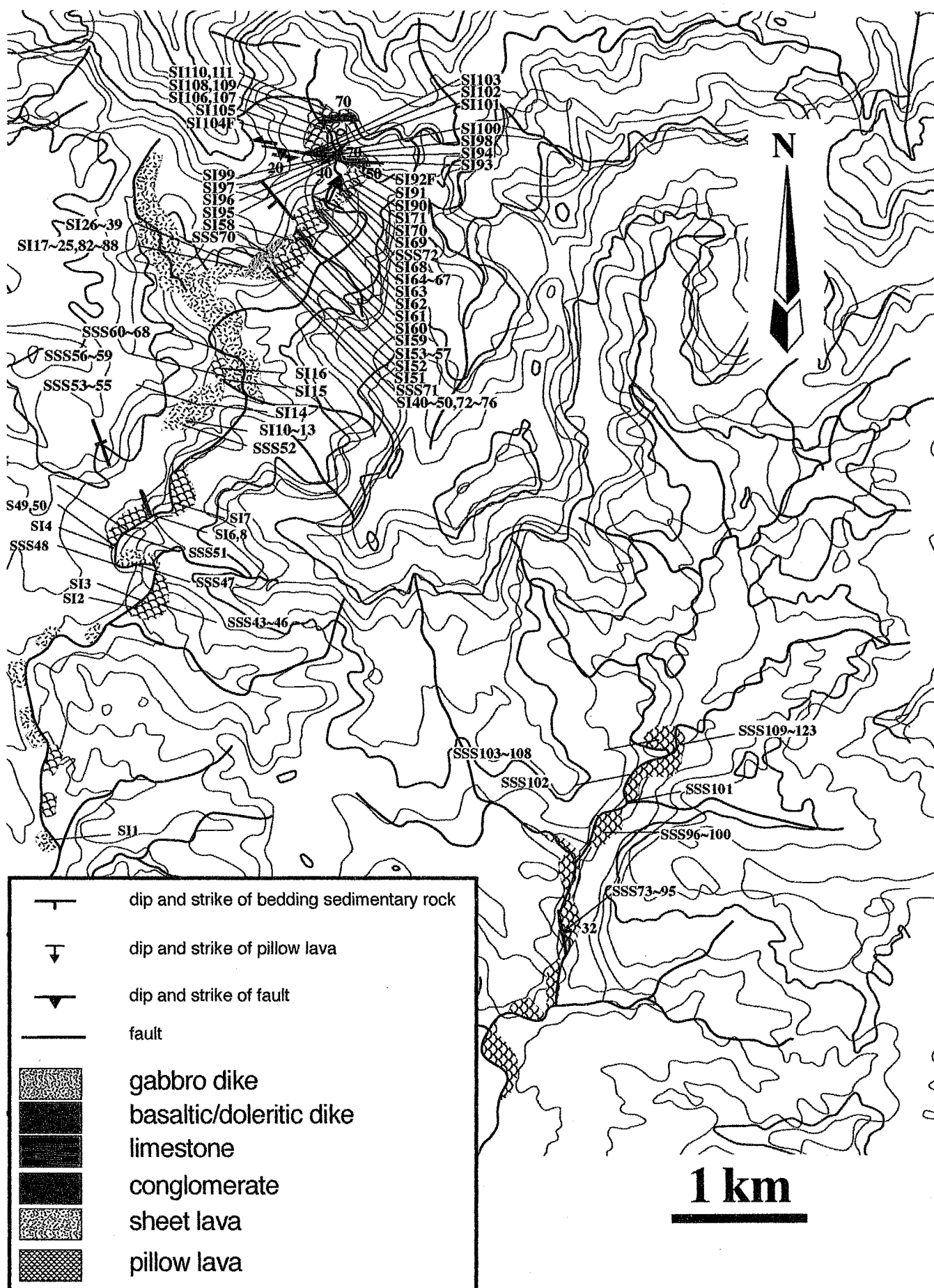


Fig. 2-8. Route map along Beamalalo River. Study area is A area shown in Fig.1-9.

Limestone lenses (St. Isabel)

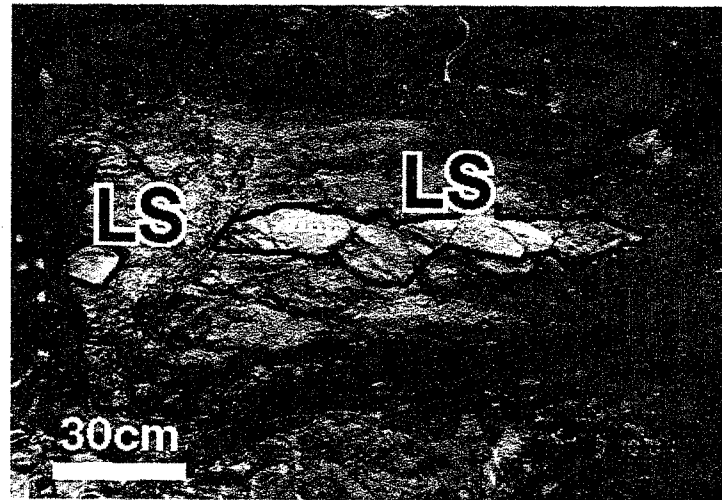
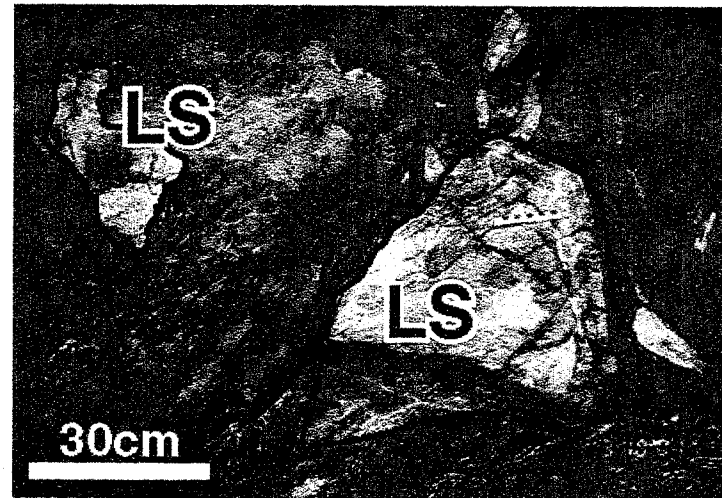
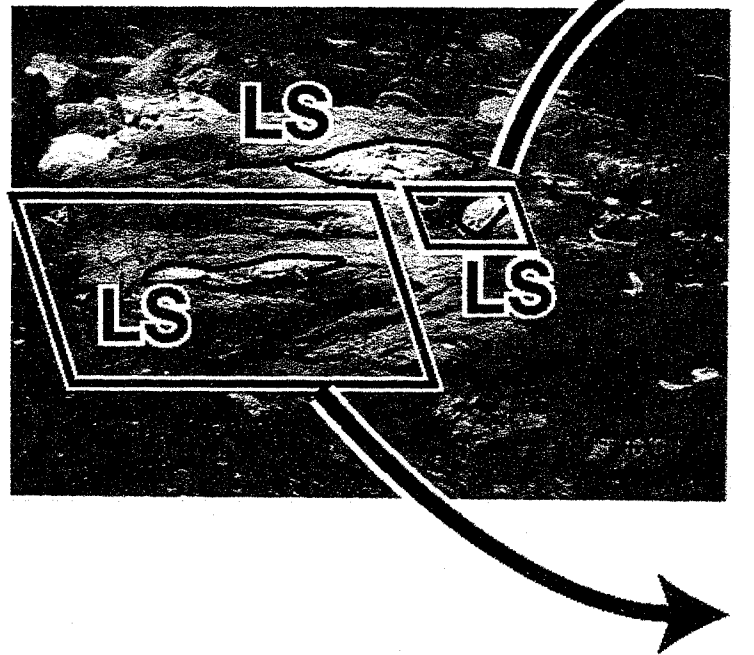


Fig. 2-9. Limestone lenses trapped by sheet lavas. Limestone have brecciated or irregular shape.

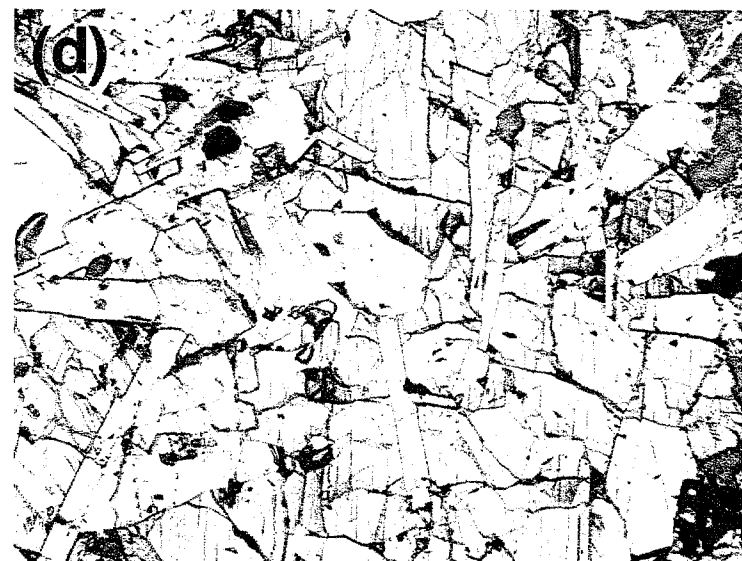
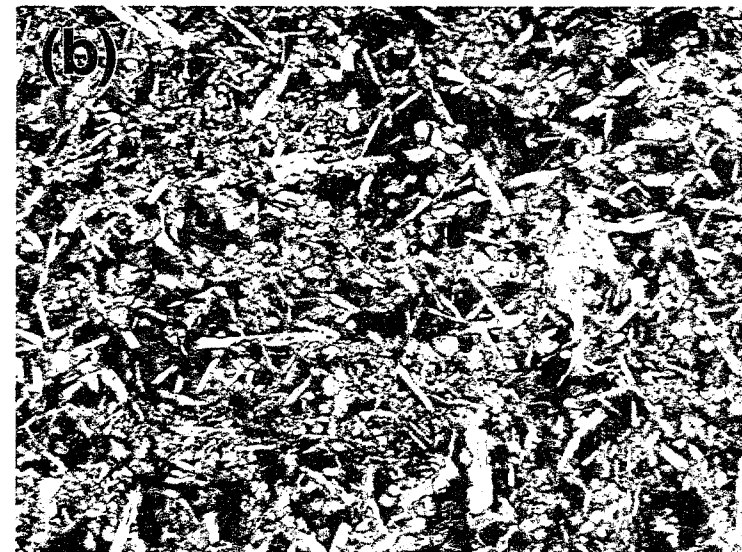


Fig. 2-10. Microphotos of the lavas from the MVG show the groundmass texture. (a) Quench valiolitic texture. (b) Intersertal texture. (c) Subophitic clinopyroxene and plagioclase bonded interstitial glass (replaced by clay mineral). (d) Ophitic texture. Scale for each photo is 2 mm in long side.

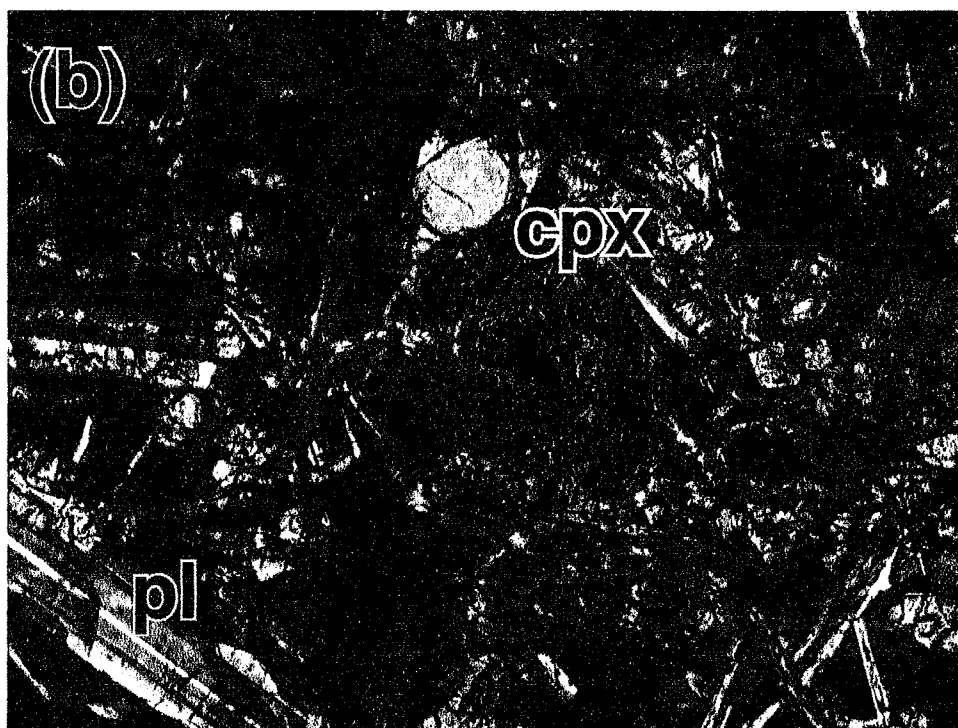
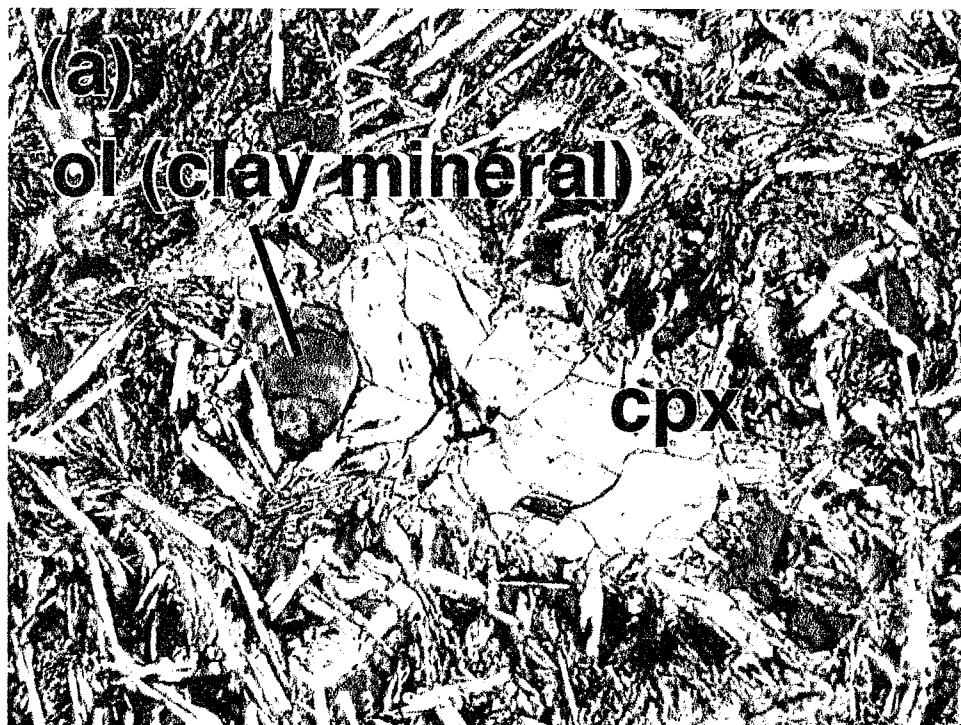


Fig. 2-11. Microphotos of lavas from the MVG show the textures of phenocrysts. (a) Euhedral olivine (ol) and subhedral clinopyroxene (cpx) make the glomeroporphyritic texture. (b) Euhedral cpx and angular plagioclase (pl). Scale for each photo is 2 mm in long side.

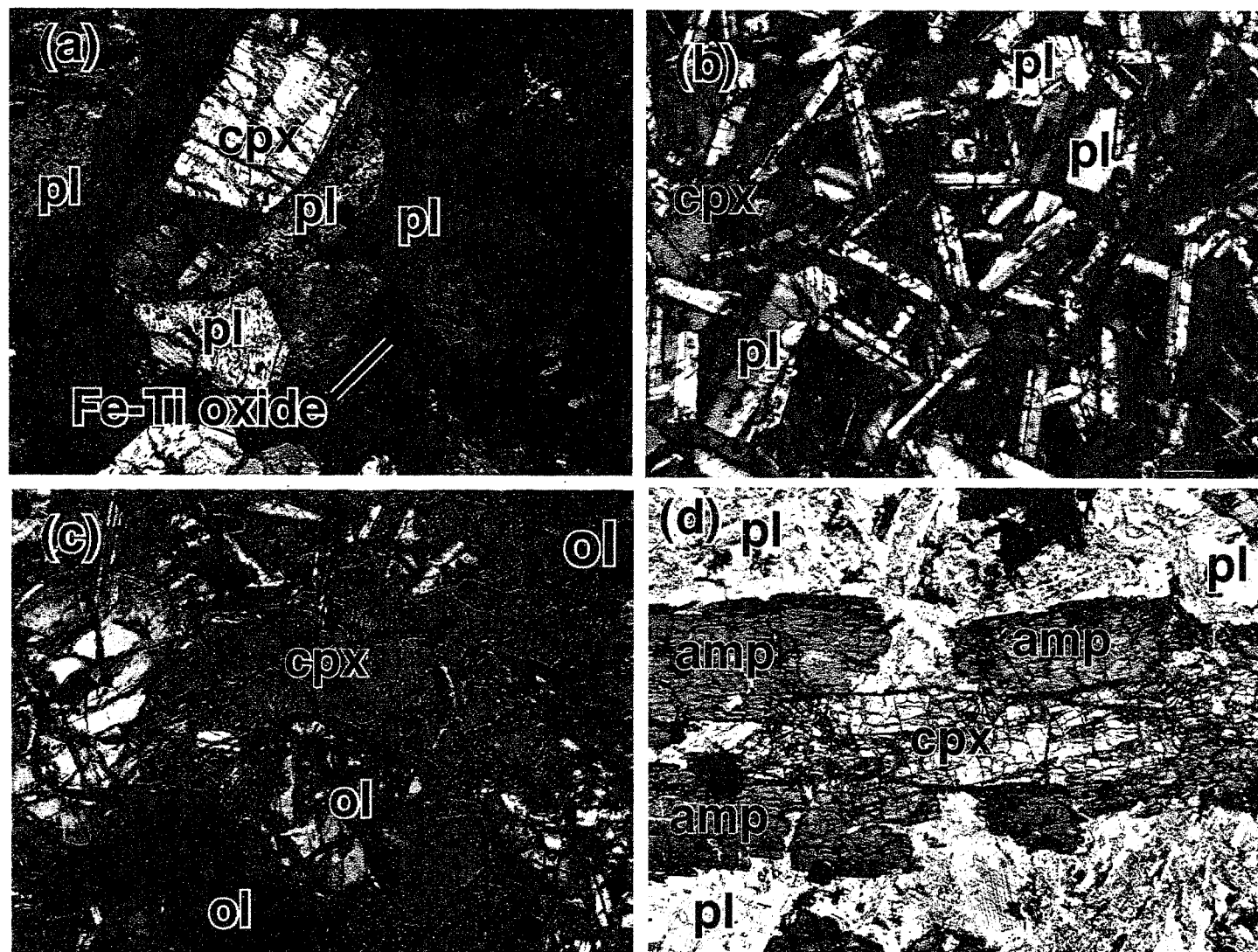


Fig. 2-12. Microphotos of the intrusion from MVG ((a)-(c)) and from Santa Isabel(d). (a) Equigranular . (b) ophitic. (c) olivine poikilitic. (d) Angular amphibole (amp) and Clinopyroxene. scale for each photo is 2 mm in long side.

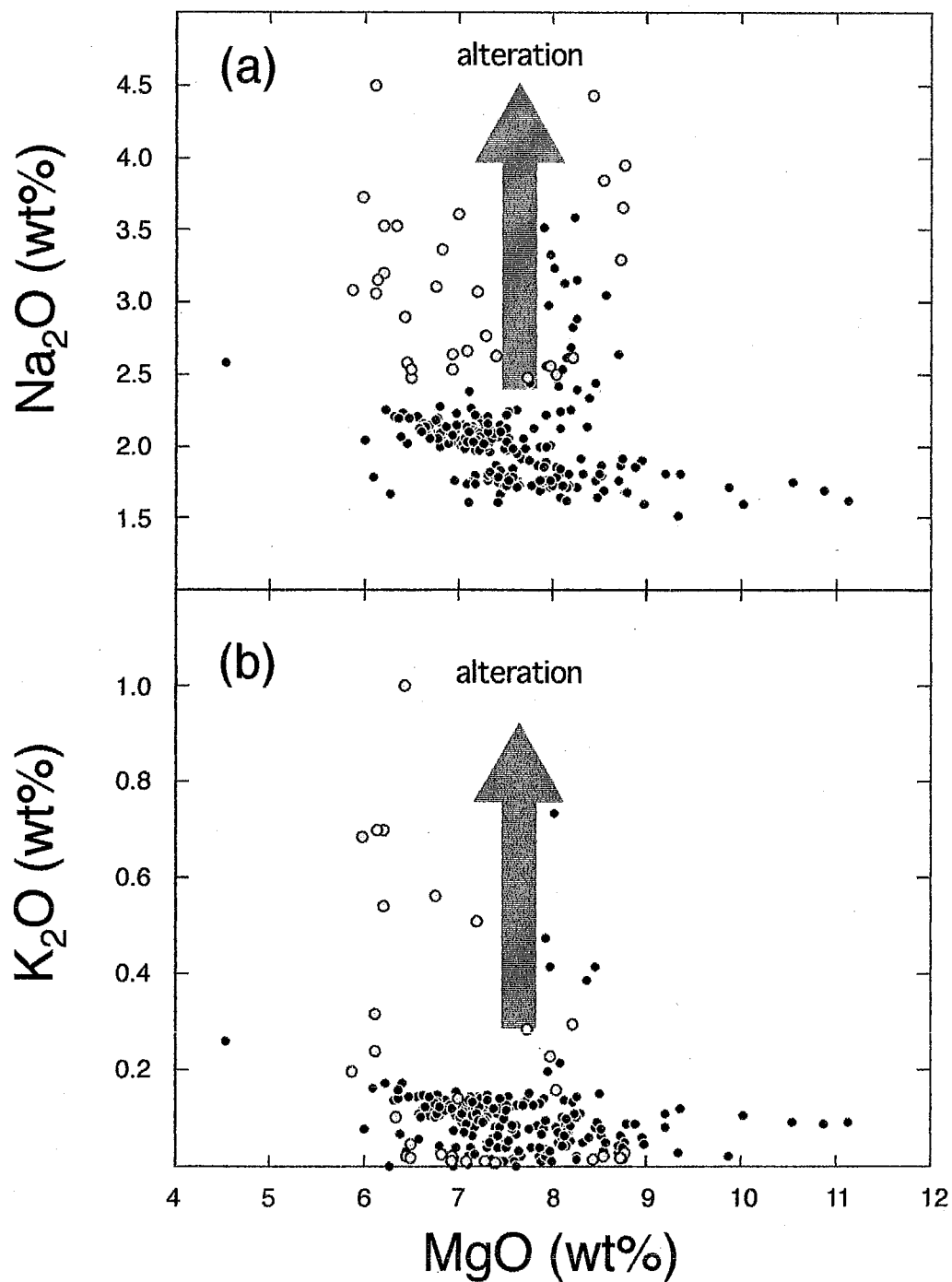


Fig. 2-13. Na₂O (a) and K₂O (b) contents against MgO variation. Black solid and grey circle show the lavas from MVG and western Santa Isabel, respectively. Na₂O and K₂O are enriched and scattered by seawater alteration.

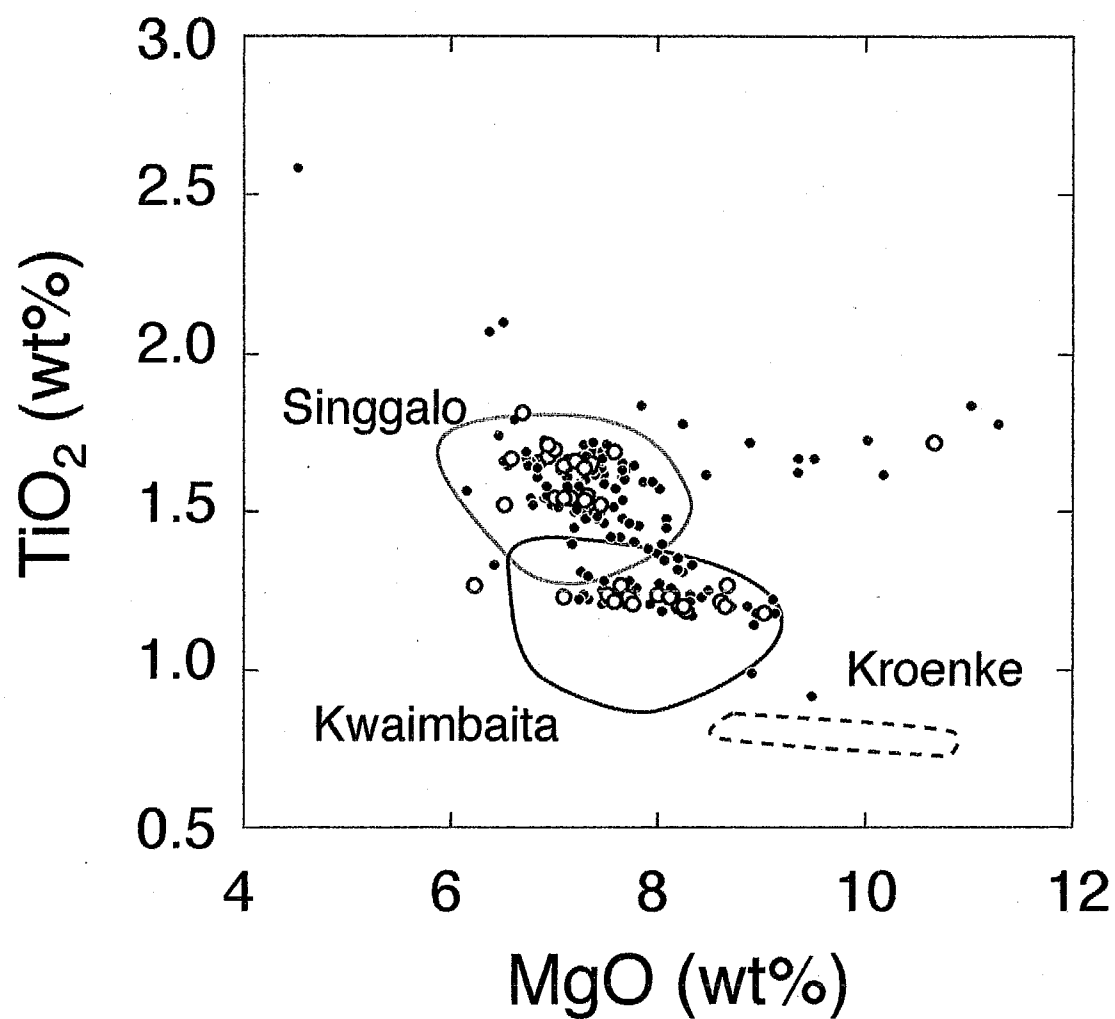


Fig. 2-14. MgO variation diagram of TiO₂ for MVG lavas. Aphyric samples are selected to plot the open circle. All data are normalized to total 100%. The fields enveloped grey and black lines show the Kwaimbaita- and Singgalo-type basalts from Tejada et al., 2002, respectively. Break line shows the Kroenke-type from Fitton and Godard, 2004.

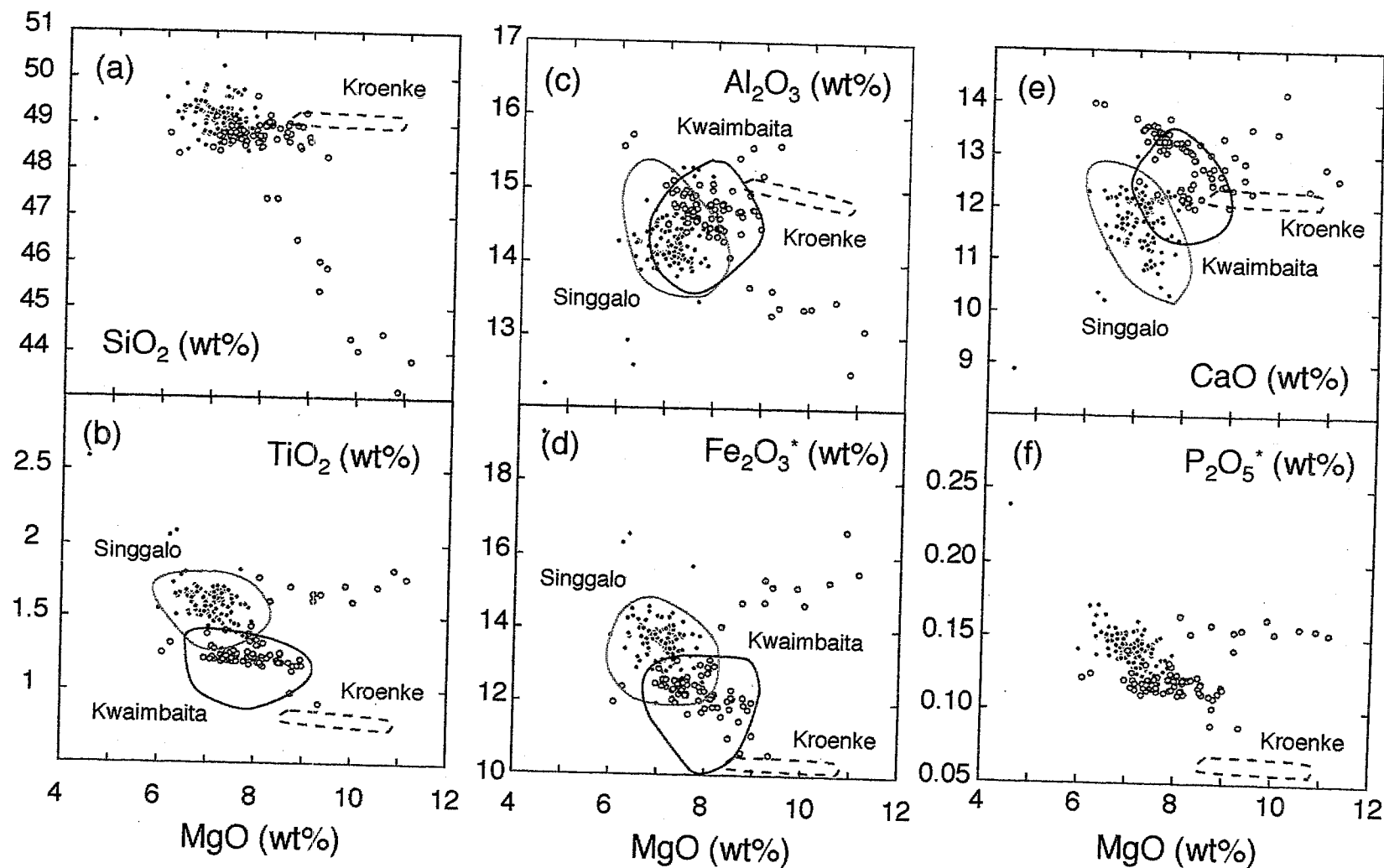


Fig. 2-15. MgO variation diagram of major elements for MVG lavas. Open and solid circles Low-TiO₂ and High-TiO₂ type lavas, respectively. Grey circles a grey and black lines show the Kwaimbaita- and Singgalo-type basalts from Tejada et al., 2002, respectively. Break line shows the Kroenke-type from Fitton and Godard, 2004.

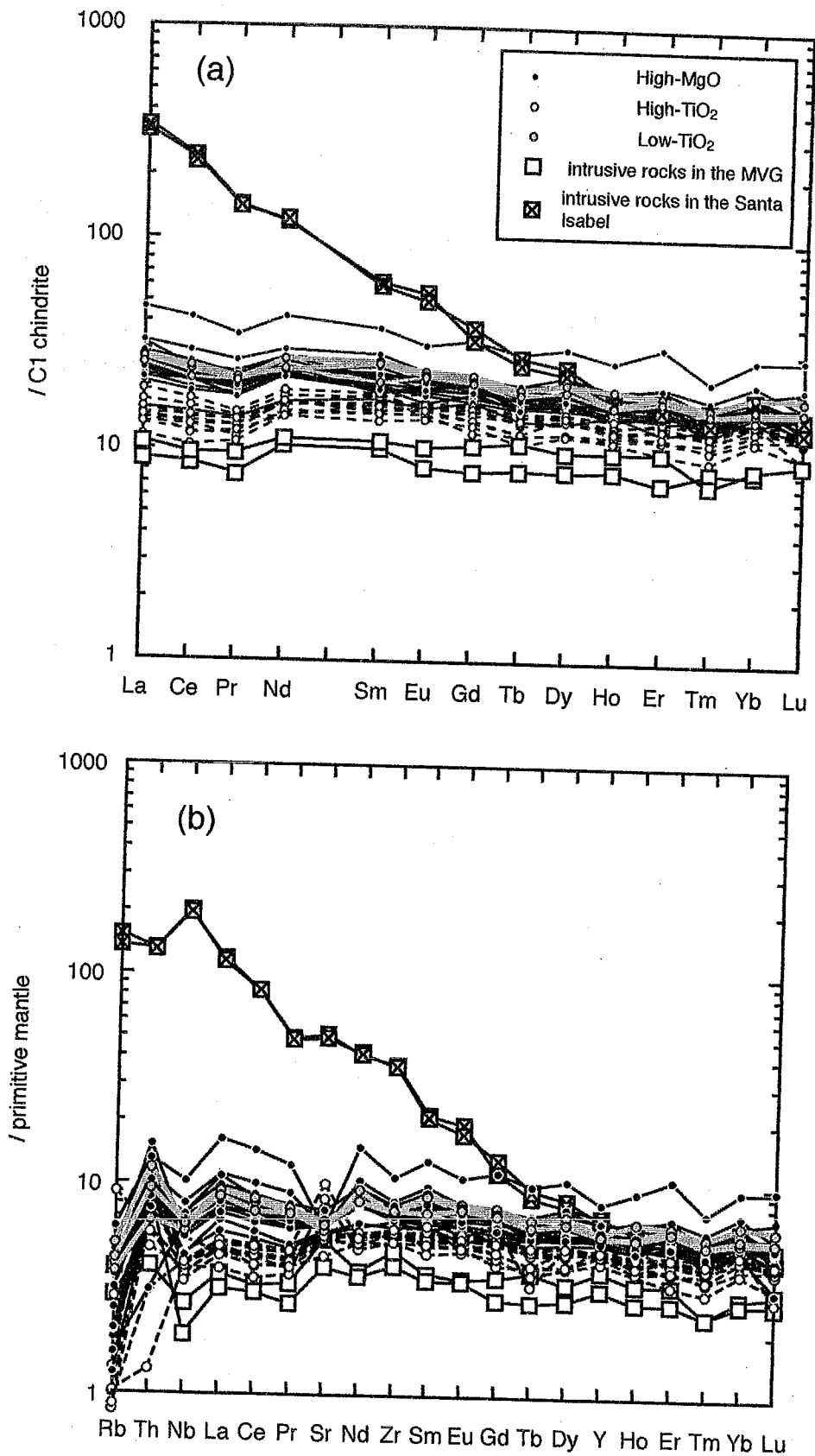


Fig. 2-16. REE (a) and trace element (b) relative abundance patterns of igneous rocks from Malaita and Santa Isabel. Symbols are shown in inset. Chondrite and primitive mantle values are from Sun and McDonough, 1989.

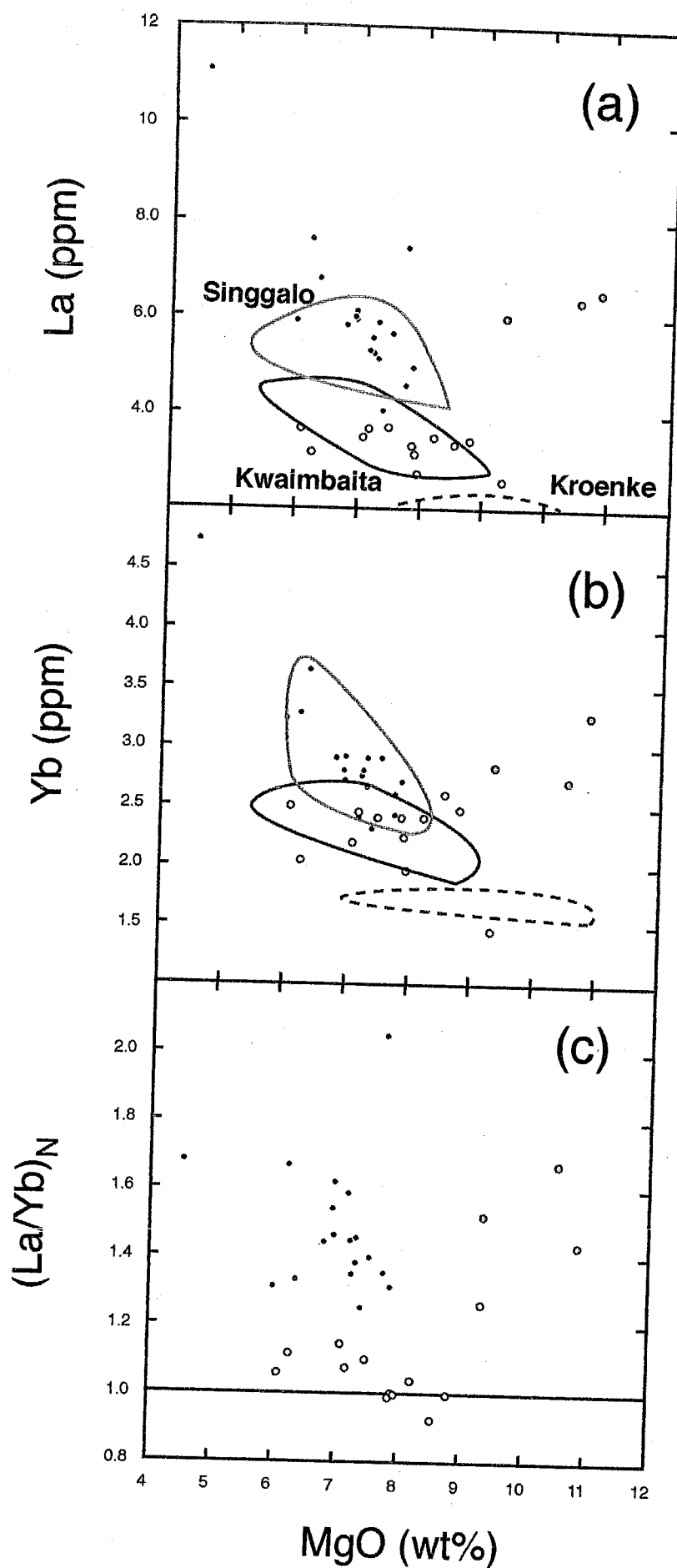


Fig. 2-17. MgO variation diagrams for REE of the lava of the MVG. Symbols and fields are shown as Fig. 2-15. $(La/Yb)_N$ means that the ratio of La nad Yb chondrite normalized value.

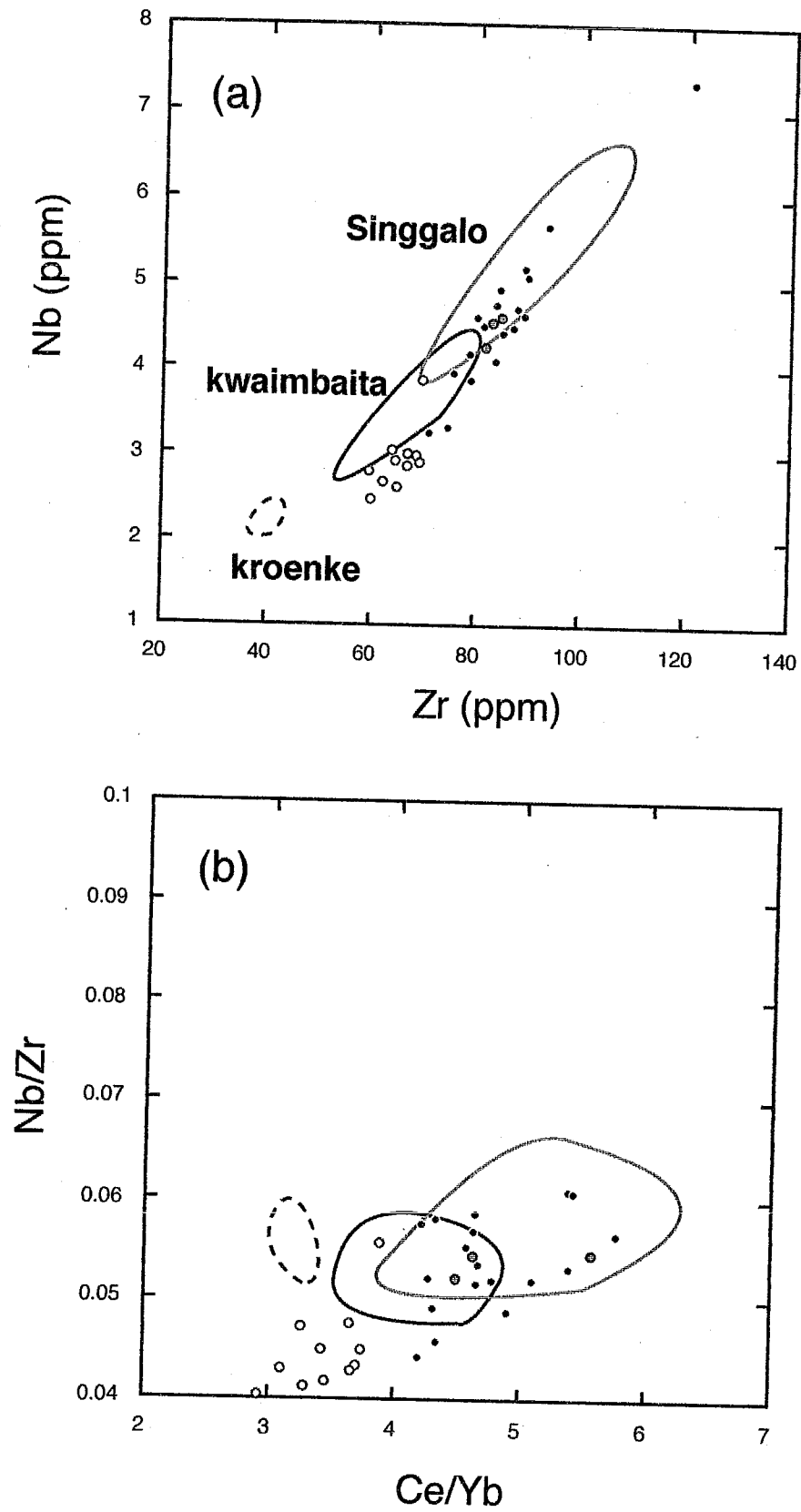


Fig. 2-18. The variation diagrams of immobile trace elements of the lavas from the MVG. Simboles and fields are shown as Fig. 2-15.

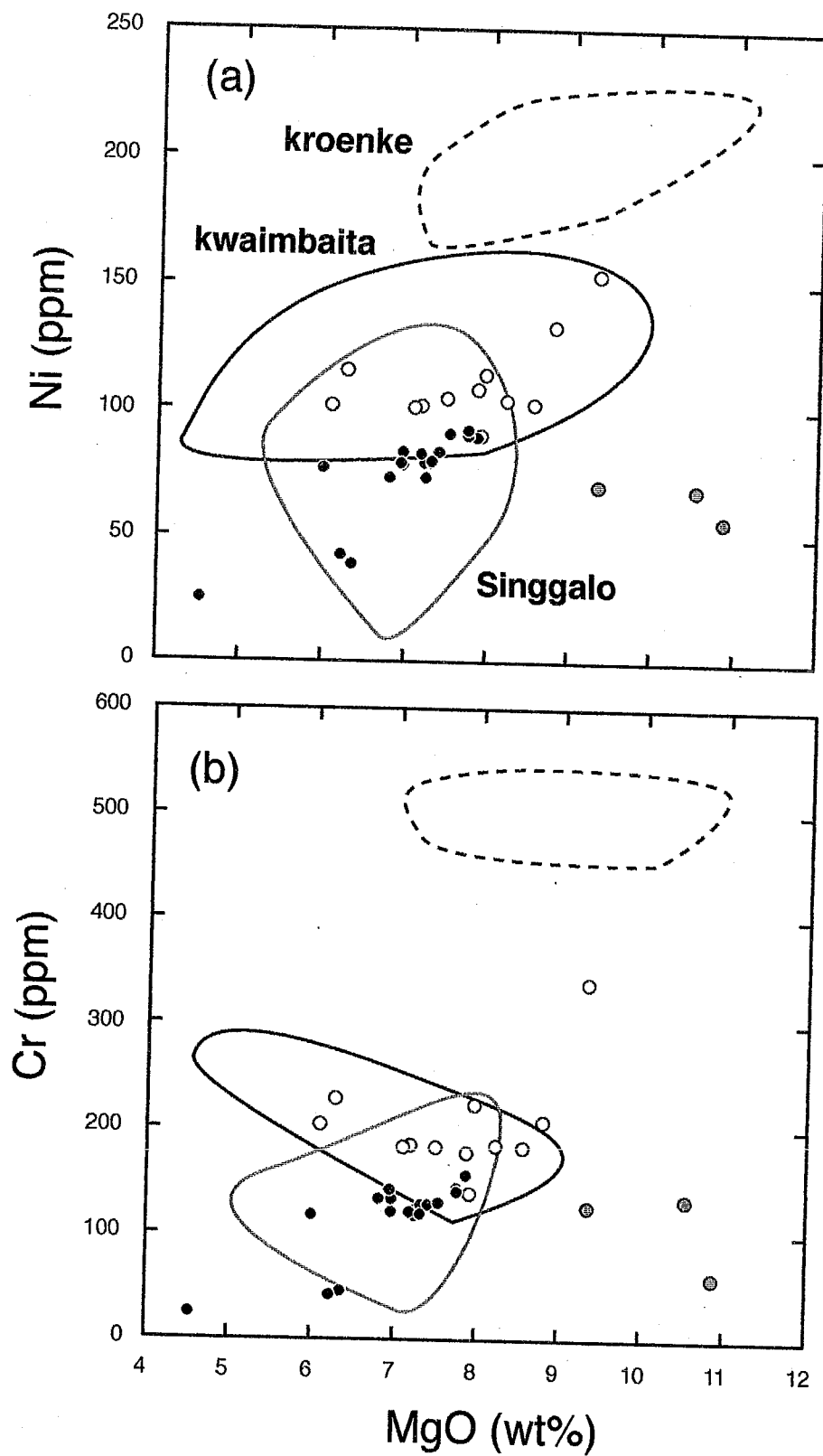


Fig. 2-19. MgO variation diagrams of compatible trace elements. Simboles and fields are shown as Fig. 2-15.

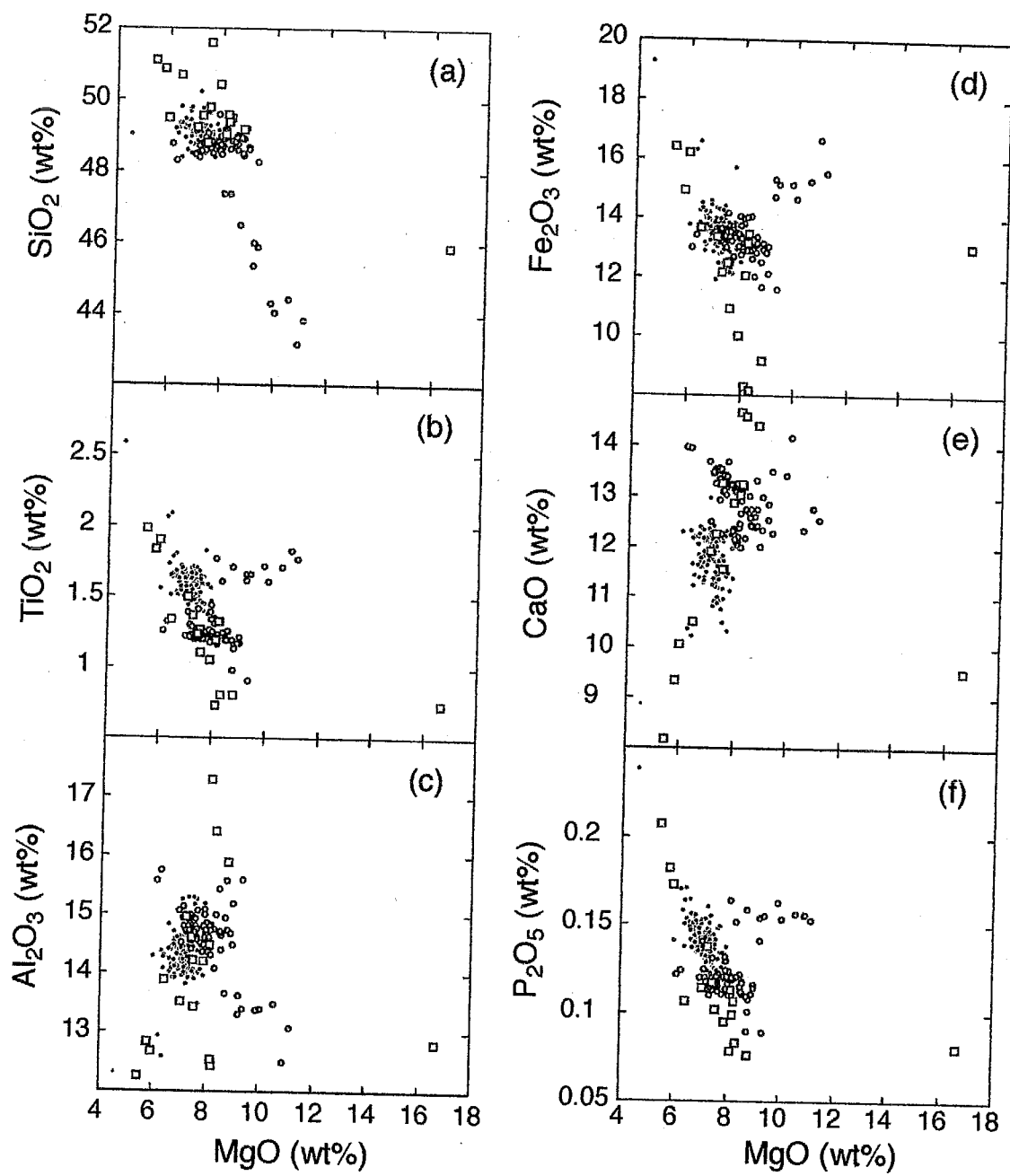


Fig. 2-20. MgO variation diagrams of major elements for intrusive rocks from MVG lavas. Open square shows the intrusive rock. Other symbols for lavas are shown as Fig. 2-15.

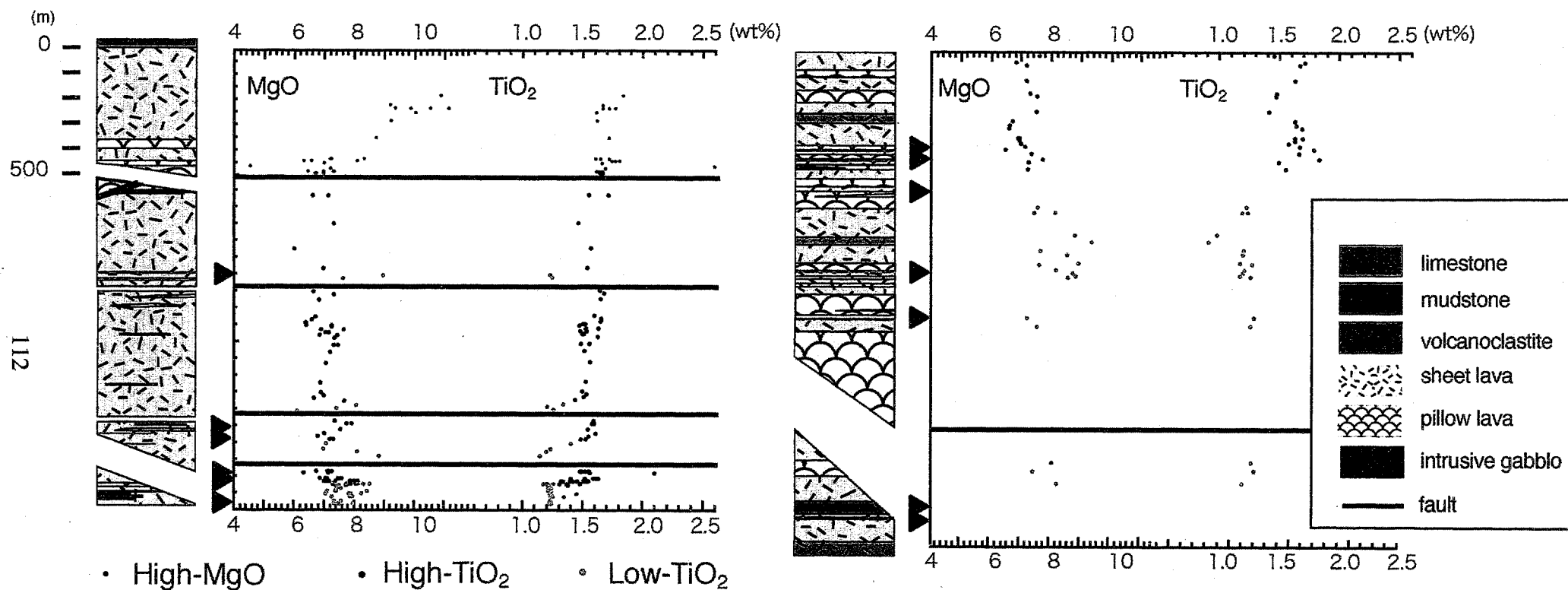


Fig. 2-21. The geochemical stratigraphy of MVG along Singgalo and Ambasi Rivers. Legends and symbols are shown in inset. Black triangles indicate the thin sedimentary layers

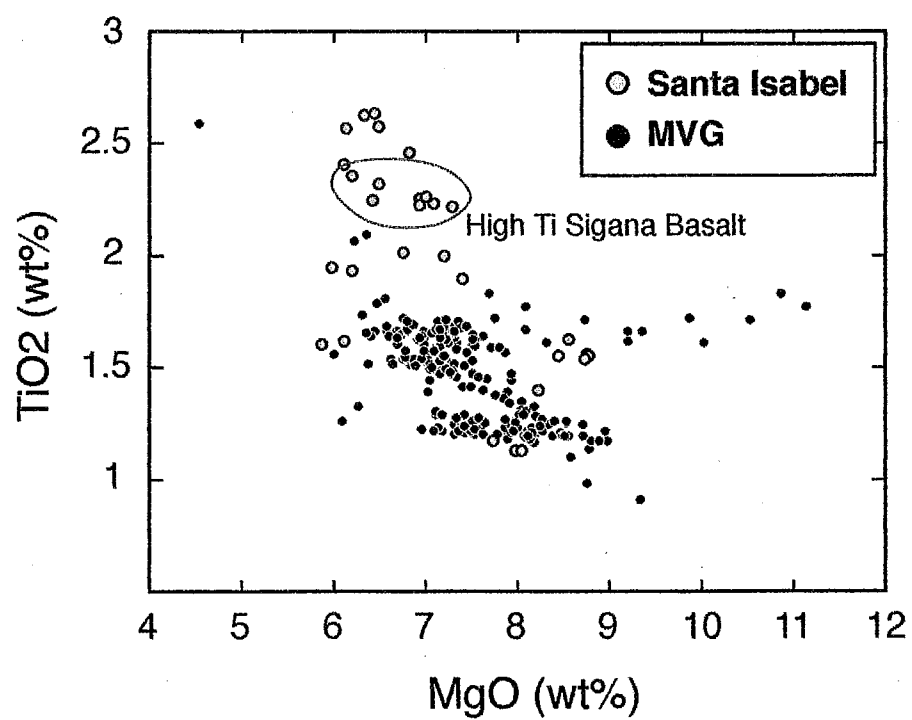


Fig. 2-22. TiO₂ vs MgO diagrams for lavas from Santa Isabel and MVG. Symbols are shown in inset. Most lavas from Santa Isabel are compared with High Ti Sigana Basalt. The data of High Ti Sigana Basalt are from Tejada et al., 1996.

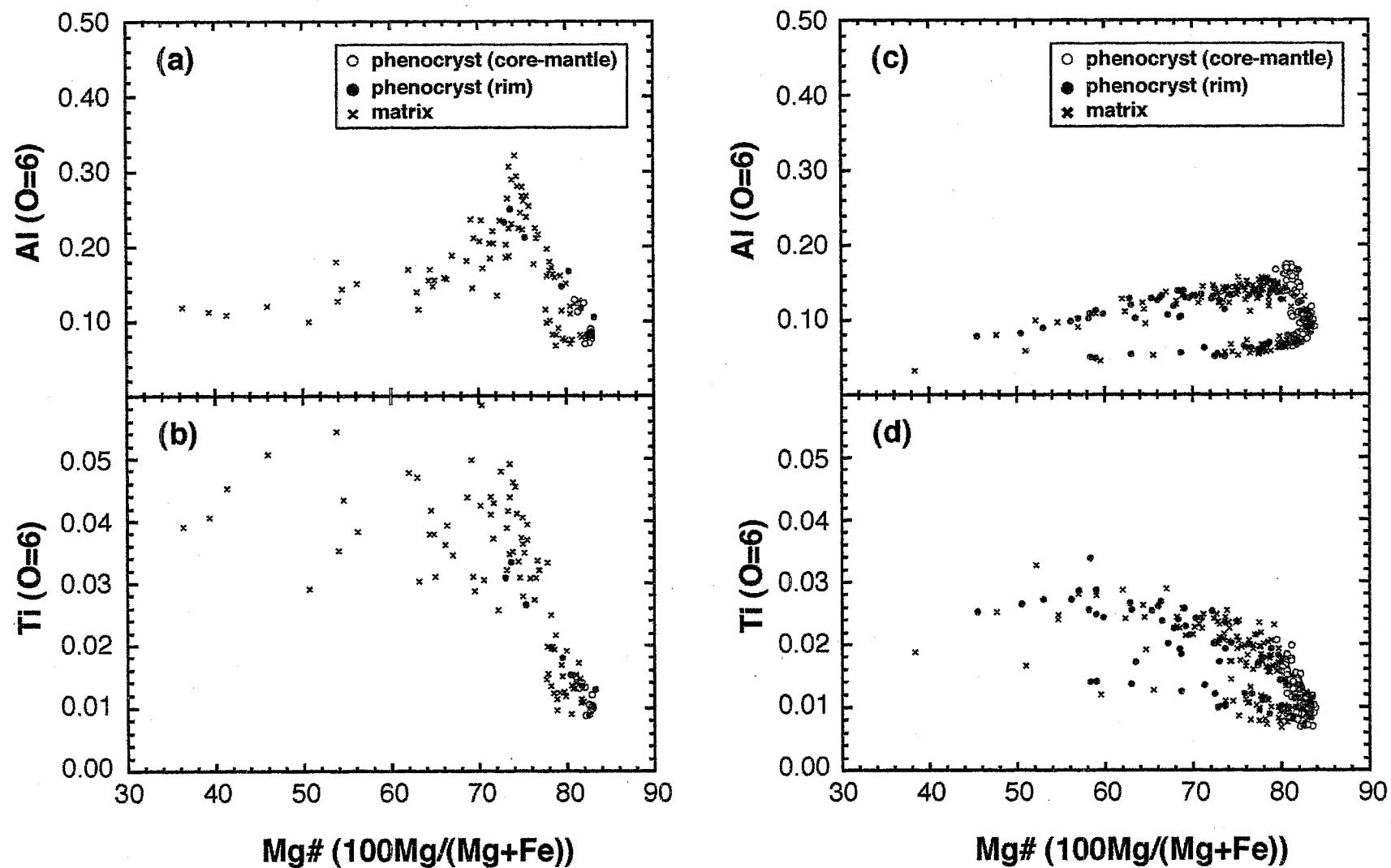


Fig. 2-23. Mg# vs Al (O=6) and Ti (O=6) for High-TiO₂ type (a and b) and Low-TiO₂ type (c and d). The data are from one thin section for each type (SN384 and SN155). Symbols are shown in inset.

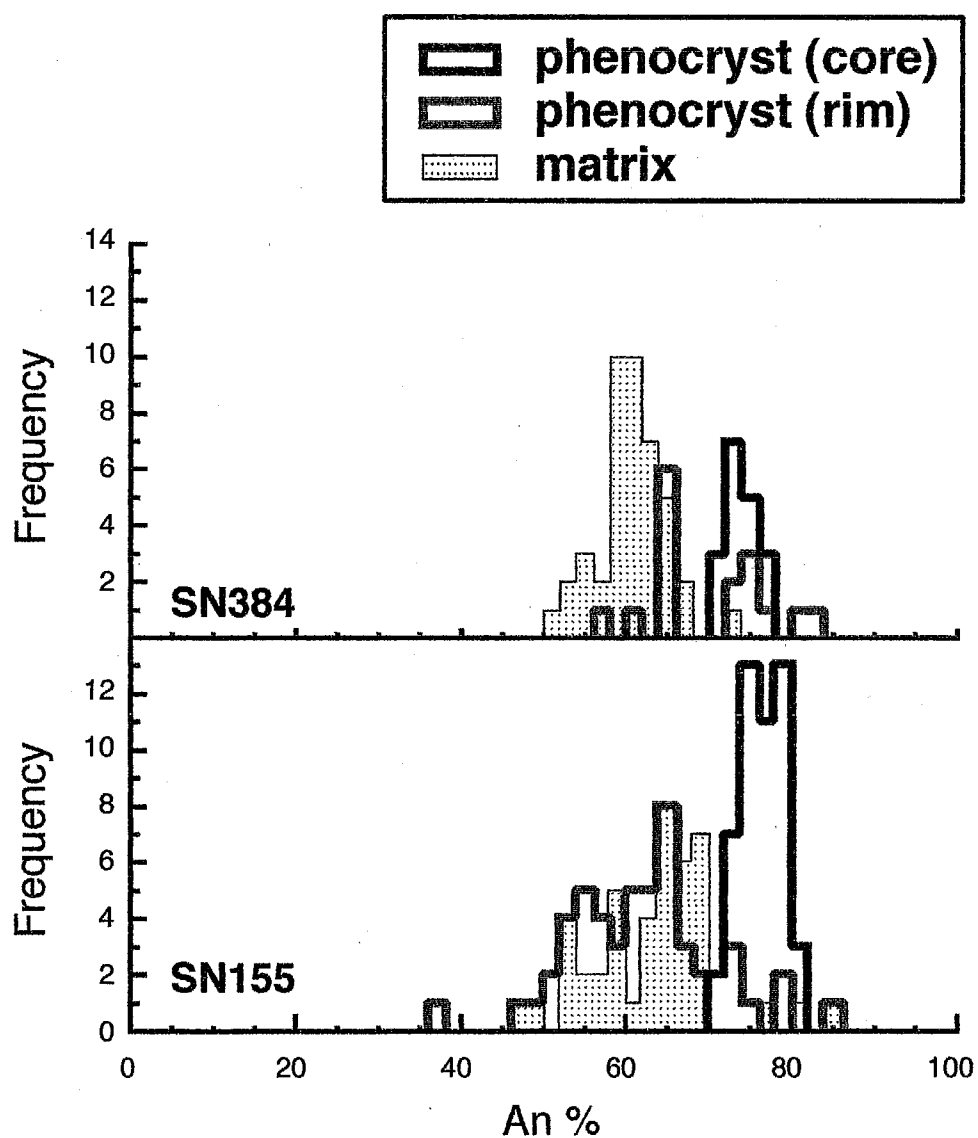


Fig. 2-24. Frequency of An content of High-TiO₂ type (SN384) and Low-TiO₂ type (SN155). Legends are shown in inset.

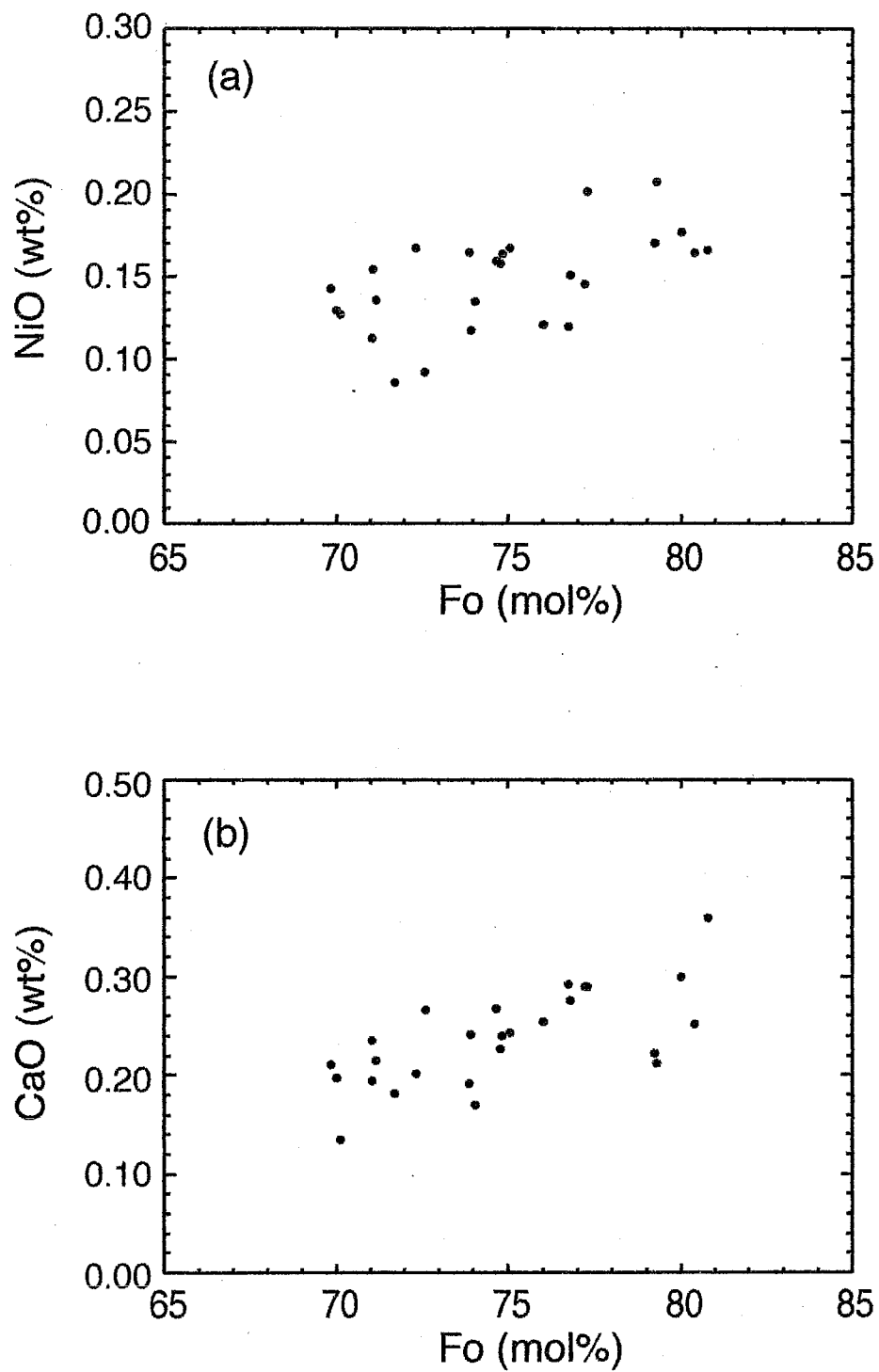


Fig. 2-25. Olivines from the olivine gabbro. Fo vs NiO_2 (a) and CaO (b).

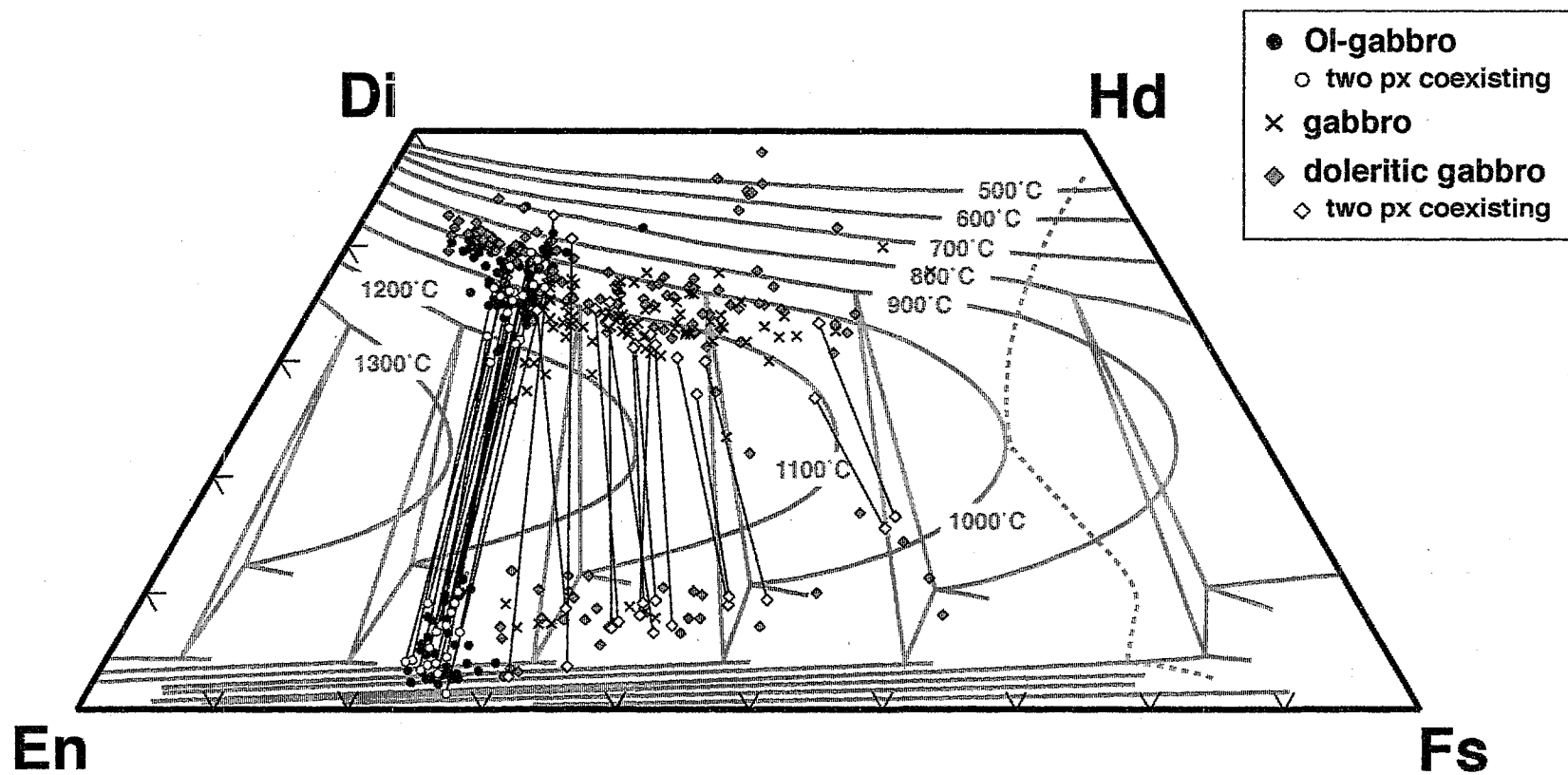


Fig. 2-26. Clinopyroxene compositions are plotted on Di-En-Fs-Hd quadrilateral diagram. Symbols are shown in inset.

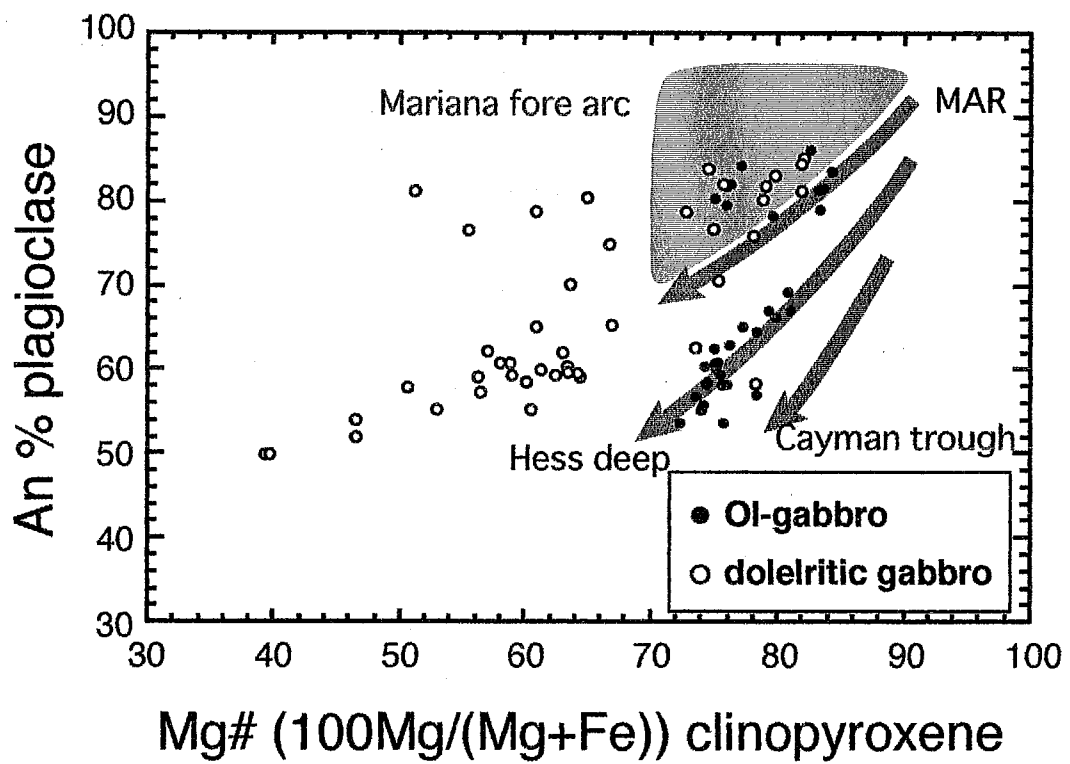


Fig. 2-27. Mg# of clinopyroxene vs An content of plagioclase. Symbols are shown in inset. Reference field and trends are from Bloomer et al., 1995, Dick and Natland, 1996.

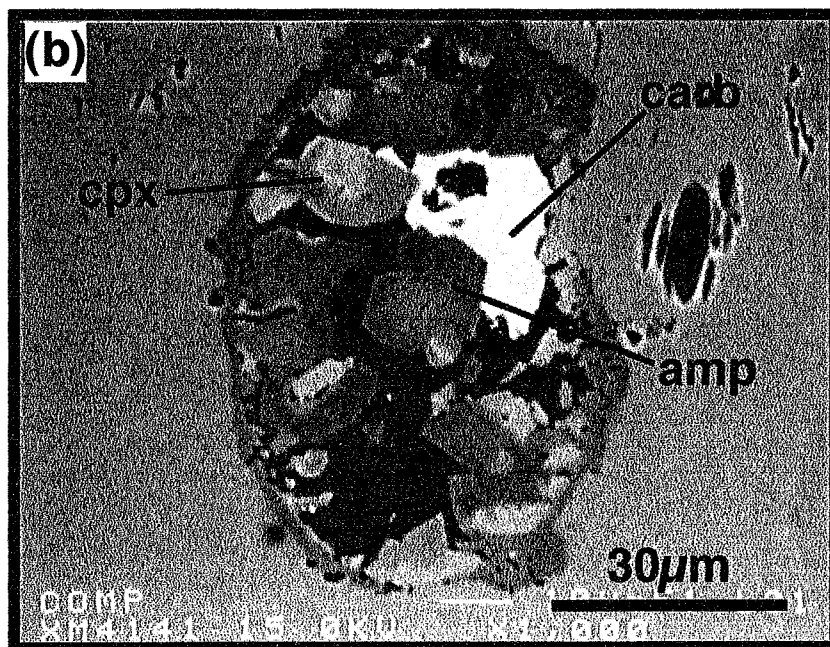
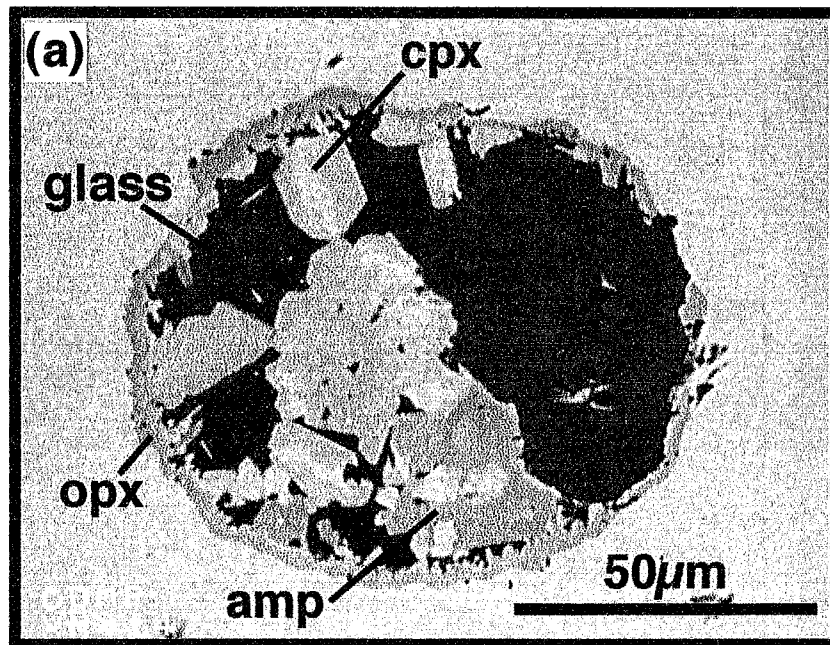


Fig. 2-28. The textures of melt inclusions. These are BEI. Scales are shown in each plate.
 cpx: clinopyroxene, opx: orthopyroxene, amp: amphibole, carb: carbonate (siderite)

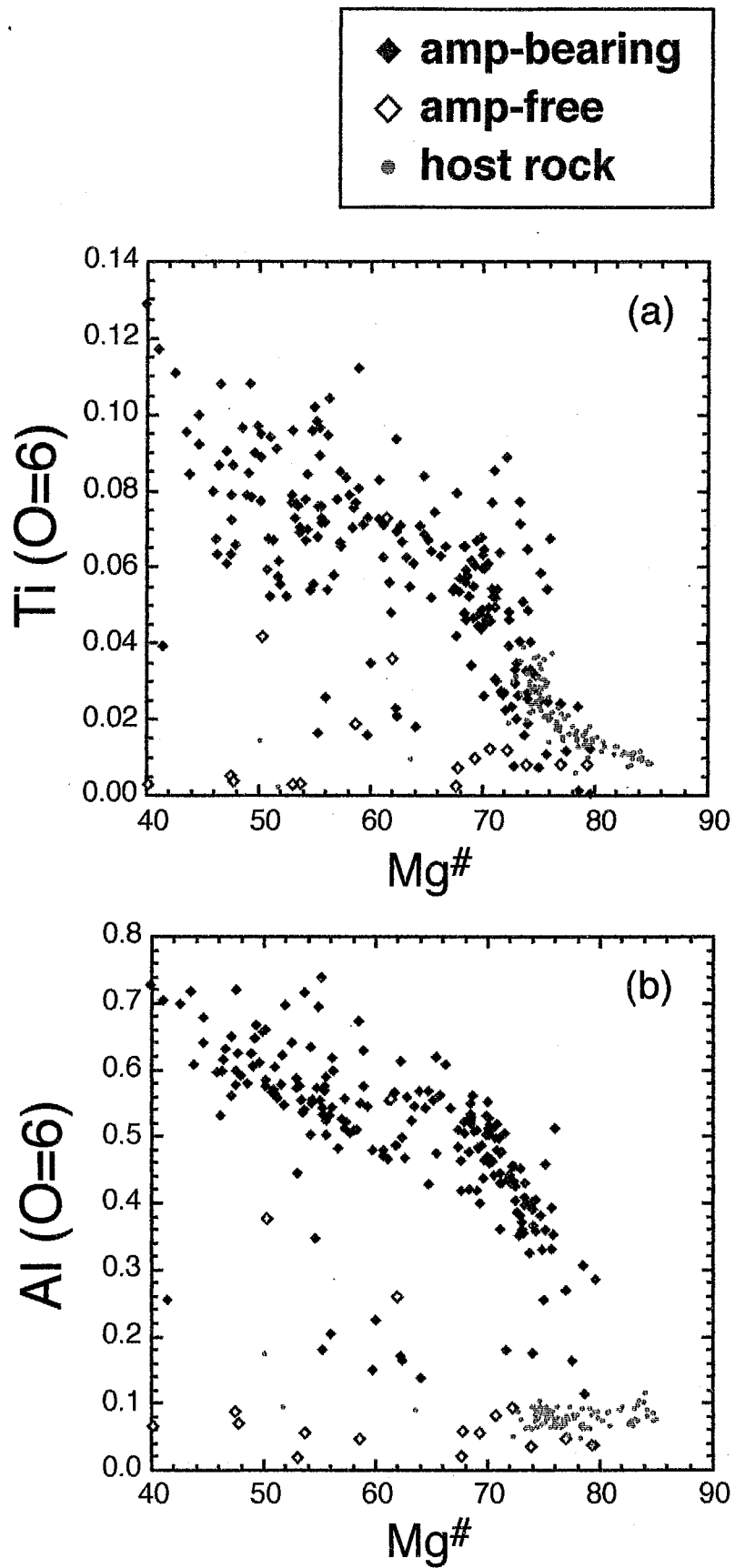


Fig 2-29. $\text{Mg\#}$ vs Ti (O=6) and Al (O=6) diagrams. Daughter pyroxenes are compared with the clinopyroxenes of host rock. Symbols are shown in inset.

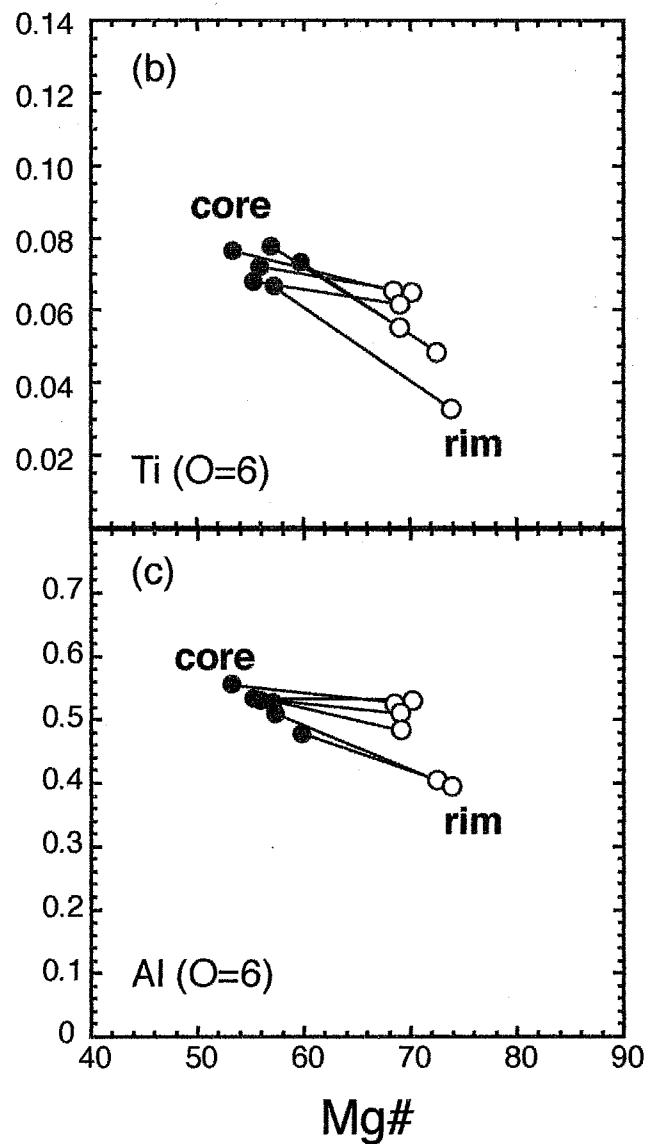


Fig 2-30. Representative composition of cpx of melt inclusions. (a) BEI of melt inclusions. plots are from it. (b) and (c): Ti (O=6) and Al (O=6) against Mg#, respectively. Tie lines show the identical daughter clinopyroxene.

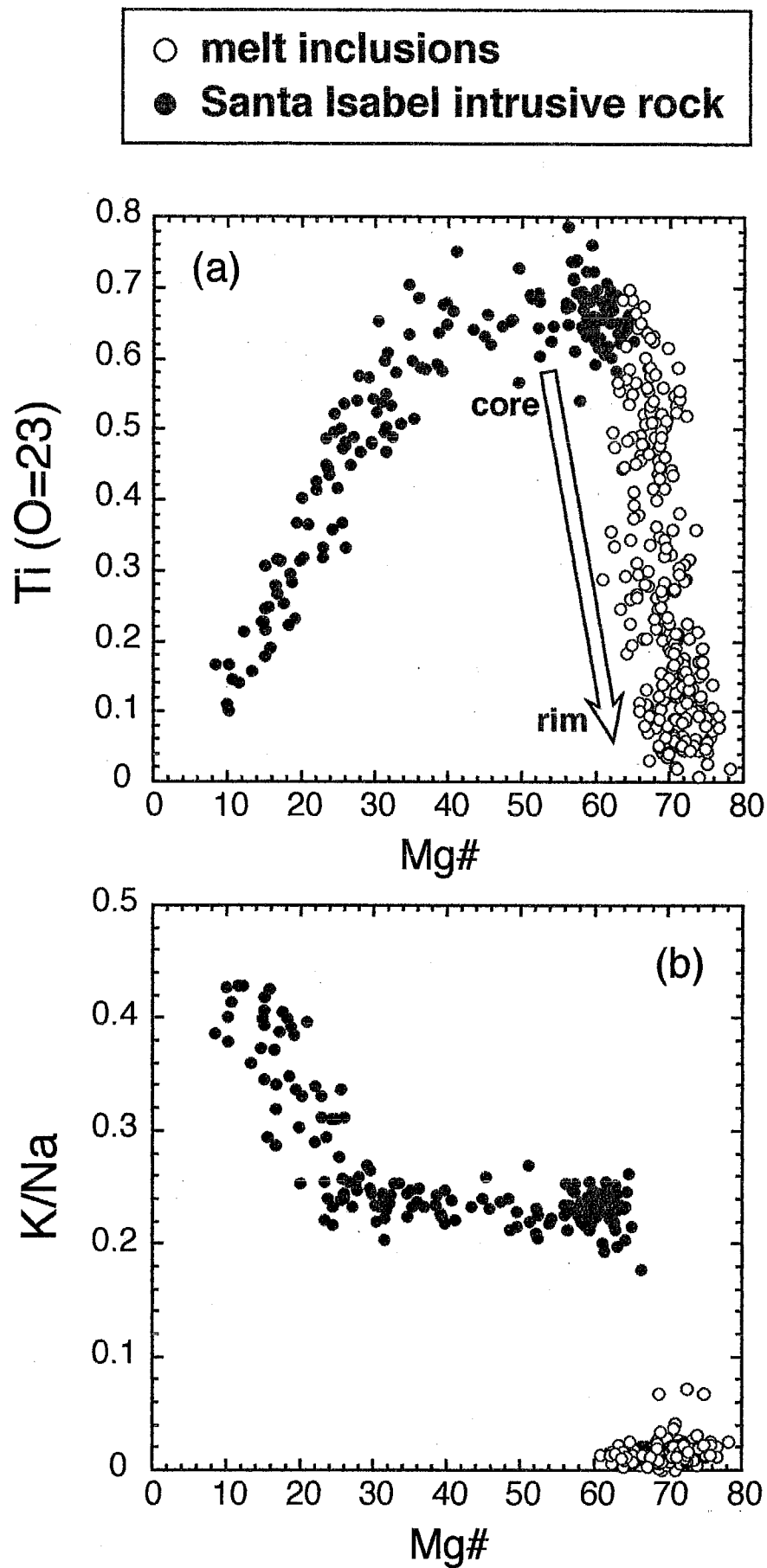


Fig 2-31. Amphiboles in melt inclusions. Ti (O=23) and K/Na against Mg# of amphibole in melt inclusions compared with those from Santa Isabel alkalic rock. Symbols are shown in inset.

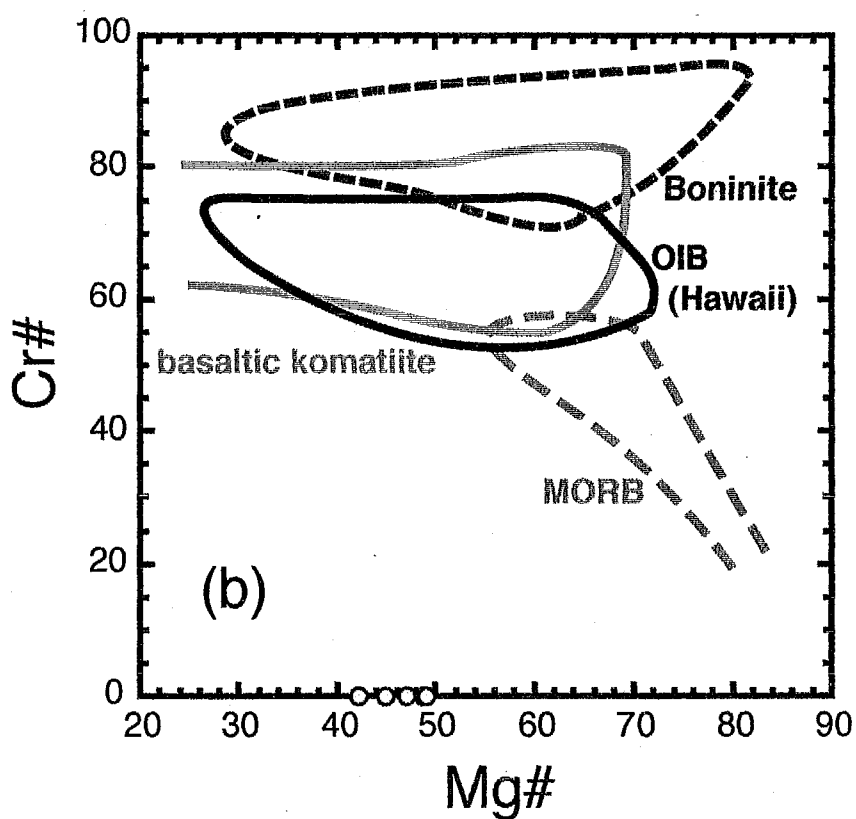
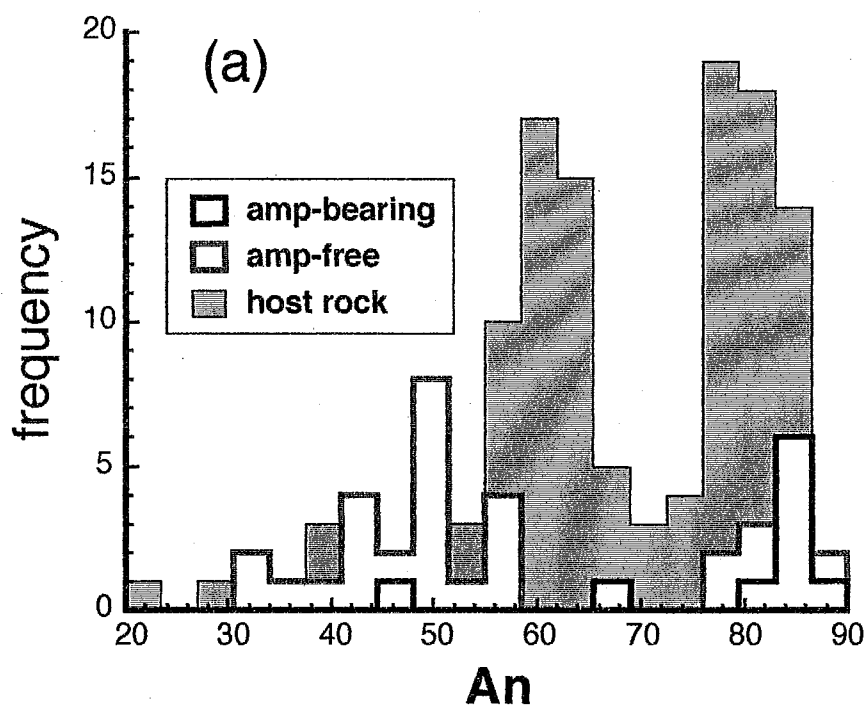


Fig 2-32. Aluminous phases in melt inclusions. (a) Frequency of An contents of plagioclase in the melt inclusion. Symbols are shown in inset. (b) The composition of spinel in melt inclusions. The fields are from Garcia and Norman, 1999.

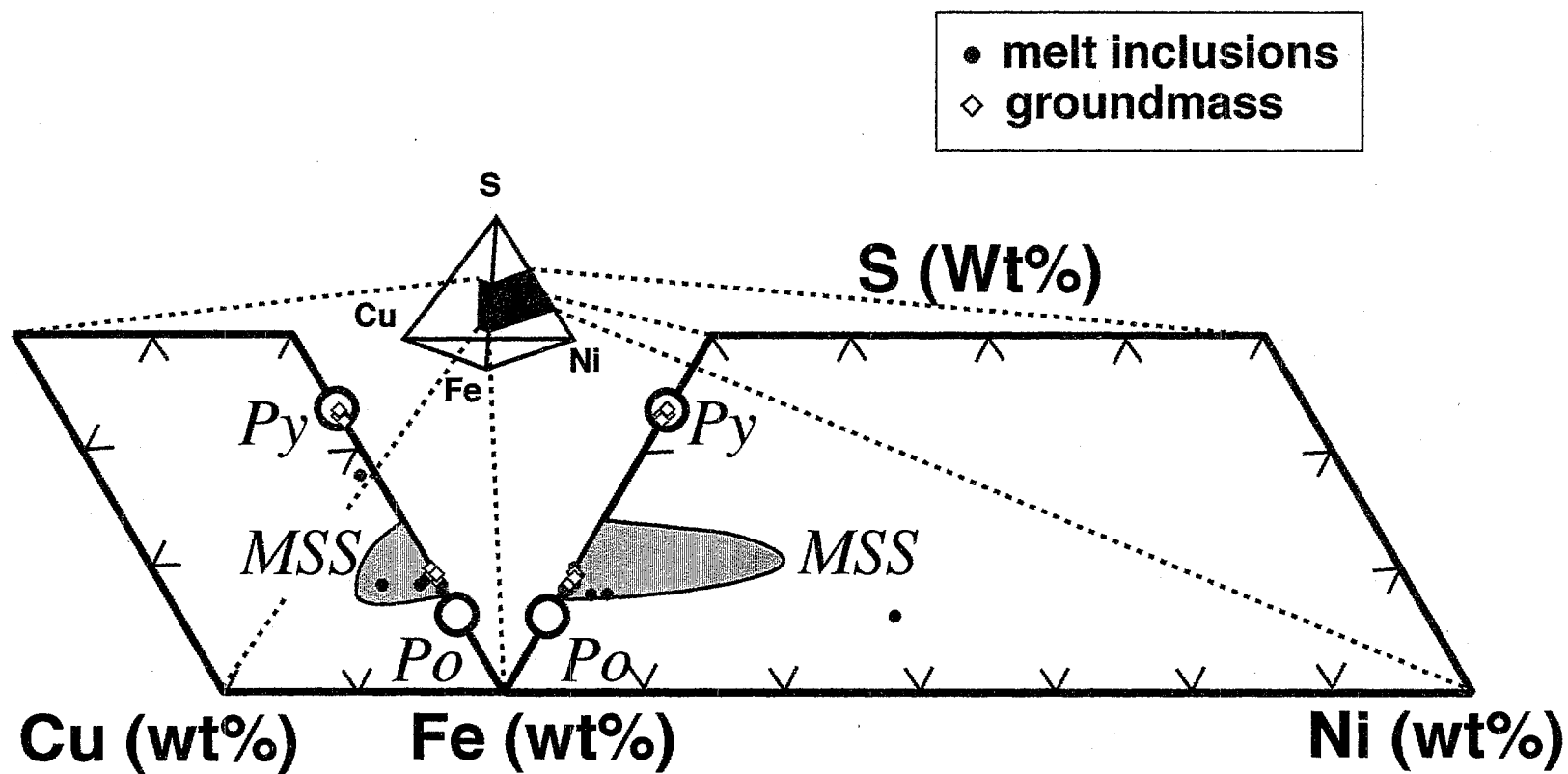


Fig 2-33. Sulfide compositions in melt inclusions. Symbols are shown in inset.
Py: pyrite, Po: pyrrhotite, MSS: monosulfide solid solution.

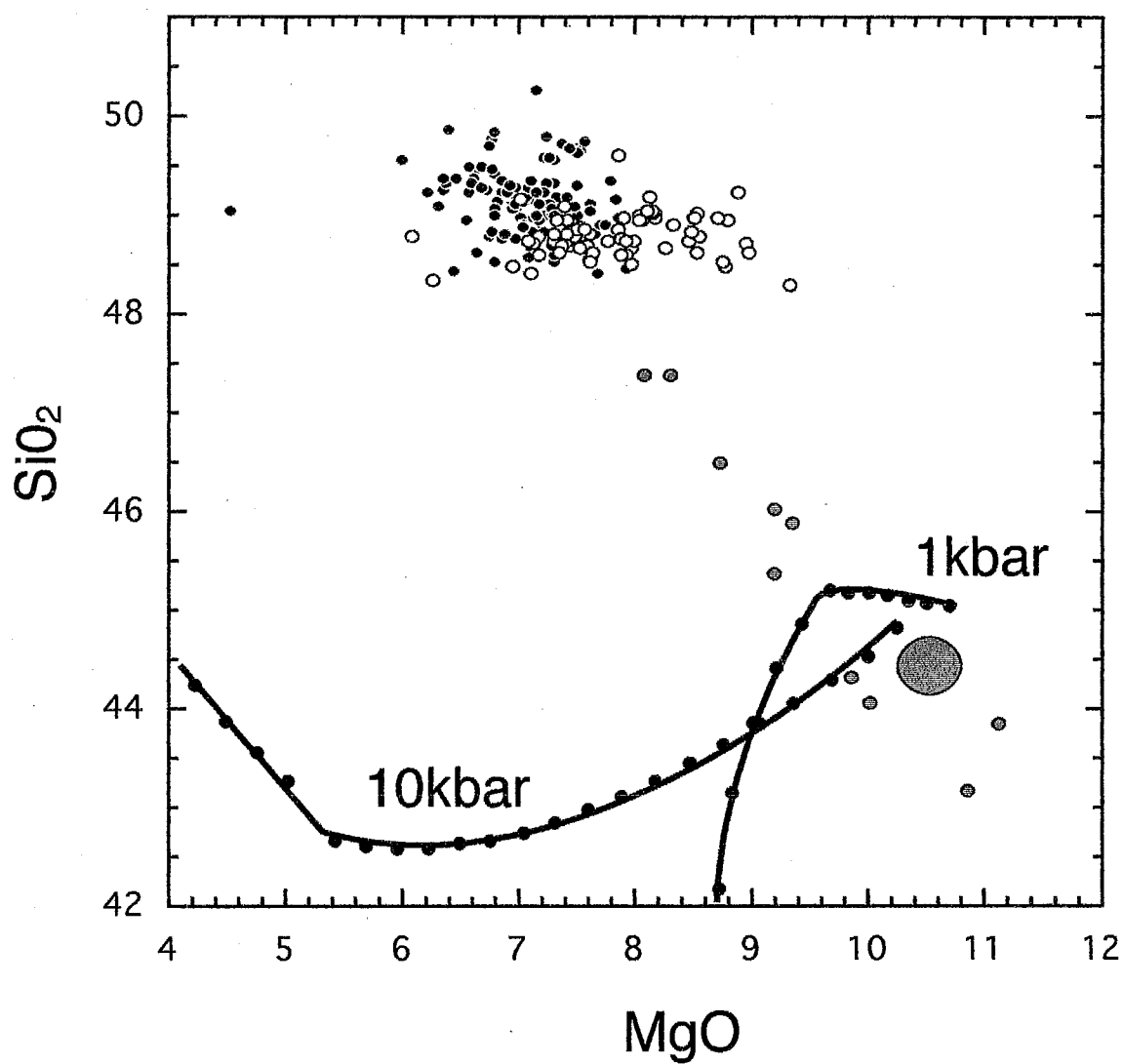


Fig. 2-34. High-MgO type lavas compared with MELTS simulations at 1 and 10 kbar. large grey circle is starting compositions of simulation. Other symbols are shown as Fig. 2-15.

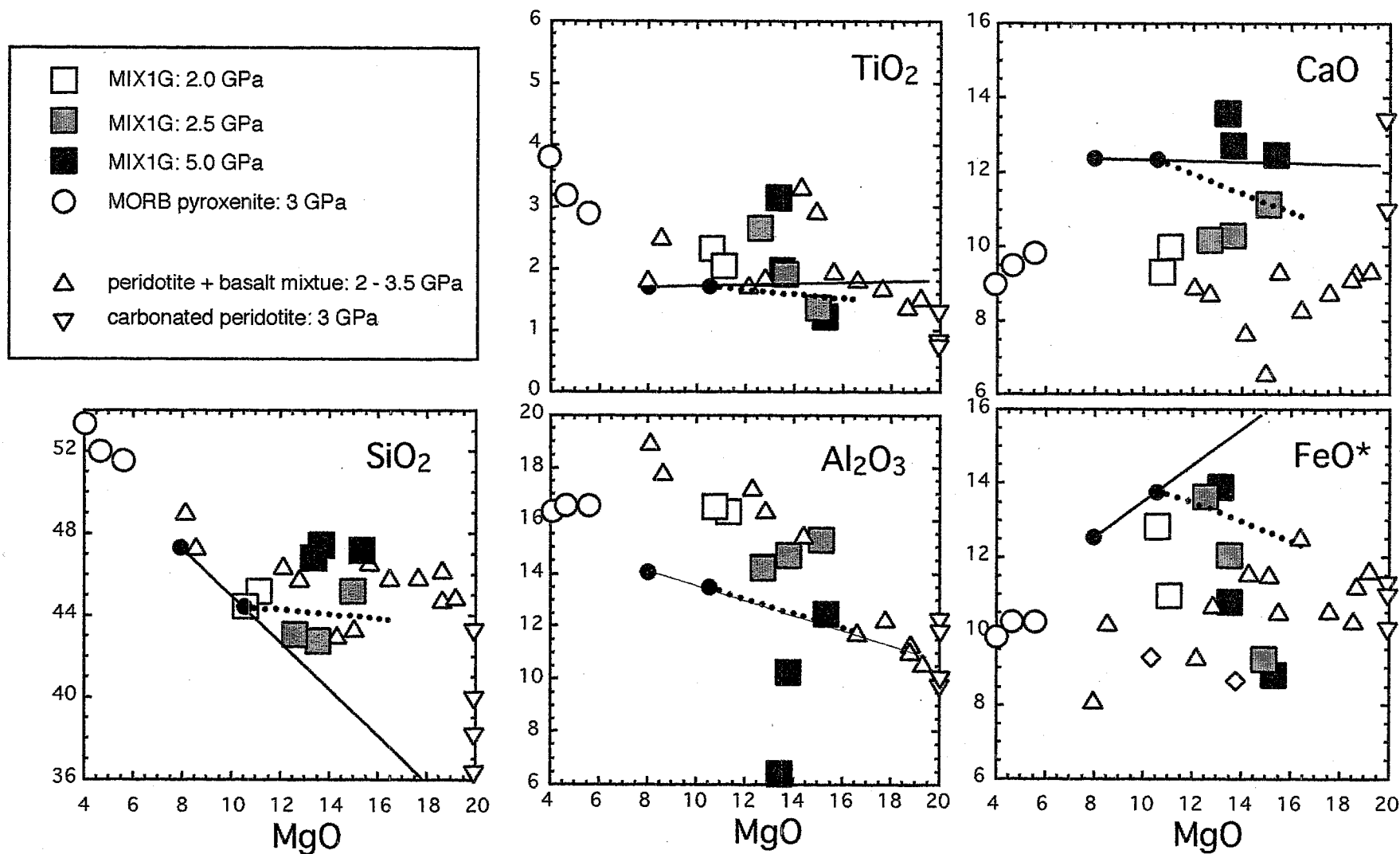


Fig. 2-35. Major element compositions of High-MgO type lavas compared with high-pressure melting experiments, large black circles shows the high-MgO type lavas and solid line show mixing line. Dot line shows the olivine accumulation. Data set of experiments are from Hirschmann et al. (2003), Pwttermann and Hirschmann (2003), Hirose (1997), Yaxley and Green (1998), Kogiso et al., (1998), and Kogiso et al., (2003). Carbonated peridotite partial melting are predicted down to 20 wt% MgO along olivine fractionation.

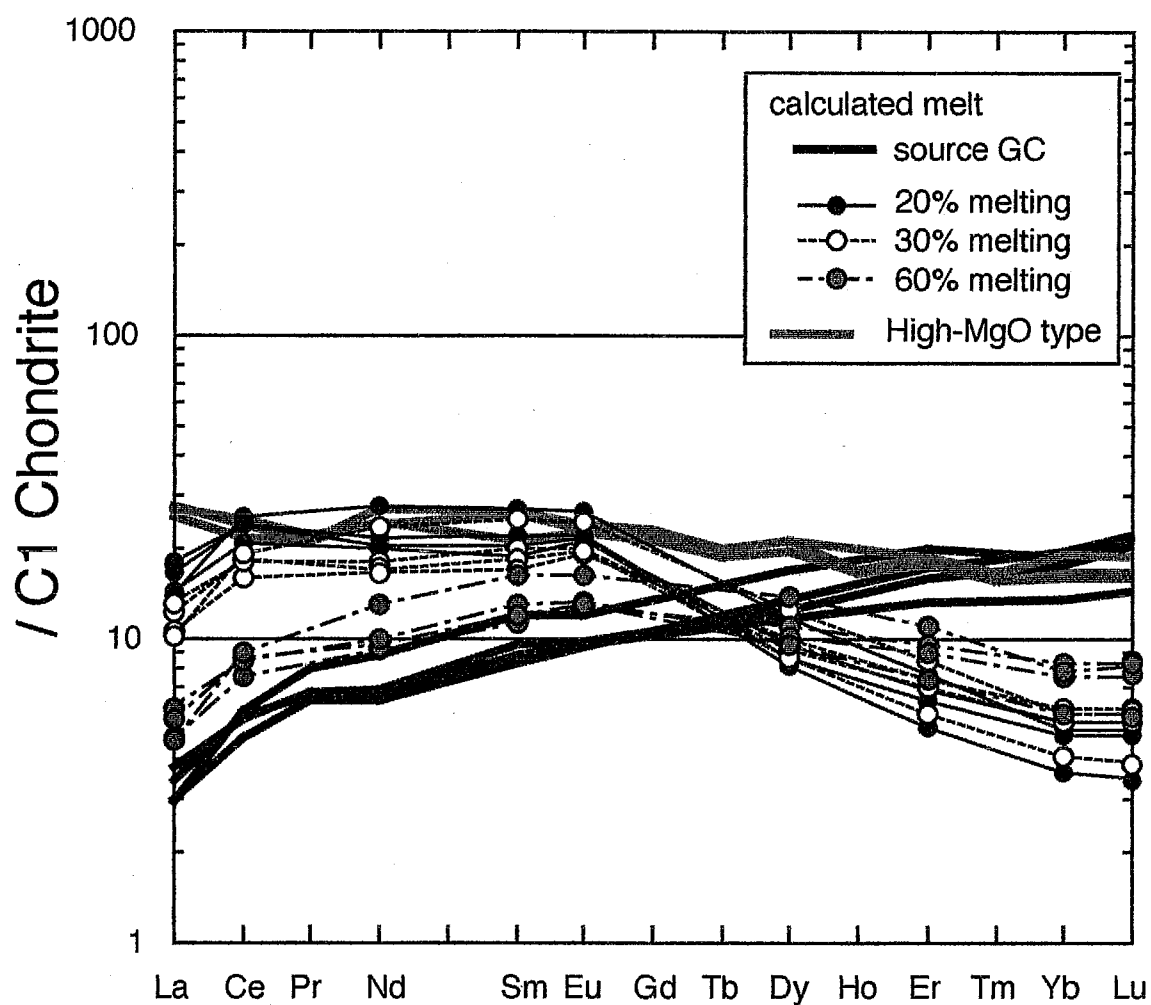


Fig. 2-36 REE patterns of calculated melts from garnet clinopyroxenite. as source garnet clinopyroxenite (GC), xenoliths in alnöite from Malaita are used and compositional data from Ishikawa et al. (2004). Detailed method and assumptions are mentioned in text. symbols are shown in inset.

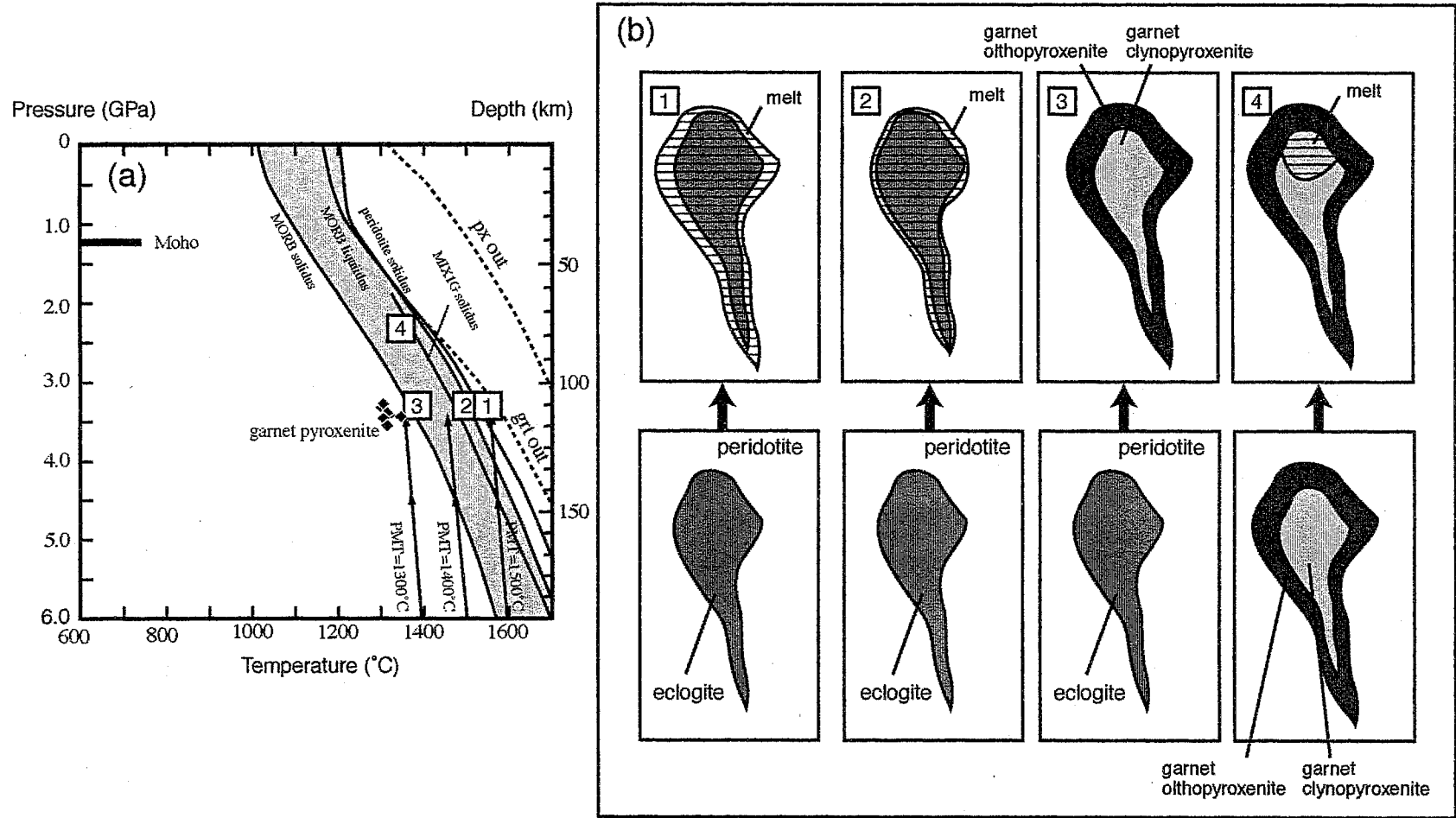


Fig. 2-37 The possible melting for OJP basalts in heterogeneous mantle plume. (a) Liquidus and Solidus for primitive MORB (Yasuda et al., 1994), solidus of fertile peridotite KLB-1 (Takahashi et al., 1993) and solidus of garnet pyroxenite MIX1G (Kogiso et al., 2003). P-T conditions of garnet pyroxenite are from Ishikawa et al., 2004. (b) Melting of eclogite enveloped peridotite at each P-T condition. The melt of 1, 2 and 4 corresponds to High-TiO₂ type, Low-TiO₂ type, and High MgO type, respectively.

Chapter III. Paleogeographic reconstruction of the Pacific domain since Late Proterozoic: History of the Pacific superplume

Abstract

Paleogeography of continents, oceans and oceanic large igneous provinces (LIPs) since 1000 Ma were reconstructed in order to decode history of the Pacific superplume activities, on the basis of the most confident recent data for accretional geology from orogens associated with ophiolite and greenstone belts, and geochronologic and paleomagnetic data from each continent in literature.

Several conclusions were led through our reconstruction; (1) Pacific superplume was born at 750 Ma to break-up supercontinent Rodinia. (2) African superplume was born at 250-200 Ma to break-up Gondwanaland. (3) Episodic activity of Pacific superplume was present with culminations at 400-500 Ma, 300-250 Ma and 150-75 Ma. (4) Presence of collage of immature oceanic island arcs at 900-600Ma in Sahara-Saudi Arabia, at 600-500 Ma in Kazakhstan Block, at 200-100 Ma in Alaska-Kamchatska region, and since Miocene in the western Pacific. These brought juvenile arc crust through convergent plate boundary process of microplates. (5) Role of Tethyan-Indian Ocean in Wilson cycle. Break-up, dispersion and amalgamation of continents was not simple reversal as suggested by Hoffman (1991). Instead, the Tethyan-Indian Ocean played an important role, which is to form double-sided convergent boundary around continental lithosphere or microplate. Opening and closing of the Iapetus Ocean were contemporaneous to break-up of Rodinia and to formation of Gondwana, respectively. At c.1000Ma, the Grenville orogeny, which formed the Rodinia, prevailed to form the Pacific-type orogens along active region of the Pacific superplume. These suggest that the numerous volumes of subducted lithospheres along the Pacific-type orogens triggered the birth of the Pacific superplume.

Key words: Paleogeographic reconstruction, supercontinents, accretional geology, Pacific superplume, plume tectonics

3-1. Introduction

The development of paleogeographic reconstruction is closely correlated with the spread of plate tectonic paradigm, however, the revolutionary ideas related with supercontinents were provided prior to the consensus that continents are practicable to drift. In 1885, E. Suess (August 20, 1831-April 26, 1914), who was an Austrian geologist, proposed the existence of supercontinent named Gondwanaland incorporating India, Africa, Madagascar, and South America based on observation of distribution of the *Glossopteris* flora and glacial sediments in the Southern Hemisphere. Wegener (1912) proposed the existence of a supercontinent in Permo-Carboniferous on the basis of the close match of opposite coastlines of continents and regional extent of the glaciations in the Southern Hemisphere. Furthermore, Wegener (1929) proposed the idea that the former supercontinent had broken up and individual continents have drifted to their present positions. Based on morphological fit and geological correlations, Du Toit (1937) proposed the existences of two supercontinents, Laurasia and Gondwanaland, separated by the Tethys, which prevented mixing of the northern and southern flora.

The concept of continental drift came back to the central of Earth dynamics after plate tectonics, and paradigm in the Earth dynamics has shifted from plate tectonics to plume tectonics (Fig. 3-1). In 1960's, discovery of a symmetric arrangement of geomagnetic banding and transform faults on oceanic crust at mid-oceanic ridges realizes the sea-floor spreading theory, which gave rise to the plate tectonic paradigm (Vine and Matthews, 1963; Wilson, 1965). With this paradigm, Wilson (1966) depicted a history of the ocean from beginning to end (Wilson cycle; Burke et al., 1976). After 1970's, studies of paleogeographic reconstruction in the Phanerozoic have been in the ascendant, with the theory of sea-floor spreading and the continental drifting (e.g., Smith et al., 1981; Scotese and McKerrow, 1990; Ziegler, 1989; Lawver and Scotese, 1987; Scotese and Sager, 1988; Smith et al., 1994). These studies were developed with morphological fitting, geological correlations among each continent, (paleo-) magnetic data and the fauna correlation, then a reconstruction of the Permo-Carboniferous supercontinent, Pangea, was established (e.g. Livermore et al., 1986). Such a reconstruction could visualize the plate tectonic theory, and

we had a deeper understanding of the theory itself by a feedback of variable information obtained during processes of the reconstruction; for example, processes of continental dispersion and convergence described by the paleogeographic reconstruction are much more complicated than those explained by the Wilson cycle. In 1990's, the reconstruction has been extended to the Proterozoic paleogeography, and the Late Proterozoic supercontinent, Rodinia, has originally been proposed as the supercontinent that begat all subsequent continents by McMenamin and McMenamin (1990). Later, the Rodinia, with a different configuration from that drawn by McMenamin and McMenamin (1990), was suggested through a correlation of the Grenville orogen in each continent (Hoffman, 1991; Dalziel, 1991). Recent years, reliable and more comprehensive paleomagnetic data than before in the Late Proterozoic have been accumulated (e.g., Weil et al., 1998), and relative positions and pale-latitudes of each continent have been determined and modified. However, reconstruction models of the Late Proterozoic supercontinent have been proposed differently by researchers (e.g., Pannotia by Powell et al., 1993; Paleopangea by Piper, 2000), and do not reach a consensus, yet.

Orogenic processes such as break-up and formation of supercontinent have been discussed within a framework of plate tectonics. However, the plate tectonic theory rules a dynamics only at subsurface domain of the Earth. On the other hand, the recent advance of seismic tomography, ultrahigh-pressure experiments, numerical simulation of mantle convection is clarifying dynamics of whole mantle, and resulted in a new paradigm (plume tectonics: Maruyama, 1994). A key of this new paradigm is plume tectonics, that is a dominant style of convection in the major part of mantle where two super upwellings and one super-downwelling prevail; it comprehensively demonstrates the whole Earth dynamics and interactions among multi-sphere including not only the solid Earth but also bio- and magneto-spheres (Maruyama, 1994; Fukao et al., 1994; Kumazawa and Maruyama, 1994).

The super upwelling in the mantle, i.e. "superplume", was first invoked in the early stages of seismic tomography (Dziewonski, 1984; Woodhouse and Dziewonski, 1984; Dziewonski and Woodhouse, 1987), which revealed two low-velocity domains of seismic waves extending continuously between the uppermost and lowermost mantle, originating under the south Pacific and Africa. These thick cylindrical structures with a girth extending

on the order of 1000 km are much thicker than a skimpy jet plume to occur so-called hot spot on the Earth's surface.

Since the early 1990's, the heterogeneity of the velocities in the mantle was more clearly drawn by seismic tomography with higher resolution than before (e.g., Fukao, 1992). The lower velocities area in lower mantle are considered to superplume that is the huge mantle plume from the core-mantle boundary (CMB) and the higher velocities area around 670 km depth are regarded as megalith that is massive stagnated slab. These images provide a new idea that is named 'plume tectonics'. This idea is that the whole Earth dynamics functionates with interactions among (1) the subsurface domain controlled by the plate tectonics, (2) the lower mantle domain by the plume tectonics, and (3) the growth tectonics in the core, which are controlled by episodic superplume activities. The plate tectonics supplies cold slab flush into the mantle domain. The cold slab stagnates in mantle transition zone for 100-400 m.y., subsequently collapses and falls on the surface of the outer core, resulting in episodic super-downwelling in whole mantle and a new down flow in the core. A counter flow of the super-downwelling, i.e., a super-upwelling of hot mantle causes continental break-up, subsequent dispersion, and formation of large igneous provinces (LIPs) including flood basalts, hotspot chains and oceanic plateaus. Fragmented continents newly form a huge composite continent through collision and subsequent amalgamation. During such a process, subducting slab descends to the mantle through trenches around the new continent, and prepares a next super-downwelling beneath the continent.

From a viewpoint of the plume tectonics, histories of supercontinents and LIPs are closely related to the superplume activities. In other words, geological records on the supercontinents and the LIPs have essential information to clarify the history of the superplume activity.

In this paper, paleogeographic reconstruction of continents and oceans since 1000 Ma is discussed, on the basis of the most confident recent data on regional geology, geochronology and paleomagnetism from each continent, in order to understand an episodic activity of the South Pacific superplume and interactions between the plume activity and tectonics on the Earth's surface. Objects in our reconstruction include distribution of not

only continents, but also mid-oceanic ridges, trenches; in addition, the ancient LIPs, which are dispersed in oceanic regions or are partly trapped in orogens at the present, are also reconstructed with consideration of accretionary geology.

3-2. Methods and data sources for reconstruction

First, we have to evaluate and rearrange a large number of data set for the paleogeography, accumulated by many previous workers. Because the accumulated data sets differ in quality from one another, and the data for paleogeography in the older periods are established with lesser confidence. In this study, we supplemented the accumulated data by previous works with the following methods and data sources.

3-2-1. Continental reconstruction

3-2-1-1. Outline of the world geology

Figure 3-2 is a compilation map showing world distribution of geologic units in both of continents and oceans with different ages, based on published data; major data sources are Liou et al., 1990, Windley (1995), Maruyama et al. (1996), and references in those. A general view of the compiled map shows that a continent is composed mainly of Precambrian nuclei (cratons) and orogenic belts with zoned growth toward their outwards.

Orogenic belts are formed at divergent and convergent plate boundaries; those in the former setting are characterized by continental rift zones. Rifting event is accompanied with the large-scale magmatism. Their traces are remained as dyke swarms on drifted continental margins. At an initial stage of continental rifting, dike swarms subvertical to the rifting axes are developed extensively, as best documented along the Mesozoic coast line in the Atlantic Ocean (Ruffman and Greenough, 1990; Hill, 1991). Their intrusive ages and

distributions have been utilized as essential information for continental reconstruction.

On the other hand, orogenic belts at convergent plate boundaries are divided into Pacific-type, which is developed by oceanic plate subduction, and subsequent collision-type associated with continental and/or arc collisions (e.g., Maruyama, 1997); they are ones of the major constituents of continents. The Pacific-type involves regional high-P/T metamorphic unit, huge TTG (tonalite-trondjemite-granodiorite) batholith with calc-alkaline volcanic rocks, accretionary complex, and B-type protolith of oceanic crust (e.g., mid-oceanic ridge basalt), trench-fill turbidite, reefal organic limestone, and pelagic bedded chert. In contrast, the collision-type is characterized by a large high-P/T metamorphic unit with ultrahigh-P relics, no TTG, foreland fold-and-thrust belt, and A-type protolith including bimodal volcanic rocks, fragments of continental basement, and platform-type peraluminous sediment and carbonate. In both types, orogenic culmination is represented by exhumation of the high-P/T units to form orogenic cores, which would result from subduction of mid-oceanic ridge in the Pacific-type and from that of microcontinent in the collision-type (Maruyama et al., 1996; 2002). In addition, the high-P/T units are exhumed towards opposite direction against plate subduction. Such a structural relation between the high-P/T units and its surroundings provide constrain to subduction polarity at the plate boundary at that time.

Therefore, compilation of orogenic belts at convergent plate boundaries was performed with attentions to the regional high-P/T metamorphic units. Consequently, spatiotemporal distribution of the Pacific-type orogens allowed us to reconstruct paleo-positions of mid-oceanic ridges and trenches.

3-2-1-2. Mesozoic-Cenozoic continental reconstruction (~200 Ma)

A general view of the compiled map of orogenic belts worldwide (Fig. 3-2) allows us a qualitative continental reconstruction. As a next step, information recorded in oceanic domains is applied in order to make it more precise reconstruction with quantitative paleo-positions and configuration.

Continental reconstruction for the recent 200 m.y., i.e., since breakup of the latest supercontinent Pangea, has been well known by numerous previous works. Classically, the continental reconstruction of this period has been established with data of the fauna and the glacial drift on continents, morphological fitting of continental outlines, and complementary correlation of orogenic belts. The geomagnetic data from oceanic crust and paleomagnetic data from continents made these discussions higher quality. Because the Pangea reconstruction corresponds to reconstruction of paleogeography before opening of the Atlantic and the Indian Oceans, applied procedures are not so complicated, compared with the reconstruction of older periods. Therefore, the distribution of major continents is practically completed. Adding to it, we corrected the distributions of microcontinents in Asia and New Zealand based on the recently insights (Maruyama et al., 1989; Malpas et al., 1994).

3-2-1-3. Paleozoic continental reconstruction (~540 Ma)

For a period before 200 Ma, there is no reliable data from oceanic domains, because ages of the oceanic plates in the present Earth's surface are younger than 200 m.y., and the older oceanic plates had been subducted into mantle by the operation of plate tectonics. Therefore, the paleogeographic reconstruction in this period depends on continental reconstruction models with correlation of orogenic belts and paleomagnetic approach with analyzing the apparent polar wander pass (APWP); correlation of paleontological data are still useful for the reconstruction during the Late Paleozoic.

Since geological inspection can merely recognize relative continental paleo-positions, paleomagnetic data (APWP) should be needed in order to obtain absolute ones. However, accumulated data for paleomagnetism differ in quality and quantity from each continent or region. Except for the best-studied North America (Laurentia), limited data for the APWP from a few regions or of relatively short time intervals are available. Therefore, correlation of orogenic belts is essential information for the reconstruction in this period, too.

A period of 200-540 Ma corresponds to the formation stage of the Pangea after

breakup of a (semi-)supercontinent Gondwana (c. 550 Ma). As described in the previous section, the Caledonian Orogeny (Middle Paleozoic), and the Hercynian and the Appalachian Orogenies (Late Paleozoic), which have been proceeded during closing processes of the Iapetus Ocean and the Tethys, respectively, are recognized over the world; central Asia, North America, Europe, and parts of South America, Africa, Antarctica and Australia. Therefore, available data sources are sufficient for the correlation of orogenic belts formed in this period, compared with those in much older periods as described below.

3-2-1-4. Late Proterozoic continental reconstruction (~1000 Ma)

For a period of 550-1000 Ma, even continental reconstruction is scarcely described except for the semi-supercontinent Gondwana and a supercontinent Rodinia (1000-750 Ma). The Gondwana has been reconstructed based on distribution of so-called Pan-African orogenic belts (c. 500-800 Ma: Hoffman, 1991) in Africa, Australia, South America, and Antarctica, and paleomagnetic data obtained from those continents. Recent years, the Pan-African orogenic belts in Africa, South America, and central Asia have been minutely (re-)examined for their structural geology, petrology, and geochronology (e.g. Katayama et al., 2001), and available data for the paleogeographic reconstruction have been newly accumulated. We took those recent data in order to modify the previous reconstruction models and to reconstruct the paleogeography of microcontinents, which have not been included in the previous models.

The Rodinia reconstruction has been based on geological evidences that include linkage of the Grenville orogen (c. 1300-1000 Ma: e.g., Moore, 1991), and on paleomagnetic data (APWP). However, to remove the Phanerozoic orogens that connect with Precambrian cratons from the present continents, over twenty of the isolated cratons appear. Procedures on the reconstruction become more complicated than those for later periods, and quite a few data on geology and paleomagnetism are available for some of the cratons. Moreover, there seem to be greater uncertainties in many of the Proterozoic paleomagnetic data than those of Phanerozoic ones (Van der Voo and Meert, 1991). Its most important dependence is poor control on the age of magnetizations. Therefore, our

reconstruction in the period from the break-up of Rodinia to the formation of Gondwana was primarily performed with geological and geochronological data of orogens and subordinately adopted the latest paleomagnetic data (e.g., Weil et al., 1998; Wingate and Giddings, 2000). In the cases that the geological and geochronological data of orogen were not consistent with and supported by paleomagnetic data, we regarded the geological records in the orogens as the most important constraint. If no reliable geological record, we adopted more reliable and inclusive data in the latest literature on paleomagnetism.

3-2-2. Paleogeographic reconstruction of oceanic domain

Based on existences of magnetic stripes on the oceanic crust, distributions of fracture zone, and assumption that hotspots in the Pacific and Atlantic Oceans are fixed with respect to one other, relative motions between major plates and distribution of mid-ocean ridges and trenches in Pacific Ocean are provided (Engelbreton et al., 1985). Paleopositions of oceanic plateaus dispersed in Pacific Ocean are reconstructed by using its relative motions model (Larson, 1991a). These reconstructions are available to the recent 140Ma, because the most part of the older oceanic crust has disappeared into the mantle at subduction zone. But there remains a few amount of the fragment of the old oceanic plate in the Pacific-type orogens. These fragments occurs as on-land ophiolite/greenstone belts (Fig. 3-3). In this study, we extracted the information of the oceanic domains from the Pacific-type orogens and reconstructed the paleogeography of oceanic domains that including the distribution of mid-ocean ridges, trenches, oceanic islands, and oceanic plateaus.

3-2-2-1. Extraction of oceanic information from Pacific-type orogens

3-2-2-1-1. Accretionary geology

Compilation of the Pacific-type orogens allowed us to reconstruct paleo-positions of trenches. Moreover, the accretionary complex in the Pacific-type orogen is a repository of information in oceanic regions, because it captures parts of materials on the oceanic plate, subducting through a trench.

Complicated structures of the accretionary complex have been decoded by microfossil mapping (Matsuda and Isozaki, 1990). Constituents of accretionary complex are divided into two groups: One is composed of sedimentary rocks derived from terrigenous clastics, and another comprises pelagic and hemi-pelagic sedimentary rocks and mafic rocks from oceanic region. Both groups are distributed as either coherent-type or imbricated-type of olistostrome or *mélange*. However, rocks and strata in the accretionary complex preserve primary stratigraphical relations, which can be reconstructed by the age of microfossils, even in a *mélange*-type (Taira et al, 1989).

Geological structures in the coherent-type accretionary complex are characterized typically by duplex structure (Fig. 3-4). The duplex structure, which is one of imbricate structures generally observed in the modern accretionary complex, has an internal structure of each faulted-package (horse) subparallel or oblique to the surface thrust (roof and floor thrusts) separating each duplex structure by link thrust. Each horse in the duplex preserves primary stratigraphical relations of oceanic materials capped by trench-fill turbidites. Ages of constituents in the horses of particular duplex become younger oceanward, because of their oceanward growth with time: moreover, the duplex develops downward with time. The primary stratigraphy of constitutive rocks of accretionary complex is called an oceanic plate stratigraphy (OPS), which is compared to stratigraphical relation of rocks and strata of the upper part of the present oceanic plate being subducted at the trench. The standard OPS is composed of greenstone with mid-oceanic ridge basalt (MORB) in origin, pelagic sediment (bedded chert), hemi-pelagic sediment (siliceous mudstone), and terrigenous clastic sediment, in ascending order. In some cases, rocks originated in off-ridge volcanism such as oceanic island, seamount and oceanic plateau associated with reef limestones are included in the accretionary complex. Such rocks in the Phanerozoic accretionary complex have been reconstructed on the basis of their lithology and microfossil ages (e.g., Akiyoshi limestones in the southwest Japan: Sano and Kanmera, 1988). Sedimentary rocks originated on the top of seamount or plateau record environmental change at shallow marine from when basaltic basements of the seamount and plateau are erupted to when they reach at trench.

Accordingly, the age of a horizon where the oceanic rocks covered conformably with

terrigenous clastic sediment approximate a timing of formation of accretionary complex at trench. And, a deposition time of the oceanic sediments, which is a time interval from the upper to the lower limits of pelagic and hemi-pelagic sediments, indicates the time range that the oceanic plate moved from the mid-oceanic ridge to the trench. That is to say, the deposition time interval of the oceanic sediments is the age of the oceanic plate, which subducts at the trench. Consequently, the OPS, which is now exposed on land, provides us information that how old the oceanic plate is, and when it formed the accretionary complex. As a high-precision method to determine these ages, microfossil dating with radiolarian stratigraphy in the pelagic and terrigenous sediments in Phanerozoic strata is valid (Isozaki and Maruyama, 1991). In the case for the Proterozoic strata, in which are difficult to obtain microfossils, the valid methods are U/Pb and Pb/Pb dating with heavy minerals such as zircon and bulk isochron. Recently, it was pointed out that the combination of the biostratigraphy and the isotopic chemostratigraphy has been enabled to correlate to the Proterozoic to Phanerozoic strata over the world. Secular variation curves of isotopic compositions of those strata were proposed based on their compilation (Kaufman and Knoll, 1995), and high precision correlations with carbon-, and strontium-isotopic stratigraphy have been become available (e.g., Veizer et al., 1997; Knoll, 2000; Walter et al., 2000; Melezhik et al., 2002).

In the case that the accretionary complex contains rocks derived from seamount or oceanic plateau, by determination of ages of carbonate rocks overlying greenstones, we can reconstruct the seamount and plateau, and obtain some information of environmental change during the oceanic plate migration.

3-2-2-1-2. Summary of the modern volcanic sequences

Information about oceanic materials, i.e., greenstones and associated sedimentary rocks, in the accretionary complexes is essential to reconstruct the oceanic paleogeography. In this study, we compared characteristics of the modern volcanic sequences in oceanic domains (Table 3-1) with those of oceanic materials extracted from the accretionary complexes, and presumed their primary tectonic settings including off-ridge (as oceanic islands, seamounts and oceanic plateaus), mid-ocean ridges, marginal basins, or island arcs.

In terms of geologic features in the modern tectonic settings, shallow-marine pelagic sediments such as limestone are suggestive of lithology conformably overlying topographic high in the mid-ocean, whereas deep-sea pelagic sediments with a significant thickness like bedded chert are characteristic in the mid-ocean floor. Around island arcs and marginal basins (including passive continental margins), terrigenous sediments are expected as predominating lithology; there can be interfingered or inter-pillowed terrigenous sediments within volcanic sequences. In these settings, chert and carbonate are also recognized, but can be distinguished from those in the mid-ocean, based on their thickness and association of terrigenous materials. It is not so common, but when the mid-ocean ridge moved to reach at a trench and subducted, the terrigenous sediment instead of the pelagic chert could directly cover mid-oceanic ridge basalts.

The other geological features of the modern volcanic sequences are modes of occurrences of their crustal constituents. Subaerial eruption is characteristic in oceanic islands and island arcs due to their topographic features, whereas pillow lavas indicate their submarine settings, i.e., seamounts, mid-ocean ridges and marginal basins. The presence of sheeted dike complexes is conclusive evidence of spreading centers like mid-ocean ridges, marginal basins and island arcs. As for crustal thickness of volcanic sequences (above the Moho depth), there is a distinctive contrast between oceanic plateaus and the others. For example, the Ontong Java plateau has an average crustal thickness about 36 km (e.g., Furumoto et al., 1970; Neal et al., 1997), and the other oceanic plateaus also have unusually thick crusts. Based on experimental studies, basaltic rocks with bulk compositions of quartz tholeiite and alkali olivine basalt can be transformed to garnet-bearing granulites under conditions above 1.2 GPa and 0.8 GPa at c. 1000 °C, respectively (e.g., Ringwood, 1975). The unusual thickness of oceanic plateaus would cause metamorphism with higher pressure and temperature conditions at the base of plateau crust rather than that at the normal mid-oceanic ridge. Moreover, it could form garnet-bearing granulites like mafic xenoliths found from the northern end of the Kerguelen plateau (Gregoire et al., 1994), depending on availability of fluids and chemical compositions of lower crusts of the plateau.

On the other hand, the thickness of common oceanic crust does not exceed c. 7 km, and the confirmed maximum grade of the modern ocean-floor metamorphism are the

amphibolite-facies conditions, except for a very rare case in the Mid Atlantic Ridge (e.g., summary in Ishizuka, 1995). However, “oceanic plateau metamorphism”, mentioned above, may differ in mechanism and process from the ocean-floor metamorphism as hydrothermal alteration at the mid-ocean ridges.

Petrography and petrology of volcanic rocks from different tectonic settings also vary, depending on their rock type and formation process. Oceanic island basalts (OIBs) and island arc basalts (IABs), predominating alkaline and calc-alkaline rocks, respectively, are porphyritic, whereas oceanic basalts in the other settings are mainly aphyric. The fractionation control of plagioclase following olivine is characteristic in mid-ocean ridge basalts (MORBs) predominating low-K tholeiite, and that of orthopyroxene and amphibole following olivine is specific feature for IABs. Recent years, it is widely known that the oceanic plateau basalts (OPBs) are generally aphyric and their fractionation is controlled by olivine and clinopyroxene, depending on depths of their generation (e.g., Neal et al., 1997), in contrast with fractionation of the MORBs controlled by olivine and plagioclase in general. In terms of mineralogy, clinopyroxene phenocrysts from the OIBs are higher in TiO_2 contents than those from the others with similar degrees of the crystallization differentiation, and generally exhibit red or brown colors, whereas those from tholeiitic rocks are commonly colorless. It is because of differences in the compositional trends, especially in increasing rate of TiO_2 , by the crystallization differentiation. The increasing rates of TiO_2 are the highest in the OIBs and moderate in the MORBs; in the IABs, the enrichment of TiO_2 by the crystallization differentiation is hardly recognized. According to recent compilation, the OPBs exhibit the compositional trends with lower increasing rates of TiO_2 than those of the MORBs (Nagahashi and Miyashita, 2002). Moreover, occurrences of high-MgO (> 14 wt.%) lava including picritic basalts are relatively common in oceanic islands, seamounts and oceanic plateaus, but are rarely recognized in the other settings.

Geochemical characteristics of basaltic rocks can distinguish their parental materials and formation processes rather than their tectonic settings. Especially, enrichment of large-ion-lithophile elements, LILEs, (e.g., La/Nb ratios, normalized by those of the primitive mantle), and light-rare-earth-elements (LREEs) are suggestive of the calc-alkaline IABs, of which sources are affected by subduction-zone metasomatism. However, chondrite-

normalized REE patterns of OIBs with the LREE-enrichment can be differentiated from those of the IABs by depletion of heavy-rare-earth-elements (HREEs). The REE patterns with depletion of LREEs are typical features of N-type MORBs.

Geochemistry and isotopic compositions of OIBs allows us to subdivide their settings and origins. Recently, Tatsumi et al. (1998) pointed out that the South Pacific superplume-related OIBs had a distinguishable trend from the other OIBs in terms of a Nb/Y-Nb/Zr diagram. Among the superplume-related OIBs, the HIMU-type detected only from the hot spot OIB in the superplume region (Cook-Austral, Polynesia; St. Helena) is enriched radiogenic Pb (e.g. Zindler and Hart, 1986). It has a specific feature in Pb isotopes ($^{206}\text{Pb}/^{204}\text{Pb} > 20.5$), which is one of the most decisive features in order to distinguish its origin.

3-2-2-1-3. Criteria for discrimination of tectonic settings of ancient volcanic sequences in accretionary complex

As described above, the modern volcanic sequences have different geological, petrochemical and isotopic features by their tectonic settings. Therefore, those features of ancient volcanic sequences trapped in accretionary complex could help us to distinguish their tectonic settings.

In terms of geologic features, sedimentary rocks such as limestone, bedded chert and terrigenous clastic rocks associated with greenstones in accretionary complexes and structural relationships between two (e.g., conformable or not) provide us useful information to reconstruct an environment where the greenstones were erupted. In a case that the oceanic plateaus were trapped in orogenic belts with plate subduction, rocks in various depths of the plateaus could be exposed after their uplift and erosion. Metamorphic petrology of such rocks, i.e., degrees of ocean floor metamorphism, would allow us to reconstruct their primary geologic structure and their original tectonic settings; if there would be any lines of evidence for the high-pressure granulite-facies by the ocean floor metamorphism, it must be derived from the lower crust of oceanic plateau.

Based on geochemical and isotopic features of the modern basaltic rocks with various tectonic settings, several discrimination diagrams have been proposed, and they have been

widely accepted (e.g., summary in Wilson, 1989; Rollinson, 1993). However, it is not always that those discrimination diagrams give us conclusions with decisive confidence for determining tectonic settings of the greenstones. The discrimination diagrams based on compilations of published data only suggest the general tendency or petrogenesis of the greenstones, and do not directly indicate their tectonic settings. Moreover, the greenstones trapped in orogenic belts have been subjected to secondary modification by hydrothermal alteration just after then and metamorphism at trench after travels from its birth place. For example, the Pb isotopic ratio of the HIMU-type OIBs is a conclusive characteristic to distinguish its origin. However, it has been thought to be difficult for accreted and metamorphosed greenstones in orogenic belts to distinguish the HIMU-type OIB by using the isotopic characteristics, because the isotopic compositions are easily modified by the secondary alteration. Due to the same reason, the characteristics on alkali elements, LILEs and some of LREEs compositions of greenstones in orogenic belts cannot be utilized. But, high-field-strength-elements (HFSEs) such as Nb, Y and Zr, and most of REEs are generally regarded as immobile against the secondary alteration, and the systematics with those elements is useful for discrimination of the greenstones. In terms of the Nb-Y-Zr systematics, Tatsumi et al., (1998; 2000) suggested that some of greenstones in accretionary complexes in the circum-Pacific orogenic belts were the South Pacific superplume in origin.

On the other hand, the Cretaceous plateaus, dispersed in the southwestern and the western Pacific at the present, have formed at the South Pacific superplume region (e.g., Larson, 1991a). However, most of basalts from the Cretaceous plateaus show tholeiite-like geochemical characteristics similar to the E-type MORBs, and are homogeneous in composition (e.g. Mahoney et al., 1993), except for some alkaline basalt with the HIMU signature (Tejada et al., 1994). Therefore, those tholeiitic OPBs exhibit different trends on the Nb/Y-Nb/Zr diagram (Tatsumi et al., 1998) from the superplume-related OIBs. Such an inconsistency would be caused by formation process of a huge amount of the plateau basalt; melting a large amount of lithospheric mantle could dilute geochemical signatures of the HIMU end-member in the resultant melt. Except for well-studied cases (Solomon islands and Ontong Java plateau), petrochemical characteristics of the oceanic plateau basalts have

been studied by using restricted samples dredged from ocean floor, which provide only limited information of subsurface of the oceanic plateaus. Because of these problems, it is controversial whether the superplume-related plateau basalts can be definitely characterized only by geochemical characteristics such as the HIMU signatures and the Nb-Y-Zr systematics. Then, we must pay attentions to the other aspects of oceanic plateaus or the OPBs, i.e., their geological, petrographical and petrological features such as crustal thickness, assemblage of phenocrysts, compositional trend by the differentiation, and so on (Table 3-1).

Another approach for greenstones with porphyritic textures is to investigate mineral compositions of clinopyroxene phenocrysts, which are more resistant against alteration and metamorphism than the other common phenocrysts like olivine and plagioclase. As mentioned above, clinopyroxenes from the OIBs have optical and compositional characteristics, i.e., red or brown colors, and high-TiO₂ contents, reflecting chemical compositions of their parental magma. Therefore, we could obtain information about origins of greenstones when we would have any fresh grains of clinopyroxenes, even if the matrix constituents would have completely been altered or metamorphosed. In addition, REEs patterns of clinopyroxene phenocrysts from the modern MORBs and OIBs have been reported with development of analytical technique such as the secondary ion mass spectrometry. Partition coefficients for REEs between clinopyroxenes and their parental magmas are high enough to evaluate the REE patterns of parental magmas by using those of clinopyroxenes (e.g., summary in Rollinson, 1993). With such ways, we can discuss about origins of altered or metamorphosed greenstones, based on petrography and (geo-) chemistry of clinopyroxenes if available.

As stated above, more comprehensive examinations enable us to determine tectonic settings of greenstones in orogenic belts. With such considerations, we determined primary tectonic settings of ophiolites and accreted greenstones in orogenic belts, listed in Table 3-2. Compiled data for the discrimination were taken from the most recent and reliable data at the present, because quality and quantity of available data are uneven by area and age.

3-2-2-2. Procedures of reconstruction of oceanic domains

3-2-2-2-1. Cenozoic-Mesozoic reconstruction

In recent 200Ma, the distribution of mid-ocean ridges and trenches has been determined well and plate reconstruction has been provided (e.g. Engebretson et al., 1985; Scotese and Sager, 1988), because the information of oceanic domain is immediately able to be got from the extant oceanic floor.

As stated above, paleopositions of the Cretaceous LIPs in the present oceanic domain including Ontong-Java and Mid-Pacific seamount chains and others can be estimated by using the relative plate motions model (Engebretson et al., 1985; Larson, 1991a); those oceanic LIPs were removed back from the present day to their erupted age, following their plate motions. For the ophiolites and greenstones in the accretionary complex, at first, we distinguished their tectonic settings on the basis of criteria and available data about geology, geochronology and petrochemistry as described above. After that, paleopositions of the ophiolites and greenstones were reconstructed by using the relative plate motion model (Engebretson et al., 1985). In the case that the determined settings were island arc or marginal basin, ophiolite or greenstone belts were generated in the continental margin at that period and migrated with it until present. In the case of oceanic setting, we determined the generated age for them on the basis of microfossil from the bottom of the sedimentary section of OPS or trapped sediments in lava, or directly isotopic age of volcanics. In the similar way, we determined the emplaced age for them on the basis of microfossil from the hemi-pelagic horizon of OPS. From the present day to their emplaced age, they were regarded to have migrated with their emplaced continents, whereas from their emplaced age to erupted age, the same method for the LIPs in the present oceanic domain were adopted.

3-2-2-2-2. Paleozoic-Neoproterozoic reconstruction

At these periods, although continental distribution has been reconstructed, that of oceanic domain has scarcely.

We arranged the trenches based on the distribution of Pacific-type orogens. The directions of dipping were decided with the growth polarity. Mid-oceanic ridges were situated on the basis of the modeling that the mid-oceanic ridges were present at a center of

small ocean surrounded by continental fragments for a period from continental break-up to initiation of plate subduction along the continental fragments. After beginning of subduction, the locations of mid-oceanic ridges are unknown because we have no method to decide the relative velocity between the oceanic plates. In these cases, we tentatively arranged the mid-oceanic ridges at anywhere in the oceans.

The ophiolite and greenstone belts in the accretionary complexes are examined their tectonic setting based on the criteria mentioned above. In the cases that these were generated at continental margin setting, these could migrate with the continents. In the cases that at oceanic setting, we could not directly use the plate motions. If paleomagnetic data were available, we would take them. But in most cases, these were not available and we can not fundamentally determine the detail positions in ocean. We are only able to arrange on pelagic environment in the ocean that orogens faced. In the cases that greenstones or ophiolites had superplume affinities and the orogens faced the Pacific Ocean, we would arrange these as oceanic islands in the Pacific superplume region at erupted age. In the cases that greenstones or ophiolites had the oceanic plateau affinities and faced the Pacific Ocean, these would be located in the Pacific superplume region at erupted age based on the analogy that Cretaceous oceanic plateaus was generated in the Pacific superplume region. Such an assumption is based on an observation on episodic ophiolite pulses (Fig. 3-5), that suggests close relation between the ophiolite and some unusual episodic volcanic activity, i.e. superplume related eruption.

In these ways, we reconstructed the ancient volcanic sequences from different tectonic settings, which have been trapped in the orogenic belts as ophiolites and greenstones. In next sections, we example some results of our reconstruction for the ancient volcanic sequences from Cretaceous and Cambrian orogenic belts.

3-2-3. Examples of reconstructions of ancient volcanic sequences in accretionary complexes

In this section, examples of reconstructions are shown based on geologic and geochemical studies in accretionary complexes in Cretaceous Sambagawa-Mikabu belt, SW-Japan and

Cambrian Gorny Altai, Russia (Fig. 3-6).

3-2-3-1. Cretaceous example: Sanbagawa-Mikabu belt, southwestern Japan

Several accreted fragments equivalent to the present oceanic LIPs have been recently recognized in the Circum-Pacific orogens (Fig. 3-3); the well-documented examples are the Latest Jurassic Sorachi oceanic plateau (northern Japan) accreted at the eastern margin of Asia in the Early Cretaceous (Kimura et al., 1994; Hori and Sakakibara, 1994; Sakakibara et al., 1999), and the Middle to Late Cretaceous Caribbean-Colombian oceanic plateau accreted at the Central Cordillera, Western Cordillera and along the Pacific coast (Kerr et al., 1996; 1997; 1998). Here, we describe our own results as the possible example from a Cretaceous metamorphosed accretionary complex, the Sanbagawa belt, southwestern Japan (Fig. 3-6).

The Sanbagawa belt is a part of accretionary complex formed by the latest Jurassic to Early Cretaceous subduction of Kula-Izanagi oceanic plate toward the eastern margin of Asia (Isozaki and Itaya, 1990), and has been metamorphosed under high-P/T conditions from the subgreenschist- to the eclogite-facies (e.g., Banno and Sakai, 1989; Enami et al., 1994; Wallis and Aoya, 2000). We have recently completed a new lithotectonic map at 1:5000 scale for the highest grade area of the Iratsu- and Higashi-Akaishi eclogite-peridotite composite mass and its surrounding region in central Shikoku with a special attention to duplex structure and protolith occurrences (Fig. 3-7) (Okamoto et al., 2001a; Maruyama et al., 2001b; Ota et al., 2002). Our mapped area consists of pelitic-psammitic, quartzose and basic schists, which enclose the eclogite-peridotite mass. Those schists and the eclogite-peridotite mass have been subjected to the high-grade metamorphism and penetrative ductile deformation; nevertheless detailed mapping of the schists and complementary geochemistry of basic schist indicated that they comprise an OPS composed of trench turbidite and its underlying pelagic chert and mid-oceanic ridge basalt formed in the Pacific domain, and have accreted or underplated as a subduction complex with a duplex structure at depths of subduction zone (Okamoto et al., 2000). The eclogite-peridotite mass is composed of ultramafic rocks, eclogitic metabasites with basalt and

gabbro origin, metacarbonate, metachert and pelitic gneiss in ascending order; metagabbro contains relics of high-pressure granulite-facies (i.e., garnet-bearing) followed by eclogite-facies metamorphism (Yokoyama, 1980). The mass is separated on the top by a roof thrust and on the bottom by a floor thrust from its surrounding schists, and is divided into four horses, which indicates a duplex structure (Okamoto et al., 2001a; Maruyama et al., 2001). The horses in the eclogite-peridotite mass are much larger than those in the adjacent schists unit, although the constituent rocks have been subjected to tectonic thinning with ductile deformation during exhumation of the Sanbagawa belt (Ota et al., 2002). The reconstructed OPS of the mass suggests its origin of huge oceanic plateau covered by pelagic limestone with minor chert on its flank (Maruyama et al., 2001; Kugimiya and Takasu, 2002); a petrological thickness of the plateau would exceed 30 km because of occurrences of the relics of high-pressure granulite in the lower horizons of the reconstructed stratigraphy (Fig. 3-8).

Along the southern margin of the Sanbagawa belt, the Mikabu greenstone belt is present with a lateral extension of about 800-km long. It is mainly composed of mafic volcanic rocks and ultramafic-mafic cumulates, and a timing of its magmatism is considered to be the post-Early Jurassic (Sakakibara et al., 1993; Ozawa et al., 1997). The Mikabu greenstones have also been subjected to the subduction-related high-P/T metamorphism, and regarded as the low-grade part of the Sanbagawa belt (e.g., Banno and Sakai, 1989; Suzuki and Ishizuka, 1998). In terms of geological and petrochemical features including HIMU signatures (Ozawa et al., 1999), the Mikabu greenstones would have been seamount chain and/or oceanic plateau in origin (Isozaki et al., 1990; Ozawa et al., 1999; Uesugi and Arai, 1999), formed by the superplume activity.

Erupted ages of the Sanbagawa and Mikabu protoliths have been estimated as 142-153 Ma by microfossils from pelagic sediments associated with greenstones (Sanbagawa: Isozaki and Itaya, 1990; Mikabu: Sakakibara et al., 1993), K/Ar dating for igneous amphiboles from the Mikabu gabbros (Ozawa et al., 1997), and U/Pb dating for zircons (inherited cores) from the Sanbagawa eclogites (Okamoto et al., 2001b). As for their accreted ages, metamorphic ages of the Sanbagawa eclogites gave constraints; U/Pb dating for metamorphic zircons yielded 110-120 Ma (Okamoto et al., 2001b). Therefore, the Latest

Jurassic Sanbagawa and Mikabu protoliths would have been arrived at subduction zone by the end of Early Cretaceous, through 20-40 million years of migrating interval.

Based on the Cretaceous plate paleogeography in the Pacific (Engebretson et al., 1985), migrating tracks of the accreted Sanbagawa and Mikabu protoliths (plateaus or seamount chains) get back to an adjacent region of the present South Pacific superplume, as the same as that of the Latest Jurassic Sorachi plateau. Thus, our reconstructed oceanic plateaus, composed of the Sanbagawa and Mikabu plateau or seamount chains, and the Sorachi plateau, might have constituted a huge oceanic plateau, formed with the South Pacific superplume activity at the Jurassic to the earliest Cretaceous (Suzuki et al., 2000). Figure 3-9 shows a paleogeographic reconstruction for these Mesozoic oceanic plateaus by this study. Although the major proportion more than 99 % of total amount of an oceanic plateau disappears downward into the mantle at the subduction zone, we can get the information from the slight fragmental remain in accretionary complex.

3-2-3-2. Cambrian example: Gorny Altai, Russia

Available data for reconstruction in the Early Cambrian and older periods are rather limited than those since the Mesozoic time, however, investigating geology, petrochemistry and geochronology of greenstones and associated rocks in accretionary complexes enable us to reconstruct their origin. Geological records during these periods have been found in intra-continental regions, rather than in the circum-Pacific and the Tethian-orogenic belts. Recent years, orogens with the Latest Precambrian to the Early Paleozoic accretionary complexes in central Asia have been well studied (e.g., summary in Maruyama et al., 1989; Sengör et al., 1993), resulting in knowledge that the Cambrian accretionary complexes and associated greenstones in the central Asia have an analogy to the Mesozoic ones in the present Pacific-rims (e.g., summary in Dobretsov et al., 1995). Here, we example our own results from the Pacific-type orogen with a Vendian-Cambrian accretionary complex in Gorny Altai, southern Russia (Fig. 3-6).

Gorny Altai Mountains in southern Siberia corresponds to the northern part of the Altai Range that continues from China through Mongolia. The Cambrian accretionary

complex is exposed widely in the northern and southeastern part of the Gorny Altai Mountains. This accretionary complex is composed of limestone, greenstone, sandstone and mudstone (Buslov et al., 1993). Limestone bodies, named Baratal limestone, have been reported as paleoseamount limestone primarily accumulated on the OIB type basaltic greenstones (Buslov et al., 1993); some of these have been metamorphosed under high-P/T conditions from the subgreenschist-facies to the eclogite-facies grades (Fig. 3-10) (Buslov and Watanabe, 1996; Ota et al., 2002b). The greenstones, conformably overlain by limestone, have characteristics similar to those of the modern oceanic island basalt and the mid-oceanic ridge basalt, in terms of petrography and the chondrite-normalized patterns of REEs in clinopyroxene phenocrysts (Fig. 3-11). This suggests that the overlying limestone has formed on and around an ancient seamount or paleo-plateau.

The primary stratigraphy of the Baratal limestone was analyzed in the southern part of the Gorny Altai Mountains (Uchio et al., 2000). Massive limestone and limestone breccia conformably overlies basaltic greenstones. Some limestones include stromatolites and ooids that suggest shallow-marine environment. Some limestone breccia contains clasts of micritic limestone, greenstone and chert. Its poor sorting and thinly bedding suggest that the limestone breccia was formed as sediment-gravity-flow (debris flow) deposits. In addition, laminated limestone and bedded micritic limestone are associated with the limestone breccia. The absence of coarse-grained terrigenous clastics in all of limestones indicates that all types of the limestone were deposited in mid-oceanic environment. These evidences suggest that these limestones were probably accumulated on the slope or at the base of slope of the paleo-seamount or plateau (Uchio et al., 2001; Uchio et al, in prep.). The depositional sites of these four types of limestone lithologies are inferred as follows; massive and micritic with shallow-water stromatolite is inferred to have formed on the top of paleo-seamount or oceanic plateau. The laminated limestone corresponds to allodapic limestone (limestone turbidite), which was accumulated at the base of slope of the paleo-seamount or plateau. The micritic limestone is also supposed to have deposited at the bottom of slope of the paleo-seamount or plateau). The limestone breccia represents sediment-gravity-flow (debris flow) deposit, which was probably accumulated on the slope or at the base of slope of the paleo-seamount or plateau (Fig. 3-12; Uchio et al., 2001).

In the northern part of the Gorny Altai Mountains, the Baratal limestone contains a few pieces of stromatolites and microphytolites, which suggest the Vendian age (Buslov et al., 1993; Afonin, 1974). Some siliceous mudstone accompanied with Baratal limestone contains siliceous Early Cambrian sponge spicules (Afonin, 1974). Thus, the age of Baratal limestone has been considered the Vendian to Early Cambrian. Their Pb isotopic compositions, analyzed by a Finnigan MAT 261E mass spectrometer at the Kumamoto University in Japan, yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ bulk isochron age of 572 ± 110 Ma with MSWD of 68 (Uchio et al., 1999; Uchio et al., in prep.). In addition, Sr isotopic data of the limestones ($^{87}\text{Sr}/^{86}\text{Sr} = 0.707940, 0.707850$; Uchio et al., in prep.) indicated the latest Vendian; c.570-580 Ma, based on the Sr isotopic stratigraphy (e.g., Verzer et al., 1997; Walter et al., 2000; Melezhik et al., 2002).

Consequently, the Cambrian OPS reconstructed with lithological assemblages and geochronology of the Gorny Altai accretionary complex suggests the following history: After the oceanic plate was born in the mid-oceanic ridge at the end of the late Proterozoic, the seamount (plateau) was formed on that at sometime before c. 570 Ma. From the latest Proterozoic (Vendian, 605-543 Ma: Bowring and Erwin, 1998) to the Earliest Cambrian, calcareous sediment was deposited on the top and around of the seamount (plateau). Finally, at the early Cambrian, the seamount (plateau) with limestone and limestone breccias on the oceanic plate arrived at trench, and parts of them were trapped into the accretionary complex (Fig. 3-12).

3-3. Reconstructions

With methods and data sources stated above, we reconstructed paleogeographic maps of not only continents but also oceans from the present back to 1000 Ma. In the following sections, paleogeographic features and their relations with geological events during the recent 200 Ma (Fig. 3-13), 200-540 Ma (Fig. 3-14), and 540-1000 Ma (Fig. 3-15) are described separately, because of differences in quality and accuracy on reconstruction methods and data sources. Abbreviations for continental fragments are shown in Table 3-3.

3-3-1. From 200 Ma to the present

Reconstruction of continental distribution from Pangea to present-day has roughly reached to the consensus (e.g. Livermore, 1989) and plate reconstruction of Pacific region has been carried out by Engebretson et al., (1985). However, the microcontinents and the microplates have been mostly neglected in the previous reconstructions of global scale. We corrected the microcontinents in the central to southern Asia and southwestern Pacific (Norfolk Plateau, Load Howe Rise, Chatham Rise, and Campbell Plateau) based on Maruyama et al. (1989) and Sutherland (1999), respectively.

At Present day to 90Ma (Fig.3-13 a-c). The oceanic plateaus in southwestern and western Pacific Ocean was tracking back to concentrate Pacific superplume region as pointed out by Larson, 1991a. LIPs in other regions were also generated. Columbia River basalt, North Atlantic Volcanic Province, and Deccan Traps are dated at 6-17 Ma (Hooper, 1997), 60 ± 1 Ma (Eldholm and Grue, 1994; Saunders et al., 1997), and 67 ± 1 Ma, respectively.

The Resurrection Peninsula and Knight Island Ophiolites from Orca Group, southern Alaska, contain pillow basalts interbedded with terrigenous sediments and sheeted dykes (Lytwyn et al., 1997). Resurrection Peninsula Ophiolite is dated to 57 ± 1 Ma by zircon U/Pb from intruded plagiogranite. Their basalts commonly have plagioclase phenocrysts and clinopyroxene microphenocrysts. Olivine phenocrysts are rare (Lytwyn et al., 1997). These have flat REE patterns and T-MORB-like geochemical features (Lytwyn et al., 1997). These seemed to be generated TTR triple junction and relict of the subduction of Kula-Farallon ridge (Lytwyn et al., 1997).

Melange units containing greenstones are common throughout Shimanto Supergroup in the Ryukyu Islands and southwestern Japan. The greenstones from some formation (Naze Formation, Makimine Formation, Okitsu Melange, Mugi Formation, Ryujin Formation, and Inui Group) are closely related with shale and absent with bedded chert (Kiminami et al., 1994). Radiolarian assemblage from the shales indicate Cenomanian-Turonian, Santonian-Middle Campanian, Coniacian-Campanian, Campanian-Early Maestrichtian, Late Campanian-Maestrichtian, and Late Maestrichtian-earliest Paleocene

for Naze Formation, for Makimine Formation, for Okitsu Melange, for Mugi Formation, for Ryujin Formation, and for Inui Group, respectively (Kiminami, et al., 1994). These have MORB-like geochemical features (Kiminami et al., 1994). These seemed to be generated TTR triple junction and relict of the subduction of Kula-Pacific ridge on the basis of these characteristics. The ridge motion by plate reconstruction (Maruyama and Seno, 1986) is consistent with the evidence that greenstones become younger eastward (Kiminami et al., 1994).

In the New Caledonia, Southwestern Pacific Ocean, the one of the largest ophiolitic complexes of the world is exposed. It consists of mainly harzburgitic allochthon, piled basaltic slices, and melange unit metamorphosed into eclogite/blueschist facies. These are named the Nappe Ophiolitique, Poya Terrane, and Pouebo Terrane, respectively (Cluzel et al., 2001). The igneous lithologies of the Poya Terrane are pillowed and massive basalts, dolerites, and fine grained gabbros. These are interbedded with or covered by abyssal black, green, or red argillite and chert (Cluzel et al., 2001). The most common radiolarian faunas extracted from these sedimentary rocks indicate Upper Cretaceous (Campanian). Upper Paleogene fauna have been also recovered from some localities (Cluzel et al., 2001). The terrane was severely folded, faulted, and disrupted structures interpreted as duplex (Cluzel et al., 2001). Adding to absence of terrigenous materials in the terrane, it is inferred to be accreted in intra-oceanic environment (Cluzel et al., 2001). The Poya Terrane in general has been metamorphosed into greenschist facies interpreted as ocean floor metamorphism (Cluzel et al., 2001). However, because the terrane was severely disrupted, primary thermal structure in the crust can not be reconstructed. The volcanic rocks from the Poya Terrane are divided into three types on the basis of REE patterns (Cluzel et al., 2001). The “undepleted type” has flat REE patterns. This type rocks have low-Ti and Al augite and labrador-andesine phenocrysts. The geochemical features are similar to typical E-MORB. This type accounts for 80% of total volcanic rocks from the Poya Terrane. The “depleted type” has slightly depleted REE patterns. This type rocks have augite microphenocrysts. The geochemical features are similar to back-arc basin basalt that have transitional features between N-MORB with depleted REE patterns and island arc tholeiite with a Nb depletion. Nd isotopic variations of “undepleted” and “depleted” types suggest mixing between DM

and EM II sources. Synthesized the features above-mentioned, “undepleted” and “depleted” types are inferred to have been generated in the plume-enhanced marginal basin (Cluzel et al., 2001). The third type is alkaline volcanic rock that has highly enriched REE patterns. This type rocks are associated with micritic interpillow sediments and have pinkish Al and Ti-riched diopside, andesine-labrador, and Fe-Ti oxides microphenocrysts. The geochemical features are similar to the OIB. It is inferred to have erupted as seamounts on the basin that had provided other type volcanic rocks. The faunas recovered from micritic interpillow material with this type rocks indicate Upper Paleocene (56-59Ma). This age is the youngest of the terrane and consistent with the interpretation of the tectonic setting (Cluzel et al., 2001). The mafic rocks from the Pouebo Terrane that is metamorphosed ophiolitic melange have similar geochemical features to the Poya Terrane and to island arc tholeiite with slightly enriched REE pattern and Nb depletion (Cluzel et al., 2001). These protoliths are seemed to be reworked Poya and island arc terranes (Cluzel et al., 2001). The ophiolitic Nappe is underthrust by Poya and Pouebo Terranes mainly composed of harzburgite and dunite (Cluzel et al., 2001). This ultramafic allochthon is extremely depleted and can not been in equilibrium with a melt of MORB composition (Eissen et al., 1998). The Ophiolitic Nappe is seemed to be the basement on which not the Poya Terrane but the arc (Loyalty Arc) has been established (Cluzel et al., 2001).

At 90-120Ma (Fig.3-13 d, e). The oceanic plateaus in southwestern and western Pacific Ocean distribute as concentrating Pacific superplume region. Age data are shown in Table 3-2.

Thick oceanic crust (up to 15-20 km) of Caribbean plate was dated to ca. 90Ma and 75Ma by bulk Ar-Ar (Sinton et al., 1998). It almost has N-MORB-like to E-MORB-like REE patterns but a few OIB-like (Kerr et al., 1997). The Cretaceous lavas that have similar geochemical features are observed on the islands in the Caribbean Sea and Eastern Cordillera in the C. America (Kerr et al., 1997). All they are named Caribbean-Colombian Cretaceous Igneous Province. It tracked back to near the Pacific superplume region on Farallon plate, when formed 90Ma (Fig.3-13 d).

East Sulawesi ophiolite, Indonesia, has the crustic layer of minimum 8km thickness (Kadarsman et al., 2001). The aphyric basalt is most common in its basaltic layer and the

porphyritic basalt has less olivine phenocryst. The mineral chemistry of cumulatic layer clearly distinct from arc setting, although geochemical features indicate suprasubduction setting. On the basis of these petrological and mineral chemical features, it was suggested that it has been generated in oceanic plateau (Kadariusman et al., 2001). It was generated at 138Ma and emplaced before 28Ma on the basis of K-Ar dating for underlying melange and metamorphic sole (Parkinson, 1998). It tracked back to near the Pacific superplume region at generated age (Kadariusman et al., 2001)(Fig. 3-13 e).

At 150-200Ma (Fig.3-13 f). Karoo-Ferra flood basalt erupted at 195Ma-. It might be related the initiation of African superplume (Daziell et al., 2000).

Iratu massif in Sanbagawa metamorphic belt, southwestern Japan is oceanic plateau origin at 150Ma above-mentioned.

The greenstones from Sorachi-Yezo belt in central Hokkaido, northern Japan, have N-MORB- to E-MORB- and OIB-like geochemical features (Sakakibara et al., 1999; Nagahashi and Miyashita, 2002). Some of OIB-like greenstones have high Nb/Y and Nb/Zr ratios (Sakakibara et al., 1999). They seem to be generated from heterogeneous sources in oceanic plateau (Kimura et al., 1994; Sakakibara et al., 1999). Radiolarian age for chert xenoliths in lavas is Latest Jurassic (ca. 150Ma) and it is regarded as erupted age (Sakakibara et al., 1999). Emplaced age is Barremian on the basis of radiolarian age for the black shale of the top layer of oceanic stratigraphy overlain by the Plateau (Sakakibara et al., 1999). It tracked back to Pacific superplume region.

The Josephine ophiolite is distributed in northern California and southern Oregon, US. Its volcanic sequence mainly consists of pillow lava and is conformably overlain by thin chert piled thick flysch (Saleeby et al., 1982; Harper, 1984). It has sheeted dyke complex (Saleeby et al., 1982; Harper, 1984). Zircon U/Pb age is 157Ma and 151Ma for plagiogranite pods in gabbro-diorite sequence and in sheeted dyke sequence and for dacite dyke cutting both volcanic and sedimentary sequences, respectively (Saleeby, et al., 1982). Former age is regarded as the generated age and latter is regarded as emplaced age (Saleeby et al., 1982). The Josephine ophiolite has short interval from generation to emplacement. Because flysch sequences include the andesitic clastics, the Josephine ophiolite is formed in proximity to the island arc (Saleeby et al., 1982; Harper, 1984). Earlier cpx crystallization

based on the phenocryst assemblage of volcanic sequences and occurrences of opx in cumulus sequences indicate island arc affinity (Harper, 1984). Geochemical features of the volcanic sequences correspond to from MORB to island arc tholeiite (Harper, 1984). The hydrothermal metamorphism that the highest grade is amphibolite facies for cumulus sequences do not indicate the thick crust (Harper, 1984). On the basis of these evidences, the Josephine ophiolite is generated by opening of marginal basin (Saleeby et al., 1982; Harper, 1984), we followed this. The Devils Elbow ophiolite in western California has similar stratigraphic and geochemical features to the Josephine ophiolite and is regarded as southern continuation of the Josephine ophiolite (Wyld and Wright, 1988). Its zircon U/Pb age is 164Ma (Wright and Wyld, 1986). The marginal basin that generate the Josephine ophiolite began to open at least this age.

The Coast Range Ophiolite is present in scattered localities over a 700-km-long segment in western California, US. The Coast Range Ophiolite is divided into two types. One type is associated with minor felsic igneous rock and has island-arc geochemical features (Shervais and Kimbrough, 1985). It is dated to 160-170Ma by zircon U-Pb dating (Shervais and Kimbrough, 1985). Another is block in serpentinite matrix and has MORB or seamount geochemical features (Shervais and Kimbrough, 1985). Chert blocks within this melange contain a Tithonian radiolarian fauna. The former seems to be generated in island arc. The latter seems to be generated as seamounts rather than as mid-ocean ridges on the basis of the field occurrences that formed the melange. It is guessed by analogy of Cretaceous-Paleocene Shimanto accretionary complex, southwest Japan, that is modified by seamounts accretions and formed the melange-like complex (Isozaki, 1990). Great Valley Group overlying the Coast Range Ophiolite has the greenstones that have similar geochemical features and radiolarian ages (MacPherson and Phipps, 1988). It seems to be related with the latter Coast Range Ophiolite and have same origin.

3-3-2. Paleozoic (200-550 Ma)

The Reconstruction of continental distribution from Gondwana to Pangea has roughly reached to the consensus (e.g. Livermore et al., 1986). However, the distribution of microcontinents has not sufficiently been discussed. Microcontinents in Asia, which are

Turkey, Iran-Afghanistan, Lhasa, Malaya, Sumatra, and Indonesia, have been positioned in northern Gondwana by 280Ma (Fig. 3-14) on the basis of floras distributions and they broke-up from Gondwana and collided to southern Laurasia one after another (Maruyama et al., 1989). The reconstruction of oceanic region was carried out as followed the method mentioned above.

At 225Ma (Fig. 3-14 a). Pangea formation is completed at this age. S. China collided N. China at 250Ma evidenced by SHRIMP dating for ultrahigh-pressure metamorphic rocks from Dabie-Sulu suture zone (Maruyama et al., 1998). Turkey and Iran-Afghanistan begun to separated from Gondwana (Maruyama et al., 1989).

Triassic large greenstone belt in Wrangellia terrane extending from Vancouver Island, British Columbia, to southern Alaska. It overlies cherts and middle Ladinian limestones and underlies late Karnian and early Norian limestones based on fauna assemblage (Jones et al., 1977). The thickness of Triassic greenstone extend to ~7000m (Lassiter et al., 1995). These overlie Pennsylvanian to early Permian volcanics and sedimentary rocks derived by island arc (Jones et al., 1977). Triassic greenstones have E-MORB-like geochemical features (David and Plafker, 1984) but suffer the contamination of arc crust. (Lassiter et al., 1995). These geological and geochemical features indicate that Triassic greenstones from Wrangellia terrane are flood basalt on the island arc. The paleolatitude of Wrangellia terrans is ca. 20° N and departed far from Laurentia on the basis of the paleomagnetic and faunal data (Keppie and Dostal, 2001). Based on it, we recognized that it is not related with the Pacific superplume but initiation event of other hotspot.

The pillow basalts from Chonos Metamorphic Complex in Aysen, southern Chile, occur with pelito-psammitic schist intercalations (Herve et al., 1999). The deposition age of the turbidite sequences point out Late Triassic. Their phenocrysts assemblage is clinopyroxene and plagioclase. The pillow basalts have E-MORB-like geochemical features (Herve et al., 1999).

At 245Ma (Fig. 3-14 b). Continental amalgamations to form Pangea were continued and minor rifting of microcontinents from Gondwana begun. N. China collided to the Central Asia fold belt like a fan closing. Microcontinents in SE Asia, which are Indonesia, Malaya, Sumatra, begun to separate from Gondwana (Maruyama et al., 1989). Siberian

Trap erupted at 251.2 ± 0.3 Ma indicated by U-Pb zircon dating (Kamo et al., 1996).

Yakuno ophiolite, Maizuru belt, inner zone of Southwest Japan, has the amphibolite sequence dated to Permian by K-Ar and Rb-Sr methods (Ishiwatari, 1985b). It seems to be low-pressure metamorphism prior to the emplacement (Ishiwatari, 1985b) and approximately correspond to the producted age. Basaltic layer contains some pillow lavas and intercalated with black shale and acidic volcanics (Ishiwatari, 1985a,b). The obvious sheeted dykes are absent (Ishiwatari, 1985b). The basalts are mostly aphyric, although olivine and plagioclase phenocrysts (both are pseudomorph) are present in some samples (Ishiwatari, 1985b). All the chemistry of whole rock, relic clinopyroxene, and relic chromian spinel indicate tholeiitic affinity (Ishiwatari, 1985a,b). Metamorphic grade increases downward the ophiolite succession and reaches pyroxene granulite facies in the middle of the cumulate layer (Ishiwatari, 1985a). It indicates that the oceanic crust is unusually thick. Although Ishiwatari (1985a) regarded it as the small oceanic basin like Black Sea, there is likelihood that it has been oceanic plateau.

Carboniferous-Permian various ophiolites occur in southern ophiolite zone in East Kunlun, western China (Yang et al., 1996). According to the Rb-Sr dating indicating an age of 260 Ma, radiolarian age, and megafossil age in sandy slate, the ophiolites are generated in Early Permian to Middle Triassic (Yang et al., 1996). The ophiolites are occurred as block in Triassic flysch (Yang et al., 1996). The volcanic sequence of Xiadawu ophiolite in this zone is composed of calc-alkalic affinity and intercalated with dacite and rhyolite (Yang et al., 1996). Their REE patterns are from LREE-depleted to enriched (Yang et al., 1996). The ophiolite seems to be generated in island arc environment. The volcanic sequence of Majixueshan ophiolite contains many thin (up to a few cm) radiolarian chert layers (Yang et al., 1996). The lavas are slightly plagioclase-phyric tholeiite and have LREE-enriched REE pattern (Yang et al., 1996). The ophiolite seems to be generated as oceanic island tholeiite.

At 280-350 Ma (Fig. 3-14 c). Its trace remains in Angayucham terrane, central Alaska. Greenstones from this region are overlain by thin chert beds (Pallister et al., 1989). The radiolaria from inter-pillowed chert point out Mississippian age (Barker et al. 1987). It seems to be initial age. They are E-MORB-like tholeiites (Barker et al. 1987; Pallister et al.,

1989). On the basis of the geological and geochemical features, its tectonic setting is oceanic plateau. After this era, they became active again at Triassic age and Jurassic age on the basis of radiolaria, finally, arrived to Brooks Range in Alaska at Middle Jurassic age (Pallister et al., 1989).

In the East Nelson area, the South Island of New Zealand, three ophiolite complexes that are named Dun Mountain ophiolites, Patuki melange, and Croisilles melange. The age of Dun Mountain ophiolite is 280Ma by zircon dating for plagiogranite. That of melange complexes is slightly older than that of Dun Mountain ophiolite by radiolarian age (Malpas et al., 1994). The greenstones from Dun Mountain ophiolite have the geochemical features of supra-subduction zone and seemed to be generated in marginal basin (Malpas et al., 1994). The greenstone blocks from Patuki melange and Croisilles melange have within plate basalt-like and E-MORB-like geochemical features (Malpas et al., 1994). Formers correspond to the OIB. However the evidences are insufficient, later might be generated in oceanic plateau.

At 350-450Ma (Fig. 3-14 d-e). It began to form Pangea. Laurentia and Baltica, Baltica and Gondwana collided each other at 400Ma and 350 Ma, respectively (Veevers, 1989). Tarim block was positioned on the basis of paleomagnetic data compiled by Chen et al. (1999).

The Paleozoic The pillow basalts from Eastern Andes Metamorphic Complex of Aysen, S. Chile, occur as blocks in turbiditic sequence (Herve et al., 1999). Marble lenses also crop out in turbiditic sequences. They are correlated with the Late Devonian Bahia de la Lancha formation in Argentina. The pillow basalts have small phenocrysts of clinopyroxene and plagioclase. They have enriched REE patterns and high Nb/Y ratios (Herve et al., 1999). Their Nb/Y ratios are comparable with Polynesian basalts. We regarded their pillow basalt as OIB related with superplume.

The northern ophiolite zone of West Kunlun, western China, is dated to be 360Ma, based on Rb-Sr isochron dating of basaltic rock (Yang et al., 1996). Kudi ophiolite in the zone contains pillow and massive lavas with radiolarian cherts and Wuyitak ophiolite includes the lavas with the limestone lenses (Yang et al., 1996). No sheeted dykes are found in this belt (Yang et al., 1996). Although the lavas are mainly basalts, a few andesitic

basalts and andesites occur (Yang et al., 1996). These are various REE patterns from LREE depleted to enriched (Yang et al., 1996). The ophiolites in this zone seem to be generated in arc environment.

At 500Ma (Fig. 3-14 f). The orogenies in Central Asia fold belt reached the height of its prosperity. They have grown since 800~ Ma. On the basis of tectonic map of central Asia (Sengeor and Natal'in, 1996), this region is complicated geological collage which has various orogenic components (mature and immature continental crust, accretionary complexes, ophiolite belts, and metamorphic belts) and structural polarity. The petrology of the metamorphic rocks from Altai region, southern Russia, indicated that it is concerned with pacific-type orogen (Ota et al., 2002b). We concluded the complicated arc systems surrounding microcontinents in this region and reconstructed the trench distributions. Microcontinents in this regions collided to form Kazakhstan block at 537Ma on the basis of zircon dating of ultra-high pressure metamorphic rocks from Kokchetav massif, northern Kazakhstan (Katayama et al., 2001). In Early Ordovician, Tarim block positioned at 20°S (Chen et al., 1999).

Accretionary geological evidences and geochemical features of greenstones from Altai region indicated that the greenstone-limestone complexes in this region were generated in pelagic setting and some of them were related with plume activity.

Qingshuiquam ophiolite in the middle ophiolite zone of East Kunlun, western China, was dated to 518Ma by single-zircon dating for the gabbro (Yang et al., 1996). This basalt has slightly enriched trace element pattern with Nb depletion and calc-alkaline affinities (Yang et al., 1996). On the basis of the distribution of Early Paleozoic granites, the subduction from south toward to north is inferred (Yang et al., 1996). This ophiolite seems to be generated at subduction zone along southern margin of Tarim block.

Cambrian greenstone belts in Victoria, southeastern Australia, are overlain by hemipelagic sediments and have calc-alkaline chemical features (Crawford and Keays, 1978). It is generated in arc or marginal sea setting.

Bay of Islands region in southwestern Newfoundland, Canada, consists of Early Ordovician ophiolite complexes. An U/Pb zircon age from intrusive trondhjemitic for Little Port Complex and a zircon and baddeleyite ages from cumulate gabbro for Bay of Islands

Complex is 505Ma and 484Ma, respectively (Jenner et al., 1991). Volcanic layers consist of pillow lavas and locally intercalate with radiolarian cherts, Although most sedimentary layers have been removed (Casey et al., 1985). Sheeted dyke units are developed well in Bay of Islands Complex and locally present in Little Port Complexes (Casey et al., 1985). These volcanic layers have tholeiitic affinities (Casey et al., 1985; Jenner et al., 1991). These have trace element patterns from slightly depleted without negative Nb anomaly through slightly depleted with negative Nb anomaly to slight enriched with negative Nb anomaly (Jenner et al., 1991). These seem to be generated in oceanic island arc – marginal basin. The Skinner Cove Formation consisting of ankaramitic lavas and trachytes underlies the Little Port Complex. These have highly enriched trace element pattern and are generated in oceanic island volcanism (Jenner et al., 1991). These have high Nb/Zr and Nb/Y ratios and are correspond to HIMU affinities. It suggests the likelihood that superplume exists in the Iapetus Ocean at Early Ordovician.

3-3-3. Late Proterozoic (550-1000 Ma)

In the period from 550Ma to 750Ma, supercontinent Rodinia broke-up followed by the formation of (semi-) supercontinent Gondwana (Hoffman, 1991; Dalziel, 1992). The formation of Gondwana has seemed to be caused by rapid fan-like closing of Mozambique Ocean that had been present between Eastern and Western Gondwana and to be dramatically completed by the 550-500Ma Mozambique orogeny that represents the final stage of Pan-African orogen (Dalziel, 1992; “Inside-out model” by Hoffman, 1991). However, recent geochronological data need to revisit the distribution of Pan-African orogen and to modify the insight into the formation of Gondwana. The one of the bases for the simple synopsis is that Eastern Gondwana has been regarded as a stable large continent in the period. On the East Antarctica that constitutes a core of Eastern Gondwana, Pan-African aged orogens have been newly discovered and subdivided into three provinces (Shiraishi, et al., 1994; Fitzsimon, 2000). On the basis of these ages, Eastern Gondwana have been never present until 530Ma (Meert, 2002). The qualitative and quantitative reinforcement of geological and geochronological data of Pan-African orogen for the Africa and South America continents provide the opportunity to discuss the detail of the formation

of Western Gondwana (Campos Neto, 2000). Based on the geochronological data, the formation of Western Gondwana has been never completed until 520Ma. These evidences prefer to the complicated synopsis that Gondwana have been formed by collisions of several pieces of cratons in sequence rather than that by simple collision between two large continents. This mode of continental amalgamation is comparable with that of the Eurasia in Phanerozoic. The births of the Pacific and Iapetus Oceans were based on the geochronology for the igneous provinces related with continental rifting (Table 3-2). Whether the Gondwana included the Laurentia or not is stipulated by the timing of the opening of Iapetus Ocean. The remnants of opening of the Iapetus Ocean on the Laurentia, Baltica, and Amazonia cratons that surrounding the Ocean were dated to ca. 560Ma (Hoffman, 1989; Kamo et al., 1989). This age indicates that the birth of the Iapetus ocean have been followed by the completion of Gondwana formation. The Baltica and Siberia cratons were also departed from Gondwana. Therefore, we regard the Gondwana as not supercontinent but semi-supercontinent.

In the period from 750-1000Ma, the capital point of contention is conformation of supercontinent Rodinia. The Reconstruction of Rodinia has been operated by distribution of Grenville orogen. At 1000Ma, the eastern margin of Laurentia is bonded with Rio de la Plata, Amazonia and Baltica cratons by Grenville orogen (e.g. Hoffman, 1991). Paleomagnetic data from each cratons support these conjunctions (Weil et al., 1998). In contrast, a consensus on the counterpart of the western Laurentia not margined by Grenville orogen has not been reached. The major concepts are SWEAT (southernwest US – East Antarctica; Moores, 1991; Dalziel, 1991; Hoffman, 1991; Young, 1992), AUSWUS (Australia -western US; Karlstrom et al., 1999; 2001; Burrett and Berry, 2000), and AUSMEX connections (Australia-Mexico (southwestern US); Wingate et al., 2002). Since East Antarctica are subdivided into three provinces on the basis of the discovery of Pan-African orogen (Fitzsimon, 2000), SWEAT connection is unlikely to explain conformation of Rodinia. Paleomagnetic data have no qualification to decide preference for AUSWUS or AUSMEX because of the uncertainty. Taken the distribution of a generation old orogens (1.6-1.8Ga) into account, WUSAUS connection is preferred rather than AUSMEX. Furthermore, Grenville and a generation old orogens in Austraria seemed to be continued to those of

South China (Li et al., 2002). The northern part of Laurentia bonded with Siberia. However, it has been discussed which site is faced to Laurentia. The paleomagnetic data have a incapability of settling the arguments at the present time, because the ages of paleomagnetic data set from Siberia and Laurentia are not well matched (Ernst, 2000).

Paleomagnetic data insist that the components of Rodinia amalgamated at 1000Ma drifted all together till 800 Ma (Weil et al., 1998). However, the traces of orogeny in 800-900Ma remains on the margin of the cratons in South America and Africa, which is components of Rodinia (Campos Neto, 2000). It is inconsistent with paleomagnetic data. The orogen related subduction could not develop inside stable continent. The continents need to face the ocean and detached from Rodinia in this period. Relative continental positions, whether any continents are attached with Rodinia or not, are based on geological evidence. In order that the inconsistency with paleomagnetic data may be made as small as possible, we decide the absolute continental positions.

At 550 Ma (Fig. 3-15 a). E. and W. Gondwana collided at this age. The geochronological data related with this event are reported as 560-630Ma, 521-553Ma, and 501Ma, for Madagascar (Paquette and Nedelec, 1998), for E. Antarctica (Shiraishi et al., 1994), and for Usagaran belt, (Moller et al., 1995), respectively. W. Africa craton has collided to western Sahala since 560Ma (Black et al., 1994). Paleomagnetic data of major continents is referred to McCausland and Hodych (1998).

The Trinity ophiolite that had been formed for a long period from Cambrian through Ordovician to Silurian is distributed in Eastern Klamath, northern California, western North America. In Lovers Leap-South China Mountain area, plagiogranite dated 565-570Ma by zircon U/Pb dating is present (Wallin et al., 1988). The volcanic sequence consists of massive flow and dykes (Brouxel et al., 1989). These lithology is basalt, andesite, and dacite (Brouxel et al., 1989). All these are plagioclase and clinopyroxene porphyritic (Brouxel et al., 1989). Orthopyroxene present in a few andesites (Brouxel et al., 1989). Hornblend phenocrysts occur in the dacitic andsites and dacites (Brouxel et al., 1989). These are divided into two types based on geochemical features. One is low-K tholeiite with depleted REE pattern and other calc-alkaline rocks with enriched REE pattern (Brouxel et al., 1989). The calc-alkaline rocks are conformably overlain by Ordovician

limestone (Brouxel et al., 1988). Both low-K tholeiite and calc-alkaline rocks have negative Nb anomaly (Brouxel et al., 1989). Well-developed sheeted dyke structure is absent in this area (Lindsley-Griffin, 1994). The pegmatitic trondhjemite dykes that dated 412Ma by zircon U/Pb dating intruded the gabbroic layer (Wallin et al., 1988). In Dunsmuir area, ophiolitic members are completed. Nd isotopic ratios measured on basalt, plagiogranite, isotropic gabbro, cumulate layered gabbro, and gabbroic pegmatite are plotted on the same 452 ± 40 Ma isochron (Brouxel and Lapierre, 1988). The peridotite layer has unusual and significant features that large amounts of lherzolite occur (Brouxel and Lapierre, 1988) and that some lherzolites and harzburgites have high plagioclase content (Lindsley-Griffin, 1994). Sheeted dyke complex occurs in this area (Brouxel and Lapierre, 1988). The basaltic pillow lavas are dominant in volcanic sequences (Brouxel and Lapierre, 1988) and unconformably overlain by Carboniferous sandstone and shale (Brouxel et al., 1988). The basalts are divided into three types based on petrographical feature; (1) plagioclase porphyritic basalts with rare clinopyroxenes; (2) intersertal basalts with clinopyroxene microphenocrysts and plagioclase laths; (3) subaphyric basalts with clinopyroxene microlites (Brouxel et al., 1989). These lavas have depleted REE patterns and negative Nb anomaly (Brouxel et al., 1989). In Redding area, massive and pillowed flows overlain rhyolite formation occur (Brouxel et al., 1988). These are unconformably overlain Early Devonian shales and limestones (Brouxel et al., 1988). The flows consist of basalts and andesites. These are highly porphyritic with plagioclase, olivine and clinopyroxene phenocrysts (Brouxel et al., 1989). The basalts and andesites show a significant geochemical heterogeneity that is indicated by variability of REE pattern from moderately depleted to strongly enriched (Brouxel et al., 1988). These have negative Nb anomaly (Brouxel et al., 1989). High-Mg andesites with $\text{SiO}_2 \geq 55\%$, $\text{MgO} \geq 7\%$, and flat REE pattern also present (Brouxel et al., 1988). On the basis of these features, the Trinity ophiolite seems to be generated in island arc-marginal basin system (Brouxel and Lapierre, 1988; Brouxel et al., 1988; Brouxel et al., 1989).

At 600–650 Ma (Fig. 3-15 b, c). Magnetisms related with rifting events occurred in eastern Laurentia and Baltica (Hoffman, 1989; Kamo et al., 1989; Ernst and Buchan, 2002). It is seemed to be corresponding to the birth of Iapetus Ocean. We reconstructed the trench

distributions in S. America and southwestern Africa on the basis of the tectonic map of Campos Netto (2000).

The Wanbaogou ophiolite in the middle ophiolite zone of East Kunlun, western China, was dated to 684Ma by Rb-Sr isochron dating for the basalts (Yang et al., 1996). These basalts are overlain by cherts and carboniferous slate, which grade up to sandstones and siltstones and finally to carbonates (Yang et al., 1996). These consist of pillow lavas that are sparsely phiric with small plagioclase and clinopyroxene phenocrysts and massive lavas that contain plagioclase phenocryst (Yang et al., 1996). These are Ti-rich tholeiites and have LREE enriched patterns (Yang et al., 1996). It indicates oceanic island basalt affinity.

At 700 Ma (Fig. 3-15 d). East African Orogen reached the height of its prosperity (Stern and Kroner, 1993). The ophiolite belts had been developed in eastern and northern Africa and Arabia peninsula since 870Ma. They have 'supra-subduction zone' chemical features and infer to be formed in island arc or arc-related setting (Pallister et al., 1988).

At 750Ma (Fig. 3-15 e). Magnetism related with rifting events occurred in western Laurentia, Australia, and S. China (Hoffman, 1989; Ernst and Buchan, 2002). They are followed by the break-up of Rodinia. It is seemed to be corresponding to the birth of Pacific Ocean. The paleolatitudes of Australia and Laurentia also indicate breaking-up then (Wingate and Giddings, 2000). The Paleoposition of S. China is depended on the paleolatitude data from Evans et al. (2000). The reconstruction of trench distribution in S. America and southwestern Africa is referred from Campos Netto (2000).

At 800-850Ma (Fig. 3-15 f, g). In Australia and South China, magnetism related rifting occurred (Li et al., 1999). They are seemed to be followed minor break-up of Pilbara-Yilgarn and Kimberly fragments. Kalahari and Congo collided in basis of Zambezi orogen age.

At 850-900Ma (Fig. 3-15 g, h). Congo craton detached from Rodinia. It is based on the age of Lufilian Arc in a Zambezi Domain (Campos Neto, 2000). East African orogen have various age of 1100-520 Ma (Stern, 1994). Their orogens also have both east and west dipping sutures. We concluded the complicated arc systems surrounding microcontinents like a present western-southwestern pacific margin exist in northeastern Congo craton. We reconstructed the trenches in eastern Africa on the basis of the tectonic map in Stern (1994).

In a same way, 870-520 Ma granitoids in the Tuareg shield (Central Sahara) were compiled by Black et al. (1994) and we reconstructed the trenches in northern Africa in the basis of their geochronological data.

At 1000 Ma (Fig. 3-15 i). Grenville orogen reached the height of its prosperity and formed the Rodinia. Distribution of trenches is based on that of Grenville orogen (e.g. Hoffman, 1991) faced to the ocean.

3-4. Discussion

3-4-1. Reconstructed paleogeography

3-4-1-1. Rodinia

On the basis of established geological, geochronological and paleomagnetic data for the (pre-) Grenville orogens and subsequently developed dike swarms on each continent, several different models for the Rodinia reconstruction have been proposed (Fig. 3-16).

In terms of the configuration (directions are relative to the present continental positions), the central block of the Rodinia is composed mainly of Laurentia, Baltica and Siberia, connecting with the southern (or eastern) segment of South American and African blocks (West Gondwana). That Siberia was connected to the northern Laurentia is widely accepted but there has been little agreement on the geometry of the fit due to a lack of the deciding paleomagnetic or geochronological data. Kravchinsky et al. (2001) pointed out that the Siberian platform, composed of Anabar, Angara and Aldan blocks, might not have fully amalgamated by Vendian to Early Cambrian time.

On the other hand, a close connection of Siberia and North China, which has been suggested with a correlation of Riphean sedimentary rocks (Li et al., 1996), is supported by a correlation of the Late Proterozoic sedimentary rocks and paleomagnetic data (Halls et al., 2000). Recently reported Grenville-age Sibao orogen (Li et al., 2002), which has been intruded by mafic dikes of c. 830 Ma (Li et al., 1999) correlated to those in the southern-central Australia, suggests that the South China has been situated between Australia and

Laurentia in the Rodinia.

For a paleo-position of East Gondwana (Antarctica, Australia and India) against Laurentia in the Rodinia, three models are proposed; first is that the East Antarctica (Transantarctic Mountains) has connected with the southwestern coast of the Laurentia, which is based on correlations of Grenville-age orogens and the Precambrian-Paleozoic sedimentary rocks (SWEAT model: Moores, 1991; Dalziel, 1991; Hoffman, 1991; Young, 1992), and the South China was placed among Australia, Laurentia-Siberia (Fig. 3-16 a; Li and Powell, 2001). Powell et al. (1993) argued the validity of the SWEAT model with the APWPs for both East Gondwana (mainly from Australia and additionally from India) and Laurentia, and Li (2000) and Wingate and Giddings (2000) also supported this model from a viewpoint of the APWPs for the western Australia and Laurentia. Second is that a suture among the eastern coast of the India, the eastern coast of the East Antarctica and the southwestern part of the Australia has continued to the Grenville orogen in the eastern coast through the southern end of the Laurentia, and Australia has been connected with the western Laurentia (AUSWUS model: Karlstrom et al., 1999; 2001; Burrett and Berry, 2000). In this model, the South China was separated by small oceans from Australia and Siberia (Fig. 3-16 b; Condie, in press, 2003). Karlstrom et al. (2001) mentioned that the AUSWUS model provided a somewhat better fit of paleomagnetic data (APWP) than the SWEAT model. Third is a model with the Australia-Mexico connection (AUSMEX), inferred from c.1070-Ma paleopoles from Australia with high-stability paleomagnetic remanence (Wingate et al., 2002).

On the other hand, Fitzsimons (2000) and Boger et al. (2001) insisted that East Antarctica should be separated into two or three provinces with Grenville-age by the Pan-African orogens, on the basis of U-Pb zircon data and geological correlations. In addition, Powell and Pisarevsky (2002) pointed out that India lay near the pole whereas Australia was at low latitudes at c. 810 Ma, and final amalgamation of East Gondwana might not have been complete until the Early Cambrian. Considering these suggestions, Pisarevsky et al. (2003) proposed a new reconstruction of Rodinia at c. 990 Ma, in which the AUSMEX model was taken, and India was excluded from the Rodinia.

However, the recent AUSWUS and AUSMEX models suggest that up to 10,000-km of

the Late Proterozoic rift margins in the northern Australia and the western Laurentia needs to be matched with other continental blocks within any proposed Rodinia. As for the paleoposition of India, Radhakrishna and Joseph (1996) placed India near the Equator at c. 1000 Ma, suggesting the Late Proterozoic East Gondwana as before, but Pisarevsky et al. (2003) disproved this reconstruction on the basis of inconsistency between paleopoles of India and Australia at c. 1070 Ma, although the APWPs of India from 1100 to 800 Ma have not been obtained yet.

The recent accumulation of more precise paleomagnetic and geochronological data yielded several constraints and models for the Late Proterozoic reconstruction, however the controversy for East Gondwana and Rodinia became more complicated. At the present time, the deciding APWPs data are not obtained yet, and it is still difficult to evaluate validities of those recent reconstruction models. Therefore in this study, the Late Proterozoic Rodinia was reconstructed on the basis of paleomagnetic data of Weil et al. (1998) with more comprehensive data set of APWPs in the recent literature, as a tentative model. Nevertheless, main points of our discussion and conclusions stated below barely yield serious contradiction, depending on the reconstruction models adopted.

3-4-1-2. Pacific Ocean

In terms of the break-up process of Rodinia, at least four substages of continental rifting were recognized on the basis of geological and geochronological data (Hoffman, 1991; Li et al., 1999; Dobretsov et al., 2003).

Based on correlations of ages of mafic dike swarms in Laurentia and adjacent cratons, Hoffman (1991) suggested two stages of continental rifting, related to the Rodinia breakup, between the eastern Laurentia, West Gondwana and Baltica at 560-620 Ma, and between the western Laurentia and the East Gondwana at 680-780 Ma; these two stages of rifting resulted in opening of Iapetus Ocean and (Paleo-) Pacific Ocean, respectively. Li et al. (1999) reported zircon SHRIMP ages of mafic dikes from South China, correlating with those from the eastern Australia, and proposed the earlier stage of rifting between South China and Australia at c. 830 Ma. Recently, Dobretsov et al. (2003) compiled geological

and geochronological data of mafic dikes and ophiolitic complexes along margins of Siberia, and suggested that a timing of the initial rifting in the Rodinia traced back to c. 950 Ma. According to the Rodinia reconstruction of Dobretsov et al. (2003), the initial stage of rifting for the Rodinia breakup is marked by mafic dikes intruding the pre-Grenvillian granite-gneiss complex in the southern margin of Siberia (Ar/Ar ages of 858-886 Ma; Sklyarov et al., 2001) and ophiolitic dike complexes in the Aldan shield and the Uchur-Maya region of the eastern Siberia (Sm/Nd ages of 900-950 Ma; Makrygina et al., 1993). In their reconstruction of Rodinia (Fig. 3-16 c), South China is not taken into account, and the northern and the eastern margins of Siberia face to the northern Laurentia (and Greenland), and the eastern Australia, respectively. Thus, a counterpart of the rift margin at 860-950 Ma of the eastern Siberia was considered to be the eastern Australia (Dobretsov et al., in press, 2003), where any contemporaneous mafic dikes have not been recognized, yet, and an ocean opened by the initial rifting was named, Paleo-Asian Ocean (Dobretsov et al., 1995). In addition, Dobretsov et al. (2003) pointed out that opening of the Paleo-Pacific Ocean might have earlier occurred at its southern part between Laurentia and Australia (740-780 Ma) than at its northern part between Laurentia (and Greenland) and Siberia (680-740 Ma).

As a summary, it is most likely that the early stages of rifting for the Rodinia breakup would have occurred at the eastern margin of Siberia (860-950 Ma: Dobretsov et al., 2003) or between South China and Australia (c. 830 Ma: Li et al., 1999), and that the Paleo-Pacific Ocean would have opened between the western Laurentia, Australia and South China, subsequently between the northwestern Laurentia and Siberia. In those periods, the Paleo-Pacific Ocean was born at RRR (ridge-ridge-ridge) triple junctions defined by Laurentia, Australia, and South China, subsequently by the northwestern Laurentia, Siberia and South China.

The RRR triple junctions are recognized in all reconstructions although the defining cratons are different by workers (Fig. 3-16). On the basis of the present analogy of African rifting system, a RRR triple junction is the expression that the continental rifting is caused by the plume up-welling (see summary by White and McKenzie, 1989). Maruyama (1994) first proposed that the Pacific superplume generated the Paleo-Pacific Ocean at 600-700 Ma on the basis that paleolatitude of the RRR triple junction at was c. 20 °S, corresponding to

an active region of the present Pacific superplume, and placed paleoposition of the Pacific superplume at the RRR triple junctions between the western Laurentia, Australia and South China, and between the northern Laurentia, South China and Siberia (Fig. 3-16 d). Li et al. (1999) and Li and Powell (2001) showed a similar style of the triple junction at c. 760 Ma, and suggested a mantle plume beneath South China; however, its paleolatitude was put at c. 30 °N (Fig. 3-16 a), due to difference in adopted paleomagnetic data set from Maruyama (1994). In our reconstruction mainly depending on paleomagnetic data set of Weil et al. (1998), the Paleo-Pacific Ocean was opened by the continental rifting with the RRR triple junction between the northwestern Laurentia-Siberia, Australia and East Antarctica at paleolatitude of c. 20 °S around 750 Ma (Fig. 3-16). It is similar to that of other workers, but differs in the paleolatitude from Li and Powell (2001), and in opening between the northern Laurentia-Greenland and Siberia at c. 750 Ma from Maruyama (1994), Condie (2003), and Dobretsov et al. (2003). Nevertheless, we have a consensus that the Paleo-Pacific Ocean would have opened by the continental rifting with the RRR triple junction to the north of the western Laurentia at c. 750 Ma, which would have been triggered by upwelling of the mantle plume, probably the Pacific superplume. Such a process of the Rodinia breakup is comparable with the discussion that break-up of Pangea and subsequent opening of the Atlantic Ocean was initiated by upwelling of the African superplume (Daziel et al., 2000).

As a summary, most probable history of the Pacific Ocean from 750Ma to 165Ma is illustrated in Fig. 3-17. The Pacific Ocean was born by break-up of supercontinent Rodinia at ca. 750Ma (Fig. 3-17 a). It was born at RRR (ridge-ridge-ridge) triple junction defined by Laurentia, Australia, and E. Antarctica. However the defining cratons are different among the Rodinia breaking-up reconstructions by other authors, RRR triple junction is recognized in all reconstructions (Fig. 3-16). On the basis of the present analogy of African rifting system, RRR triple junction is the expression that the rifting is caused by the plume upwelling. The paleolatitude of RRR triple junction in our reconstruction is ca. 20°S. It is similar to the latitude of the present Pacific super plume. These lead us to speculate that the superplume generating the Pacific Ocean at 750Ma is corresponding to the present Pacific superplume same as Maruyama (1994). It is comparable with the discussion that break-up

of Pangea is caused by African superplume (Daziel et al., 2000). At ca. 550 Ma, subduction zones begun to develop along the Pacific margin (Fig. 3-17 b). It occurred along both eastern E. Gondwana and western Laurentia. The Mozambique orogen begun to develop at this timing. The subduction along the Pacific margin seems to be demanded on the lack of the plate consuming because of the termination of Pan-African orogen. The Pacific Ocean passed through the superocean - only one continent and one ocean exist on the earth- at ca. 230 Ma (Fig. 3-17 c) and begun to close at ca.180 Ma (Fig. 3-17 d). These are coincided with the formation of supercontinent Pangea and the birth of the Atlantic Ocean, respectively. The history of the Pacific Ocean has recorded not only on the Pacific Ocean itself but also in the circum-Pacific orogens since 550 Ma.

3-4-2. Impacts of the Pacific superplume to the surface environment

Our paleogeographic reconstruction suggests several implications in relationships between the paleogeography (the surface geology), the global environmental changes, and the Pacific superplume activities in the Earth's history.

In the present western Pacific, Cretaceous oceanic plateaus are widespread (e.g., Manihiki Plateau, Shatsky Rise, Hess Rise, Mid-Pacific Mountain), and most of them have been formed at 90-130 Ma (Table 3-4). They were traced back and concentrated to the present active region of the Pacific superplume (Fig. 3-17). These suggest that those were once formed a huge composite volcano as large as 4000 km x 2400 km across at the Cretaceous by superplume activity. Estimated crustal thickness of this composite volcano is c. 26 km; its estimated size ($9.5 \times 10^6 \text{ km}^2$) is larger than Australia continent ($7.6 \times 10^6 \text{ km}^2$). Total volume is estimated as c. $2.5 \times 10^8 \text{ km}^3$, which is comparable with the Tharsis bulge volcanic complex on Mars ($3 \times 10^8 \text{ km}^3$) as the largest volcano on planets in the solar system at the present. These estimates are about 30 % larger than those estimated by Larson (1991b). This difference resulted from addition of accreted plateaus around the circum-Pacific orogenic belts (Kimura et al., 1994; Kerr et al., 1996; 1998; Ozawa et al., 1999; Suzuki et al., 2000; Ota et al., 2002a). We estimated height of the composite volcano from Cretaceous to present day, using the Sclater curve (e.g., Parsons and Sclater, 1977; Sclater

et al., 1980) and current sea depth of oceanic plateaus and seamounts. It is c. 2000 m (max. 3600 m) above sea level and would be a vast plateau when it was formed, and major part of this plateau sank below sea level after 50 million years (Fig. 3-18).

It has been pointed out that relationship between tectonic events related to superplume activities and the surface environmental change, such as sea-level change, deposition of black shale and global warming with CO₂ releasing. It was suggested that such environmental changes affect the biological evolution, including diversification and extinction.

As the oldest mention, Valentin et al. (1970) showed a model that the diversity of marine fauna and fluctuations in sea level could be related to patterns of continental fragmentation and reassembly. Moores (1993) noticed relationships among changes of surface environment, thickness of oceanic crust and mantle activity, and suggested a model that the Neoproterozoic oceanic crust thinning caused lower average sea level, increasing erosion, sediment deposition; biological productivity and burial led to increased atmospheric oxygen, which permitted the evolutionary development of animals. Condie et al. (2001) mentioned prominent peaks of black shale abundance and its ratio in the shale, peaks of chemical index of alteration of shale, and the carbon isotope anomalies at Precambrian strata, and suggested that these signs might reflect the combined effects of mantle plume events and formation and breakup of supercontinents. The mantle plume events at 2.7 and 1.9 Ga may have introduced large amounts of CO₂ into the atmosphere-ocean system, increasing depositional rates of carbon and increasing global warming, and the breakup of supercontinents at 2.2-2.0 Ga and again at 800-600 Ma, which resulted in increased numbers of partially closed marine basins, disruption of ocean currents, and increased hydrothermal vents at ocean ridges, all of which led to widespread anoxia.

These studies suggest that a chain of the tectonic events occurred by the superplume activity likely cause the environmental change on the Earth surface, and the change leads to evolution of the life, too.

In the mid-Cretaceous time, most of global changes on the surface environment could be comprehensively explained by the activities of the Pacific superplume (Larson, 1991a; 1991b); the Cretaceous global transgression would be caused by a rise of sea level due to

increasing spreading rates of mid-oceanic ridges and the resultant young oceanic lithospheres with the Pacific superplume activity. The Cretaceous global warming also would result from increasing rates of degassing by an extensive volcanism related to the superplume activity. Jaharen (2002) also noticed that the mid-Cretaceous carbon isotope records of marine carbonate and organic carbon, and terrestrially photosynthesized carbon showed a 3-5 % negative excursion, as evidence of the methane release. Rates of land uplifting by upwelling of superplume would result in the liberation of hundreds of Gt of carbon from methane clathrates and the associated eruption of the Ontong-Java plateau basalts would additionally destabilized hundreds of Gt of carbon from methane hydrates during the mid-Cretaceous. The oxidation of these methane, released in the biosphere, would lead to disrupt the chemical cycling, including oceanic anoxia, and allow angiosperms to displace conifer at mid- and high-latitudes (Jaharen, 2002). In our estimation, a total amount of CO₂ derived from magmatism at that time is c. 7.7×10^{21} g. The amount of CO₂ produced by our reconstructed composite volcano, MORB and arc volcanic rocks, are estimated as 2.3×10^{21} g, 4.5×10^{21} g and 0.9×10^{21} g, respectively, where CO₂ contents in plateau basalt, in arc basalt and in MORB were assumed to be 0.3, 0.2 and 0.1 wt% (Sano and Williams, 1996; Marty and Jambon, 1987, Pineau et al., 2004), respectively. The amount of CO₂ deduced from our composite volcano occupies as much as 30 wt% of the total estimation in the Cretaceous time, suggesting a significant impact for the Cretaceous global warming. In addition, this period corresponds to the Cretaceous superchron; it is no doubt that the Pacific superplume activity in a whole mantle scale would have affected to convection in the outer core (e.g, Larson, 1991a; Tatsumi et al., 1998).

Such an episodic activation in the whole mantle, i.e., the extensive volcanism both at the spreading center and the hot spot seems to have been a specific phenomenon in the Pacific domain (e.g., Larson, 1991a; 1991b), although its mechanism has been not clear, yet. Recent knowledges by ultrahigh-P experiments, however, enable us to speculate the following model on the Cretaceous active volcanism in the Pacific (Fig. 3-19). Phase relations in a mantle peridotite system (Zhang and Herzberg, 1994; Hirose, 2002) suggests that the plume geotherm passes the majorite/perovskite transition curve with a positive

slope, indicating that a mantle upwelling can be accelerated at c. 670-km depth to transport fertile lower mantle materials to the upper mantle (Fig. 3-19 a). Hot spot volcanism related to the mantle upwelling indicates that a window between the upper and the lower mantles is opened in the Pacific. However, the mantle convection at a normal stage is double-layered and the mantle materials are independently circulated within the upper and the lower mantles (Fig. 3-19 b). The downwelling beneath a subduction zone due to plate subduction since c. 450 Ma in the circum-Pacific is stopped once at 670-km depth by a negative Clausius-Clapeyron slope along the cold geotherm along a subducting plate, resulted to form a megalith (e.g., Irifune and Ringwood, 1993). Subsequently, the megalith collapses down to the lower mantle, and accelerates a single-layered convection at a whole mantle scale (Fig. 3-19 c); the resultant mantle upwelling activates episodically. As a result, material- and heat-supply from the lower to the upper mantles through the window activate the volcanism both at the mid-oceanic ridge and the hot spot in the Pacific domain. No window between the upper and lower mantles would be opened beneath the Indian and the Atlantic domains at that time. Because of this, the volcanism both at the spreading center and the hot spot would be activated only in the Pacific, and would not be in the Indian and the Atlantic.

In the previous section, we assumed that oceanic plateaus would have been formed at periods of the ophiolite pulses (Fig. 3-5) before the mid-Cretaceous, too. Those pulses also might have caused the global environmental changes similar to those in the Cretaceous time. Around the break-up period of the supercontinent Pangea (c. 200 Ma), the extensive volcanism with continental flood basalts has been occurred at South Africa (Karoo) and East Antarctica (Ferra). As a possible cause of the end-Permian extinction, it has been pointed out that an intensive magmatism at the initial stage of the break-up of Pangea, predating the extensive volcanism formed the flood basalts, might have strongly influenced redox states in the ocean-atmosphere, resulted in the global mass extinction around the Permian-Triassic boundary (Fig. 3-20; Isozaki, 1997b; 2002). Around the Proterozoic-Phanerozoic boundary in the broad sense, environmental changes such as the global glaciation (snow-ball Earth: Hoffman et al., 1998; Hoffman, 1999) and mass extinction have been recognized. This period corresponds to the break-up period of the supercontinent

Rodinia with the birth of Pacific superplume. The superplume activities at these periods with break-up of the supercontinents might have caused catastrophe, affecting biological evolution such as the diversification and extinction, of higher order than that in the Cretaceous pulse period.

Based on the paleogeography of the Grenvillian orogens within the Rodinia, we can speculate on the site of subduction zones when the Rodinia was formed, and on Pacific-type subduction zones afterward developed around its margin. During the formation process of the Rodinia, a large amount of oceanic lithosphere would have been subducted beneath the proto-Rodinia along paleo-trenches of the proto-Grenville orogens. About 200 m.y. after that, the Pacific superplume would be born at the center of the Rodinia, where was surrounded by the Grenville orogens (Fig. 3-15 i).

Moreover, the present position of the African superplume also overlaps in paleo-latitude with the region where the 750-500 Ma Pan-African orogens have been present (Figs. 3-15 a-e). The southern half of the Late Paleozoic Pangea, Gondwana was formed by extensive collision-amalgamation processes around Paleo-Africa, which was completed by the collision of East and West Gondwana at c. 520 Ma (Fig. 3-15). Since 520Ma, subduction zones developed along the western, southern and eastern continental margins of Gondwana with the resultant formation of V-shaped Pacific-type orogens (Fig. 3-14). The northern half of Pangea; Laurasia was completed by extensive collision-amalgamation of more than seven continents during the Permo-Triassic (Fig. 3-14), suggesting the presence of large amounts of subducted lithospheres underneath the present Asia.

These correlations suggest that the birth of superplume would be caused by something during formation processes of the supercontinents, which include the Pacific-type orogeny along the V-shaped subduction zones with the resultant collage of island arcs (Fig. 3-21). Concentration of oceanic plate subduction during the formation process of supercontinent could have greatly accelerated to make physical instabilities and chemical heterogeneities in deep mantle, either or both of which must deeply be concerned in the birth of superplume. In terms of this, there is no reliable model with consensus, however transportation of oceanic lithospheres into the deep mantle most likely plays an essential role in the birth of superplume. Unresolved problems in terms of the birth of superplume,

including its initial driving force and why the superplume is born only beneath the supercontinent, are the ultimate and essential subjects to establish the plume tectonics theory and beyond, the whole Earth dynamics.

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Table 3-1, Geological and petrochemical characteristics of volcanic sequences from different tectonic settings (after Miyashiro, 1975; Ishiwatari, 1985a; Wilson, 1989; Rollinson, 1993; Ishizuka, 1995; Tatsumi et al., 1998; Kerr et al., 2000). Grades of ocean floor metamorphism (*) are estimations inferred from crustal thickness of volcanic sequences, except in a case of the mid-ocean ridge with accumulated data (summary in Ishizuka, 1995). Abbreviations: pmn, primitive mantle normalized; cn, chondrite normalized; Amp, amphibole; Cpx, clinopyroxene; Ol, olivine; Opx, orthopyroxene; Pl, plagioclase.

tectonic setting	oceanic plateau	oceanic island /seamount		mid-ocean ridge	marginal basin	island arc
predominating sediment	shallow-marine pelagite	shallow-marine pelagite		deep-sea pelagite	hemipelagite, terrigenous clastics	terrigenous clastics
crustal thickness	25~40 km	7~15/7~10 km		< 7 km	< 7 km	7~20 km
ocean floor metamorphism (gabbro layer)	possibly granulite facies	~ amphibolite facies		~ amphibolite facies	~ amphibolite facies	possibly granulite facies
dike complex	not sheeted	not sheeted		sheeted mainly aphyric	sheeted	no
texture of volcanics	mainly aphyric	porphyritic			mainly aphyric	porphyritic
fractionation control	Ol, Cpx, Pl	Ol, Cpx		Ol, Pl	Ol, Cpx, Pl	Ol, Opx
type of basalt	TH, A	A, TH		low-K TH	TH, A	(Cpx), Amp
high-MgO (>14%)						CA, TH, A
lava	yes	yes		rare	rare	rare
(La/Nb)pmn	< 1	< 1		< 1	< 1	>> 1
(REE)cn pattern	flat	LREE enriched + HREE depleted		LREE depleted	flat	LREE enriched
		normal hot spot	superplume-related			
Nb/Y		< 5.5(Nb/Zr)+0.5	> 5.5(Nb/Zr)+0.5			
Nb/Zr			non-HIMU	HIMU		
²⁰⁶ Pb/ ²⁰⁴ Pb			< 0.23	> 0.23		
			< 20.5	> 20.5		
modern analogue	Ontong Java	Hawaii	Society	Cook-Austral	-	-

Table 3-2, Summary of compiled ophiolites, greenstone belts and the large igneous provinces (LIPs). Their localities are shown in Fig. 3-3.

Sym- bol	Name	Region	Type	Erupted age, Ma		Accreted age, Ma		Reference
a	Sorachi Plateau	E.Asia	OP	132	-	161	-	132 Kimura et al., 1994; Hori and Sakakibara, 1994; Sakakibara et al., 1999
b	Sanbagawa-Mikabu Greenstones	E.Asia	OP	142	-	153	110 -	120 Isozaki and Itaya, 1990; Sakakibara et al., 1993; Ozawa et al., 1997; Okamoto et al., 2001
c	Gorny Altai	Central Asia	SM		Vend-E.Camb		E.Cambrian	Buslov et al., 1993; Buslov and Watanabe, 1996; this study
d	Kazakhstan	Central Asia	IA		Vend-Ord			Yakubchuk, 1994; Yakubchuk et al., 1994
e	Arabian-Nubian Shield ophiolite	N.Africa	IA	694	-	850		Stern, 1994; Yakubchuk et al., 1994
f	East-Sulawesi Ophiolites	Indonesia	OP	79	-	137		Kadurusman et al., 2001
g	Victorian greenstone belts	SE. Australia	IA		Cambrian			Crawford and Keays, 1978
h	Northland Ophiolite	New Zealand	MO R		L.Creta-Olig			Brothers and Delaloye, 1982; Malpas et al., 1994
i	Dun Mountain Ophiolite	New Zealand	IA	275	-	285		Kimbrough et al., 1992; Malpas et al., 1994
j	Brooks Range Ophiolite	N.America	OP	165	-	185		Wirth et al., 1993; 1994
k	Angayucham Ophiolite	N.America	OP		Missisipian			Barker et al., 1987; Murchey, 1989
l	Resurrection Peninsula ophiolite	N.America	MO R	56	-	58		Lytwyn et al., 1997; Kusky and Young, 1999
m	Wrangellia Flood Basalt	N.America	CFB		L.Triassic			Jones et al., 1977; Lassiter et al., 1995
n	Trinity ophiolite	N.America	MB	556	-	579		Linsley-Griffin, 1994
o	Coast Range ophiolite	N.America	MO R	163	-	169		McLaughlin et al., 1988; Hagstrum and Murchey, 1996; Giaramita et al., 1998
p	Josephine Ophiolite	N.America	MB	157	-	162		Harper, 1985; 1994; Harper et al., 1994
q	Caribbean-Columbian Cretaceous Igneous Province + Central/Western Cordilleran ophiolite	S.America	OP	69	-	94		Schubert and Sandwell, 1989; Kerr et al., 1996; 1998; Sinton et al., 1998
r	Chilean Ophiolite	S.America	OP		Ordovician			H_rve et al., 1987

Table 3-3. Abbreviations for continental fragments described in this study.

Abbreviation	Craton
Af	Africa
Am	Amazonia
Ant	Antarctica
Aus	Australia
Ba	Baltica
Ch	China
Co	Congo
EA	East Antarctica
In	India
Ka	Kalahari
Kz	Kazakhstan
La	Laurentia
N.Am	North America
NC	North China
PY	Pilbara-Yilgarn (Western Australia)
RP	Rio de la Plata
S. Am	South America
SC	South China
Si	Siberia
Ta	Tarim
WA	West Africa

Table 3-4. Summary of the Cretaceous LIPs and their accreted fragments. AOP, accreted oceanic plateau; OBFB, oceanic basin flood basalt; OP, oceanic plateau; SM, seamounts. Data sources are the same as Table 3-2.

Name	Type	Erupted age, Ma	Area, km ²	Volume, km ³	Thickness, km
Ontong Java Plateau	OP	119-125	1.9E+06	6.7E+07	36
Nauru Basin	OBFB	121-124	3.6E+05	3.6E+06	-
Manihiki Plateau	OP	112-125	6.5E+05	1.6E+07	25
Marcus Wake Seamounts	SM	106-120	1.5E+06	3.1E+07	-
East Mariana Basin	OBFB	115-119	3.8E+05	3.8E+06	-
Magellan Seamounts	SM	-	9.0E+04	1.8E+06	-
Mashall Gillbert Seamounts	SM	-	2.0E+04	4.0E+05	-
Magellan Rise	OP	140-150	2.4E+05	3.6E+06	-
Mid-Pacific Mountains	SM	75-130	1.8E+06	4.3E+07	-
Pigafetta Basin	OBFB	113-119	2.5E+05	2.5E+06	-
Line Islands	SM	75-85	3.2E+05	1.0E+07	-
Shatsky Rise	OP	130-150	5.4E+05	9.9E+06	-
Hess Rise	OP	95-120	5.9E+05	7.8E+06	-
Caribbean Plateau	OP	80-120	8.0E+05	2.0E+07	10-20
Sorachi Ophiolite	AOP	110-160	6.0E+04	5.0E+05	30
Mikabu Greenstone Belt	AOP	142-153	-	4.0E+05	-
East Sulawesi Ophiolite	AOP	120-140	3.0E+04	4.5E+05	15

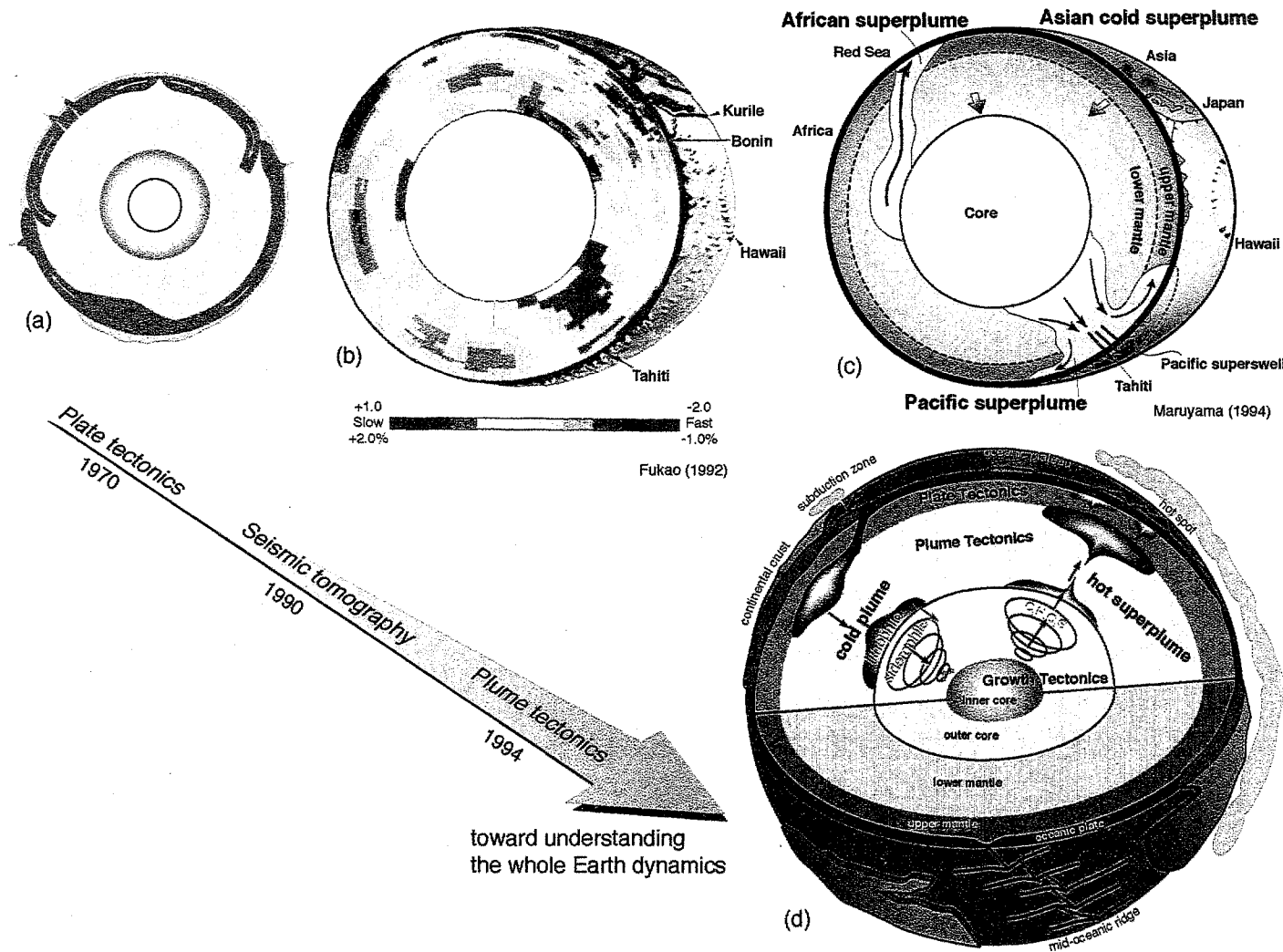


Fig. 3-1. Cartoons showing transition of paradigm on the Earth dynamics from plate tectonics to plume tectonics, and scheme of plume tectonics theory. (a) Plate tectonics. Orogenic processes at subsurface domains of the Earth are controlled by motion of several rigid plate. (b) Mantle tomographic image (Fukao, 1992), and (c) its interpretation (Maruyama et al., 2002a). There are two super-upwellings beneath the Pacific Ocean (Pacific superplume) and African continent (African superplume), and one super-downwelling beneath the Asian continent (Asian cold superplume), with the whole mantle scale. (d) The whole Earth dynamics from a view point of the plume tectonics (modified after Maruyama, 1994). The solid Earth is divided into three domains; the subsurface domain controlled by the plate tectonics, the mantle domain by plume tectonics, and the core by the growth tectonics. The whole Earth dynamics functions with interactions among them, which are triggered by episodic superplume activities.

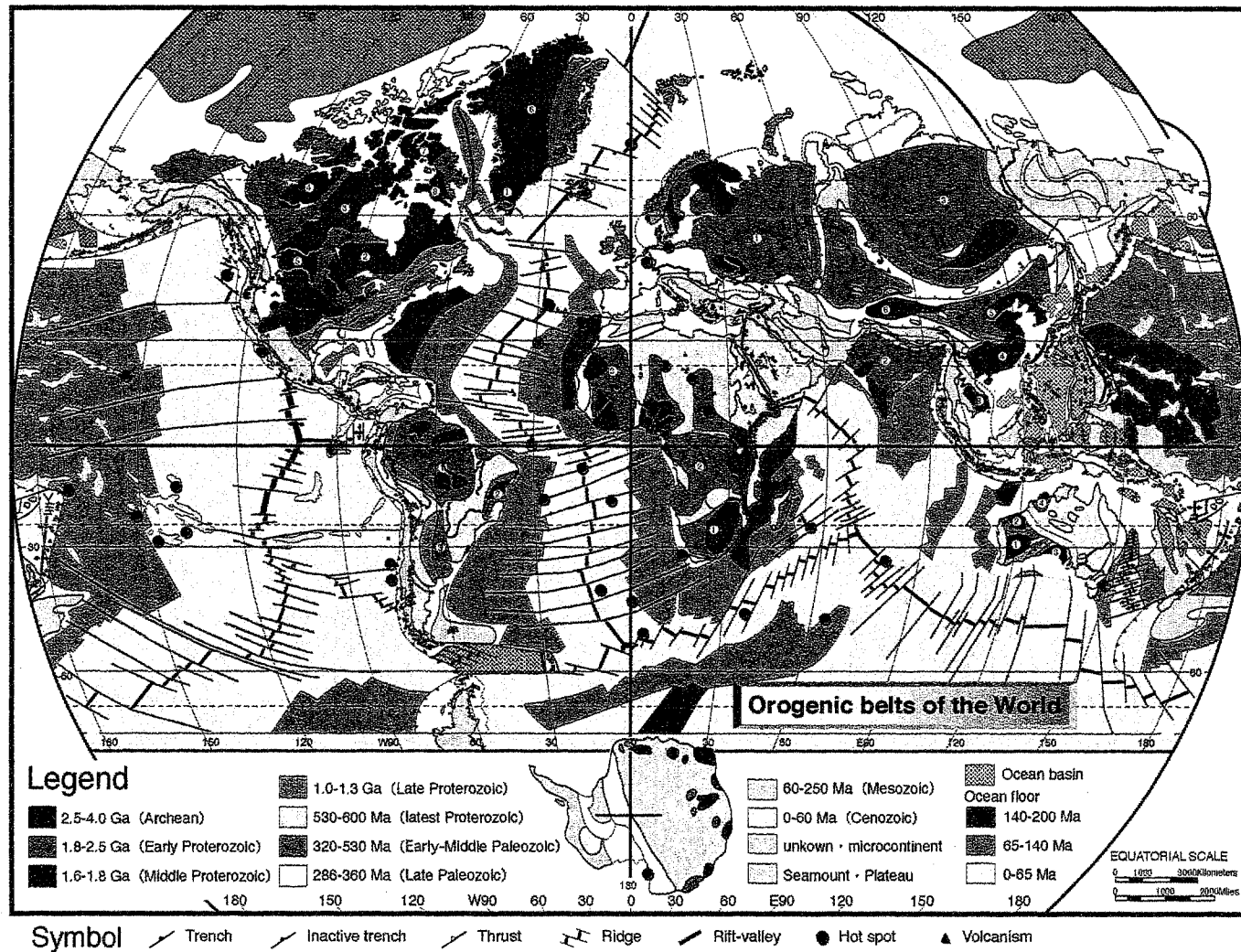


Fig. 3-2 ☒ Compilation map of orogenic belts over the world. Major cratons in each continent are numbered: (Asia-Europe) 1, Russian platform; 2, Indian platform; 3, Siberian platform; 4, Yangtze craton; 5, Sino-Korean craton; 6, Tarim block; 7, Arabian shield; 8, Indochina block; (Australia) 1, Yilgarn; 2, Pilbara; 3, Gawaler; 4, Kimberley; (North America) 1, Nain; 2, Superior; 3, Hearne; 4, Slave; 5, Wyoming; 6, Nagssugtoqidian; 7, Rae; (South America) 1, Amazon; 2, San Francisco; 3, Rio de La Plata; (Africa) 1, Kalahari; 2, Congo; 3, West Africa. See text for further explanations.

Large igneous provinces (LIPs) and ophiolite/greenstone belts over the world

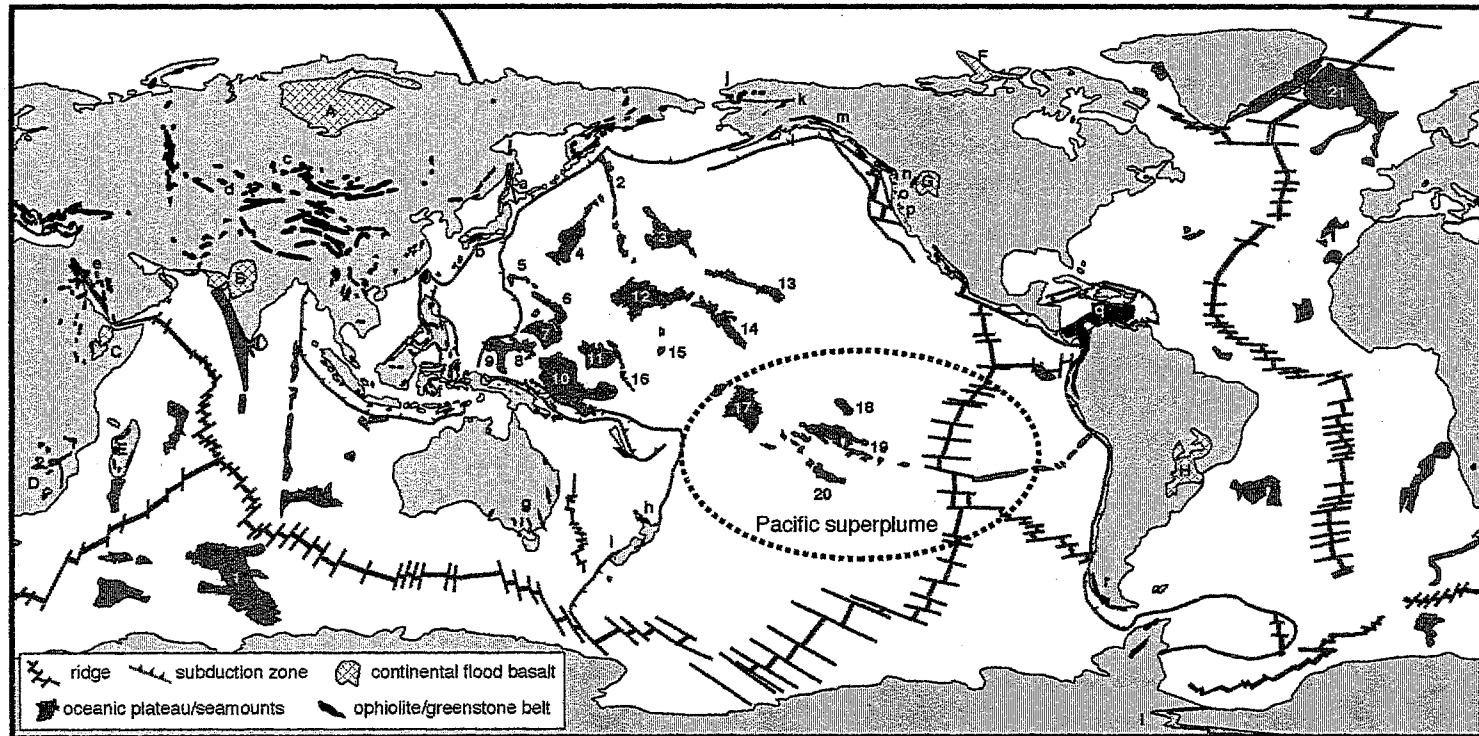


Fig. 3-3. Distribution map of large igneous provinces (LIPs), ophiolites and greenstone belts over the world. Numbers and alphabetic symbols show LIPs, ophiolites and greenstone belts listed in Table 3-2.

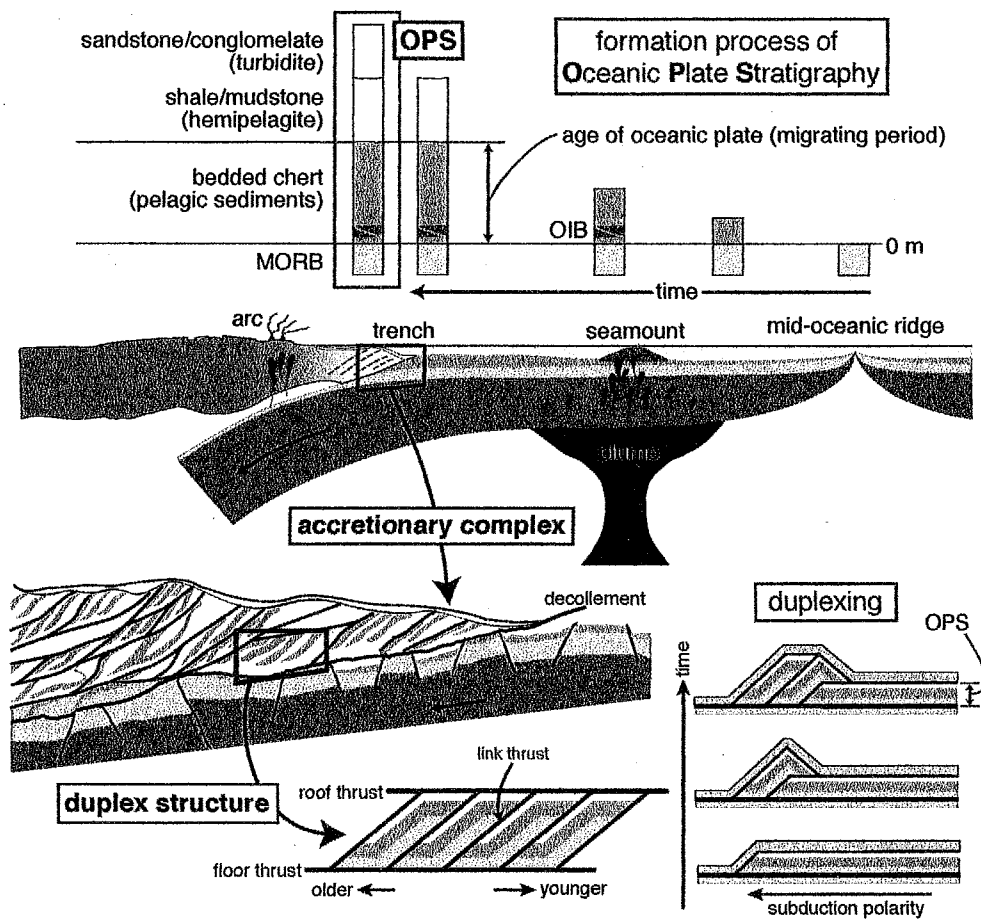


Fig. 3-4. Formation process of oceanic plate stratigraphy and accretionary complex. See text for explanations.

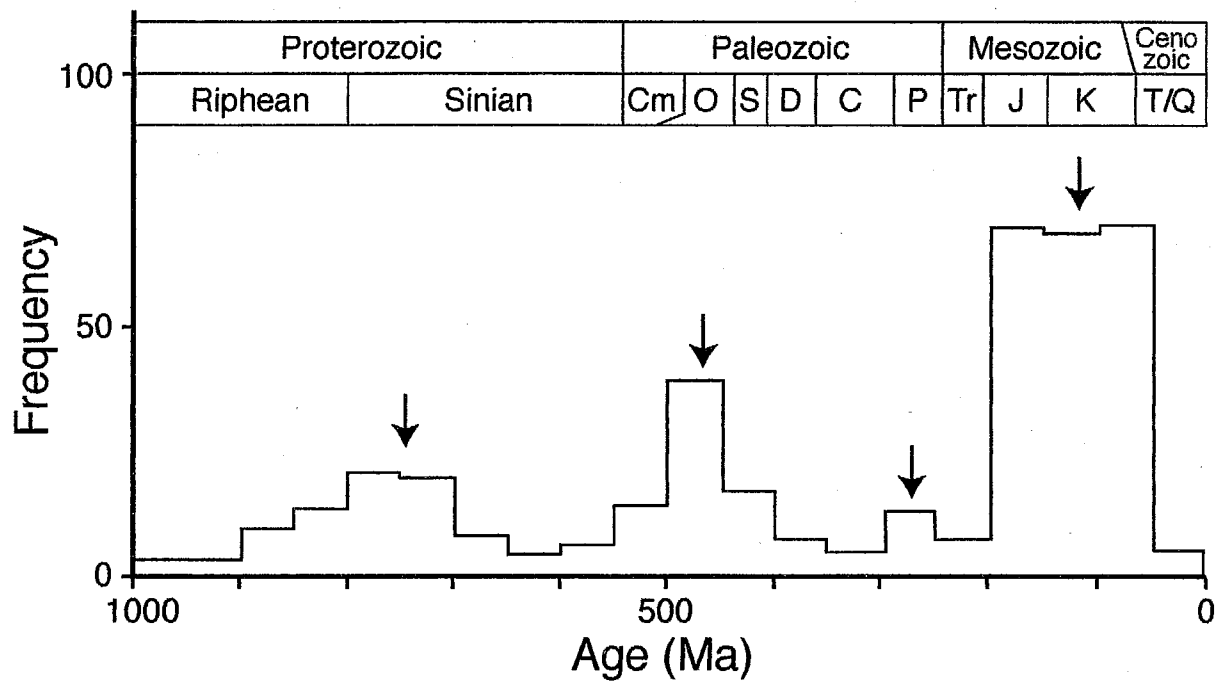


Fig. 3-5. Frequency diagram on ages of ophiolites and greenstone belts in the Pacific-type orogens (modified after Ishiwatari, 1994). Arrows indicate periods of ophiolite pulses.

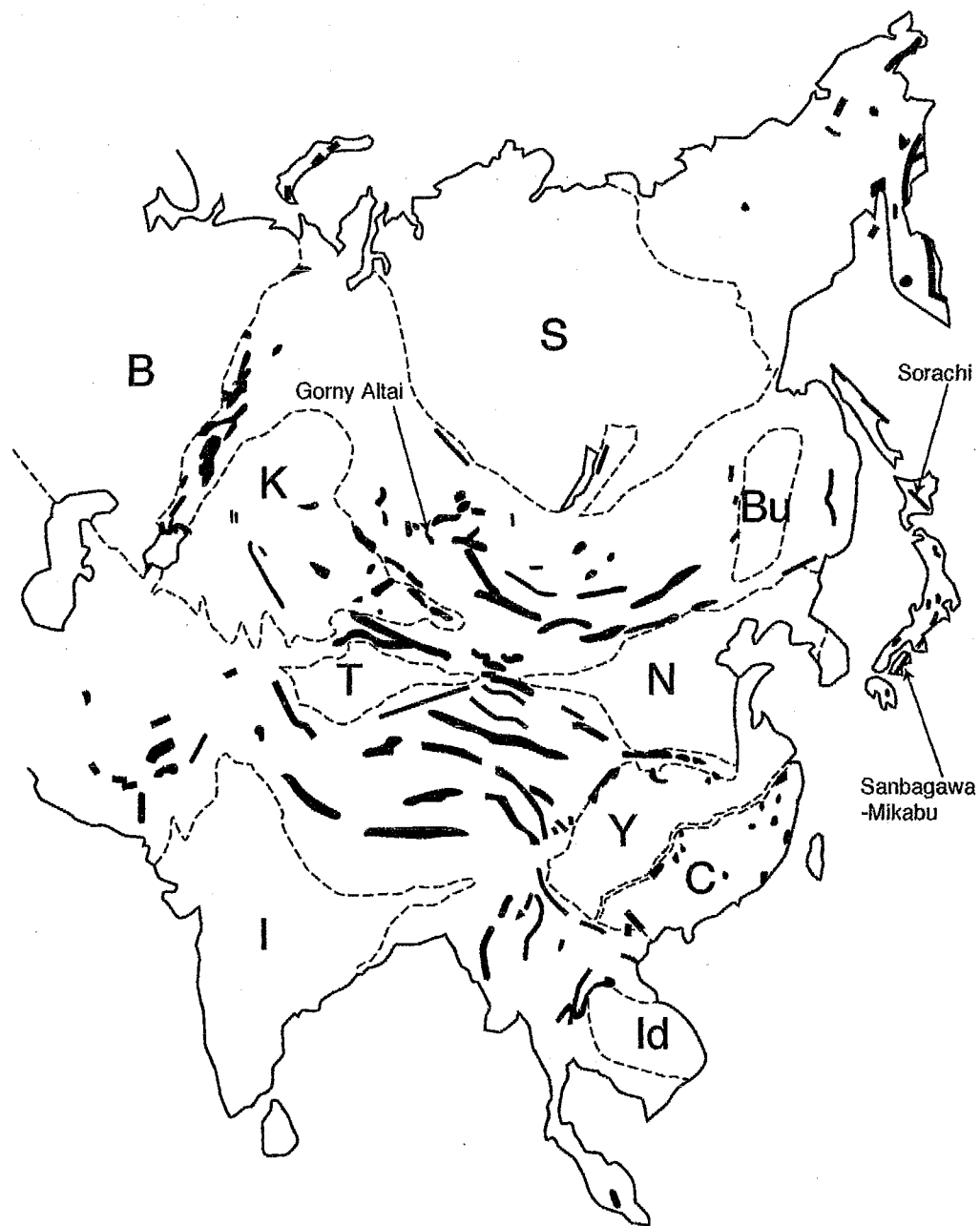


Fig. 3-6. Distribution map of ophiolites and greenstone belts in Asia. Abbreviations used are: S: Siberian Platform; B: Baltica; K: Kazakhstan; Bu: Bureya; T: Tarim; N: North China Platform; Y: Yangtze Platform; C: Cathaysia; I: Indian Platform; Id: Indonesia Block.

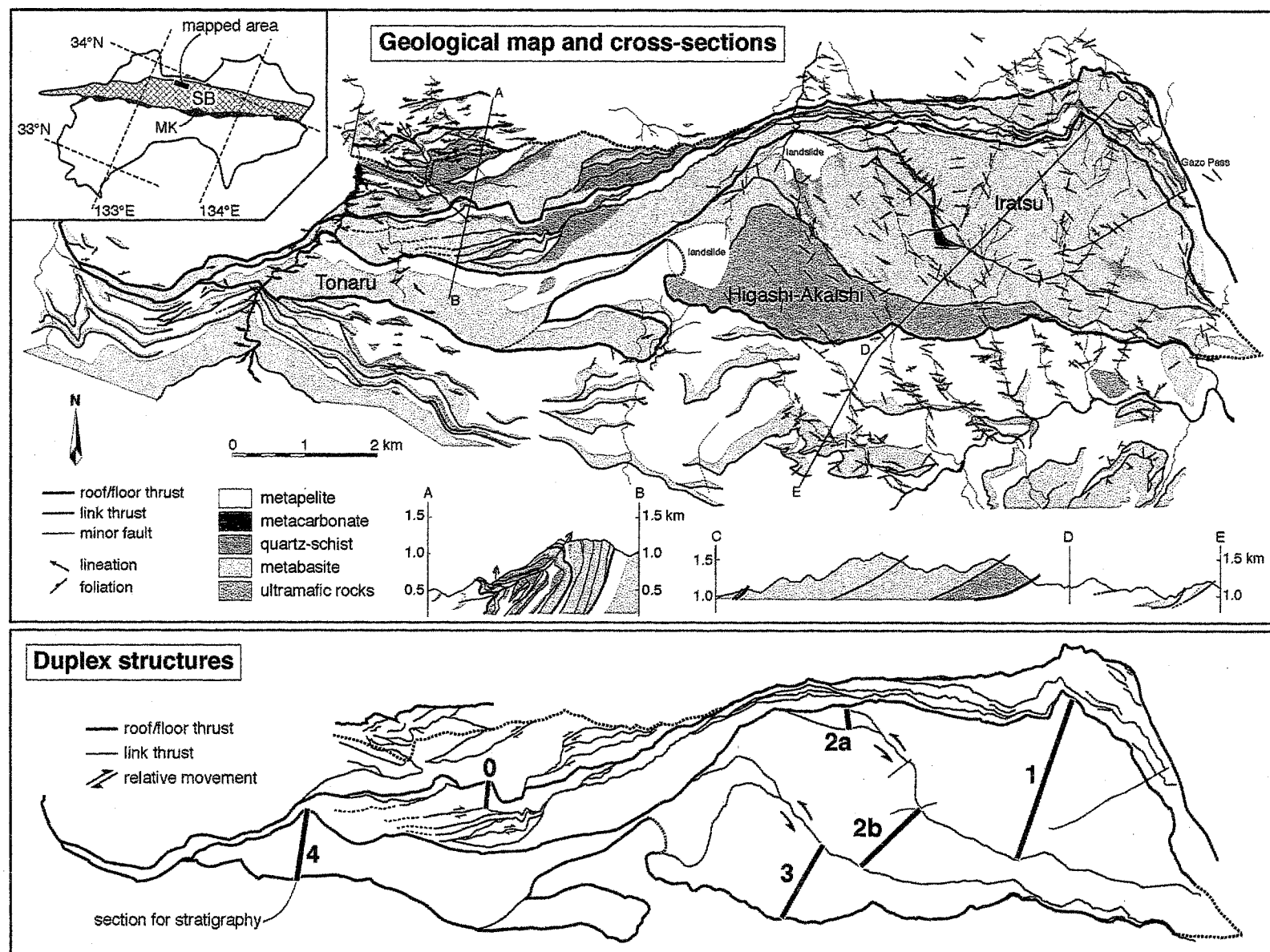


Fig. 3-7. Geologic map around the Iratsu- and the Higashi-Akaishi eclogite-peridotite mass in the Sanbagawa belt, central Shikoku, southwestern Japan (above) and simplified map showing duplex structure in study area (below).

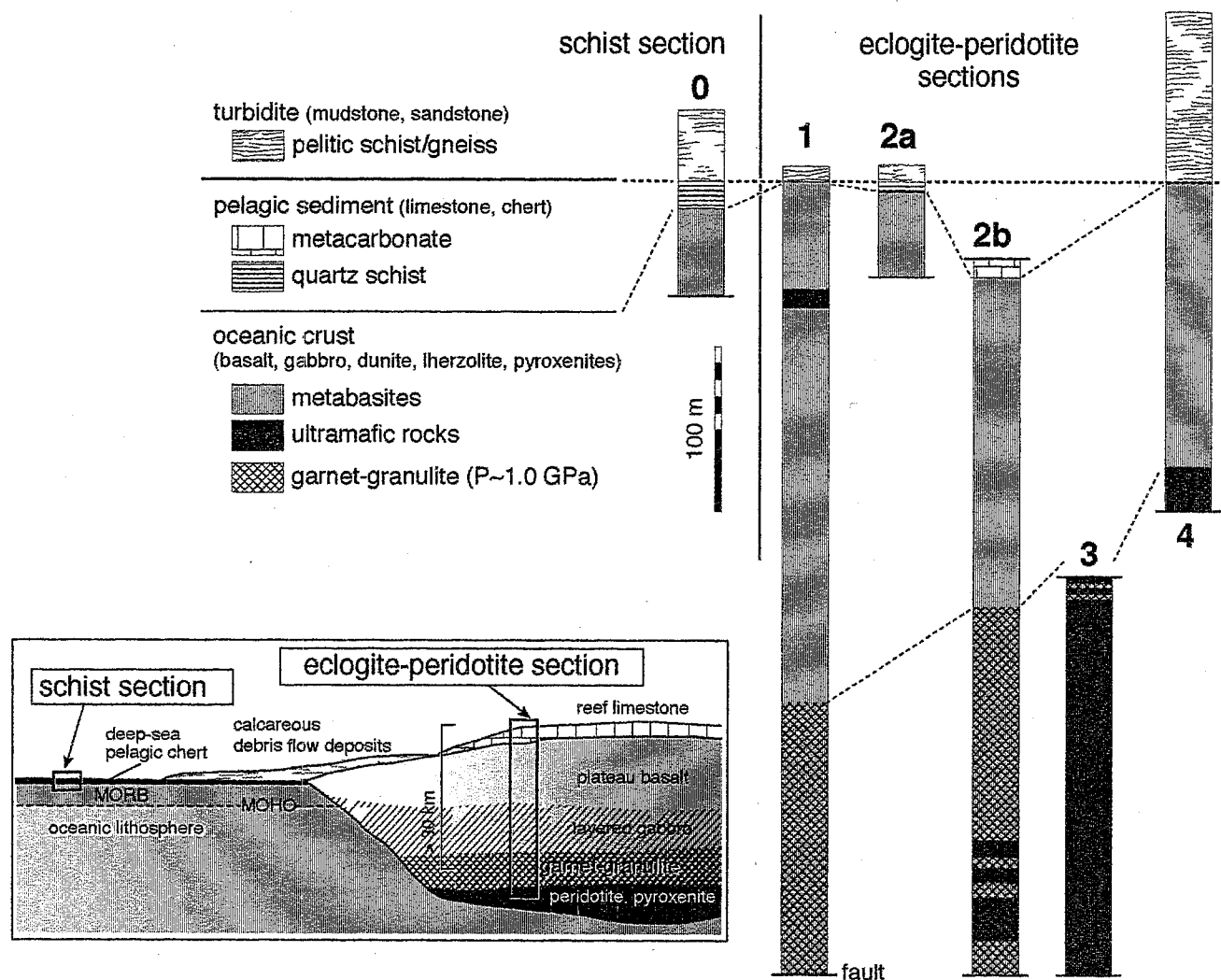
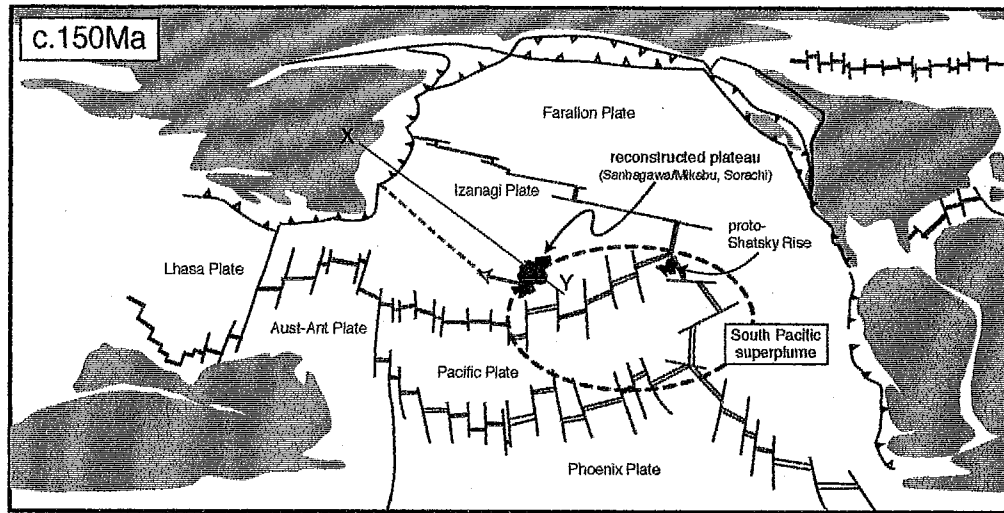
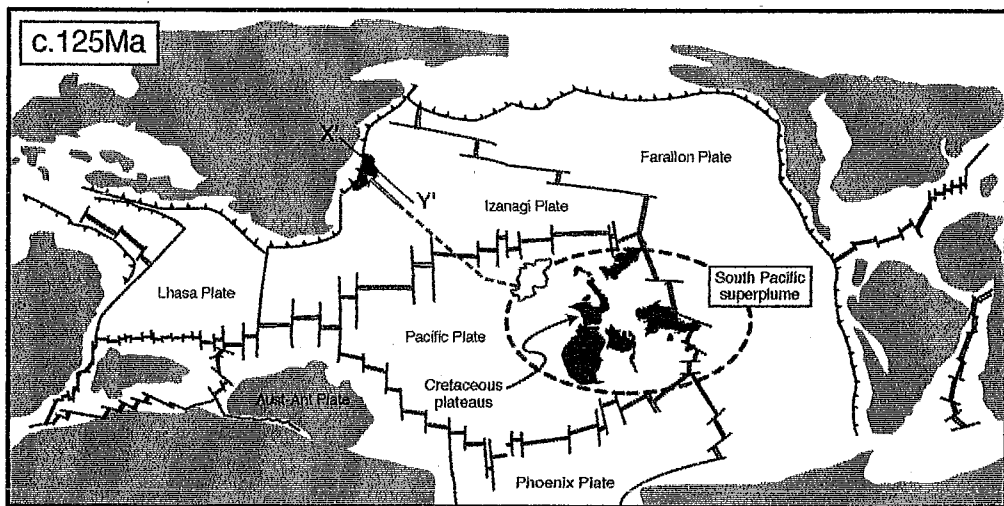
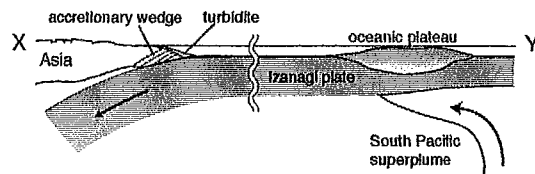


Fig. 3-8. Reconstructed oceanic plate stratigraphy for the Sanbagawa protoliths, deduced from duplex structure around the Iratsu- and the Higashi-Akaishi eclogite-peridotite mass in the Sanbagawa belt. Garnet-granulites, formed prior to subduction-related metamorphism, are overlain by pelitic gneiss of hemipelagic origin and by metacarbonate of pelagic sediment, suggesting that their original thickness over 30 km, although the present thickness is less than a hundredth of the original one due to tectonic thinning during exhumation process.



Formation of oceanic plateau related to the South Pacific superplume activity



Accretion and underplating of oceanic plateau, subsequent regional high-P/T metamorphism

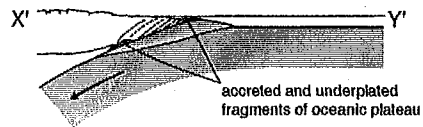


Fig. 3-9. Paleogeographic reconstruction for the Sanbagawa-Mikabu-Sorachi plateau at the late Jurassic to the early Cretaceous. Schematic cross-sections along X-Y (c. 150 Ma) and X'-Y' (c. 125 Ma) profiles are also shown.

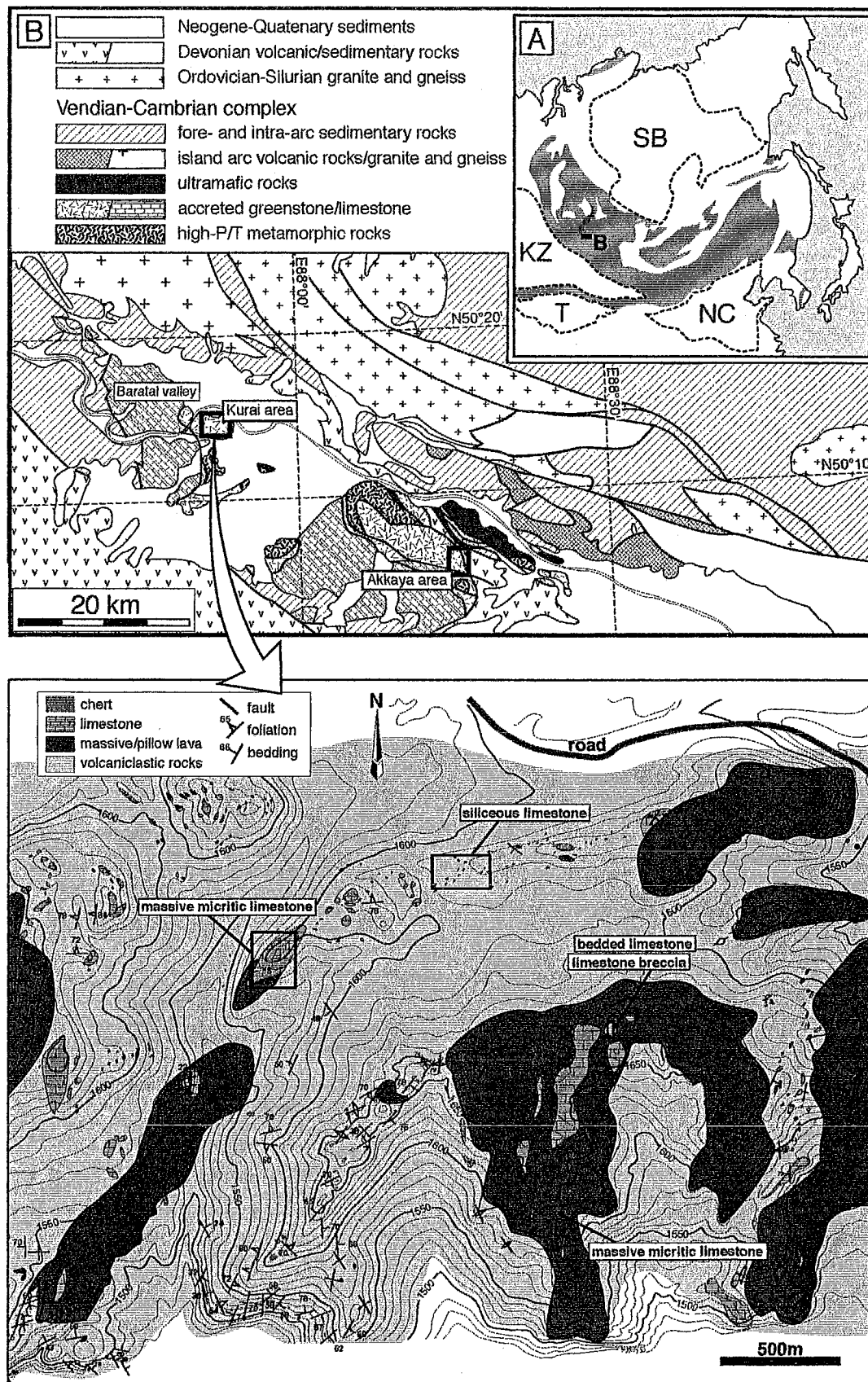


Fig. 3-10. Simplified unit map of the southeastern Gorny Altai, southern Russia (above; modified after Buslov et al., 1993), and lithological map of the Kurai area in the southeastern Gorny Altai (below).

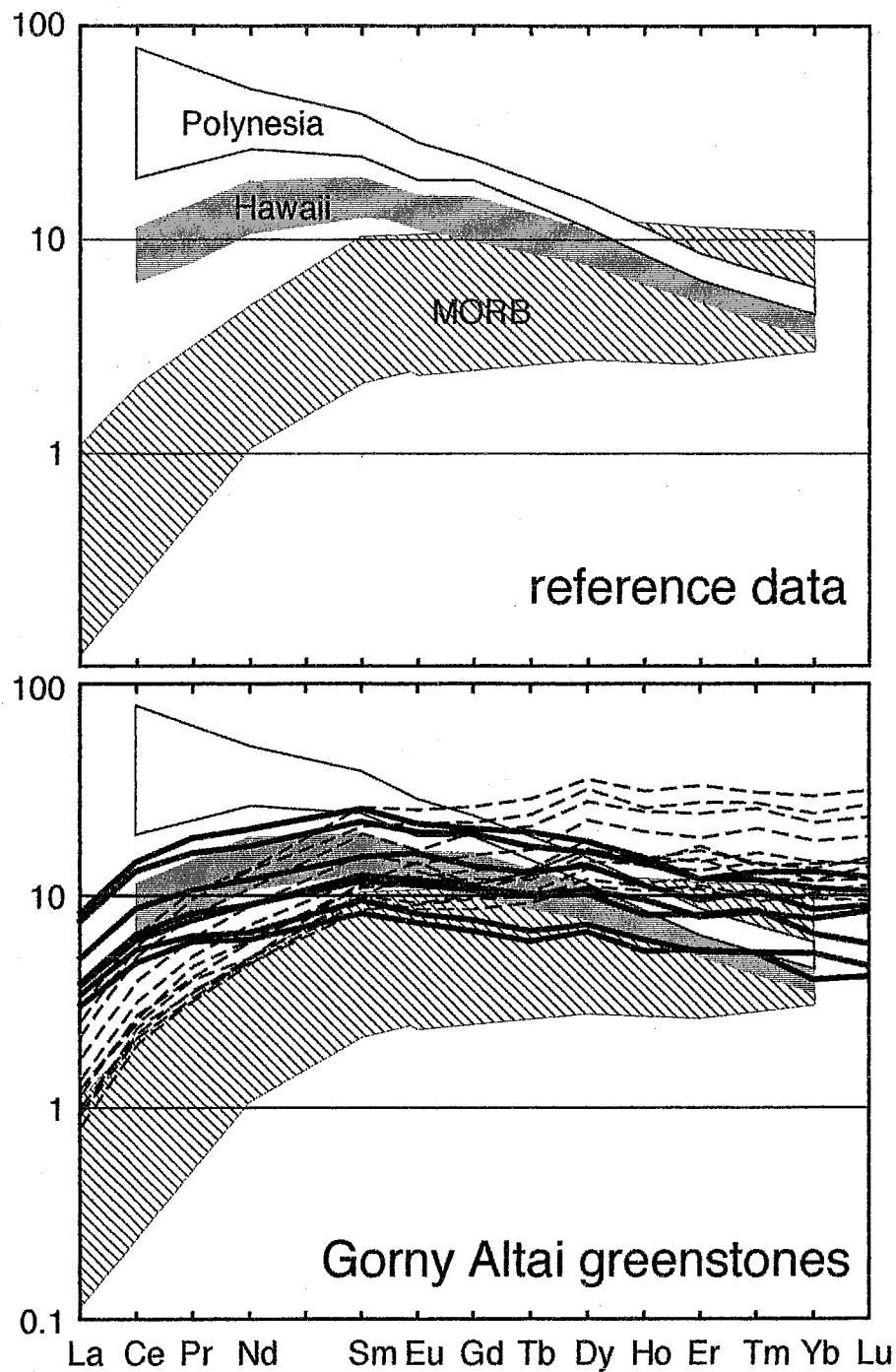
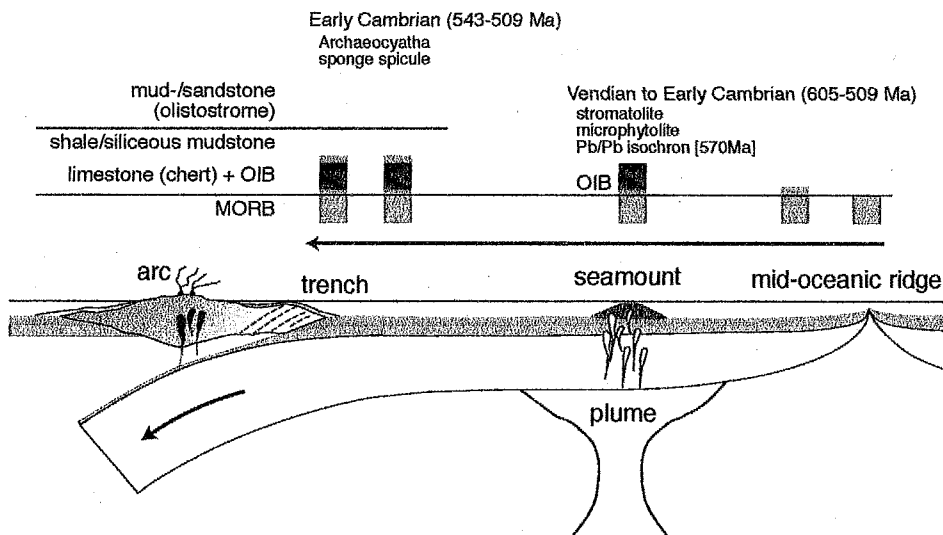
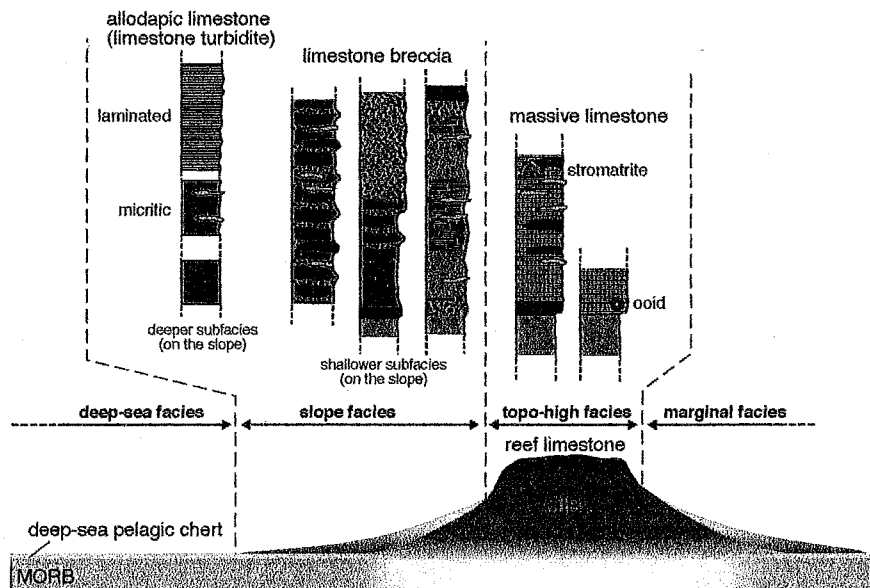


Fig. 3-11. Chondrite-normalized patterns of rare earth elements (REEs) in clinopyroxene phenocrysts from the oceanic island basalts (OIB), the mid-oceanic ridge basalt (MORB), and the Gorny Altai greenstones. Data sources: chondrite (ref), Polynesia (HIMU OIB: ref), and Hawaii (non-HIMU OIB: ref). The REEs concentrations in clinopyroxene phenocrysts from the Gorny Altai greenstones were analyzed by a secondary ion mass spectrometer (Cameca-3f), housed in Tokyo Institute of Technology. Note that two kinds of origins are recognized for the Gorny Altai greenstones; the Hawaii-type OIB (solid line) and the MORB (dashed line).

travel history of Gorny Altai protoliths



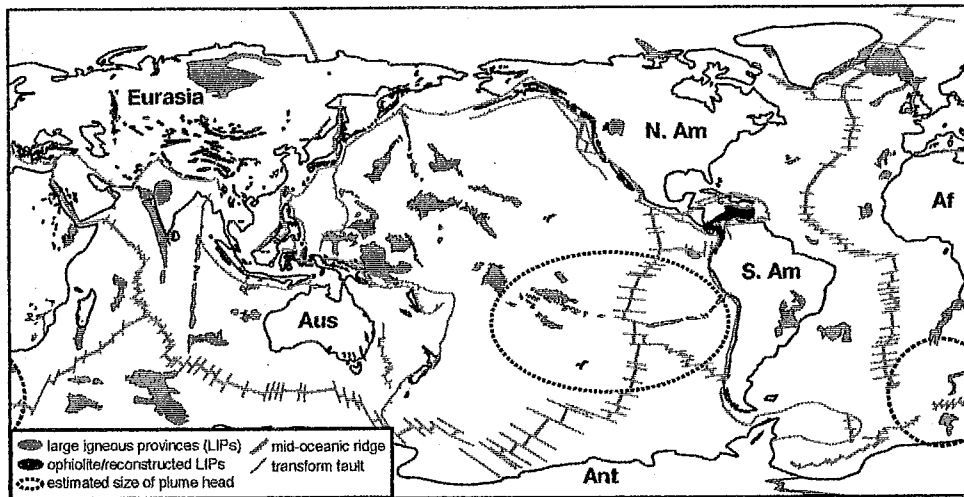
Columnar sections of Gorny Altai limestones and reconstruction model for their primary sites on a paleo-seamount



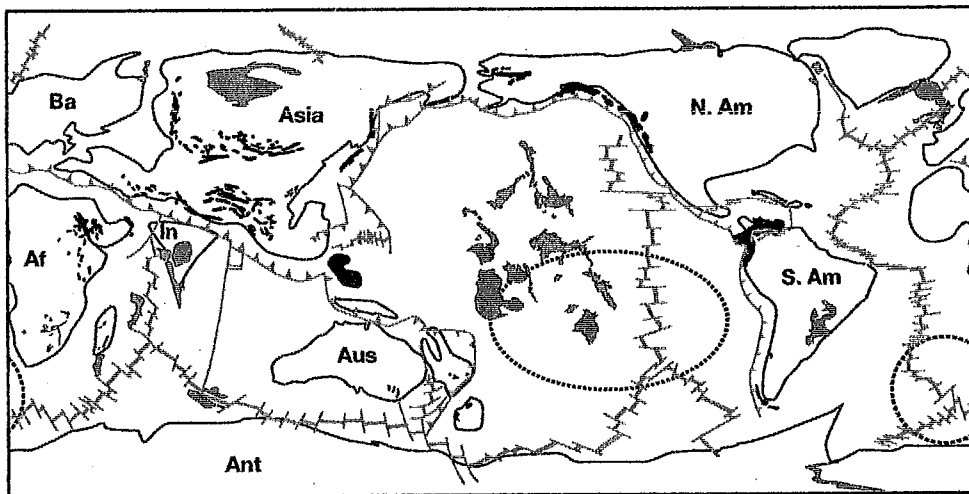
Uchio et al. (2004)

Fig. 3-12. Cartoons showing the Gorny Altai protolith stratigraphy, and its formation and emplacement processes, deduced from the Vendian-Cambrian accretionary complex, northern Altai Mountains, southern Russia. (a) Travel history of the Gorny Altai protoliths. The age of the seamount limestone was presumed by the occurrence of stromatolites and fossils of microphytolites, which show Vendian (605-543 Ma), and by the bulk Pb-Pb isochron age (572 ± 110 Ma) for the limestone. The late Vendian to Late Cambrian time scale was referred from Bowring and Erwin (1998). On the other hand, the accretion age of the seamount was estimated based on occurrences of the Archaeocyatha and the siliceous sponge spicules (Early Cambrian: 543-509 Ma). (b) Columnar sections of Gorny Altai limestones and a reconstruction model for their primary sites on a paleo-seamount. See text for explanations.

(a) Present day



(b) ca. 35 Ma



(c) ca. 65 Ma

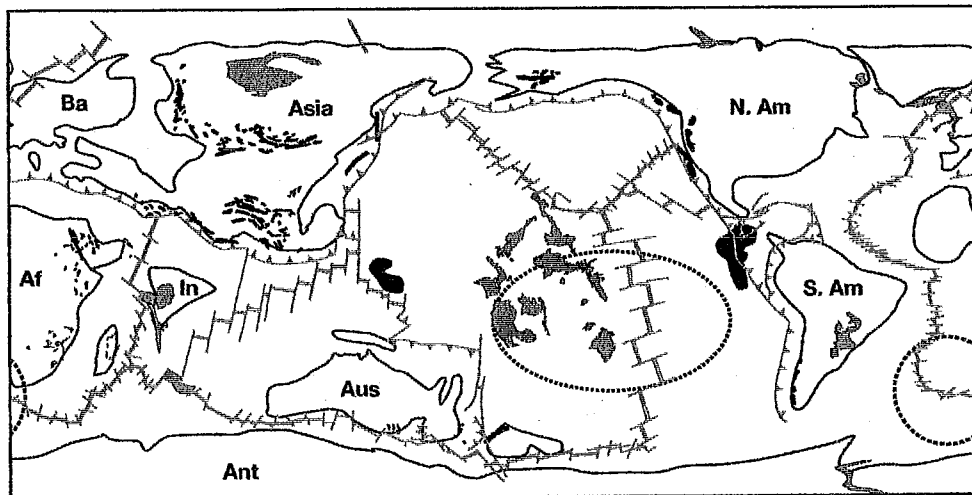
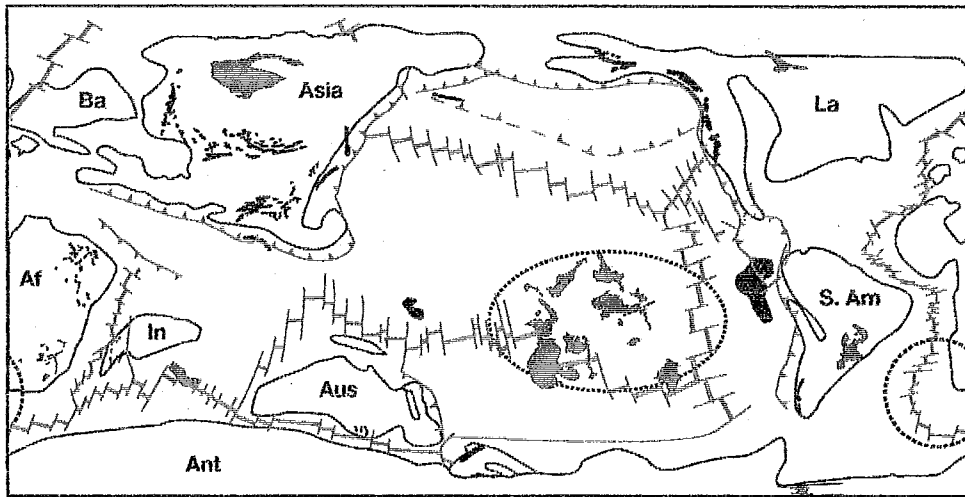
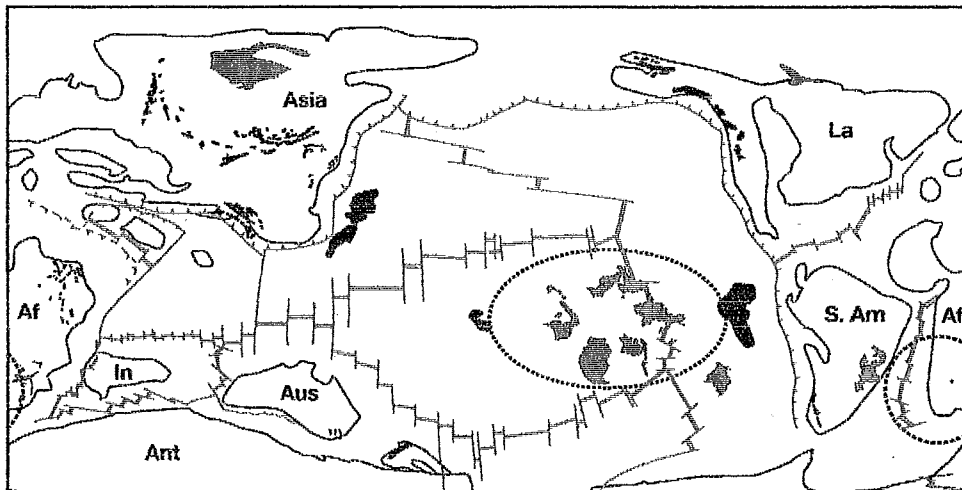


Fig. 3-13. Reconstructed paleogeographic maps at 0-165 Ma. (a)-(f). See Table 3-3 for abbreviations.

(d) ca. 85 Ma



(e) ca. 120 Ma



(f) ca. 165 Ma

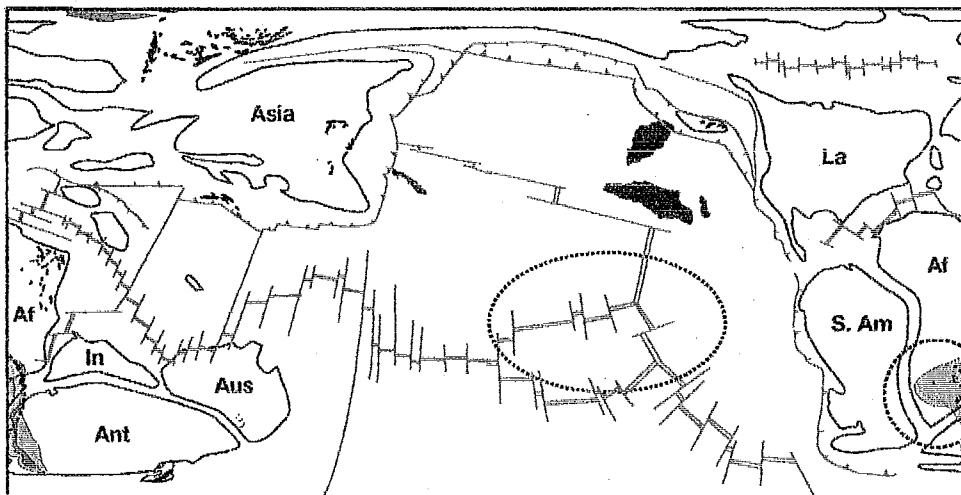
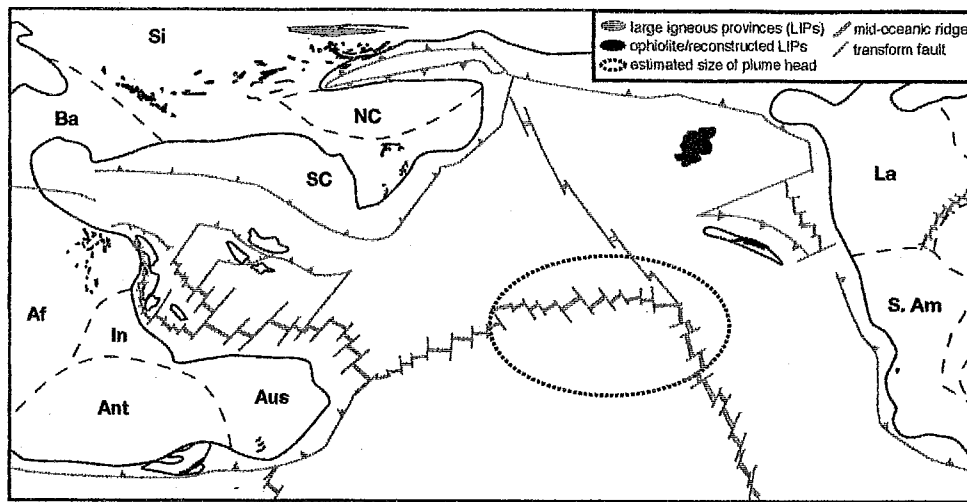
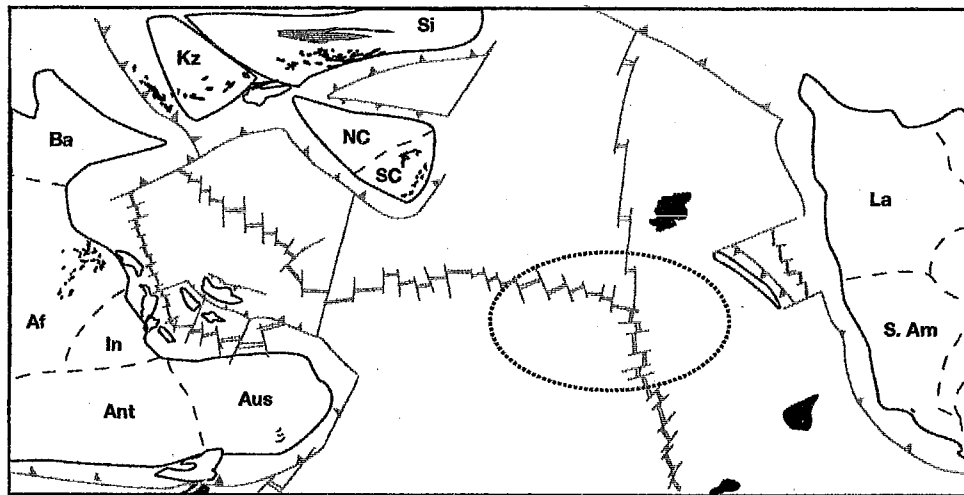


Fig. 3-13. (continued)

(a) ca. 225 Ma



(b) ca. 245 Ma



(c) ca. 280 Ma

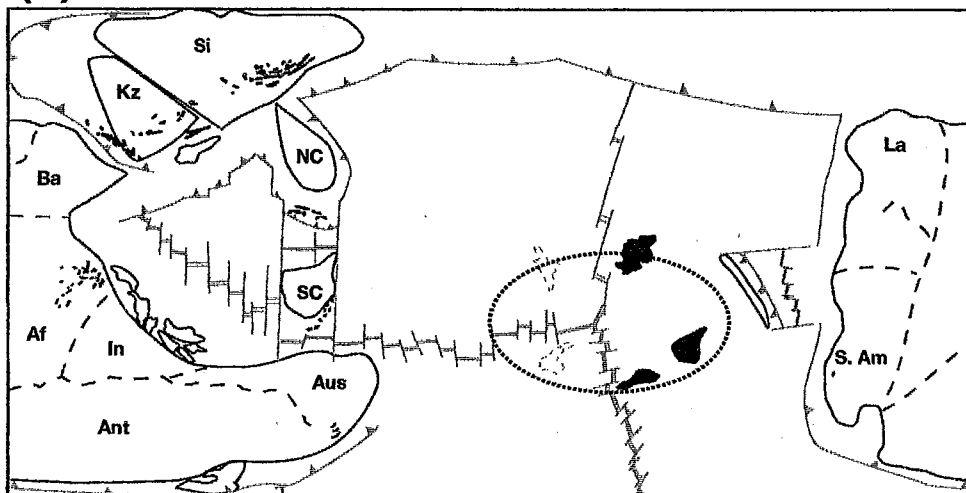
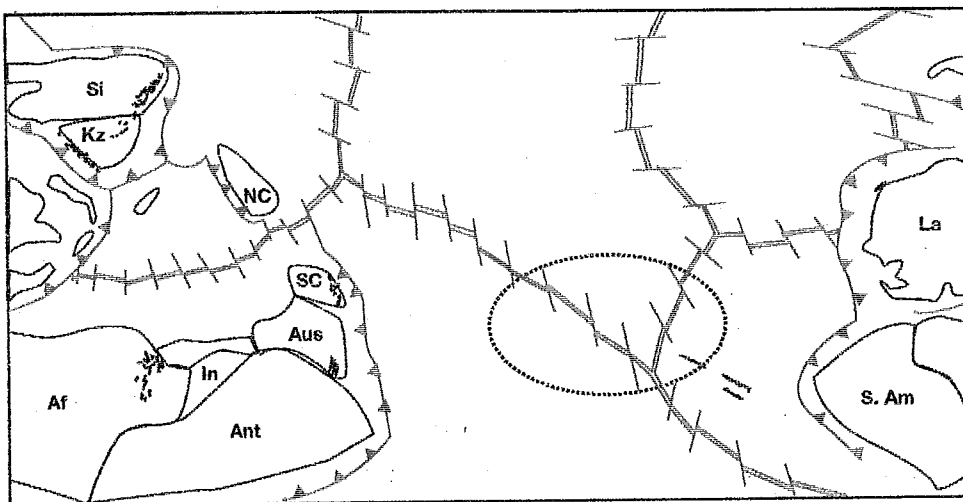
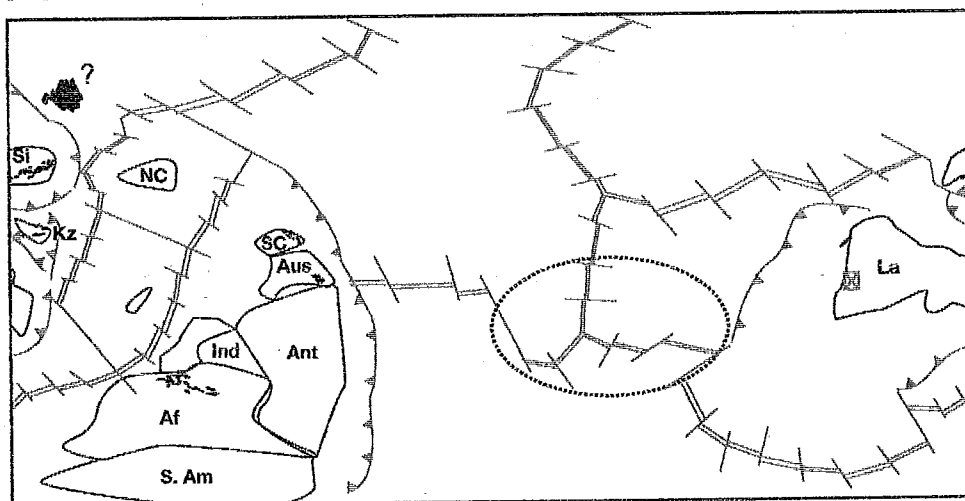


Fig. 3-14. Reconstructed paleogeographic maps at 225-550 Ma. (a)-(f). See Table 3-3 for abbreviations.

(d) ca. 390 Ma



(e) ca. 450 Ma



(f) ca. 500 Ma

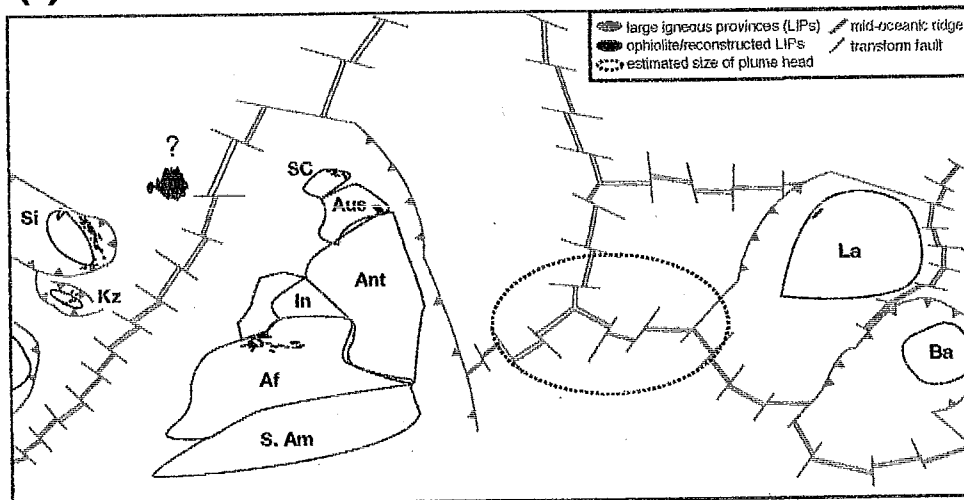
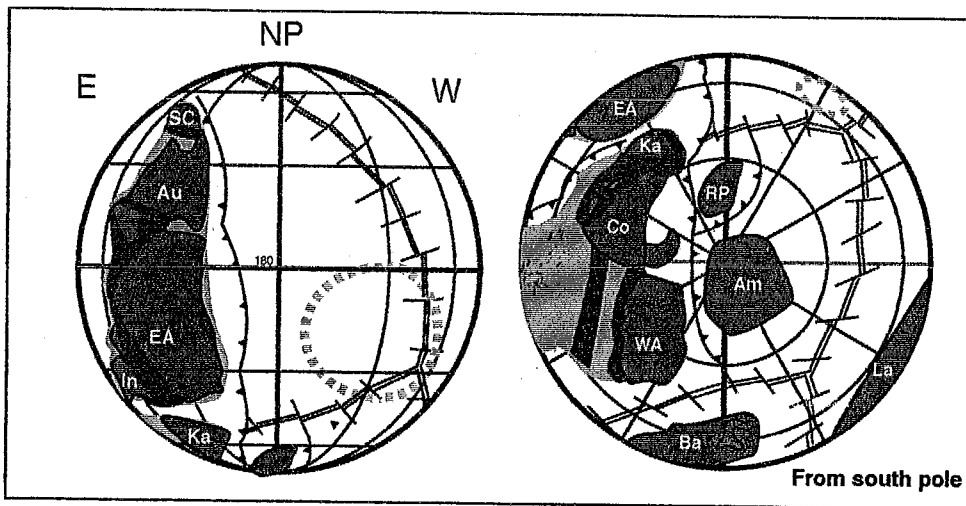
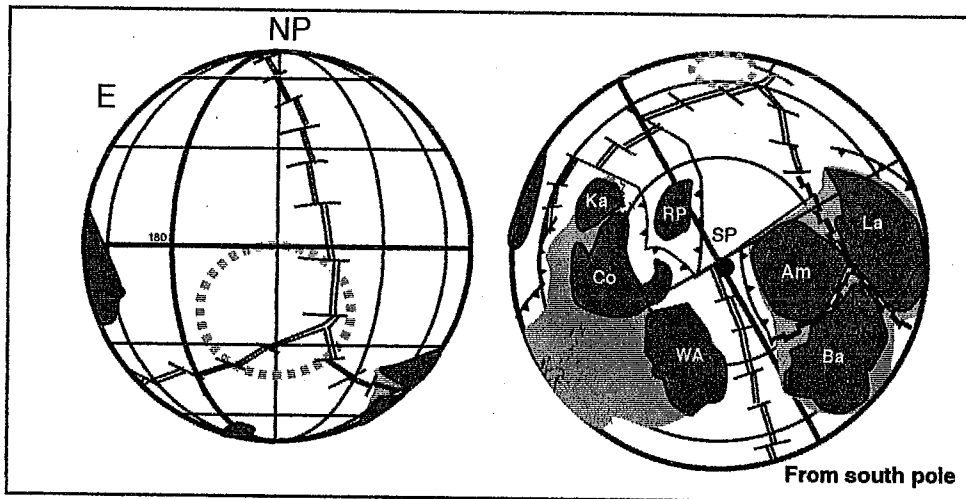


Fig. 3-14. (continued)

(a) ca. 550 Ma



(b) ca. 600 Ma



(c) ca. 650 Ma

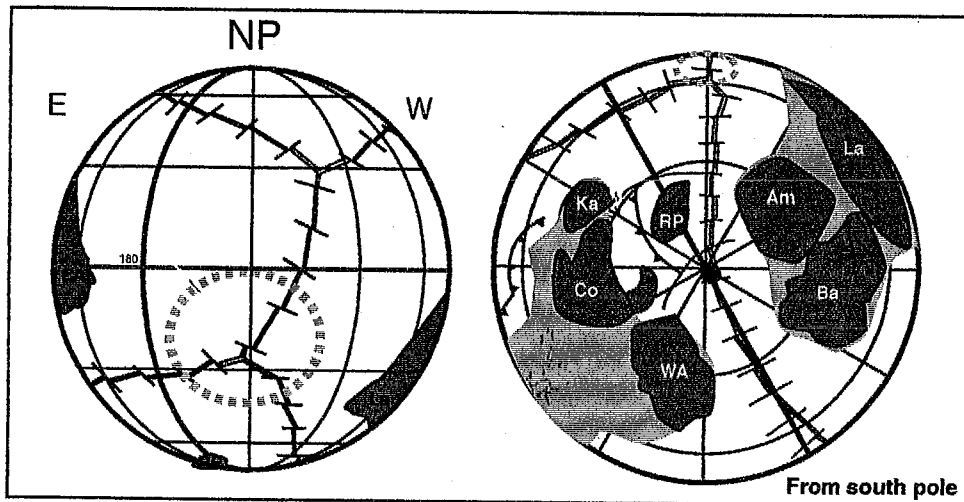
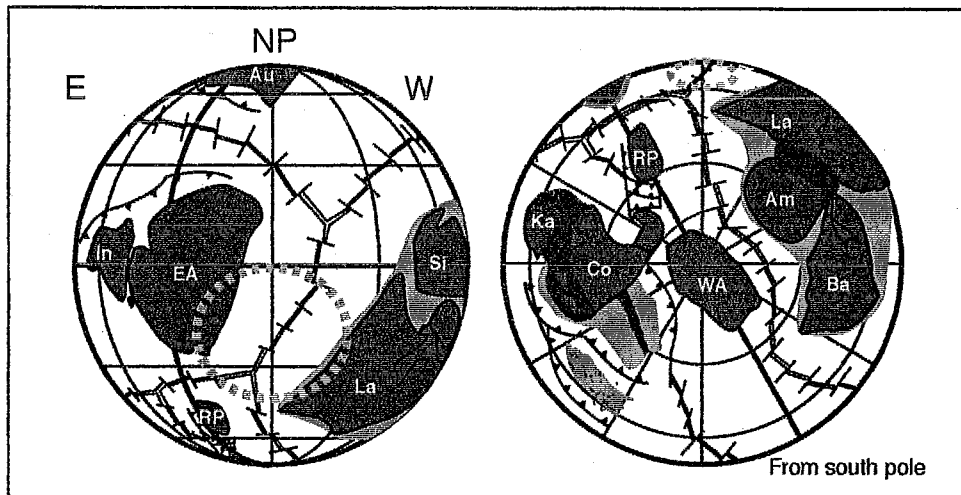
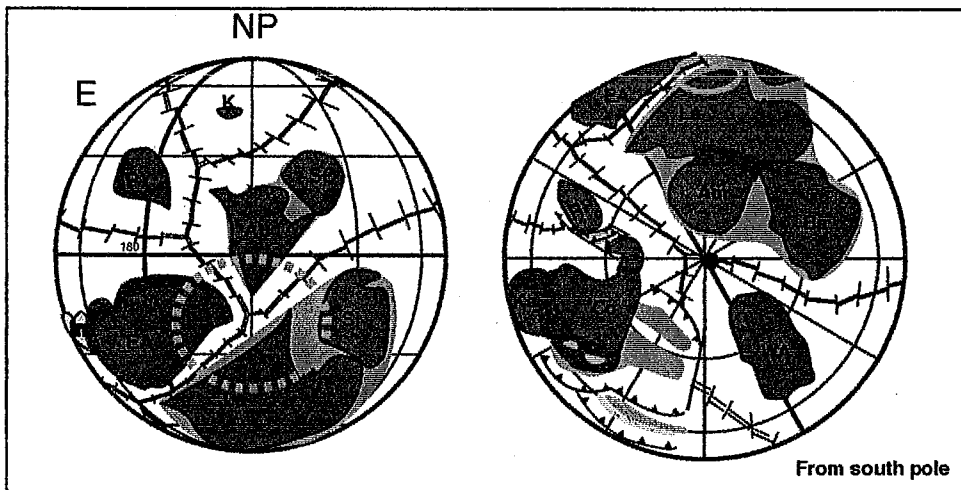


Fig. 3-15. Reconstructed paleogeographic maps at 550-1000 Ma. (a)-(i). See Table 3-3 for abbreviations.

(d) ca. 700 Ma



(e) ca. 750 Ma



(f) ca. 800 Ma

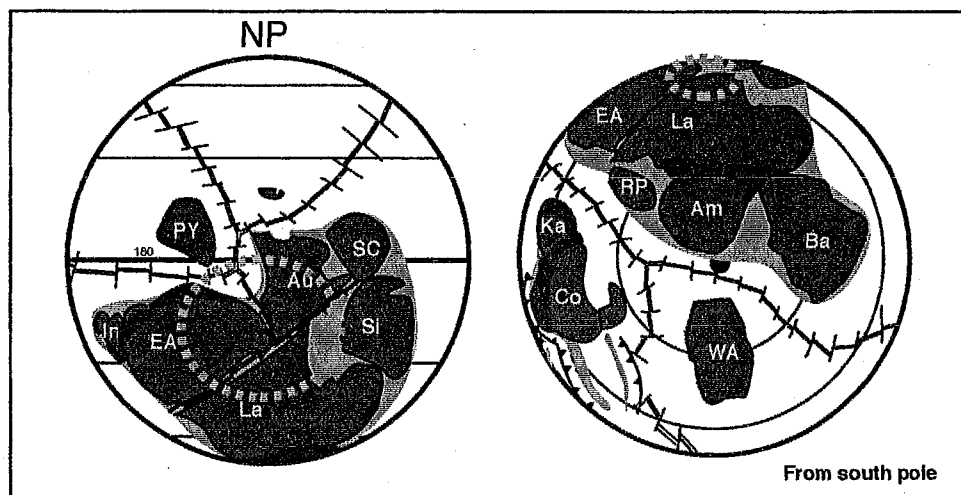
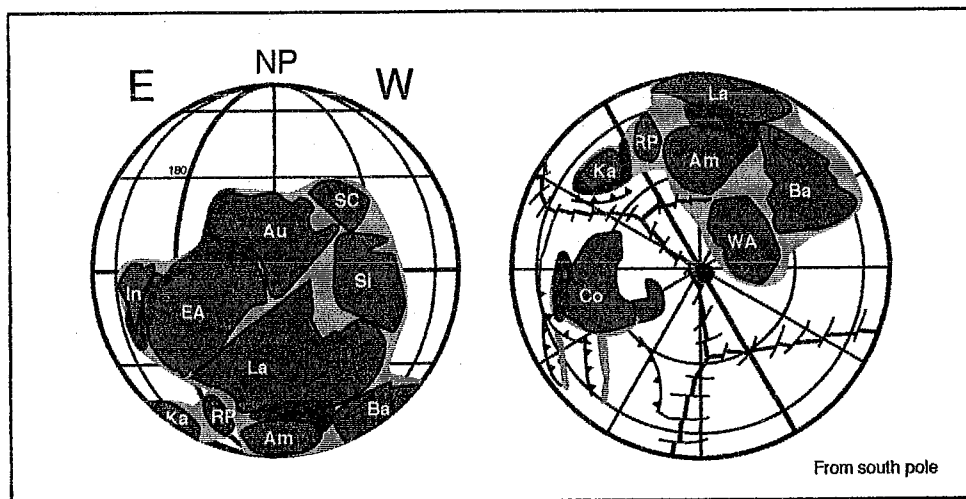
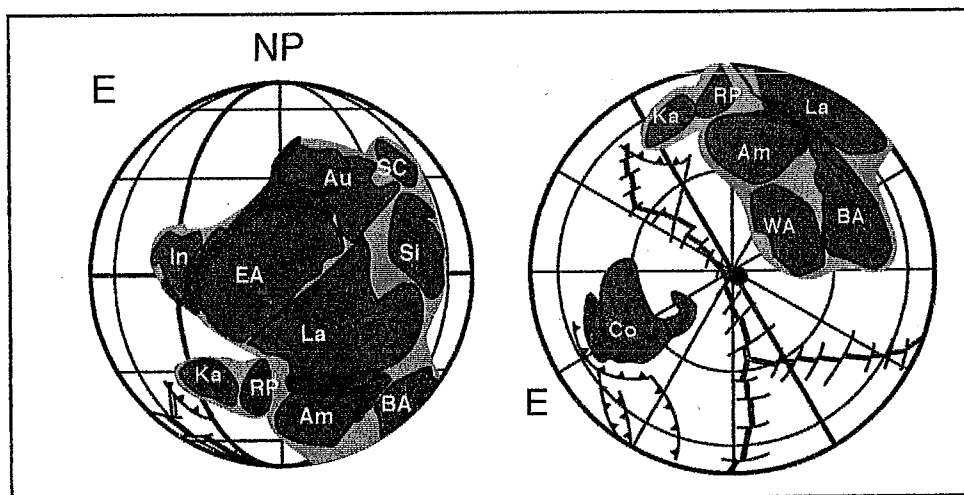


Fig. 3-15. (continued)

(g) ca. 850 Ma



(h) ca. 900 Ma



(i) ca. 1000 Ma

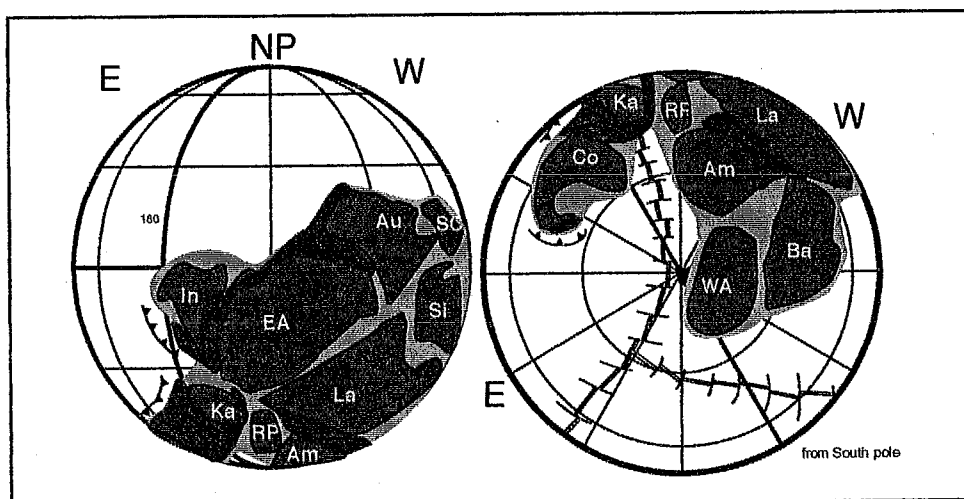


Fig. 3-15. (continued)

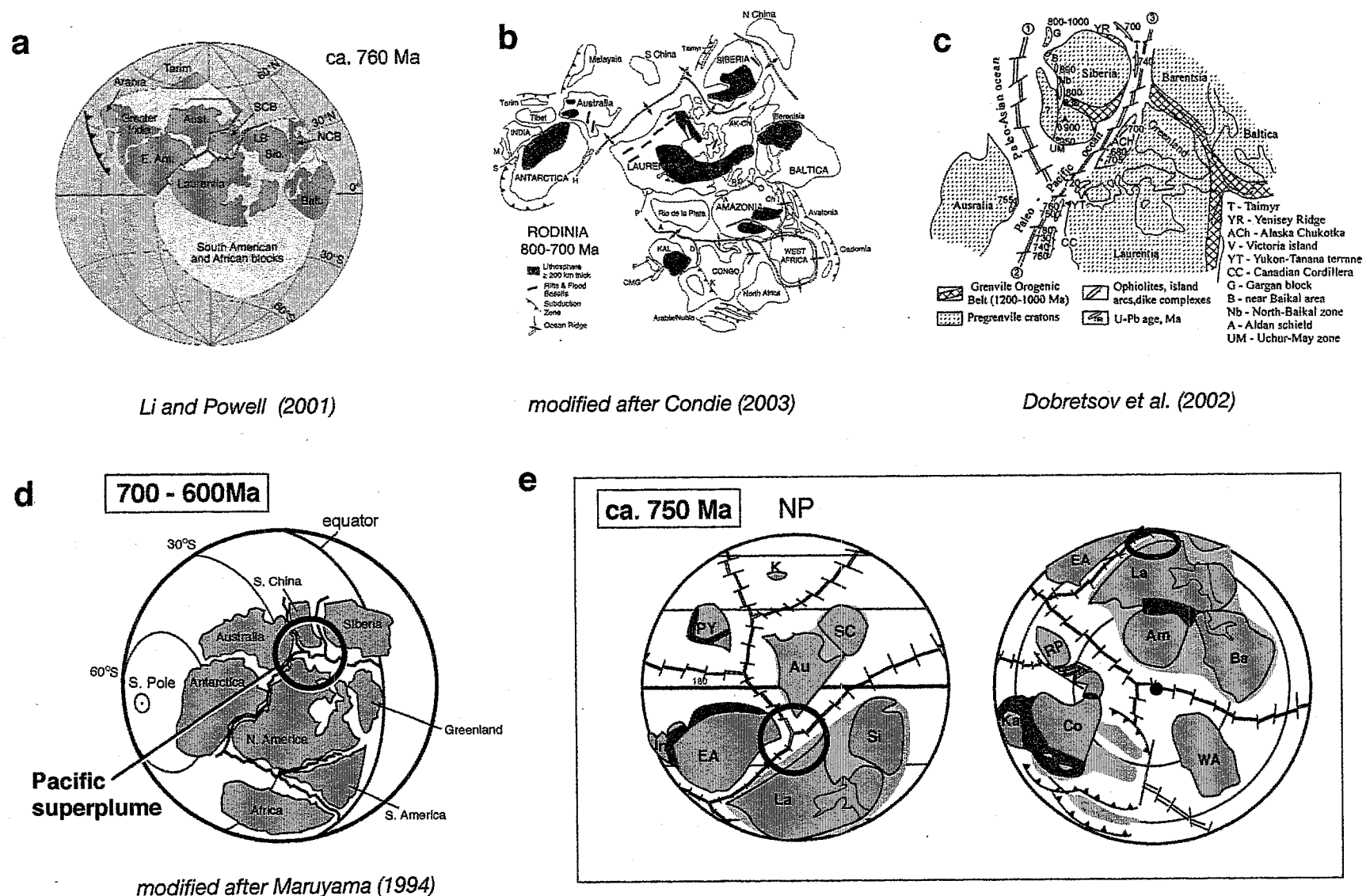
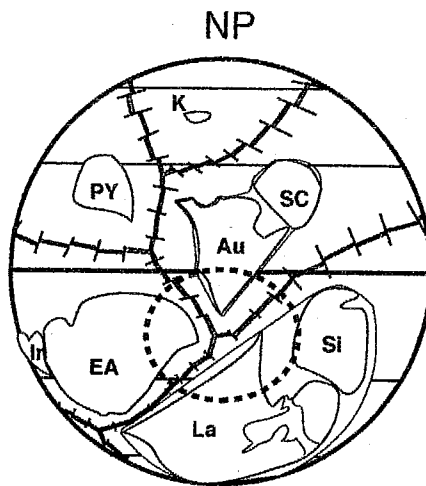
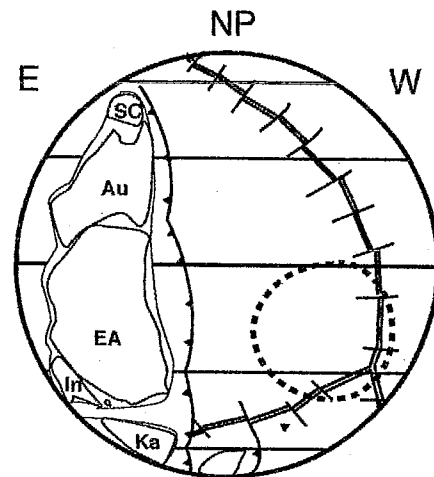


Fig. 3-16. Paleogeographic reconstruction models of the supercontinent Rodinia. a) Li and Powell (2001), b) Condie (2003), c) Dobretsov et al. (2002), d) Maruyama (1994), e) this study.

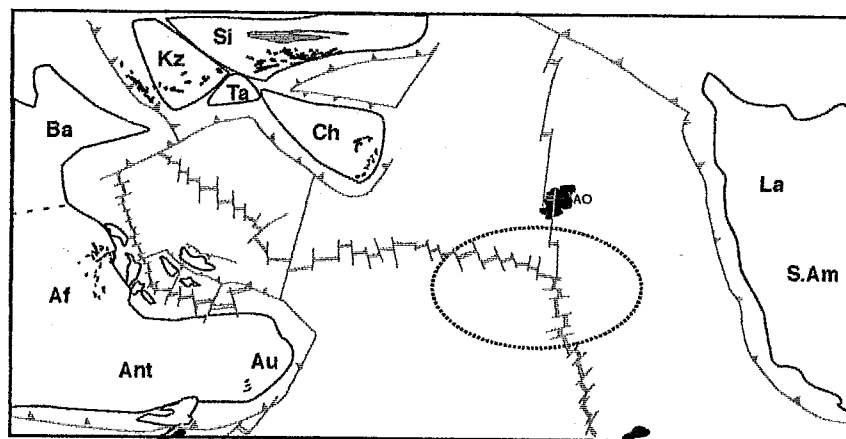
(a) ca. 750 Ma



(b) ca. 550 Ma



(c) ca. 245 Ma



(d) ca. 165 Ma

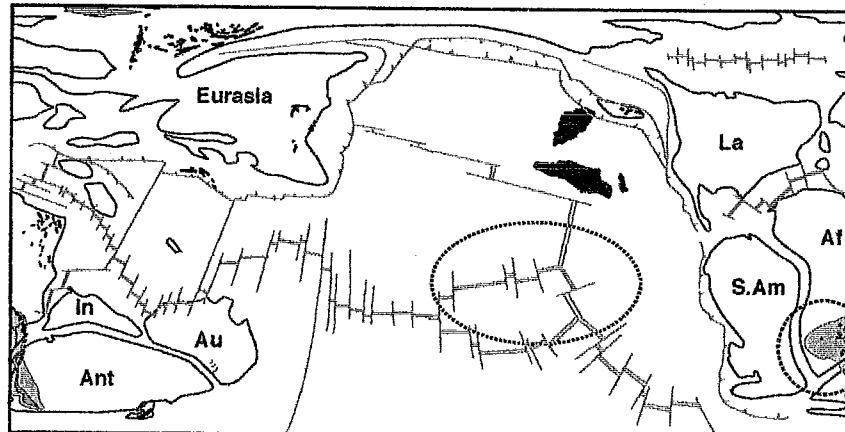


Fig. 3-17. Cartoons showing history of the (paleo-) Pacific Ocean. Symbols and acronyms are the same as Fig. 3-13. (a) Opening with the birth of Pacific superplume at c. 750 Ma. (b) Beginning of subduction around margins of eastern Gondwana and western Laurentia (not shown in this figure) at c. 550 Ma. (c) Formation of super-ocean coincided with the supercontinent Pangea. (d) Beginning of closing with the Atlantic opening, and continued to present day. See Table 3-3 for abbreviations.

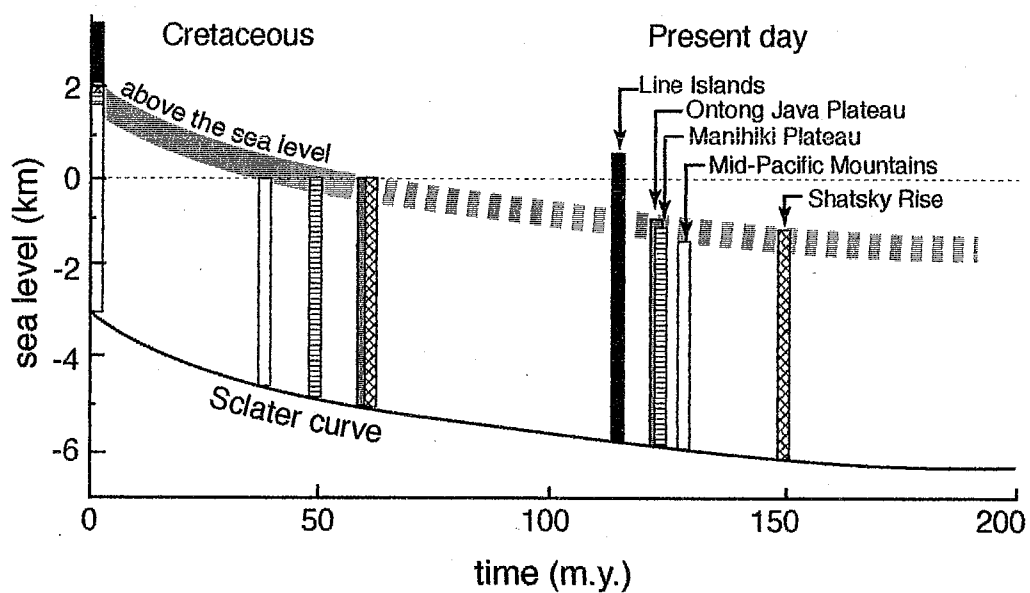


Fig. 3-18. Diagram showing height of the Cretaceous LIPs in the Pacific, calculated using the Sclater curve (Parsons and Sclater, 1977; Sclater et al., 1980). See text for explanations.

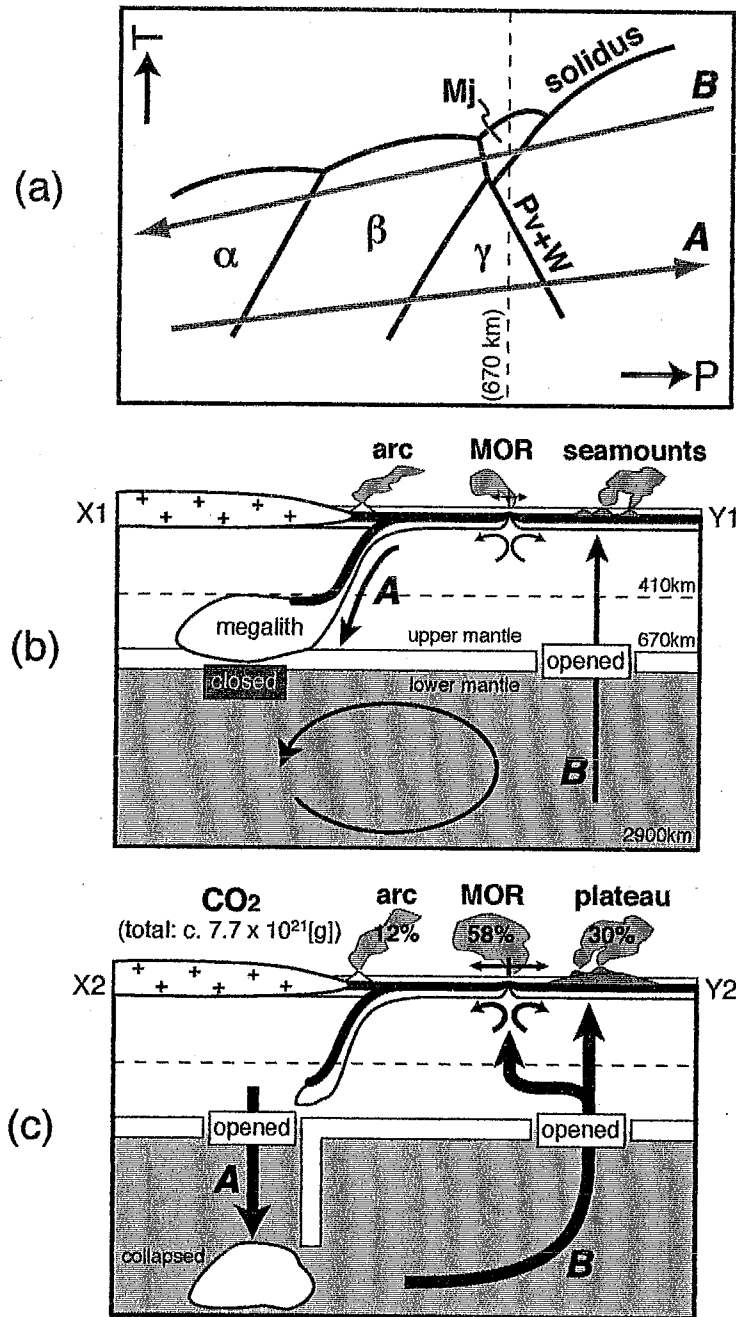


Fig. 3-19. Cartoons showing the Cretaceous plate tectonics and superplume in the Pacific. (a) Schematic phase diagram of mantle peridotite in P-T space (modified after Zhang and Herzberg, 1994; Hirose, 2002) with subduction zone geotherm A and plume geotherm B. The plume geotherm passes the majorite/perovskite transition curve with a positive slope, suggesting that a mantle upwelling can be accelerated at c. 670-km depth to transport fertile lower mantle materials to the upper mantle (See also Kumazawa and Maruyama, 1994). (b) Schematic cross-section showing mantle convection and its related volcanism at a normal stage. (c) Schematic cross-section showing a whole mantle convection and the resultant activated volcanism at a pulse stage. The collapsed megalith accelerated a single-layered convection at a whole mantle scale.

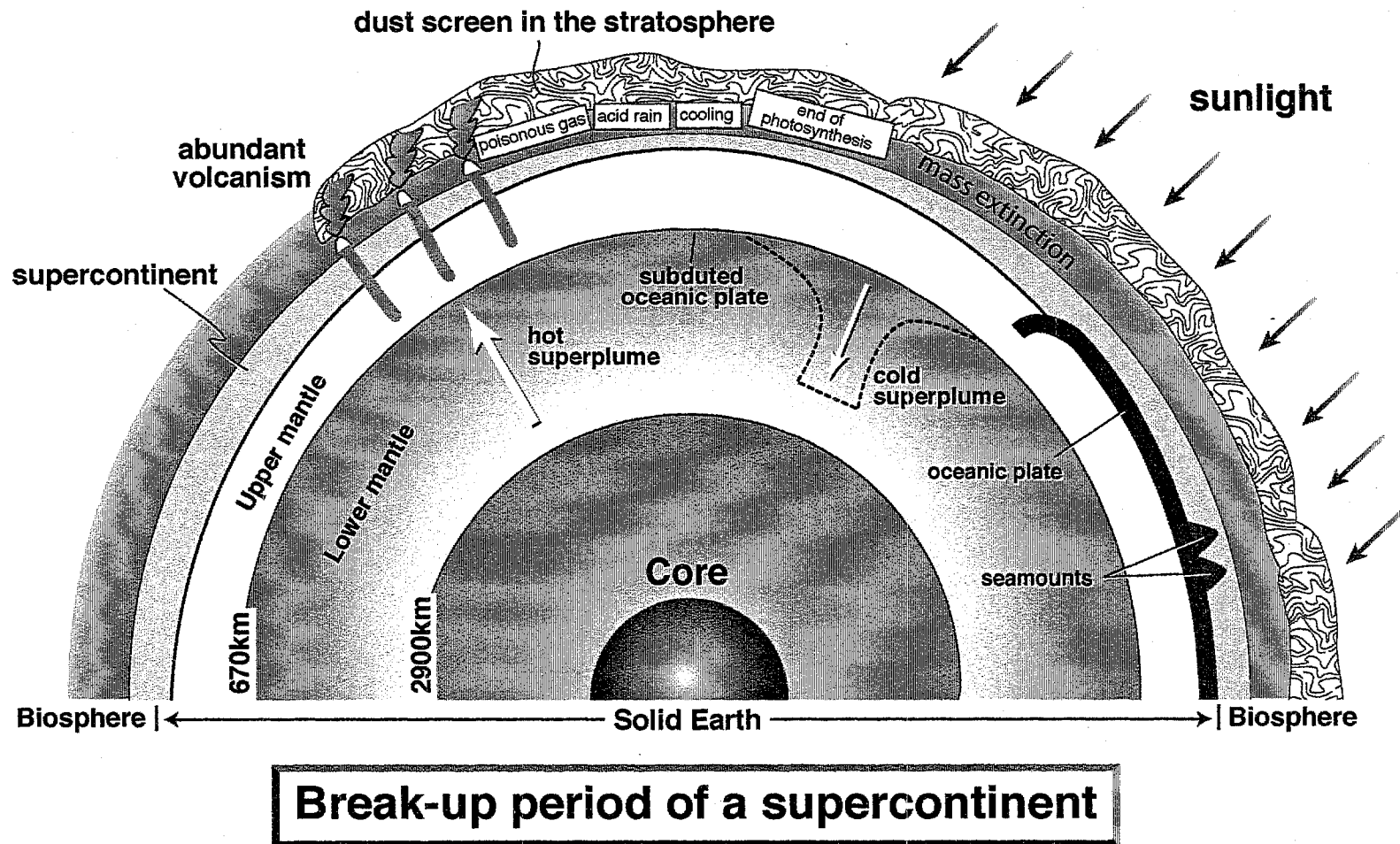


Fig. 3-20. Schematic cartoon showing interactions between the solid Earth and the surface environment at a break-up period of supercontinent. See text for explanations.

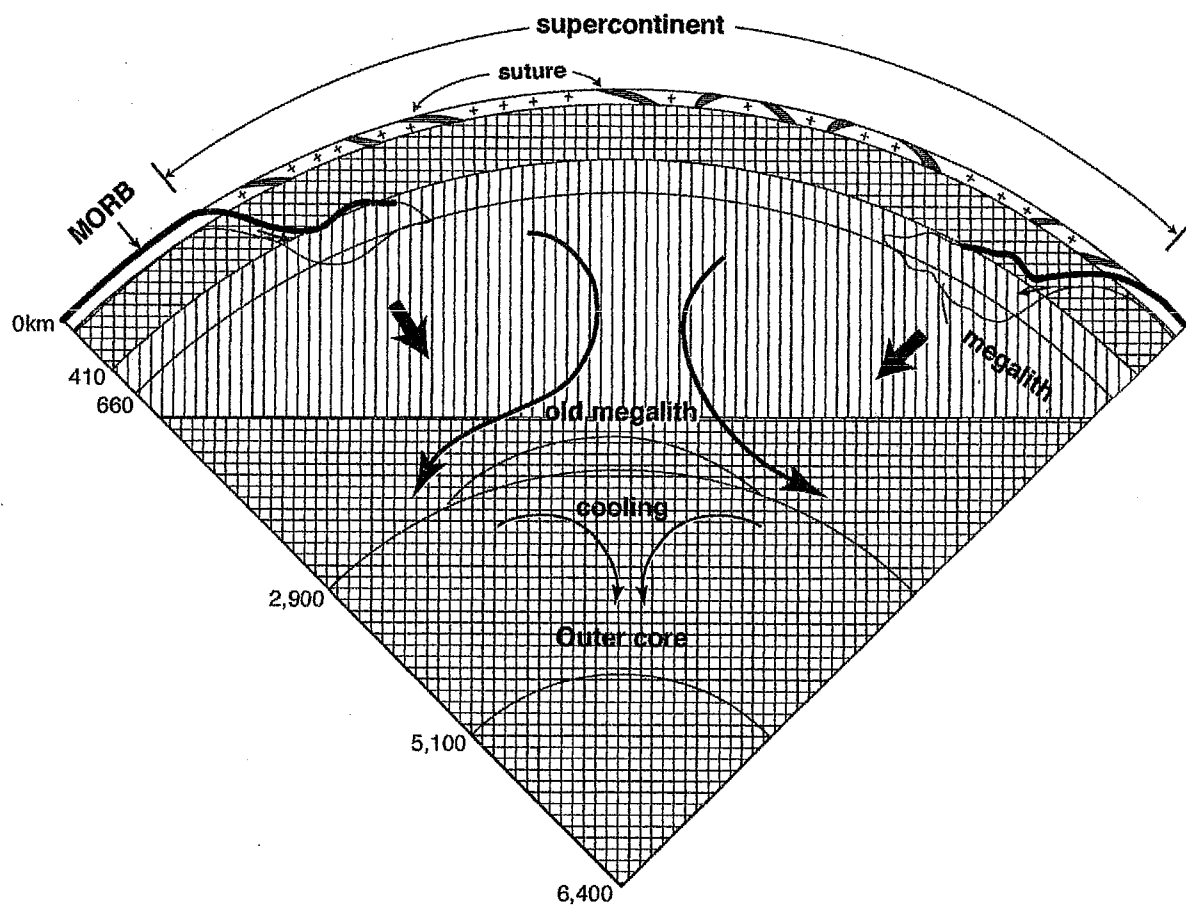


Fig. 3-21. Cartoons showing formation process of a supercontinent. Numerous micro-plates and subduction zones contributed to form a large amount of juvenile crust and collage of immature island arcs resulted in a composite huge continent. See text for explanations.

Appendix

Pb isotopic composition of olivine-hosted melt inclusions from Koolau and Kilauea Volcanoes, Hawaii

Abstract

Pb isotopic compositions of olivine-hosted melt inclusions from Hawaiian Islands have been determined by secondary ion mass spectrometry. We selected samples from Koolau and Kilauea volcanoes representing the end-members of the Hawaiian Pb isotopic mixing array. 78 and 33 melt inclusions from Koolau and Kilauea were analyzed, respectively. Melt inclusions from Koolau have higher $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{208}\text{Pb}/^{206}\text{Pb}$ ratios than those from Kilauea. These Pb isotopic compositions are equivalent to those of whole rocks for each volcano. These features suggest that Pb isotopic compositions of Hawaiian volcanoes have been mainly determined by undifferentiated primary magma generated in the source region for each volcano. The Pb isotopes of the Hawaiian melt inclusions show much limited variation range than those found in melt inclusions from Mangaia, French Polynesia. Present results suggest that the isotopic variation in the Hawaiian magma source region is within the limit of analytical resolution by SIMS. These Melt inclusions from Kilauea and Koolau record simpler magma process than those from Mangaia that contributed by the complicated metasomatized material.

Introduction

Hawaiian islands are located in the middle of the Pacific Ocean, currently active and one of the largest basaltic volcanism setting on the earth. Naturally, Hawaii attracted many researchers to study mantle plume and origin of basaltic magma. Numerous Sr, Nd, Pb, and Os isotopic data on Hawaiian lavas have been reported and the nature of their sources has been discussed (e.g. Tatsumoto, 1978; Stille et al., 1983; Staudigel et al., 1984; Hegner et al., 1986; Stille et al., 1986; Kennedy et al., 1991; Kurz and Kammer, 1991; Roden et al., 1994; Yang et al., 1994; Bennett et al., 1996). The Hawaiian isotopic compositions extend from compositions similar to Pacific mid-oceanic-ridge basalts (MORB) toward the end-member component called "Enriched Mantle 1 (EM1)" (e.g. Zindler and Hart, 1986; Hart et al., 1992). These end-members appear in Kea trend volcanoes and in Koolau volcano, respectively. Based on the major elements and isotopic compositions, Hauri (1996) estimated that mafic components are recycled ancient oceanic crusts entrained in the Hawaiian plume. The recycled oceanic crusts may represent the enriched component in the Hawaiian plume.

Most previous discussions on Hawaiian magma source are based on isotopic chemistry of bulk rocks. Because lavas may suffer magma mixing and homogenization processes in magma chamber, it is not clear that to which extent the bulk rock isotopic compositions represent that of mantle derived primary magmas. Recently, technical development of in-situ Pb isotope analysis by secondary ion mass spectrometry (SIMS) and its application to the melt inclusions opened a new insight of isotopic feature of the mantle source. Saal et al. (1998) have demonstrated that melt inclusions from only two rock samples in Mangaia, Cook-Austral islands, have Pb isotopic variability larger than

the entire range of this island by bulk rock analyses. Furthermore, the isotopic range they found from the two rocks is equivalent to the entire Pb isotope range reported from Cook-Austral islands chain and this leads to reconsider the origin of mantle heterogeneity (Saal et al., 1998).

In this study, we determined Pb isotopic compositions of olivine-hosted melt inclusions from two Hawaiian volcanoes, Kilauea and Koolau. In contrast to the Mangaia samples, the Pb isotopes in Hawaiian melt inclusions are relatively homogeneous but show intershield variations similar to those found from the bulk rocks. Based on the new measurements, we discuss the nature of source heterogeneity in the Hawaiian plume.

Samples

We used 7 lava samples from Koolau volcano and 2 lava samples from Kilauea volcano. KOL- series samples and MH-5a were collected at Makapuu Head of Koolau volcano. WAF-4 and WAF-5 (212f) are drilling cores in Wheeler Air Force Base of Koolau volcano, Oahu Island. Lithologies of KOL-3, KOL-9, KOL-16, and MH-5a are olivine basalts and other samples are picrites. The rock samples from Koolau volcano were taken from the portion representing the last stage (Makapuu stage, Tanaka et al, 2002) in the growth history of this volcano (See Figure 8 of Shinozaki et al, 2002). KIL-5 was collected from surface of the Kilauea Iki lava lake formed in 1959 and KIL-9 from Mauna Ulu lava along Chain of Craters Road. Lithology of KIL-5 and KIL-9 is picrite and olivine basalt, respectively.

Selected samples of 30 to 200 cm³ in volume were roughly crushed in a stainless steel mill. Olivine crystals were picked up from these crushed samples by hand. The

picked olivine crystals were examined under optical microscope to seek for melt inclusions. Melt inclusions connected with crack or vein in host olivine crystals were excluded from analyses. Olivine crystals hosting the melt inclusions were mounted in acrylic resin and ground until the melt inclusions exposed onto the olivine surface. The mounted olivine crystals were polished with plastic films with 3 μm and 1 μm corundum grains, successively. Prior to electron-probe microanalysis (EPMA), polished samples were coated with carbon film so as to eliminate electrostatic charge-up on the sample surface. For the same purpose, before analysis by SIMS, samples were coated with gold film of 40 nm in thickness after removing the carbon film.

Analytical Procedures

Host olivine and melt inclusions were observed by scanning electron microscope (SEM) and major compositions of them were determined by EPMA (JEOL JXA-8800) at the Tokyo Institute of Technology (TiTech). Host olivines were analyzed at 25 kV accelerating voltage and 24 nA beam current. Counting time was 30 s for Ni, Mn, and Ca and 20 s for Si, Mg, and Fe. Melt inclusions that consist of glass and daughter minerals were analyzed at 15 kV accelerating voltage and 12 nA beam current. Counting time was 10 - 20 s. Beam diameter was focused into a spot of 1 to 3 μm in diameter. Data were reduced by oxide ZAF correction.

Pb isotope analysis was performed by TiTech Cameca ims 1270. $^{16}\text{O}^-$ with the energy of 23 keV was used for primary ion. The current was 30 to 60 nA and was focused into a spot of 20 to 50 μm in diameter. Mass resolution power was adjusted to 3600 to 4000 so as to exclude the interference of molecular ions (Fig. 1). Each analysis

consisted of 50 cycles of cyclic measurements of ^{204}Pb , ^{206}Pb , ^{207}Pb , and ^{208}Pb in peak jumping mode. Counting time for each cycle was 10 s for ^{204}Pb , ^{206}Pb , and ^{207}Pb and 5 s for ^{208}Pb . Because counts of ^{204}Pb from melt inclusions were less than detection limit, we discussed $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{208}\text{Pb}/^{206}\text{Pb}$ ratios in this study.

Using the standard samples (JB-2, JB-3, JA-2, JR-1, and JGb-1 from Geological Survey of Japan, GSJ Rock References), the analytical precision, repeatability and instrumental mass fractionation were carried out. These standard samples except the JR-1 were prepared as the synthetic quenched glasses from the rock powder prepared by GSJ and the JR-1 was as natural glass. These have the wide variability of major elements composition (from basaltic to rhyolitic compositions) and Pb concentrations (1.92 to 19.30 ppm). Koide and Nakamura (1990) determined Pb isotopic ratios of the JB-2, JB-3, JA-2, and JR-1 by thermal ionization mass spectrometry (TIMS). We used their isotopic ratios to evaluate the instrumental mass fractionation and matrix effects generated by variations of major elements compositions for the SIMS analysis. Appropriate recommended values of Pb isotopic compositions have not been provided for JGb-1. However, because Hawaiian tholeiites usually contain ca. 1 ppm of Pb, we used JGb-1 glass (Pb = 1.92 ppm) to evaluate repeatability of analyses for low Pb concentration.

The JB-2 and JGb-1 were repeatedly analyzed 20 and 11 times, respectively (Fig. 2). Each analysis was successively carried out on the same day. The mean values among analyses are approximately repeatable within standard error of $2\sigma_m$. Degree of typical $2\sigma_m$ relative to the average values of $^{207}\text{Pb}/^{206}\text{Pb}$ and of $^{208}\text{Pb}/^{206}\text{Pb}$ is about 0.7 % for JB-2 and about 1.2 % for JGb-1, respectively.

Same experiments were carried out for other standard glasses. These average

value normalized by the reference value are shown in Fig. 3. All the average values are shifted to smaller side from the recommended values. It is well known that the lighter isotopes are released easier from in surface during spattering and counted more than the heavier ones. The shifts of Pb isotope ratios are consistent to the mass fractionation. Degrees of mass fractionations for $^{207}\text{Pb}/^{206}\text{Pb}$ ratio are 2.6, 0.5, 0.8 and 0.3 ‰ for JB-2, JB-3, JR-1 and JA-2, respectively. Those for $^{208}\text{Pb}/^{206}\text{Pb}$ ratio are 3.7, 2.7, 5.7, and 5.7 ‰, respectively. The mass fractionations are along the mass dependent fractionation line (Fig. 3). However, such small instrumental mass fractionation among samples (i.e. matrix effect) can be negligible, because the matrix effect is smaller than the estimated error for each analysis (Fig. 4). Thus, we used JA-2 glass as a standard sample for every independent analysis. There are two advantages to use of JA-2 as a standard: first is accurate Pb isotopic compositions determined by Koide and Nakamura (1990). The second is the high Pb concentration (19.20 ppm). In order to evaluate stability of SIMS condition during a continuous session, the JA-2 glass was analyzed at the start and at the end of the session. All Pb isotope ratios of unknown samples were normalized to the JA-2 value, in order to correct possible inconsistencies among independent SIMS session.

Petrology of the melt inclusions

Olivine compositions around melt inclusions from Koolau volcano are Fo_{83-89} (Fo: $100 \text{ Mg} / (\text{Mg} + \text{Fe})$) (Table 1). Most olivine grains have no significant chemical zoning, but some have normal zoning from Fo_{89} to Fo_{81} . Host olivine crystals from Kilauea volcano are also Fo_{83-89} (Table 2) but they have no significant chemical zoning. Host

olivine crystals from Koolau have higher contents for NiO and lower contents for CaO than those from Kilauea (Fig. 4). These features are consistent with those reported by previous studies (e.g. Norman and Garcia, 1999).

The melt inclusions analyzed in this study are amount to 78 and 33 globules for Koolau and Kilauea volcanoes, respectively. The analyzed melt inclusions are oval shape with 25 to 150 μm in diameter. Mineral assemblages of melt inclusions from Koolau volcano consist of glass (Fig. 5(a)), glass + clinopyroxene (cpx) (Fig. 5(b)), glass + cpx + orthopyroxene (opx) (Fig. 5(c)), glass + chromian spinel (Cr-spinel), glass + cpx + Cr-spinel (Fig. 5(d)), and glass + cpx + opx + Cr-spinel. Some of the melt inclusions include small sulfide globules (Fig. 5(e)). Pyroxene crystals are euhedral, subhedral or dendritic. Curved faces of subhedral crystals are contacted to edge of the inclusion. These features show that pyroxene crystals were crystallized after entrapment of the melt inclusions. The dendritic texture reveals the crystallization under supercooled conditions (Fig. 5 (b) and (d) pyroxenes). Cr-spinel crystals are euhedral. Some spinel crystals are partly embedded into the wall of melt inclusions (Fig. 5 (d)). Volumes of most Cr-spinel are too large to have crystallized in the melt inclusions indicating that they are trapped crystals when host olivine grew.

The melt inclusions from Kilauea volcano consist of glass, cpx, and Cr-spinel. Opx was not observed in studied samples. Mineral assemblages are glass (Fig. 5(f)), glass + cpx (Fig. 5(g)), glass + Cr-spinel (Fig. 5(h)), and glass + cpx + Cr-spinel. Some inclusions include small sulfide globules. Most cpx crystals have dendritic texture. The features of the Cr-spinel crystals are similar to those from Koolau volcano.

The melt inclusions without daughter pyroxenes are considered to preserve more primitive chemical compositions of the trapped magma than those with daughter

pyroxenes (Tables 3 and 4). However, even the crystal-free inclusions suffered significant chemical modifications by the overgrowth of host olivine. MgO variations of major elements for the glasses in pyroxene-free melt inclusions from Koolau and Kilauea volcanoes are shown in Fig. 6. Maximum MgO content is 6.63 wt% and 6.99 wt% for Koolau and Kilauea, respectively. Their large variation in the MgO diagrams can be understood by olivine overgrowth. Glasses from Koolau have higher contents for SiO₂ and lower contents for TiO₂ and CaO at any MgO contents than those of Kilauea. These features in melt inclusion chemistry are consistent with those in major elements of whole-rocks (Frey et al. 1994) and the result of other melt inclusions study (Norman et al., 2002).

Pb isotope chemistry of the melt inclusions

²⁰⁷Pb/²⁰⁶Pb and ²⁰⁸Pb/²⁰⁶Pb ratios of the melt inclusions from Koolau and Kilauea are shown in Tables 5 and 6, respectively. These are plotted in Fig. 7. Because ²⁰⁴Pb is not available in our SIMS analysis, we adopted ²⁰⁸Pb/²⁰⁶Pb-²⁰⁷Pb/²⁰⁶Pb diagram, although Pb isotopic compositions are usually plotted on ²⁰⁷Pb/²⁰⁴Pb-²⁰⁶Pb/²⁰⁴Pb and ²⁰⁸Pb/²⁰⁴Pb-²⁰⁶Pb/²⁰⁴Pb ones. We have demonstrated that olivine-hosted melt inclusions from Koolau and Kilauea volcanoes are systematically different in Pb isotopic values. Considering the typical error of individual analysis, the Pb isotopic compositions for each volcano are consistent with those of whole-rock and the variation is roughly at the same level as that of the whole-rocks (Tatsumoto, 1978; Stille et al., 1983; Newsom et al., 1986; Roden et al., 1994; Bennett et al., 1996; Chen et al., 1996; Garcia et al., 1996; Norman and Garcia, 1999; Pietruszka and Garcia, 1999; Garcia et al., 2000; and Tanaka et al., 2002). The variation between the host rock samples was not detected. There is a clear

systematic difference in Pb isotopes in melt inclusions from Koolau and Kilauea volcanoes. Accordingly, the olivine phenocrysts have the same values and variability of Pb isotopic composition to the magma. These agreements indicate that compositions of magma did not change from olivine crystallization stage on their way to or in the magma chamber to eruption stage for each volcano.

Discussion

We have demonstrated that olivine-hosted melt inclusions from Koolau and Kilauea volcanoes are systematically different in Pb isotopic values but their variations are within those of the whole rocks in each volcano. The fact that they show limited variation typical to each volcano (Fig. 7) supports either, 1) the magma homogenization took place in a reservoir prior to crystallization of the host olivines, or 2) isotope variations in the source rocks in Hawaii is much smaller than Mangaia and detection limits of our analysis.

Trace element compositions of melt inclusions from Kilauea and Koolau are greater diversity than those of whole-rocks (Norman et al., 2002). Because the diversity of Kilauea is explainable by the different degree of melting of compositionally homogeneous source, it requires no contribution from the distinct source (Norman et al., 2002). It is similar to the melting model that account for the diversity of trace element compositions of the melt inclusions from Iceland (Gurenko and Chaussidon, 1995; Slater et al., 2001) and FAMOUS area (Shimizu, 1998). In the case of Kilauea, although Pb isotopic composition is rather homogeneous, it is consistent with the observation of trace element. In contrast with the trace element compositions of melt inclusions of

Kilauea, Mauna Loa, Loihi, and Hualalai produced by systematic melting of compositionally homogeneous source, those from Koolau deflect from the other Hawaiian trend and reflect the contributions of unique source (Norman et al., 2002). But the diversity of Koolau is not explainable by the melting of single unique source and requires the mixing with other Hawaiian source. The other Hawaiian sources are compositionally homogeneous but isotopically heterogeneous (Norman et al., 2002). The homogeneous Pb isotopic feature of Koolau indicates that the contributing the other components are isotopically Koolau-like beyond the realm of possibility to distinguish from Koolau source by our analyses. Most of melt inclusions from Mauna Loa have the trace element compositional feature that the diversity is explainable by the systematic melting as MORB and other Hawaiian volcanoes except Koolau, however, a few is extremely enriched in Sr and severely depleted in highly incompatible elements that are not generated by the systematic melting (Sobolev et al., 2000). These features are inherited from gabbroic eclogite as recycled lower crust (Sobolev et al., 2000). Because Koolau basalts are seemed to represent the component of recycled crust in Hawaiian plume (e.g. Hauri, 1996), the Sr-excess melt inclusions from Mauna Loa have the possibility that these have Koolau like Pb isotopic feature. On the basis of these diversity of trace element compositions, the homogeneous Pb isotopic feature are more explainable by originally small isotope variations in the source regions than by the magma homogenization took place in a reservoir prior to crystallization of the host olivines.

In contrast with Kilauea and Koolau, Pb isotopic compositions of the melt inclusions from only two rock samples of Mangaia Island, (one of type localites of HIMU basalt), have so larger variations that overlap with those of whole-rocks from the

entire Cook-Austral Islands chain between Mangaia Island and Macdonald seamount (Saal et al., 1998). On the basis of examinations of metasomatized mantle xenoliths, Cook-Austral basalts are inferred to have experienced the complex process such as reaction of carbonatite with previously crystallized phases (Hauri et al., 1993). Some melt inclusions from Mangaia that contain coexisting silicate glasses, sulfide, and carbonate globules document direct evidence of such complex magmatic processes (Saal et al., 1998) or the influence of the sulfide-bearing fluid from core (Yurimoto et al., 2004). On the other hand, melt inclusions from Koolau and Kilauea have rather simple history. The mineral phases found in melt inclusions are consistent with simple crystal fractionation. Although sulfide- and carbonate-bearing melt inclusions from Mangaia show end-member HIMU characters (Saal et al., 1998; Yurimoto et al., 2004), melt inclusions with sulfide from Koolau and Kilauea do not reach the end-member compositions of Hawaiian Islands (Fig. 7). These differences of the process generating source mantle reflect the Pb isotopic features of the melt inclusions. The source of Hawaiian basalts is mixture of depleted mantle and enriched mantle or depleted mantle and recycled oceanic crusts.

Conclusions

Pb isotopic compositions of olivine-hosted melt inclusions from Koolau and Kilauea volcanoes are equivalent to those of the whole-rocks in each volcano, respectively. These results indicate that the isotopic variation in the magma source region is within the limit of analytical resolution by SIMS.

Melt inclusions from Kilauea and Koolau record simpler magma process than

those from Mangaia, French Polynesia that contributed by the complicated metasomatized material. These are reflected in the each Pb isotopic feature.

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Figure captions

Fig. 1. Mass spectrum around Pb isotopes from the JA-2 glass standard under mass resolution power of 3800.

Fig. 2. Reproducibility of Pb isotope analyses. (a) JB-2. (b) JGb-1. Solid and broken lines denote average and reference value, respectively. Each error bars show standard errors ($2\sigma_m$) corresponding to about 0.7 % and 1.2 % for JB-2 and JGb-1, respectively.

Fig. 3. Comparison between average of secondary ion ratios for Pb isotopes and the reference values of rock standards. Symbols are shown in inset. $(^{207}\text{Pb}/^{206}\text{Pb})_N$ denotes analytical $^{207}\text{Pb}/^{206}\text{Pb}$ ratio normalized to each reference value (Koide and Nakamura, 1990). $(^{208}\text{Pb}/^{206}\text{Pb})_N$ denote as same as $(^{207}\text{Pb}/^{206}\text{Pb})_N$. Error bars show standard errors ($2\sigma_m$). The solid line shows mass dependent fractionation line. N: number of independent analyses for averaging.

Fig. 4. Compositional variations of lavas from Koolau and Kilauea volcanoes versus forsterite content of the olivine phenocrysts. Symbols are shown in inset.

Fig. 5. Back-Scattered electron images of typical olivine-hosted melt inclusions from Koolau ((a)-(e)) and Kilauea ((f)-(h)) volcanoes. Scales are shown in inset.

Fig. 6. Compositional variations of major elements for the glasses in the melt inclusions from Koolau (solid circle) and Kilauea (open circle) volcanoes. Melt inclusions without daughter minerals are selected to plot.

Fig. 7. $^{208}\text{Pb}/^{206}\text{Pb} - ^{207}\text{Pb}/^{206}\text{Pb}$ diagrams (a) Pb isotopic compositions of whole-rocks from Hawaiian Islands and Cook-Austral Islands. The dark gray field indicates basalts from Hawaiian Islands. Bold solid and broken lines show field of Koolau

(Makapoo stage) and Kilauea volcanoes, respectively. The Hawaiian data are from Tatsumoto (1978), Stille et al. (1983), Staudigel et al. (1984), Hegner et al. (1986), Newsom et al. (1986), Kennedy et al. (1991), Kurz and Kammer (1991), Roden et al. (1994), Yang and Fey (1994), Bennett et al. (1996), Chen et al. (1996), Garcia et al. (1996), Norman and Garcia (1999), Pietruszka and Garcia (1999), Garcia et al. (2000), and Tanaka et al. (2002). The light gray field indicates whole-rock basalts from Cook-Austral Islands between Mangaia Island and McDonald seamount. The bold solid line shows field for Mangaia based on compiled data by Saal et al. (1998). Large open circles are the end-members of the mantle tetrahedron (Hart et al., 1992) projected on the $^{208}\text{Pb}/^{206}\text{Pb} - ^{207}\text{Pb}/^{206}\text{Pb}$ plane (Saal et al., 1998). (b) Pb isotopic compositions of melt inclusions. Each plot corresponds to individual melt inclusion from Koolau (solid diamond) and Kilauea (open diamond) volcanoes. Error bars denote typical standard errors ($2\sigma_m$). Error bars ($2\sigma_m$) of melt inclusions with sulfide globules (gray square for Koolau and open square for Kilauea, respectively) are shown independently. Large open circles and the fields are same as those in (a). The striped field indicates the melt inclusions from Mangaia (Saal et al., 1998).

Table 1 Major elements compositions of host olivine crystals from Koolau

sample No.	KOL-3									KOL-4
olivine No.	ol2	ol3	ol4	ol6	ol7	ol10	ol13	ol14	ol15	ol1
SiO ₂	40.81	39.95	39.77	39.14	39.75	41.59	40.08	40.27	40.14	40.33
MgO	46.98	46.19	48.99	49.18	47.85	45.61	44.17	47.91	45.57	45.778
CaO	0.18	0.14	0.18	0.17	0.17	0.18	0.14	0.18	0.16	0.145
FeO	10.69	11.50	10.38	10.37	10.93	11.60	13.85	10.78	12.53	11.951
MnO	0.15	0.15	0.12	0.13	0.16	0.15	0.17	0.14	0.17	0.178
NiO	0.48	0.42	0.49	0.49	0.50	0.42	0.40	0.39	0.53	0.407
Total	99.28	98.36	99.94	99.49	99.35	99.54	98.80	99.66	99.10	98.789
Fo	88.68	87.74	89.38	89.42	88.64	87.51	85.04	88.80	86.64	87.22

Table 1 (continued)

sample No.	KOL-4					KOL-9				
olivine No.	ol2	ol3	ol6	ol7	ol9	ol11	ol7	ol9	ol10	ol11
SiO ₂	39.858	39.87	38.243	41.031	41.107	39.116	40.02	41.72	41.44	41.17
MgO	46.557	45.128	48.179	46.119	46.268	48.101	48.44	45.02	47.40	47.28
CaO	0.178	0.167	0.186	0.164	0.162	0.17	0.15	0.16	0.18	0.13
FeO	12.78	13.818	11.843	12.331	11.6	11.858	10.64	11.06	10.76	11.62
MnO	0.168	0.161	0.131	0.159	0.161	0.148	0.10	0.13	0.13	0.15
NiO	0.363	0.362	0.433	0.412	0.5	0.475	0.47	0.44	0.44	0.40
Total	99.904	99.506	99.015	100.216	99.798	99.868	99.82	98.52	100.34	100.75
Fo	86.66	85.34	87.88	86.96	87.67	87.85	89.03	87.89	88.70	87.88

Table 1 (continued)

sample No.	KOL-9					KOL-16				
olivine No.	ol13	ol17	ol19	ol20	ol21	ol25	ol26	ol2	ol3	ol4
SiO ₂	40.23	40.56	38.42	39.32	39.16	39.48	39.73	40.08	39.31	42.09
MgO	48.32	49.35	47.09	48.22	48.15	46.58	46.82	48.82	49.32	43.62
CaO	0.16	0.17	0.14	0.17	0.17	0.16	0.17	0.13	0.16	0.16
FeO	11.11	10.80	12.17	11.06	10.98	11.80	11.11	11.34	10.98	12.11
MnO	0.00	0.06	0.00	0.11	0.17	0.15	0.16	0.00	0.00	0.15
NiO	0.41	0.47	0.41	0.43	0.45	0.43	0.45	0.44	0.49	0.44
Total	100.22	101.40	98.24	99.30	99.08	98.59	98.44	100.82	100.26	98.56
Fo	88.57	89.06	87.33	88.60	88.66	87.56	88.25	88.47	88.89	86.52

Table 1 (continued)

sample No.	KOL-16					MH-5a				
olivine No.	ol5	ol8	ol10	ol13	ol1	ol2	ol3	ol4	ol5	ol6
SiO ₂	39.89	39.83	40.05	40.43	40.80	39.73	38.23	39.47	38.73	39.83
MgO	49.03	49.29	49.33	47.13	46.60	46.06	48.63	49.20	44.68	46.98
CaO	0.15	0.15	0.14	0.17	0.15	0.15	0.14	0.11	0.15	0.15
FeO	10.78	10.38	10.61	10.78	11.78	12.53	12.53	10.58	15.86	11.71
MnO	0.14	0.12	0.11	0.15	0.16	0.17	0.18	0.13	0.17	0.16
NiO	0.48	0.48	0.50	0.47	0.41	0.41	0.36	0.47	0.33	0.41
Total	100.46	100.25	100.74	99.12	99.91	99.05	100.08	99.96	99.93	99.25
Fo	89.02	89.44	89.23	88.63	87.58	86.76	87.37	89.24	83.39	87.73

Table 1 (continued)

sample No.	MH-5a					WAF-4			WAF-5	
olivine No.	ol8	ol10	ol13	ol20	ol22	ol23	ol3	ol6	ol3	ol4
SiO ₂	39.07	38.55	39.55	38.92	38.77	39.90	40.24	40.17	38.78	40.85
MgO	43.49	44.40	44.06	44.74	46.32	48.20	47.14	45.26	47.98	49.55
CaO	0.18	0.17	0.18	0.20	0.17	0.14	0.16	0.18	0.18	0.16
FeO	17.30	15.81	16.17	15.81	12.66	11.60	11.22	12.16	11.06	10.81
MnO	0.25	0.48	0.24	0.00	0.19	0.14	0.13	0.16	0.00	0.13
NiO	0.34	0.34	0.33	0.33	0.38	0.43	0.44	0.46	0.43	0.43
Total	100.61	99.74	100.52	100.01	98.49	100.39	99.34	98.39	98.41	101.94
Fo	81.76	83.35	82.93	83.45	86.70	88.11	88.22	86.90	88.55	89.09

Table 1 (continued)

sample No.	WAF-5						
olivine No.	ol10	ol11	ol12	ol14	ol16	ol17	ol19
SiO ₂	39.23	39.80	38.63	40.22	39.83	38.94	39.26
MgO	48.43	47.43	47.30	48.39	48.53	47.78	47.29
CaO	0.16	0.14	0.16	0.16	0.15	0.15	0.16
FeO	11.02	12.33	12.37	11.41	10.74	11.74	11.01
MnO	0.13	0.15	0.27	0.14	0.13	0.04	0.01
NiO	0.43	0.40	0.42	0.39	0.42	0.35	0.59
Total	99.39	100.27	99.15	100.70	99.79	99.00	98.32
Fo	88.68	87.27	87.20	88.32	88.96	87.88	88.44

Table 2 Major elements compositions of host olivine crystals from Kilauea

sample No.	KIL5					KIL-9				
olivine No.	ol2	ol3	ol5	ol6	ol10	ol11	ol12	ol13	ol14	ol17
SiO2	39.52	39.94	39.09	39.16	39.75	39.41	39.75	38.84	39.21	38.63
MgO	47.73	48.99	46.59	47.27	49.02	48.43	46.47	44.80	47.51	44.90
CaO	0.19	0.23	0.23	0.21	0.20	0.23	0.24	0.23	0.23	0.21
FeO	12.17	11.35	13.18	12.45	10.46	10.85	14.38	15.05	12.22	15.03
MnO	0.13	0.08	0.15	0.14	0.09	0.00	0.35	0.08	0.04	0.14
NiO	0.35	0.38	0.35	0.36	0.40	0.39	0.29	0.24	0.41	0.35
Total	100.09	100.97	99.58	99.60	99.93	99.31	101.48	99.24	99.61	99.26
Fo	87.48	88.50	86.30	87.13	89.31	88.84	85.20	84.14	87.39	84.19

Table 2 (continued)

sample No.	KIL-9									
olivine No.	ol18	ol19	ol23	ol24	ol26	ol29	ol30	ol31	ol32	ol33
SiO ₂	40.56	39.76	39.68	39.54	39.04	37.58	38.69	38.57	38.22	38.78
MgO	49.25	45.40	47.31	46.45	47.62	43.58	46.25	46.89	45.91	45.21
CaO	0.21	0.15	0.20	0.20	0.21	0.22	0.21	0.20	0.20	0.18
FeO	11.06	16.17	11.84	14.81	10.87	15.71	13.82	11.80	13.36	15.34
MnO	0.04	0.00	0.23	0.00	0.31	0.79	0.39	0.56	0.46	0.21
NiO	0.40	0.34	0.41	0.32	0.41	0.28	0.32	0.39	0.35	0.25
Total	101.52	101.81	99.67	101.31	98.45	98.16	99.67	98.42	98.50	99.97
Fo	88.81	83.35	87.69	84.83	88.65	83.18	85.65	87.63	85.97	84.01

Table 2 (continued)

sample No.	KIL-9					
olivine No.	ol34	ol35	ol39	ol40	ol42	ol43
SiO ₂	39.68	39.68	39.95	39.84	39.38	40.10
MgO	47.82	45.75	49.49	48.32	47.43	48.13
CaO	0.16	0.21	0.21	0.21	0.20	0.22
FeO	13.49	15.93	11.38	11.18	12.39	12.77
MnO	0.17	0.00	0.00	0.00	0.00	0.14
NiO	0.37	0.26	0.41	0.40	0.38	0.39
Total	101.68	101.84	101.44	99.97	99.77	101.74
Fo	86.34	83.66	88.58	88.51	87.22	87.04

Table 3 Major elements compositions of glasses in olivine hosted melt
inclusions without daughter pyroxines from Koolau

sample No.	KOL-4 KOL-9				KOL-16			MH-5a
olivine No.	ol9	ol13	ol15	ol19	ol4	ol8	ol10	ol1
inclusion No.	1	1	1	1	1	1	1	1
SiO ₂	61.96	60.78	59.98	66.89	65.93	62.62	61.72	54.41
TiO ₂	1.85	2.15	1.94	1.09	2.01	2.16	2.03	2.02
Al ₂ O ₃	16.75	17.70	17.75	16.75	16.00	16.98	17.43	13.89
Cr ₂ O ₃	0.06	0.10	0.01	0.05	0.02	0.01	0.07	0.03
FeO	2.04	1.77	1.89	1.77	3.15	2.32	2.28	7.02
MnO	0.03	0.04	0.01	0.04	0.03	0.06	0.01	0.13
MgO	1.19	1.04	1.15	0.92	1.25	1.44	1.49	6.63
CaO	9.91	11.21	10.98	8.19	8.30	10.51	10.42	10.07
Na ₂ O	3.13	3.19	3.57	3.27	3.13	3.04	3.41	2.57
K ₂ O	0.43	0.89	0.88	0.37	0.72	0.64	0.68	0.41
NiO	0.03	0.01	0.00	0.03	0.00	0.04	0.05	0.00
Total	97.39	98.88	98.14	99.37	100.53	99.83	99.59	97.17

Table 3 (continued)

sample No.	MH-5a							
olivine No.	ol2	ol3	ol5	ol6	ol10	ol12	ol13	ol20
inclusion No.	1	1	1	1	1	1	1	1
SiO ₂	58.30	57.61	54.14	58.19	55.01	54.34	55.66	53.15
TiO ₂	1.36	1.93	2.23	1.93	2.21	2.04	2.22	2.08
Al ₂ O ₃	14.90	14.33	14.07	14.98	14.01	14.65	16.22	14.88
Cr ₂ O ₃	0.06	0.05	0.07	0.01	0.04	0.03	0.06	0.01
FeO	6.69	7.51	9.89	6.33	9.51	9.15	8.34	9.43
MnO	0.08	0.07	0.07	0.09	0.13	0.20	0.16	0.14
MgO	5.28	5.34	4.80	4.29	5.24	5.64	2.89	3.91
CaO	9.95	9.94	10.85	10.64	9.79	9.76	10.60	11.02
Na ₂ O	2.41	2.33	2.40	2.13	2.54	2.70	2.88	2.58
K ₂ O	0.37	0.34	0.40	0.25	0.55	0.48	0.53	0.48
NiO	0.02	0.05	0.07	0.02	0.03	0.08	0.00	0.04
Total	99.42	99.51	98.99	98.85	99.06	99.05	99.54	97.71

Table 3 (continued)

sample No.				
olivine No.	ol21	ol22	ol23	ol24
inclusion No.	1	1	1	1
SiO ₂	56.50	57.16	59.55	55.84
TiO ₂	1.83	2.10	1.82	2.20
Al ₂ O ₃	13.85	15.15	16.46	17.04
Cr ₂ O ₃	0.07	0.05	0.00	0.07
FeO	8.00	6.94	5.11	6.34
MnO	0.14	0.10	0.07	0.10
MgO	5.86	4.13	3.02	2.84
CaO	9.95	10.55	10.26	10.52
Na ₂ O	2.48	2.94	2.32	3.51
K ₂ O	0.45	0.59	0.61	0.65
NiO	0.00	0.00	0.00	0.03
Total	99.14	99.71	99.22	99.14

Table 4 Major elements compositions of glasses in olivine hosted melt

Inclusions without daughter pyroxenes from Kilauea

sample No.	KIL-5 KIL-9								
olivine No.	ol10	ol17	ol23	ol26	ol27	ol33	ol34	ol35	ol39
inclusion No.	1	1	1	1	1	1	1	1	1
SiO ₂	54.53	57.29	53.24	51.36	51.96	51.16	51.72	51.81	52.96
TiO ₂	2.87	2.16	2.65	2.83	3.04	2.79	3.23	2.54	2.99
Al ₂ O ₃	15.41	18.00	15.61	14.76	14.54	13.93	15.25	13.88	15.28
Cr ₂ O ₃	0.09	0.00	0.05	0.07	0.03	0.01	0.03	0.06	0.05
FeO	6.18	2.23	6.06	6.52	7.15	11.05	7.75	10.26	6.37
MnO	0.14	0.02	0.08	0.04	0.04	0.25	0.18	0.18	0.11
MgO	3.01	1.45	3.29	6.78	6.62	5.04	5.63	5.68	6.95
CaO	14.26	11.48	12.91	12.87	12.97	13.46	12.47	12.06	12.28
Na ₂ O	2.55	3.28	2.69	2.22	2.26	2.20	2.46	2.57	2.26
K ₂ O	0.55	0.29	0.30	0.30	0.31	0.36	0.58	0.52	0.55
NiO	0.00	0.06	0.09	0.04	0.06	0.00	0.07	0.00	0.05
Total	99.58	96.27	96.96	97.79	98.97	100.24	99.36	99.55	99.86

Table 4 (continued)

sample No.	KIL-9			
olivine No.	ol40	ol41	ol42	ol43
inclusion No.	1	1	1	1
SiO ₂	51.57	49.67	52.01	52.77
TiO ₂	3.05	2.79	2.83	2.68
Al ₂ O ₃	15.67	15.31	15.23	15.25
Cr ₂ O ₃	0.09	0.05	0.09	0.09
FeO	6.87	7.56	6.79	6.27
MnO	0.07	0.17	0.11	0.12
MgO	4.32	6.90	6.52	6.99
CaO	14.05	12.32	12.35	12.87
Na ₂ O	2.27	2.15	1.73	2.24
K ₂ O	0.49	0.55	0.50	0.65
NiO	0.01	0.06	0.00	0.03
Total	98.45	97.53	98.14	99.96

Table 5 Pb isotopic compositions of olivine-hosted melt inclusions from Koolau

sample No.	olivine No.	inclusion No.	$^{208}\text{Pb}/^{206}\text{Pb}$	in-run error (1 σ)	$^{207}\text{Pb}/^{206}\text{Pb}$	in-run error (1 σ)
KOL-3	ol2	1	0.856	0.007	2.072	0.017
	ol3	1	0.865	0.007	2.083	0.017
	ol4	1	0.858	0.011	2.138	0.031
	ol4	2	0.875	0.011	2.105	0.023
	ol6	1	0.862	0.011	2.156	0.026
	ol7	1	0.859	0.010	2.139	0.022
	ol10	1	0.848	0.008	2.100	0.016
	ol13	1	0.850	0.014	2.092	0.025
	ol14	1	0.868	0.008	2.110	0.017
KOL-4	ol15	1	0.863	0.011	2.080	0.023
	ol1	1	0.893	0.013	2.153	0.029
	ol1	2	0.861	0.009	2.149	0.024
	ol2	1	0.858	0.009	2.075	0.022
	ol3	1	0.863	0.008	2.125	0.018
	ol6	1	0.859	0.009	2.067	0.021
	ol7	1	0.874	0.006	2.118	0.017
	ol9	1	0.868	0.013	2.137	0.034
	ol11	1	0.866	0.008	2.104	0.035
KOL-9	ol7	1	0.871	0.008	2.142	0.028
	ol7	2	0.867	0.010	2.135	0.032
	ol9	1	0.862	0.007	2.090	0.015
	ol9	2	0.862	0.015	2.132	0.038
	ol10	1	0.868	0.012	2.060	0.027
	ol11	1	0.885	0.006	2.135	0.020
	ol13	1	0.858	0.010	2.133	0.026
	ol15	1	0.871	0.008	2.129	0.023
	ol16	1	0.858	0.010	2.118	0.025
	ol17	1	0.859	0.008	2.110	0.021
	ol19	1	0.861	0.006	2.115	0.017
	ol20	1	0.862	0.010	2.147	0.026
	ol21	2	0.872	0.013	2.129	0.031
	ol22	1	0.863	0.006	2.127	0.015
	ol22	2	0.871	0.009	2.104	0.022
KOL-16	ol25	1	0.858	0.005	2.101	0.012
	ol26	1	0.881	0.010	2.172	0.023
	ol2	1	0.862	0.008	2.099	0.019
	ol2	2	0.869	0.005	2.121	0.012
	ol3	1	0.853	0.009	2.117	0.025
	ol4	1	0.855	0.005	2.088	0.017
	ol5	1	0.870	0.010	2.127	0.027
	ol8	1	0.873	0.011	2.050	0.022
	ol10	1	0.873	0.011	2.155	0.028
MH-5a	ol13	1	0.869	0.002	2.108	0.007
	ol1	1	0.874	0.010	2.108	0.027
	ol2	1	0.846	0.006	2.083	0.019
	ol3	1	0.873	0.007	2.134	0.019
	ol4	1	0.887	0.008	2.149	0.023
	ol5	1	0.867	0.010	2.158	0.027
	ol6	1	0.854	0.008	2.082	0.024
	ol8	1	0.867	0.005	2.114	0.011
	ol10	1	0.854	0.002	2.095	0.005
WAF-4	ol12	1	0.847	0.011	2.093	0.023
	ol13	1	0.868	0.011	2.156	0.027
	ol20	1	0.866	0.007	2.126	0.018
	ol21	1	0.865	0.007	2.108	0.019
	ol22	1	0.876	0.008	2.168	0.021
	ol23	1	0.840	0.011	2.112	0.026
	ol24	1	0.851	0.007	2.116	0.021
	ol2	1	0.856	0.003	2.103	0.006
	ol3	1	0.844	0.005	2.080	0.017
WAF-5	ol3	2	0.840	0.012	2.082	0.030
	ol3	3	0.864	0.011	2.094	0.029
	ol6	1	0.857	0.010	2.111	0.019
	ol3	1	0.868	0.007	2.113	0.017
WAF-5	ol3	2	0.871	0.003	2.095	0.009

Table 5 (Continued)

sample No.	olivine No.	inclusion No.	$^{208}\text{Pb}/^{206}\text{Pb}$	in-run error (1 σ)	$^{207}\text{Pb}/^{206}\text{Pb}$	in-run error (1 σ)
WAF-5	ol3	3	0.871	0.007	2.124	0.015
	ol3	4	0.880	0.012	2.086	0.030
	ol4	1	0.874	0.008	2.094	0.020
	ol5	1	0.883	0.009	2.127	0.027
	ol10	1	0.855	0.008	2.076	0.020
	ol11	1	0.880	0.013	2.192	0.031
	ol12	1	0.867	0.011	2.135	0.027
	ol13	1	0.859	0.010	2.092	0.030
	ol14	1	0.865	0.008	2.148	0.021
	ol15	1	0.884	0.008	2.090	0.019
	ol16	1	0.860	0.009	2.124	0.024
	ol17	1	0.873	0.007	2.125	0.022
	ol19	1	0.869	0.003	2.118	0.013

Table 6 Pb isotopic compositions of olivine-hosted melt inclusions from Kilauea

sample No.	olivine No.	inclusion No.	$^{208}\text{Pb}/^{206}\text{Pb}$	in-run error (1 σ)	$^{207}\text{Pb}/^{206}\text{Pb}$	in-run error (1 σ)
KIL5	ol2	1	0.846	0.009	2.069	0.021
	ol2	2	0.838	0.009	2.073	0.021
	ol3	1	0.839	0.006	2.083	0.018
	ol4	1	0.845	0.003	2.061	0.011
	ol5	1	0.863	0.011	2.100	0.022
	ol6	1	0.831	0.005	2.067	0.013
	ol10	1	0.840	0.008	2.053	0.022
	ol11	1	0.849	0.005	2.086	0.017
	ol15	1	0.851	0.009	2.067	0.024
KIL9	ol12	2	0.829	0.006	2.056	0.018
	ol13	1	0.857	0.006	2.112	0.015
	ol14	1	0.853	0.007	2.076	0.018
	ol14	3	0.845	0.008	2.109	0.023
	ol17	1	0.840	0.008	2.076	0.024
	ol18	1	0.834	0.003	2.071	0.006
	ol19	1	0.842	0.012	2.055	0.027
	ol23	1	0.824	0.007	2.019	0.015
	ol24	1	0.825	0.009	2.050	0.023
	ol26	1	0.848	0.008	2.063	0.026
	ol27	1	0.829	0.007	2.080	0.024
	ol29	1	0.846	0.007	2.099	0.020
	ol30	1	0.844	0.006	2.092	0.014
	ol31	1	0.844	0.006	2.101	0.015
	ol32	1	0.856	0.009	2.091	0.025
	ol33	1	0.845	0.008	2.084	0.018
	ol34	1	0.832	0.006	2.042	0.018
	ol35	1	0.839	0.009	2.052	0.025
	ol39	1	0.855	0.007	2.054	0.018
	ol40	1	0.844	0.009	2.067	0.016
	ol41	1	0.846	0.008	2.077	0.016
	ol42	1	0.854	0.006	2.091	0.013
	ol43	1	0.841	0.008	2.073	0.022

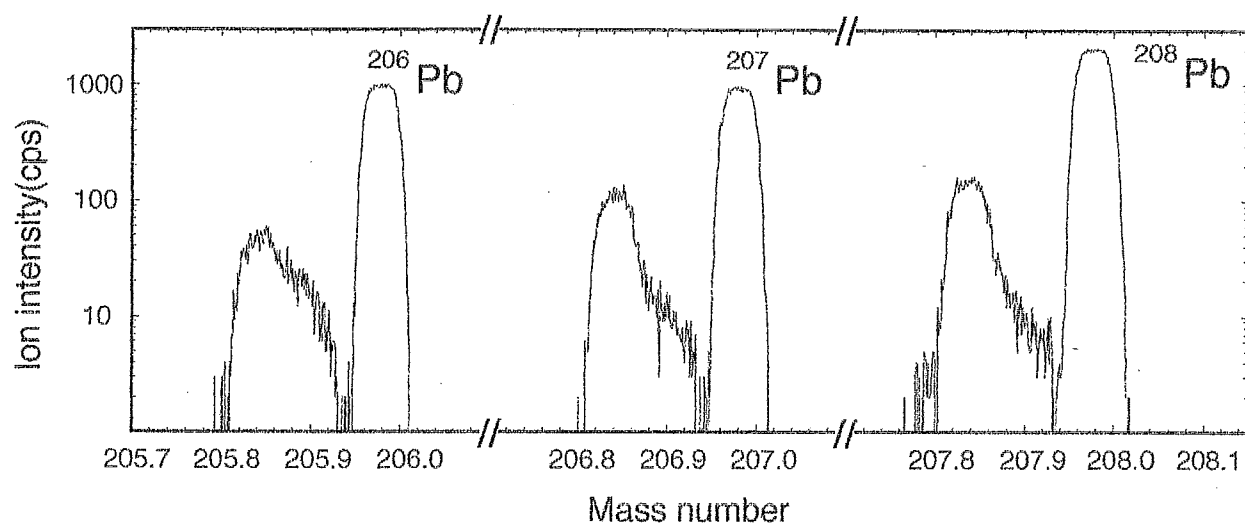


Fig. 1

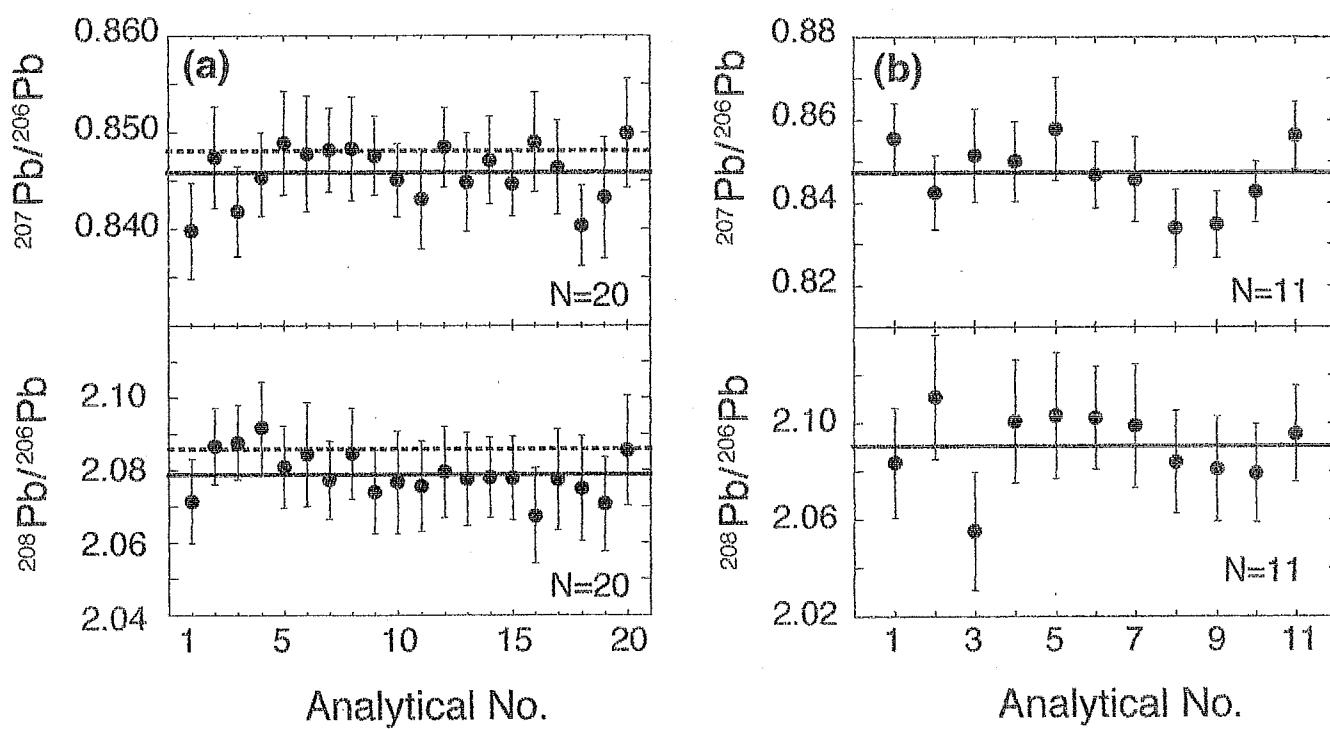


Fig. 2

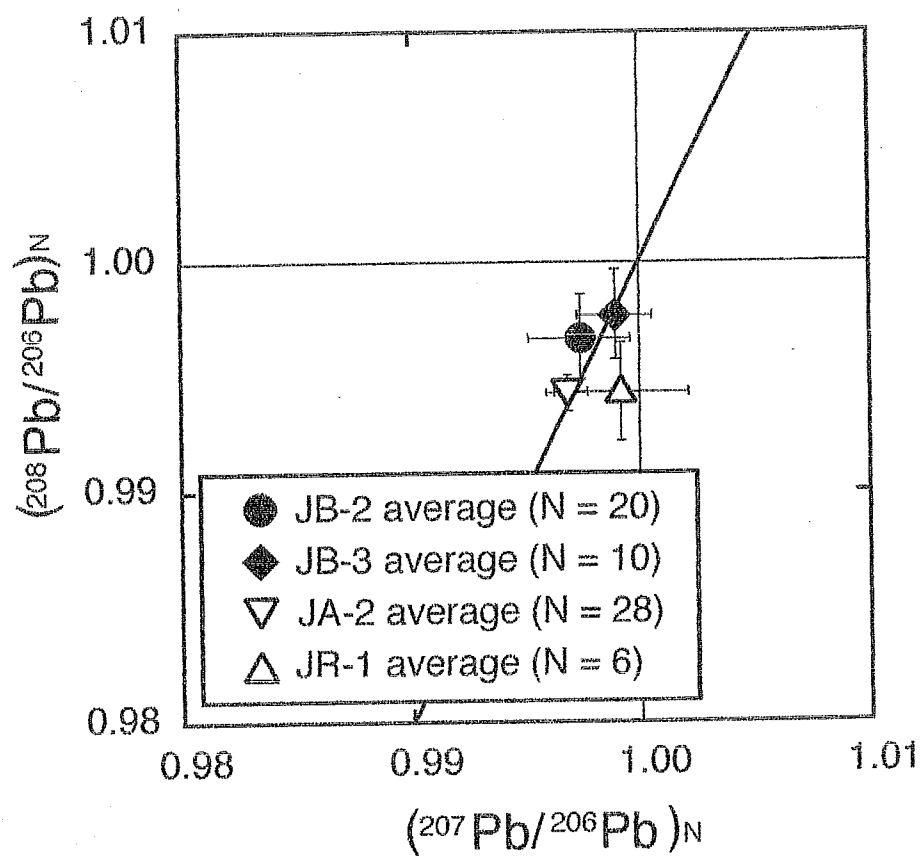


Fig. 3

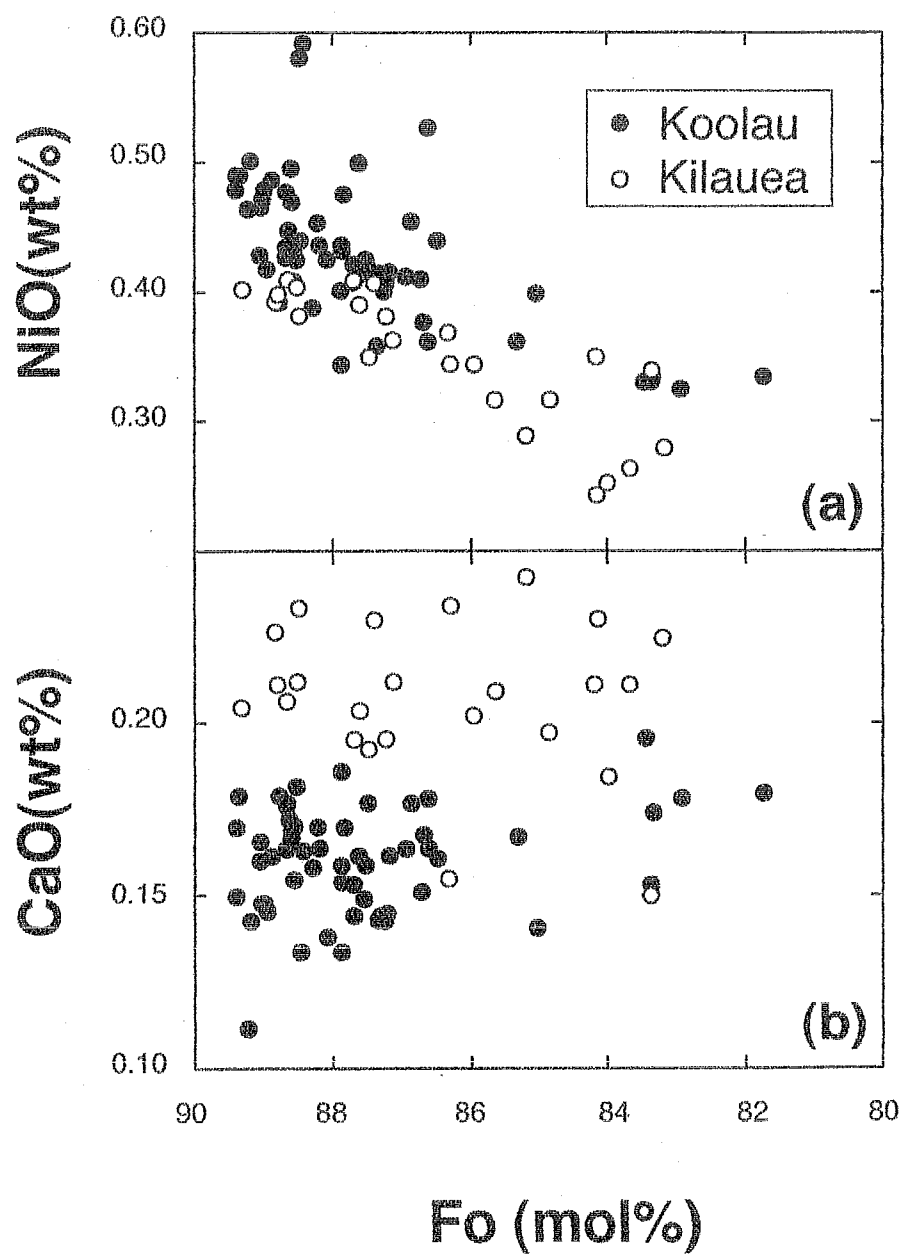
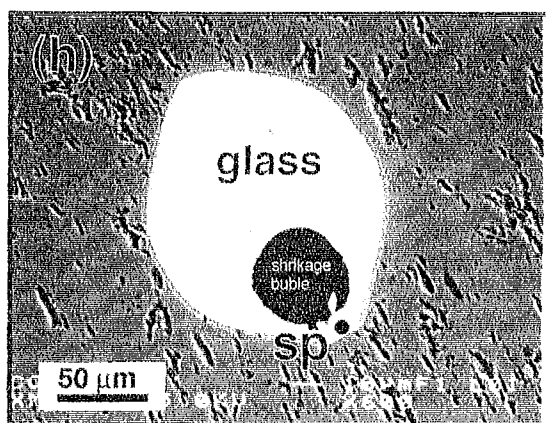
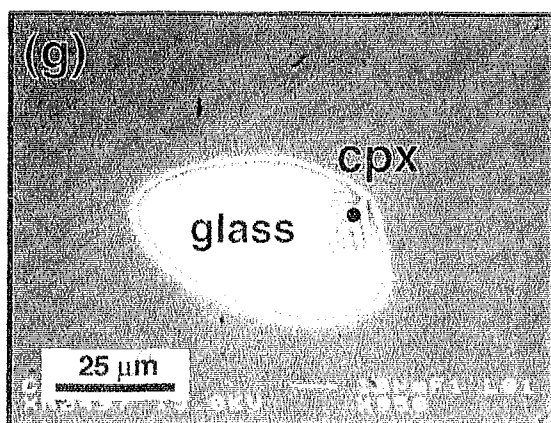
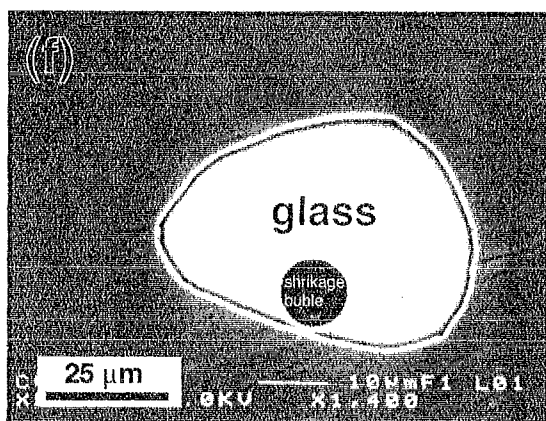
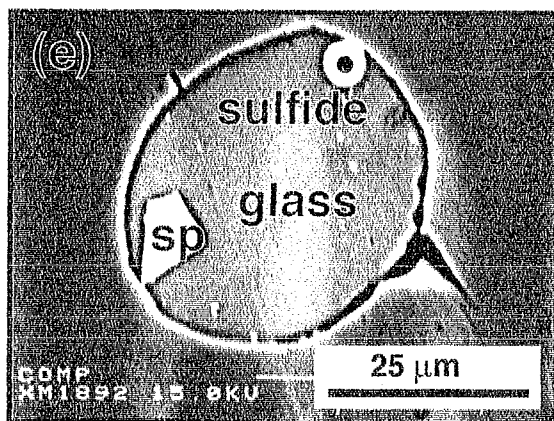
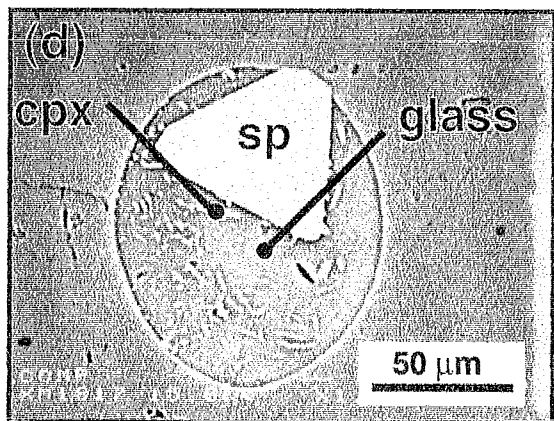
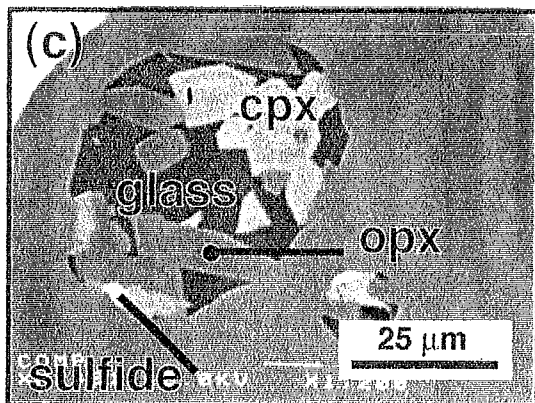
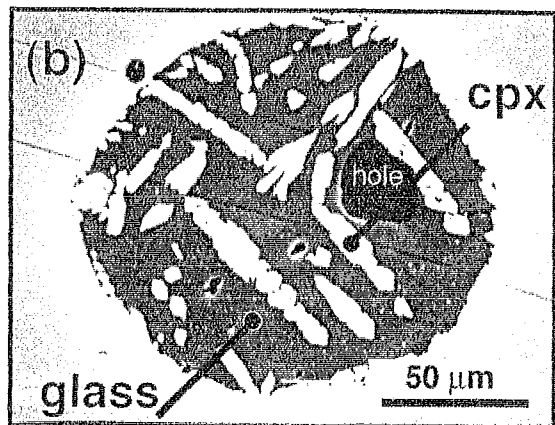
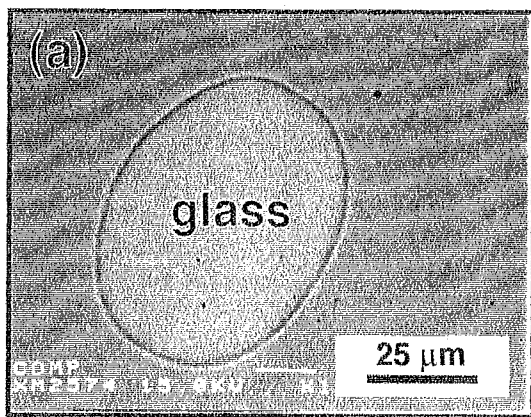


Fig. 4



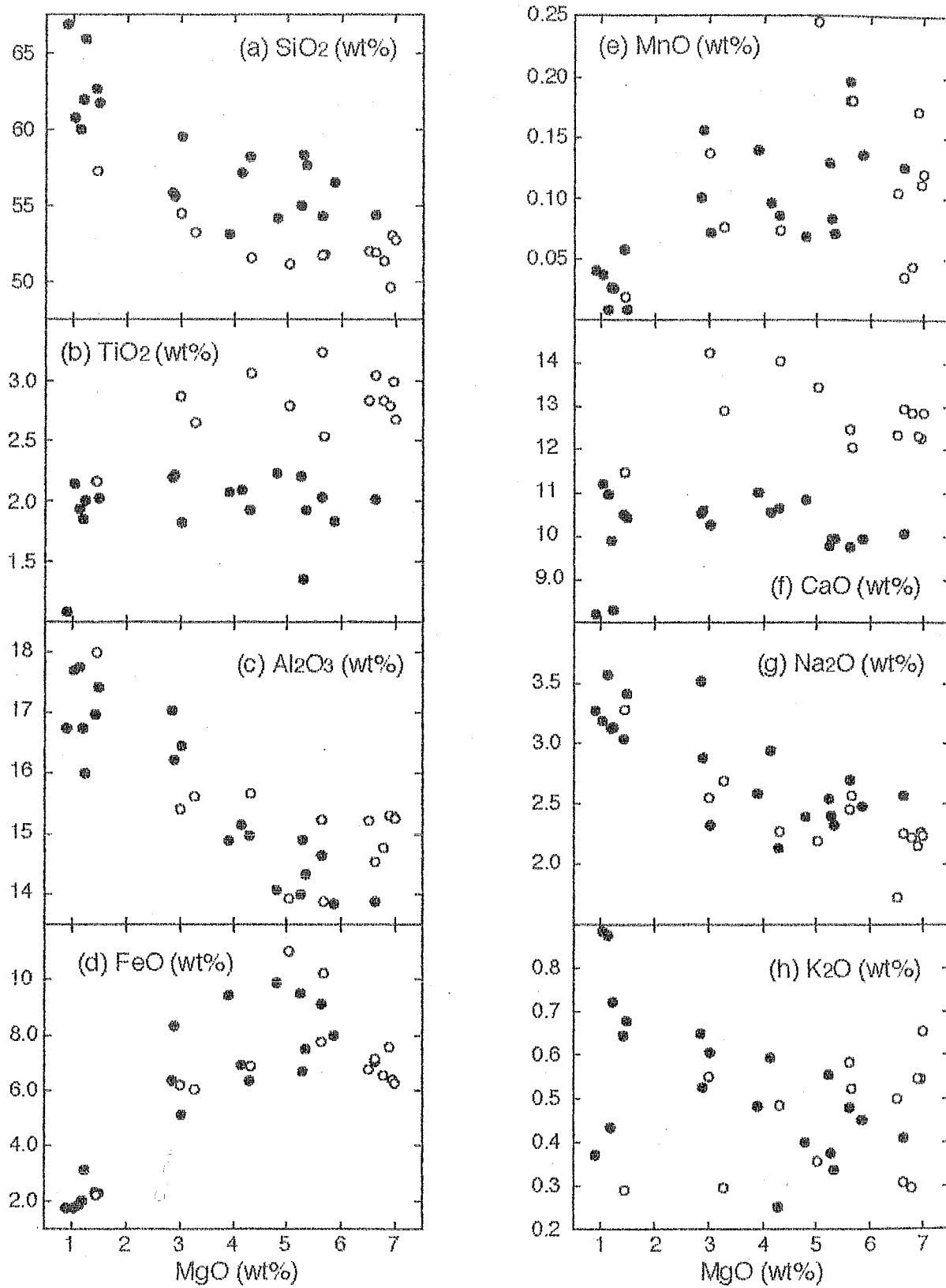


Fig. 6

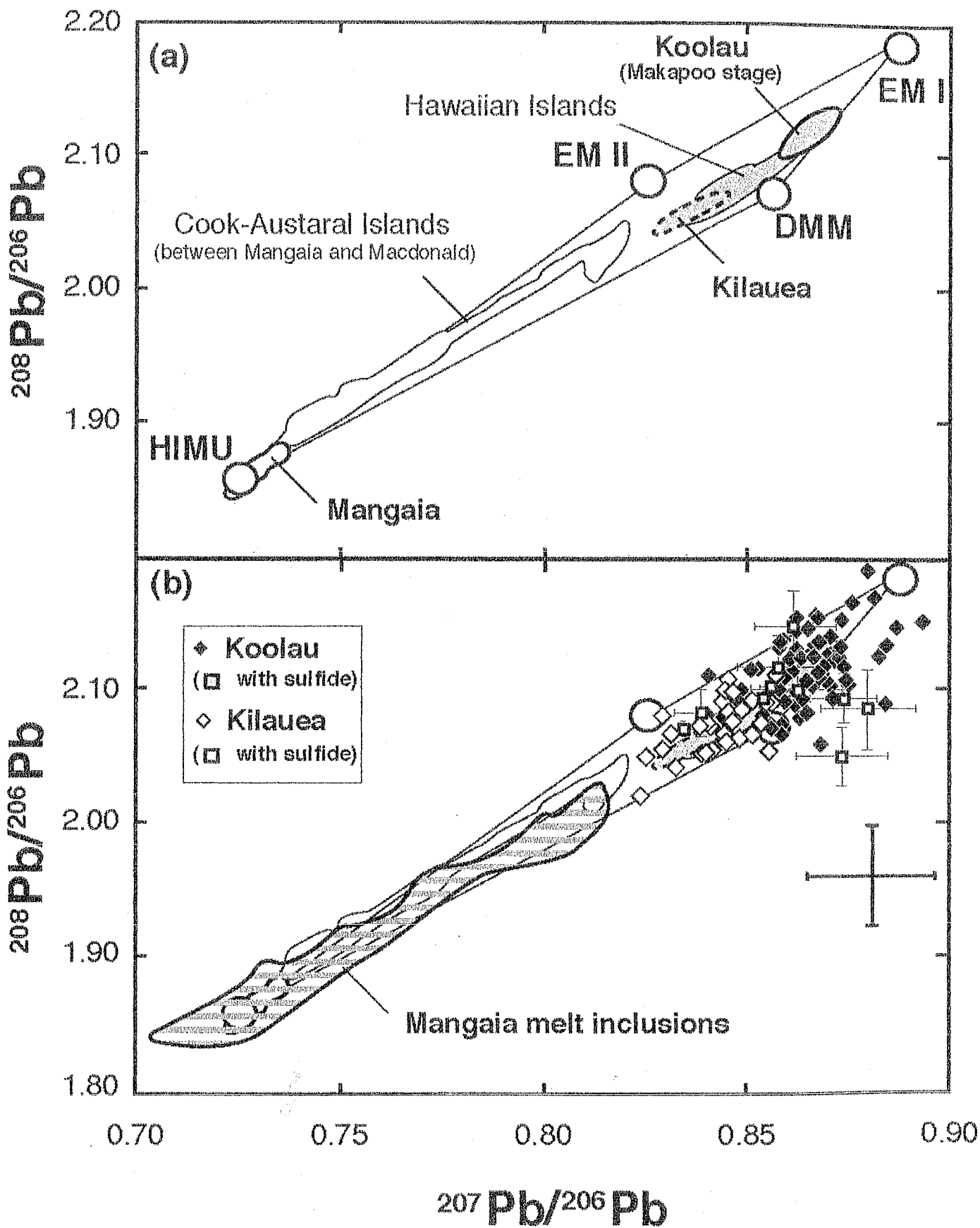


Fig. 7

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