

Geology of the Gorny Altai subduction–accretion complex, southern Siberia: Tectonic evolution of an Ediacaran–Cambrian intra-oceanic arc-trench system

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Abstract

The Gorny Altai region in southern Siberia is one of the key areas in reconstructing the tectonic evolution of the western segment of the Central Asian Orogenic Belt (CAOB). This region features various orogenic elements of Late Neoproterozoic–Early Paleozoic age, such as an accretionary complex (AC), high-*P/T* metamorphic (HP) rocks, and ophiolite (OP), all formed by ancient subduction–accretion processes. This study investigated the detailed geology of the Upper Neoproterozoic to Lower Paleozoic rocks in a traverse between Gorno-Altai city and Lake Teletskoy in the northern part of the region, and in the Kurai to Chagan-Uzun area in the southern part. The tectonic units of the studied areas consist of (1) the Ediacaran (=Vendian)–Early Cambrian AC, (2) ca. 630 Ma HP complex, (3) the Ediacaran–Early Cambrian OP complex, (4) the Cryogenian–Cambrian island arc complex, and (5) the Middle Paleozoic fore-arc sedimentary rocks. The AC consists mostly of paleo-atoll limestone and underlying oceanic island basalt with minor amount of chert and serpentinite. The basaltic lavas show petrochemistry similar to modern oceanic plateau basalt. The 630 Ma HP complex records a maximum peak metamorphism at 660 °C and 2.0 GPa that corresponds to 60 km-deep burial in a subduction zone, and exhumation at ca. 570 Ma. The Cryogenian island arc complex includes boninitic rocks that suggest an incipient stage of arc development. The Upper Neoproterozoic–Lower Paleozoic complexes in the Gorno-Altai city to Lake Teletskoy and the Kurai to Chagan-Uzun areas are totally involved in a subhorizontal piled-nappe structure, and overprinted by Late Paleozoic strike-slip faulting. The HP complex occurs as a nappe tectonically sandwiched between the non- to weakly metamorphosed AC and the OP complex. These lithologic assemblages and geologic structure newly documented in the Gorny Altai region are essentially similar to those of the circum-Pacific (Miyashiro-type) orogenic belts, such as the Japan Islands in East Asia and the Cordillera in western North America. The Cryogenian boninite-bearing arc volcanism indicates that the initial stage of arc development occurred in a transient setting from a transform zone to an incipient subduction zone. The less abundant of terrigenous clastics from mature continental crust and thick deep-sea chert in the Ediacaran–Early Cambrian AC may suggest that the southern Gorny Altai region evolved in an intra-oceanic arc-trench setting like the modern Mariana arc, rather than along the continental arc of a major

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continental margin. Based on geological, petrochemical, and geochronological data, we synthesize the Late Neoproterozoic to Early Paleozoic tectonic history of the Gorny Altai region in the western CAOAB.

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1. Introduction

The distinction of two types of orogeny, i.e., oceanic subduction-related, accretionary-type and continent–continent collision-type, has been widely accepted within the plate tectonic framework since Dewey and Bird (1970). The former, also called Pacific-type (Matsuda and Uyeda, 1971) or Miyashiro-type (Maruyama, 1997), is characterized by the formation of subduction–accretion complexes involving high-*P/T* metamorphic (HP) rocks, and extensive calc-alkaline magmatism that usually results in a voluminous increase of continental crust. At the orogenic climax, the HP rocks are tectonically exhumed from mantle depths to the surface to form a slab-like occurrence, and then the primary structure of an orogen is completed (Maruyama et al., 1996; Maruyama, 1997). During a long-term convergent orogeny, intermittent collisions of minor arcs, microcontinents, oceanic plateaus, or seamounts, and related re-organization of plate boundaries may cause secondary modification on a smaller scale on the primary larger-scale structures. In contrast, a collision-type orogeny contributes very little to continental growth, because it involves no more than the reworking of pre-existing continental material, primarily formed through the above-mentioned accretionary orogeny. Thus, the subduction-related accretionary orogens are most important in understanding continental growth through time, as emphasized by Maruyama (1997) and Isozaki (1996).

Central Asia, surrounded by the Siberian, North China, and Kazakhstan continental blocks, occupies a huge area within the modern Eurasian continent (Fig. 1a). Its orogenic belts have long been studied with respect to the Paleo-Asian and Paleo-Pacific oceans (e.g., Maruyama et al., 1989; Berzin and Dobretsov, 1994; Maruyama, 1994; Dobretsov et al., 1995, 2003; Sengör and Natal'in, 1996; Buslov et al., 2001; Yakubchuk, 2002; Khain et al., 2003; Xiao et al., 2003). The tectonic evolution of central Asia provides a record of the long-term convergent history of the Paleo-Pacific Ocean, because it mainly comprises orogenic complexes that formed by subduction–accretion and collisional processes (Zonenshain et al., 1990; Windley, 1992; Sengör et al., 1993; Berzin and Dobretsov, 1994; Berzin et al., 1994; Dobretsov et al., 1995; Sengör and Natal'in, 1996; Buslov et al., 2001; Khain et al., 2003; Kheraskova et al., 2003; Jahn, 2004; Windley et al., 2007). Over the last two decades, the resultant orogen in central Asia, one of the largest in the world, has been called the Central Asian fold belt (Zonenshain et al., 1990; Mossakovsky et al., 1993), the Altaid Tectonic Collage (Sengör et al., 1993; Sengör and Natal'in, 1996; Yakub-

chuk, 2002, 2004), or the Central Asian Orogenic Belt (CAOB, Jahn et al., 2000; Xiao et al., 2003; Windley et al., 2007); we use the last.

Gorny Altai in southern Siberia, located in the western segment of the CAOAB (Fig. 1a), is a key region for understanding the tectonic evolution of central Asia. It contains an exceptionally well-preserved Late Neoproterozoic–Early Cambrian subduction–accretion orogen that formed prior to final continental collision between Siberia and Kazakhstan in the Late Paleozoic, as pointed by Buslov et al. (1993), Buslov and Watanabe (1996) and Buslov et al. (1998, 2001).

During the last decade since 1997, in a joint research project between the Tokyo Institute of Technology and the Institute of Geology and Mineralogy, Siberian Branch of Russian Academy of Sciences, we conducted intensive field mapping in the Gorny Altai region at scales of 1:5000 to 1:6250 in order to document the primary tectonic framework of the Late Neoproterozoic–Early Paleozoic AC, HP rocks, OP, and arc complex. Some preliminary results of our research, e.g., petrochemistry of the accreted basaltic rocks, stratigraphy and structure of the accreted paleo-atoll limestone complex, petrology of the HP rocks, and radiometric dating of the accreted limestones, were reported in Utsunomiya et al. (1998), Uchio et al. (2001, 2003, 2004), Ota et al. (2002) and Nohda et al. (2003). On the basis of our research results accumulated in the last decade, this article describes the tectonic units of the Gorny Altai region and their structures and discusses the pre-collisional evolution of the Late Neoproterozoic–Early Paleozoic accretionary orogen that developed as a mid-oceanic arc-trench system in the western segment of the CAOAB.

In this article, we follow the latest geologic timescale by Gradstein et al. (2004), using Ediacaran (542–630 Ma) and Cryogenian (630–850 Ma) for the late Neoproterozoic, instead of traditionally-used Vendian and Sturtian.

2. Geological outline of the Gorny Altai region

The Gorny Altai region in southwestern Siberia (Fig. 1a) forms a triangle-shaped tectonic domain that is fault-bounded with two neighboring orogenic units; i.e., the West Sayan terrane to the east and the Altai-Mongolian terrane on the southwest (Fig. 1b). The northern extension of the Gorny Altai region is extensively covered by Quaternary sediments. The West Sayan terrane and the Altai-Mongolian terrane with a micro-continental nucleus are individual intra-oceanic arc systems developed in the Paleo-Asian Ocean, contemporaneously with the Gorny Altai region. These terranes were docked together with

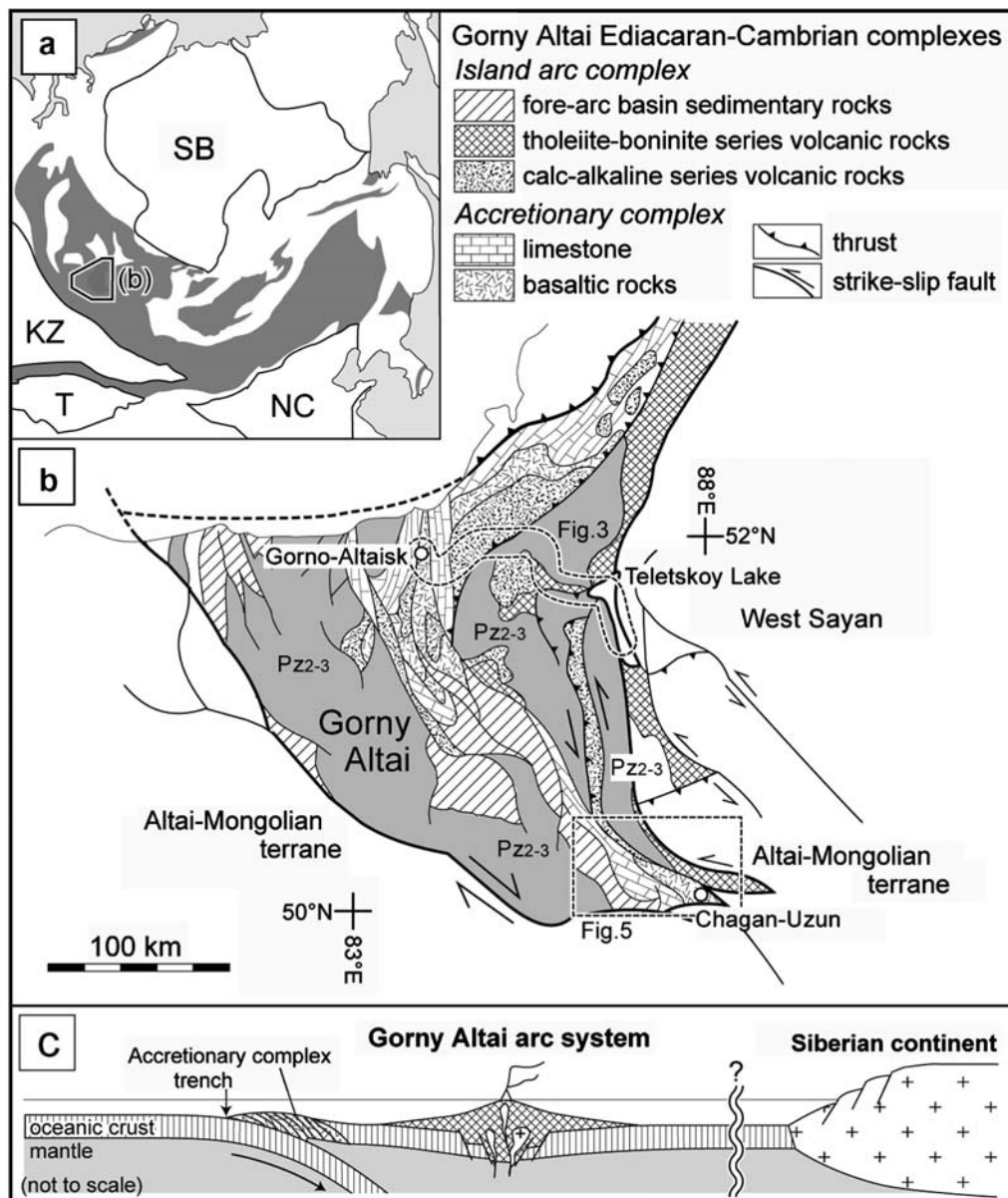


Fig. 1. (a) Index map of subduction-accretion complexes (grey) in the Central Asian Orogenic Belt, surrounded by Siberian (SB), Kazakhstan (KZ), North China (NC) and Tarim (T) continental blocks (modified after Sengör and Natal'in, 1996). (b) Geological sketch map of the Gorny Altai region (modified after Buslov et al., 1993, 2001, 2004), surrounded by the West Sayan and Altai-Mongolian terranes. Pz2-3, Middle to Late Paleozoic unit. Localities of Figs. 3 and 5 are shown. (c) Schematic profile showing tectonic setting of the Ediacaran-Cambrian Gorny Altai arc system with subduction-accretion complexes.

the Gorny Altai region during final closure of the ocean in the Late Paleozoic, giving rise to the western segments of the CAO between Siberia and Kazakhstan (Buslov et al., 2001, 2002, 2003, 2004).

The Gorny Altai region is mainly composed of several Neoproterozoic-Cambrian geological units; i.e., non- to weakly-metamorphosed AC, HP and OP rocks, arc volcano-sedimentary rocks, and Paleozoic cover. All these units were likely formed through Late Neoproterozoic to Paleozoic convergent tectonics that took place off the southern margin of the Siberian craton (Fig. 1c). The Ediacaran-Cambrian complexes were intruded by Middle-Late Paleozoic plutons and overlain by a Middle Paleozoic sedimentary cover as the arc grew with time (Buslov et al.,

1993, 2001; Dobretsov et al., 1995). These units are closely associated, particularly in the eastern half of the Gorny Altai region; Sengör and Natal'in (1996) once called them the Eastern Altai unit.

The Ediacaran-Cambrian complexes generally show an east-dipping imbricate structure that suggests a westward tectonic vergence during the Paleozoic (Fig. 2). In the southern Gorny Altai region, a subhorizontal to gently east-dipping nappe-pile structure is dominant. The secondary strike-slip faults modified the primary subhorizontal nappe structure. Nonetheless, the arc complex generally occurs on the eastern (Siberian) side, whereas the AC, the HP and the OP complexes are on the west. Although geological units in the northern part can be regionally

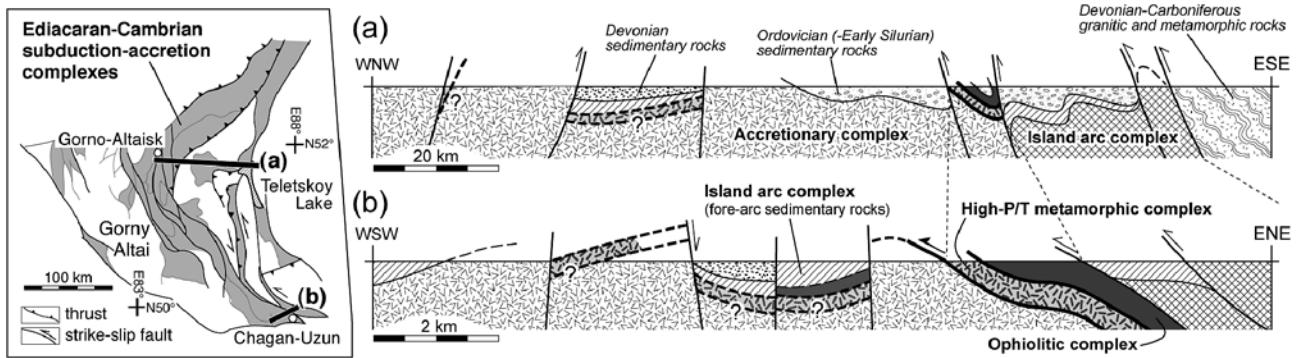


Fig. 2. Schematic geotectonic profiles of the Gorny Altai subduction-accretion complexes, based on Buslov et al. (1993) and this study. The inset shows locations of the geotectonic profiles (a) and (b). (a) Profile from Gorno-Altai to Lake Teletskoy, northern Gorny Altai (see also Figs. 1b and 3); (b) Profile in the Chagan-Uzun area, southern Gorny Altai (see also Figs. 1b, 5 and 6a). The Gorny Altai subduction-accretion complex represents a subhorizontal piled nappe structure, composed of the Ediacaran–Cambrian accretionary complex (AC), ca. 630 Ma high- P/T metamorphic (HP) complex, ophiolite (OP) complex, and Cryogenian–Ediacaran arc complex, in ascending order. The island fore-arc sedimentary rocks unconformably cover the AC plus HP and OP complexes. The primary subhorizontal structure has been modified by secondary strike-slip fault systems related to the Late Paleozoic collisional events.

correlated with those in the south (Fig. 2a), thick Mid-Upper Paleozoic sedimentary rocks and later strike-slip deformation has concealed the mutual geological relations and structural contacts among the units. Therefore, our field mapping mainly focused on the southern Gorny Altai, supplemented by a minor study in the north. First, we describe briefly the regional geology in the northern Gorny Altai region, and then geological details of the southern areas.

3. Northern Gorny Altai

In the northern Gorny Altai region (Fig. 1b), pre-Cenozoic rocks expose continuously across a traverse between Gorno-Altai city and Lake Teletskoy (Fig. 3). In this transect, (1) Cambrian AC, (2) HP rocks, (3) OP complex, (4) Ordovician–Devonian clastics, (5) Ediacaran to Early Cambrian tholeiite–boninite-bearing island arc complex with, and Early–Middle Cambrian calc-alkaline island arc complex, and (6) Devonian–Carboniferous gneiss–schist complex (Teletsk complex) occur as north–south trending belts. The Ordovician (–Early Silurian), and Devonian terrigenous and volcanic rocks unconformably overlie the AC, the HP rocks and the OP complex (e.g., Buslov et al., 1993).

The AC widely occurs to the south of Gorno-Altai city along the Katun river (Fig. 3). This unit is composed mostly of non- to weakly metamorphosed basalt, limestone, mudstone with minor amount of chert and sandstone. On the outcrop scale, these rocks occur as several fault-bounded slices (Fig. 4). Within individual units, basaltic rocks are directly covered by limestones with sedimentary contact. The basaltic rocks include massive and pillow lavas, lava breccias, dikes or sills, and often contain thin limestone intercalations. The limestones are bedded, micritic, and interbedded with thin massive chert layers immediately above the basaltic lavas. Judging from

an analogy with modern and ancient examples, these lithological assemblages correspond to those of the sediments on and around a mid-oceanic seamount, particularly those of the slope facies transient to the deep-sea floor facies (Uchio et al., 2004). Both basalt and limestone are in contact with surrounding mudstone with sandstone lenses, suggesting a block-in-matrix relationship for the mid-oceanic rocks enveloped within the mudstone matrix. The block-in-matrix relationship suggests secondary mixing of these mid-oceanic rocks (basalt and limestone) and continent-derived terrigenous clastics (mudstone and sandstone) probably in an active trench (e.g., Isozaki, 1987, 1997; Sano and Kanmera, 1991). Thus, field observations on the lithological assemblage and the mode of occurrence indicate that these rocks represent an ancient AC with paleo-seamount fragments.

From the AC in the northern Gorny Altai, “Vendian-type” stromatolites occur with microphytolites, calcareous algae and sponge spicules in limestones, whereas siliceous mudstones adjacent to the limestones yield Early Cambrian sponge spicules (Afonin, 1976; Zybin and Sergeev, 1978; Terleev, 1991; Buslov et al., 1993, 2004). As the Ordovician and Devonian clastic rocks unconformably cover the AC, the formation age of the AC at the ancient trench is estimated to be middle–late Cambrian.

Both the HP and the OP complexes occur as thin tectonic slices. However their structural relationships with other units are not clear. The HP complex is composed mostly of basic schists of the (sub-) greenschist-facies grade. The OP complex is composed of layered gabbros and pyroxenites, sheeted dikes, pillow lavas and breccias, and siliceous and argillaceous sedimentary rocks in ascending order. The basal gabbros and pyroxenites are in fault contact with the greenschist- to the lower amphibolite-facies basic schists.

Near Lake Teletskoy, there is a small exposure of the island arc complex composed of the Ediacaran–Early Cambrian tholeiite–boninite series basaltic lavas and dikes,

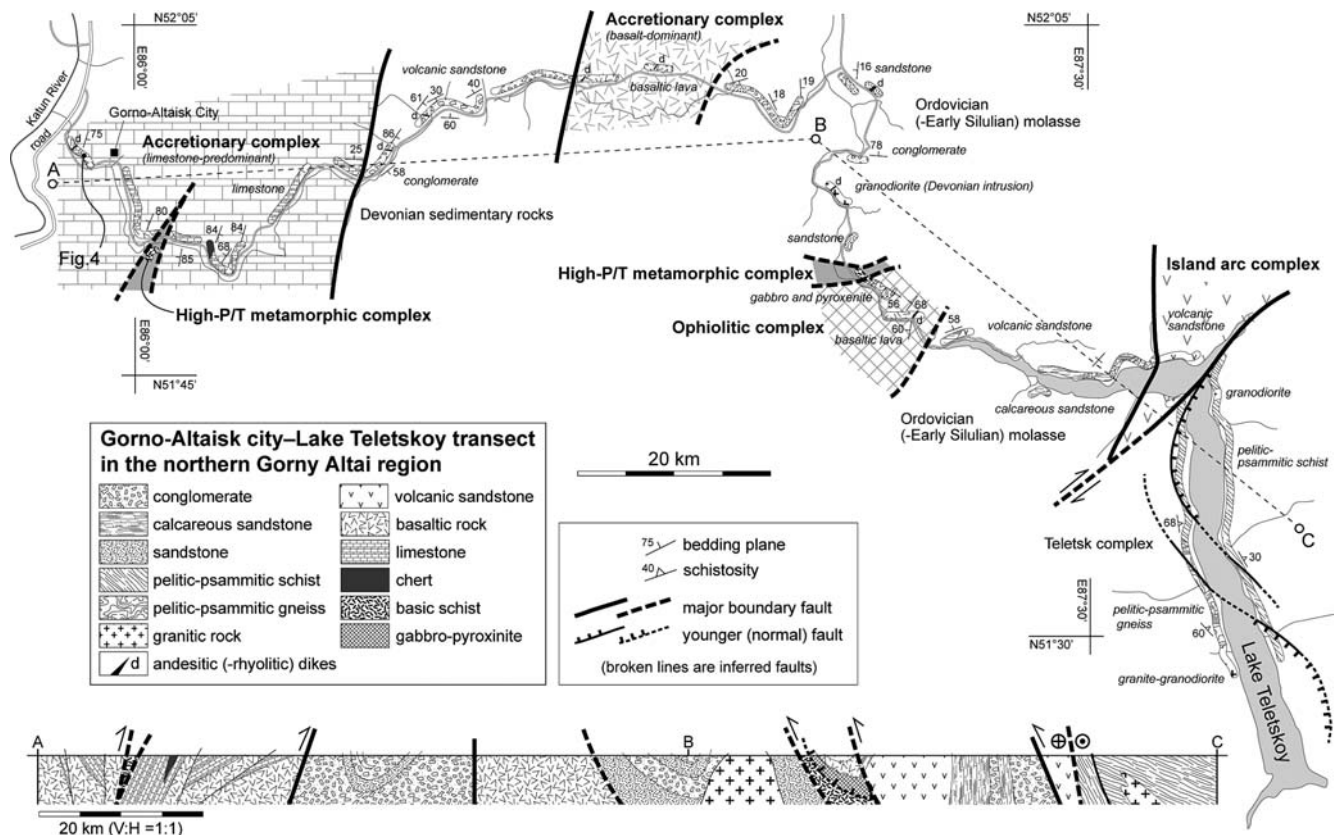


Fig. 3. The geotransverse and profile of the northern Gorny Altai between Gorno-Altai and Lake Teletskoy. Locality of Fig. 4 is also shown. See text for explanations.

Middle–Late Cambrian calc-alkaline series andesitic dikes, lavas, tuffs, diorites, plagiogranites, and sedimentary rocks. Dikes of the calc-alkaline series intrude the tholeiite–boninite series rocks and the OP complex to the east, and the AC in the west (Fig. 3). The Early–Middle Cambrian age (Botomian–Amgian) of the island arc complex has been confirmed by numerous archaeocyathan and trilobites from intercalated sedimentary rocks (Buslov et al., 1993).

This island arc complex is bounded by a strike-slip fault from the Devonian–Carboniferous gneiss–schist complex (Teletsk complex). The Teletsk complex forms a domal structure and is composed of schists and gneisses of greenschist- to the lower amphibolite-facies grades associated with two-mica granites (Buslov et al., 1993; Buslov and Sintubin, 1995; Smirnova et al., 2002). K–Ar ages of muscovite, biotite, and amphibole from the Teletsk granites and metamorphic rocks range from 318 Ma (Carboniferous) to 390 Ma (Devonian) (Buslov and Sintubin, 1995).

4. Southern Gorny Altai

In the southern Gorny Altai region (Figs. 1 and 2), the pre-Cenozoic rocks are extensively exposed and cover a much greater area than the above-described northern part near Gorno-Altai city. We investigated the area along the main highway between Aktash and Chagan-Uzun

(Fig. 5). Intensive field mapping was conducted particularly in two areas; i.e., Chagan-Uzun area to the east and Kurai area to the west.

The pre-Cenozoic rocks of the southern Gorny Altai region belong to three distinct geotectonic units; i.e., the Teletsk complex, the Altai–Mongolian terrane, and the Ediacaran–Cambrian subduction–accretion complexes (Fig. 5). The Teletsk complex is a southern extension of the same unit in the northern Gorny Altai region described above (Fig. 3). The Altai–Mongolian terrane is composed of shelf-type Ediacaran–Early Cambrian sedimentary and volcanic rocks, and was intruded by the Late Devonian two-mica granites, and thermally metamorphosed under greenschist- to amphibolite-facies conditions (Buslov et al., 1993, 2001; Monie et al., 1998; Plotnikov et al., 2001). In this section, we describe the lithologic assemblages and structures of the Ediacaran–Cambrian subduction–accretion complexes.

The Ediacaran–Cambrian subduction–accretion complexes consist of (1) Ediacaran–Cambrian AC, (2) Cryogenian (late Neoproterozoic) HP complex, and (3) OP complex. They are associated with (4) Ediacaran–Cambrian island arc complex and (5) Early–Middle Devonian fore-arc sedimentary rocks (Figs. 5 and 6a). The Ediacaran–Cambrian AC is the most dominant unit both in the Kurai and Chagan-Uzun areas, and it structurally forms the lowest unit in this region. Both the OP complex and

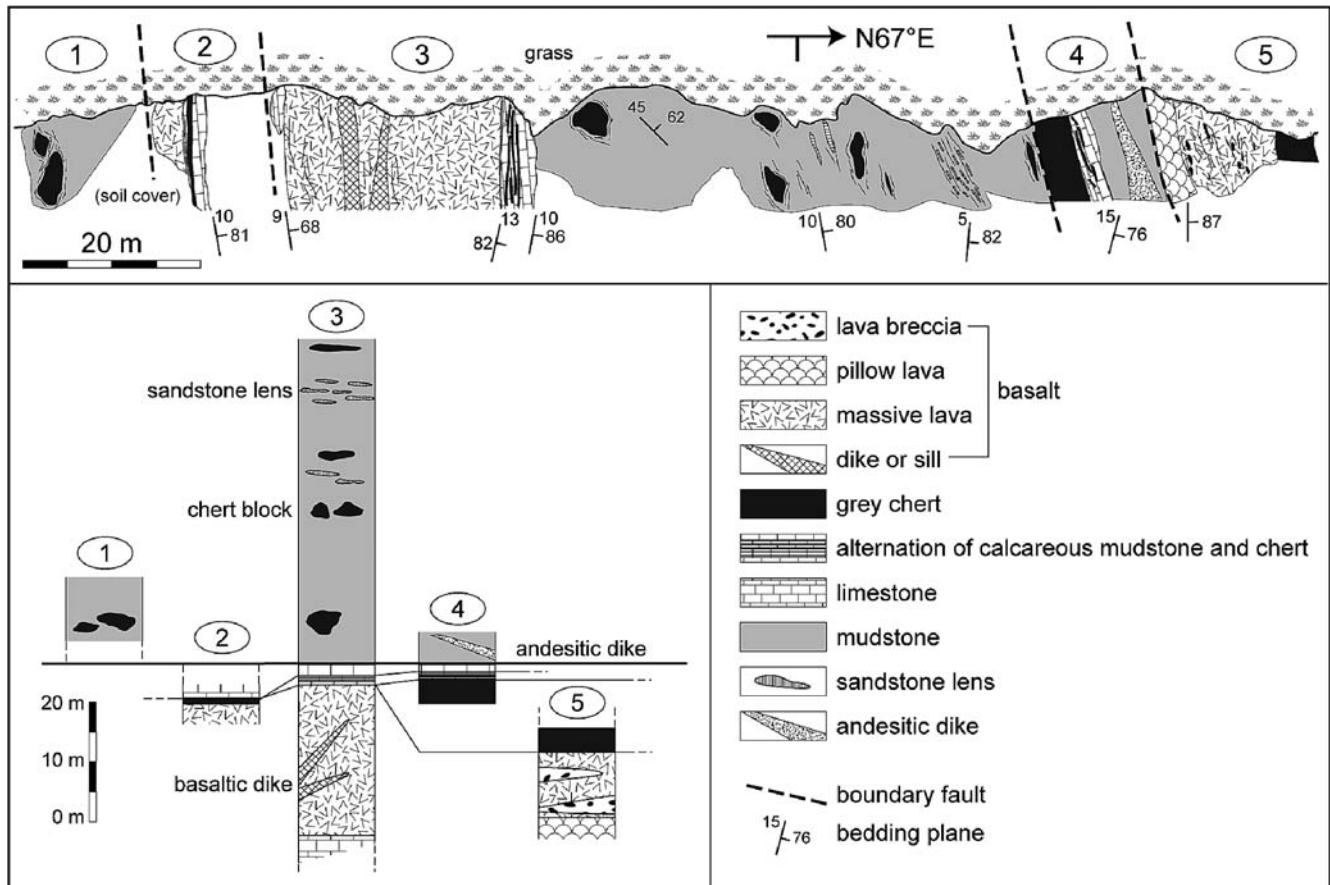


Fig. 4. Field sketch map (above) and stratigraphic columns (bottom) of the Cambrian accretionary complex near Gorno-Altai, northern Gorny Altai. Numbers of columnar sections correspond to those of fault-bounded slices in the sketch map. See text for further explanations.

HP complex occur as subhorizontal nappes, tectonically overlying the AC. The Ediacaran–Cambrian island arc complex occurs in the east, occupying the highest structural level over the OP complex. As all these rocks occur as a subhorizontal to gently east-dipping nappe, they form a piled nappe structure as a whole. These four units are unconformably overlain by the Devonian fore-arc sedimentary rocks. Later extensional and sinistral strike-slip faults disorganized the pre-existing piled nappe structure and distribution of Devonian covers into NW–SE oriented mosaic structures in the Southern Gorny Altai region (Fig. 1).

4.1. Accretionary complex

The AC is exposed extensively from the west of Chagan-Uzun village to the east of Aktash village (Fig. 5). The AC is weakly deformed and metamorphosed and comprises two major distinct units, i.e., limestone-dominant and basalt-dominant ones. The former tectonically overlies the latter by a low- to moderate-angle fault (Fig. 6a and d).

4.1.1. Limestone-dominant unit

The limestone-dominant unit is widely exposed near Chagan-Uzun (Fig. 6c and e). This unit is composed mostly

of limestone with minor amounts of basaltic lenses. These rocks occur as large exotic blocks in a matrix of cataclastic serpentinite, mostly composed of chrysotile. The serpentinite matrix is weakly foliated in the margins of the lenses, but is clearly less recrystallized than the serpentine (antigorite) schist in the HP complex. Penetrative deformation is concentrated in the matrix rather than interior of the blocks and lenses. Folds of various scales with northwest-trending axes were again refolded with northeast-trending axes (Fig. 6e).

Despite the block-in-matrix and multiple deformation overprints, the primary stratigraphy of the limestone is preserved solely within a single block (column in Fig. 6e). The limestone conformably overlies basaltic rocks that include pillowed and massive lavas and volcanoclastic rocks with lava clasts. The limestone is more than 80 m thick. The thickest section in the area comprises, from bottom to top, micritic limestone with basaltic lens, massive grey chert, bedded grey limestone with siliceous lenses, black carbonaceous limestone with siliceous lenses, and alternations of black carbonaceous limestone and siliceous rock (Fig. 6e). All the limestones lack coarse-grained terrigenous clastic material. The stratigraphic relation with the basalt and absence of terrigenous clastics suggest a mid-oceanic

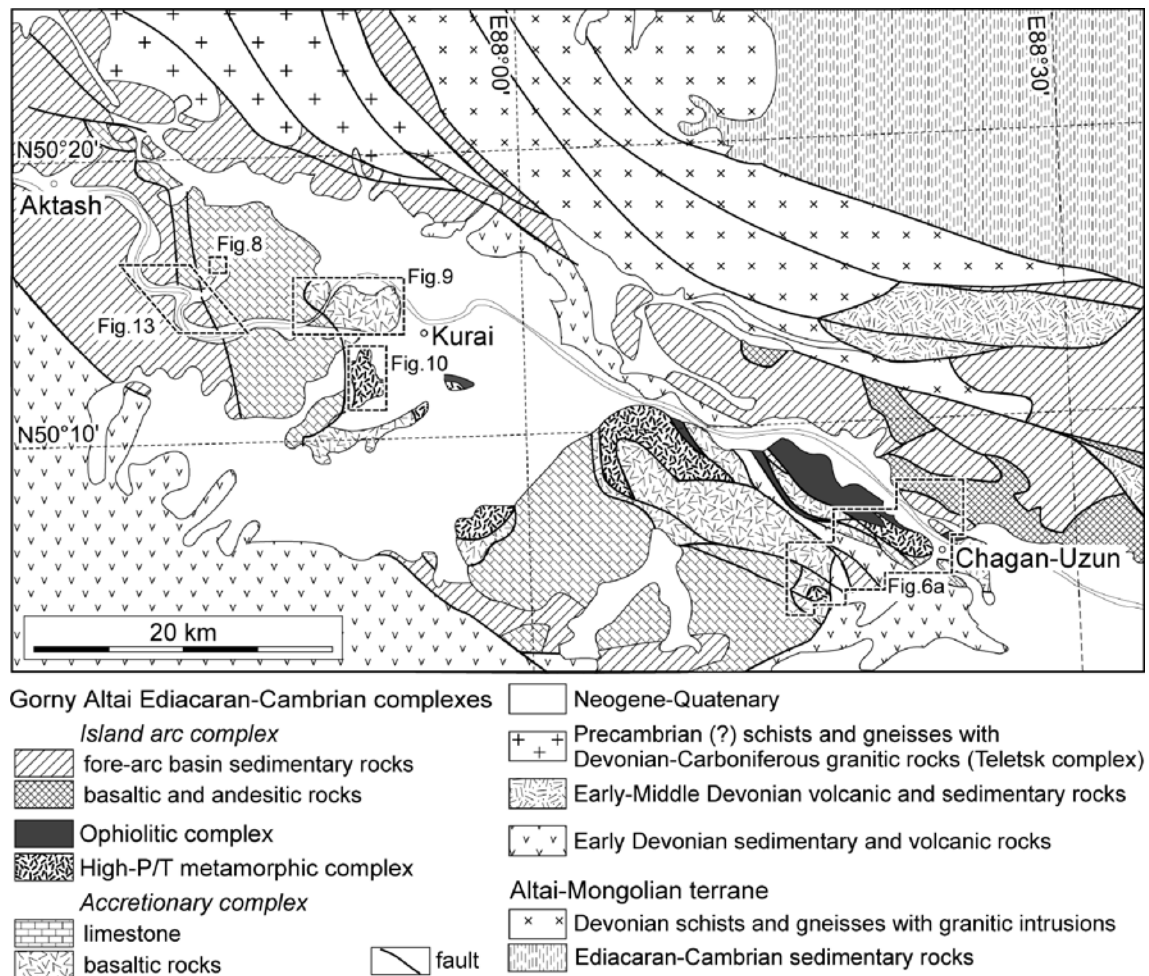


Fig. 5. Geological sketch map of the southern Gorny Altai region (modified after Buslov et al., 1993, 2002, 2004). Localities of Figs. 6a, 8–10 and 13 are also shown.

seamount/plateau origin for the limestone. Refer to Uchio et al. (2003, 2004) for details of limestone in the Kurai and Chagan-Uzun areas.

Elsewhere in the studied area (Fig. 6d), the limestone-dominant unit is composed solely of massive and bedded limestones with limestone breccia and massive grey chert. The massive and bedded limestones are often dolomitized in part and contain siliceous nodules or thin layers. Black carbonaceous limestone is occasionally associated with the bedded limestone (Fig. 8). The limestone breccia contains subangular to subrounded clasts of limestone and minor basaltic rocks and chert within a matrix of lime mud. All these limestones are micritic, and lack coarse-grained terrigenous clastic material (Uchio et al., 2003, 2004).

4.1.2. Basalt-dominant unit

The basalt-dominant unit usually occurs as a *mélange*, in which discrete or composite, variable-sized blocks or lenses of basaltic lavas and limestones occur in a fine-grained matrix of basaltic volcanoclastics (Fig. 6c and d). Lenses of amphibolite are similar to those in the OP com-

plex (Fig. 6c). In Kurai, the predominant basaltic clastics surround lenses of basaltic lava, micritic limestone, and grey chert (Fig. 9). Both the lenses and their matrix are sheared and often folded with randomly-oriented foliations. However, the primary stratigraphy can be recognized in one well-preserved large lens as shown in Fig. 6e.

The basaltic lavas are mostly massive but sometimes contain pillows, suggesting submarine eruption. These lavas are often intruded by basaltic dikes. A pillow lava with chilled margins at Kurai is directly covered by limestone, and inter-pillows are filled with micritic limestone (Uchio et al., 2003, 2004).

The basaltic clastics are poorly sorted and contain various kinds of subrounded or subangular clasts, ranging in diameter from 0.1 to 6 mm, in a fine-grained matrix with chlorite, opaque minerals and calcareous material. Mineral fragments include quartz, plagioclase, clinopyroxene, and epidote, and the lithic clasts are plagioclase-porphyritic or ophiolite basalt, micritic limestone with or without ooids, dolomitized limestone, subarkose and grey chert. In the Akkaya river area (Fig. 6d) there are lithic clasts of quartz-chlorite-phengite schist with or without garnet,

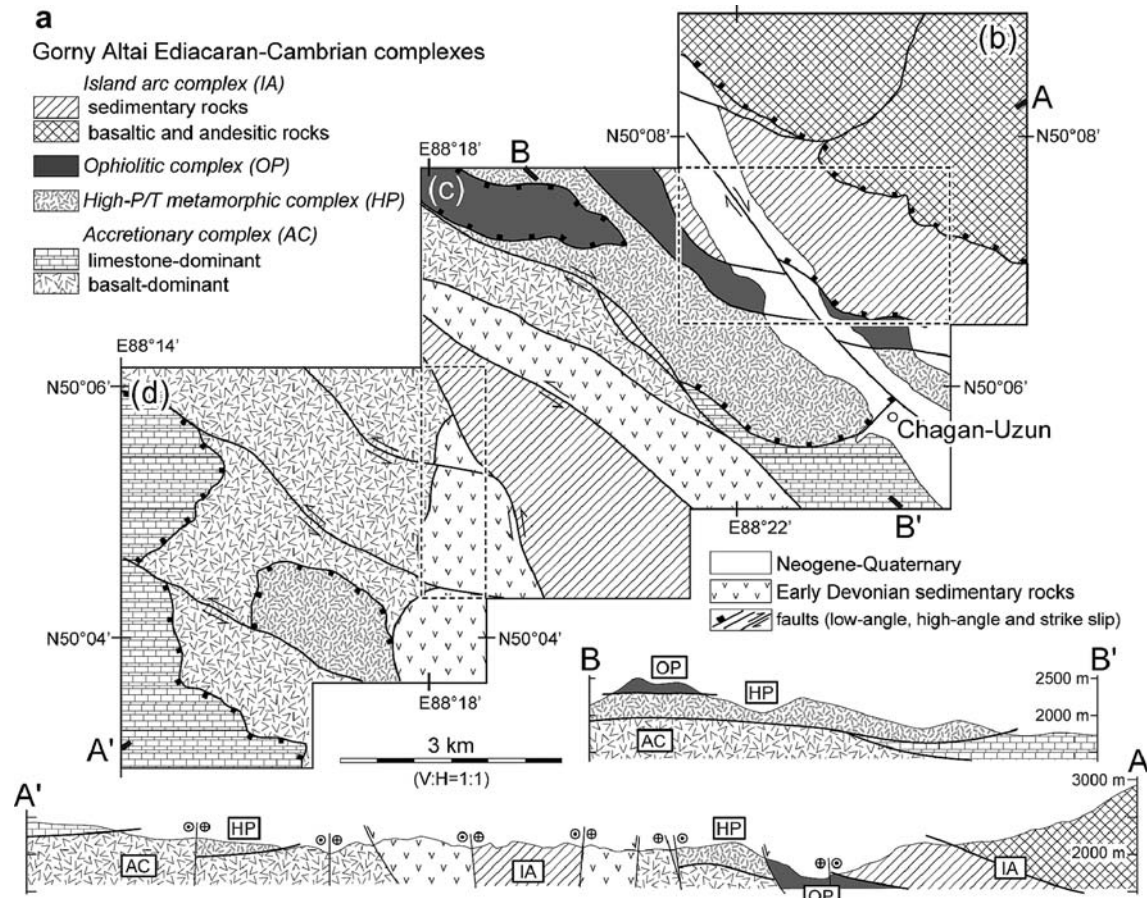


Fig. 6. (a) Geologic sketch map, and profiles A–A' and B–B' of the Chagan-Uzun area in southern Gorny Altai. Localities of labels (b), (c) and (d) are also shown. (b) Geologic map of the northern part of the Chagan-Uzun area. (c) Geologic map around Chagan-Uzun. A locality of label (e) is also shown. (d) Geologic map of the western part of the Chagan-Uzun area. Note the location of two columnar sections of Fig. 7b and c are shown. (e) Geologic map and profile of the southwestern Chagan-Uzun. See the primary stratigraphy of basalt and limestone (inset column) preserved in a large block within the Cambrian accretionary complex.

and chlorite–epidote–albite schist with or without actinolite. Many of these fragments are similar to constituent lithologies of the island arc and HP complexes, suggesting that the AC is composed of accreted oceanic materials together with olistostromes derived from the island arc and with HP complexes that had already formed and been exhumed.

The limestones in the basalt-dominant unit include laminated limestone, bedded micritic limestone, limestone breccia and massive limestone; the massive and bedded limestones often have a depositional contact with underlying pillow lavas (Uchio et al., 2003, 2004). The “Vendian-type” stromatolite and microphytolite in the bedded limestone is comparable with those in a Siberian continental shelf facies and constrain the depositional age of the limestone. In addition, the earliest Cambrian microphytolite (*Epiphyton*) occurs in the matrix micrite of the limestone breccias (Buslov et al., 1993). The black carbonaceous limestones associated with the bedded limestone in the limestone-dominant unit (Fig. 8) have a bulk Pb–Pb isochron age of c.570 Ma (middle Ediacaran) that suggests the depositional age (Uchio et al., 2001). A massive limestone in the basalt-dominant unit (Figs. 7a and 9) has a

bulk Pb–Pb isochron age of 598 ± 25 Ma (Nohda et al., 2003; Uchio et al., 2004). These radiometric ages are consistent with the above-mentioned fossil ages of the bedded limestones.

All these limestones lack coarse-grained terrigenous clastic material. Based on their lithological features in comparison with present-day sediments (Cook and Mullins, 1983; Halley et al., 1983), Uchio et al. (2003, 2004) classified these limestones roughly into three groups, i.e., (1) massive, (2) bedded or brecciated, and (3) thinly laminated, explaining their different depositional environments within the same mid-oceanic setting as follows (Fig. 7). The massive limestone containing stromatolites and ooids directly overlies pillowed basalt and thus was probably deposited on top of an ancient mid-oceanic topographic high (seamount or plateau). The bedded or brecciated limestone often shows slump structures. The limestone breccia includes poorly-sorted subangular clasts of micritic limestone, basaltic rocks and chert. This type of limestone was probably formed as submarine slide or sediment-gravity-flow (debris flow) deposits that accumulated on the slope of the mid-oceanic topographic high. The thinly

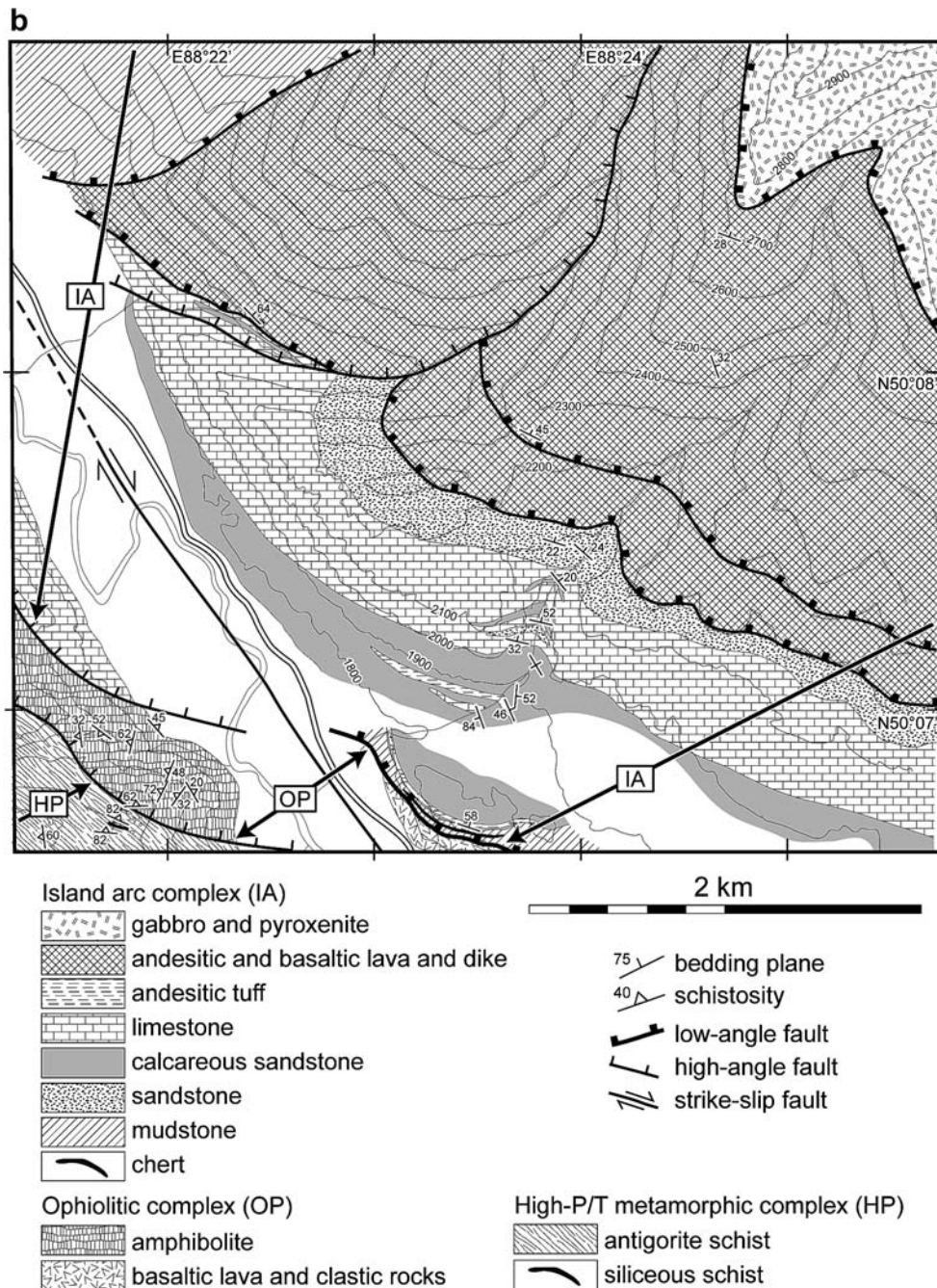


Fig. 6 (continued)

laminated micritic limestone is comparable with an allodapic limestone (limestone turbidite) deposited at the base of slope of a mid-oceanic topographic high.

4.1.3. Accretion age

In general, rocks of the AC in the Gorny Altai region are not fossiliferous; even limestone is mostly barren of fossils. It is difficult to identify a precise age of AC formation by using an oceanic plate stratigraphy (Matsuda and Isozaki, 1991; Isozaki, 1996) in such a chert-poor AC. Nonetheless, there are some clues to constrain the AC formation age. For example, some limestone blocks yield “Vendian-

type” stromatolite and the earliest Cambrian microphytolite (Afonin, 1976; Buslov et al., 1993). In addition, a bulk Pb–Pb isochron age of 570–598 Ma (middle Ediacaran) was determined for the basal limestone immediately above the pillowed OPB (Uchio et al., 2001, 2004; Nohda et al., 2003). These data indicate that the limestone ranges in age from at least middle Ediacaran to the earliest Cambrian. Thus, the formation of the AC at a trench should have been younger than the ages of oceanic rocks. On the other hand, the conglomerate of the Early–Middle Cambrian arc complex unconformably covering the AC can cap the uppermost age limit. Although the depositional

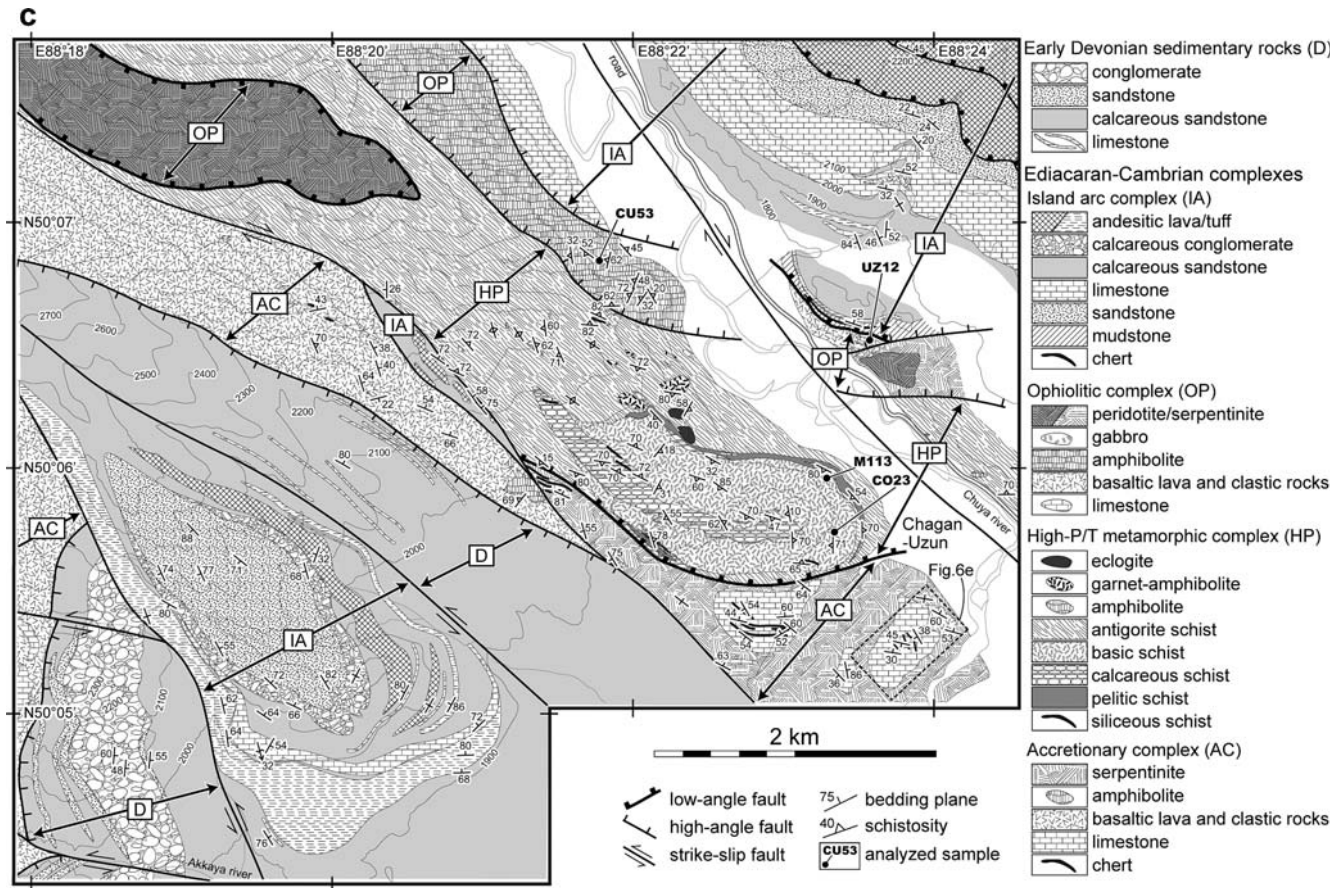


Fig. 6 (continued)

age is not yet certain, the cover conglomerate contains a limestone clast with Early Cambrian Archaeocyathids (Buslov et al., 1993, 1998) probably derived from the AC. Thus at present, we consider that the AC formed in an ancient trench some time in the early–middle Cambrian.

4.2. High-*P/T* metamorphic complex

Near Chagan-Uzun village (Fig. 6a and c), a HP complex occurs as a subhorizontal thrust sheet bounded by low-angle faults with overlying peridotite of the OP complex and an underlying serpentinite with limestone lenses of the AC. It is composed of well-recrystallized and -foliated serpentinite (antigorite schist), with lenticular intercalations of metabasites, and pelitic, calcareous and siliceous schists. The metabasites include eclogite, garnet-amphibolite, amphibolite and lower-grade metabasalt. The schists are highly deformed in close and tight folds at various scales.

Recent Ar–Ar dating of amphiboles in the Chagan-Uzun eclogite indicate that most of the previously reported K–Ar ages (535–567 Ma; Buslov and Watanabe, 1996) were mixing ages affected by later events. The Ar–Ar plateau ages of 627–636 Ma (late Cryogenian, Late Neoproterozoic) represent a more reliable age for the eclogite-facies metamorphism (Buslov et al., 2001, 2002).

Additionally, amphiboles in the Kurai GH metabasite that preserve its metamorphic peak assemblage have an Ar–Ar plateau age of 629 ± 9 Ma (Fig. 11). On the other hand, a highly deformed and retrograded greenschist-facies metabasite does not yield a plateau age (Buslov, unpublished data; Fig. 11); its total gas age of c.564 Ma suggests that the younger K–Ar ages were affected by the retrograde metamorphism.

Eclogites occur in lenses and layers less than a few centimeters thick, are partly amphibolitized, but preserve a granoblastic texture with a mineral assemblage of garnet + omphacite + barroisite + epidote + quartz + rutile. Garnet-amphibolites occur as intercalations within the eclogite bodies and as discrete intercalations in antigorite schist; they contain an assemblage of garnet + barroisite + epidote + titanite $\pm$ rutile $\pm$ winchite $\pm$ quartz $\pm$ albite $\pm$ phengite. Elongated barroisite, prismatic epidote and rutile lie on schistosity planes. Rare amphibolite lenses are heterogeneously foliated and contain assemblages of hornblende + albite + quartz + titanite $\pm$ epidote. Green hornblendes are porphyroblastic and their grain boundaries are filled with albite and minor quartz. These higher-grade metabasites were partly retrograded under greenschist-facies conditions, as indicated by replacement of hornblende-, barroisite- and omphacite-rims by actinolite, and garnet rims by chlorite. Lower-grade metabasalts are heteroge-

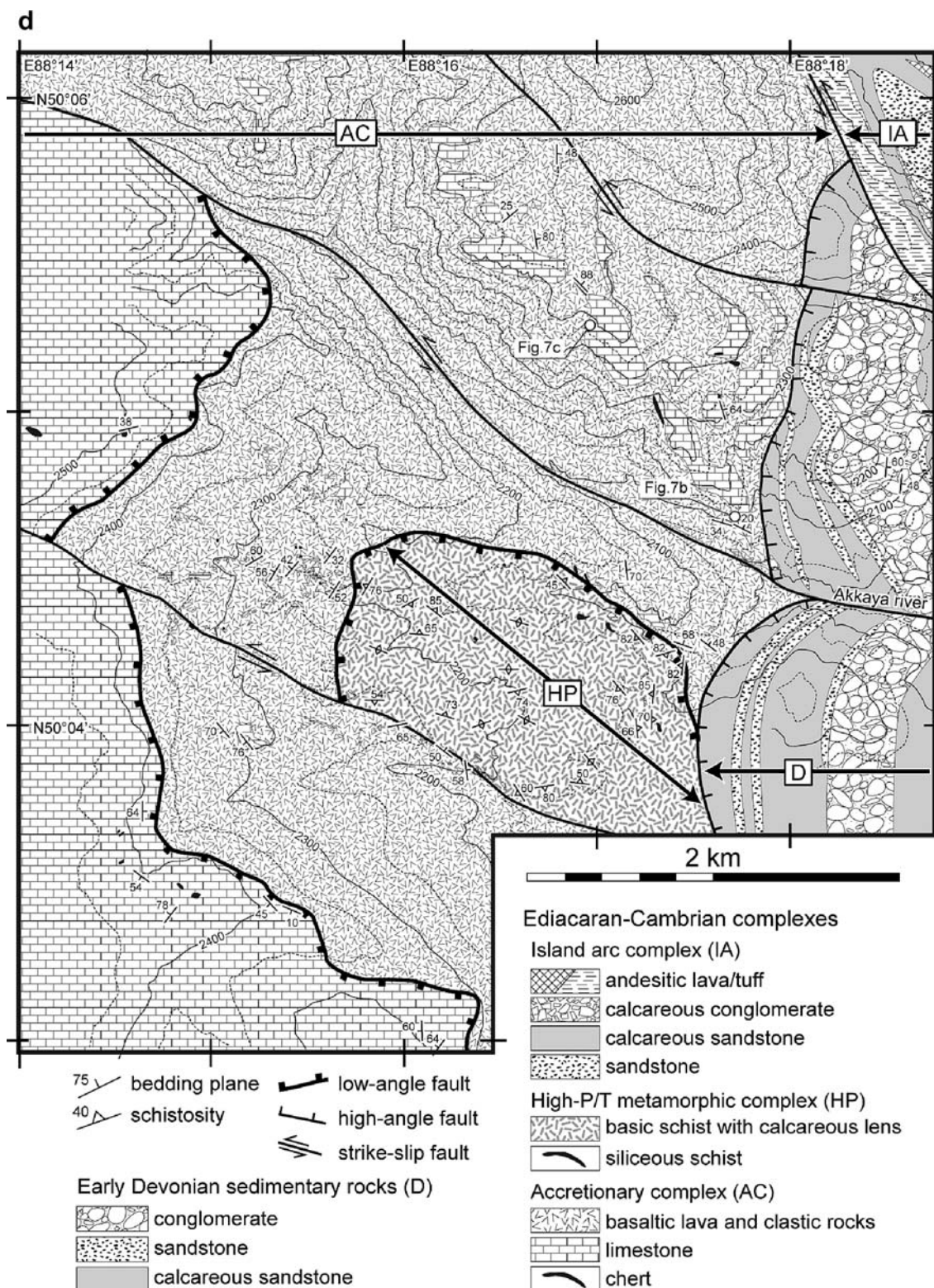


Fig. 6 (continued)

neously recrystallized. Some are less foliated and contain relict igneous clinopyroxenes. Their characteristic assemblages are; actinolite + chlorite + pumpellyite + stilpnomelane, magnesioriebeckite + actinolite + chlorite, and phengite + chlorite + epidote; all these contain quartz, albite and titanite.

Near Akkaya river to the southwest of Chagan-Uzun village (Fig. 6d), the HP complex is composed of metabasites with minor lenses of siliceous and calcareous schists. It occurs as a slab bound by a low-angle fault from underlying basaltic rocks of the AC (see A'–A cross-section in Fig. 6a). The HP complex also occurs at Kurai (Fig. 10)

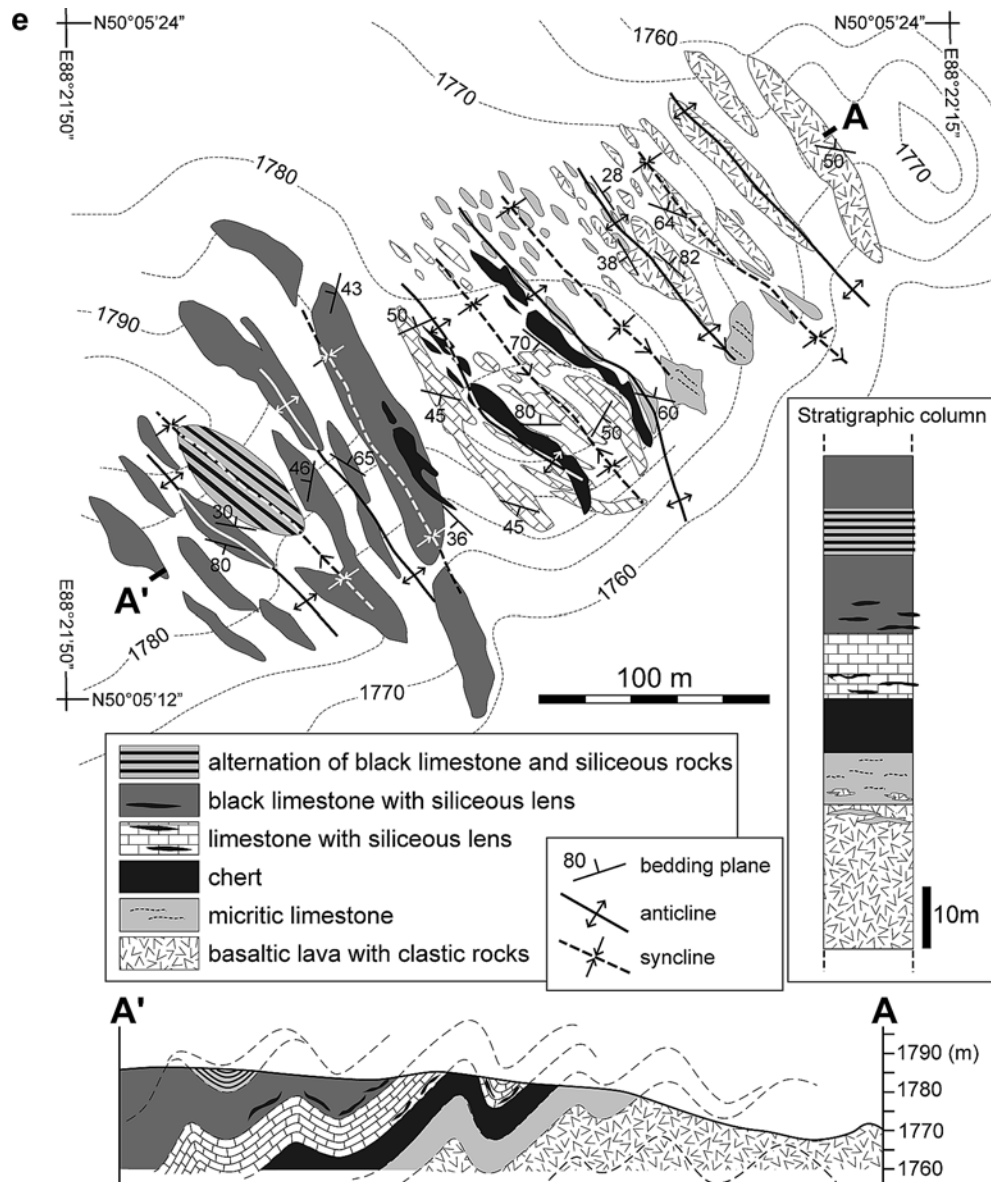


Fig. 6 (continued)

and has a similar lithology as that in the Akkaya river area to the southeast. The metabasites in these two localities frequently preserve igneous textures, suggesting that they were basaltic lavas, clastic rocks, and dolerites in origin.

The metabasites change in mineral assemblage with increasing metamorphic grade from actinolite + chlorite ± hematite, through hornblende + chlorite ± ilmenite ± actinolite, to barroisite + garnet ± actinolite ± hornblende ± rutile; all these include epidote + quartz + albite + titanite ± phengite ± calcite. The actinolite + chlorite metabasites are fine-grained and show weak schistosity formed by acicular actinolites with chlorite flakes. Less recrystallized samples preserve basaltic clasts ranging in size from 1 to 5 mm. Hornblende-bearing metabasites are generally foliated and amphiboles, epidotes and chlorites lie on the schistosity planes. The garnet-bearing metabasites are foliated, and barroisite and epidote often form a mineral linea-

tion. Fine-grained garnets range in size from 0.1 to 0.3 mm. Micro-folds and related axial cleavages are developed in well-foliated samples. The garnet-bearing metabasites are often gneissose with garnet porphyroblasts ranging from 1 to 5 mm in diameter. These higher-grade metabasites are overprinted by greenschist-facies mineralogy, giving rise to actinolites at hornblende- and barroisite-rims, and chlorites at garnet rims. Metadolerites contain relict plagioclase, clinopyroxene phenocrysts, and minor magnetite. Brown hornblendes and biotites occur around clinopyroxene phenocrysts and are further replaced by chlorite, titanite, epidote and actinolite. Plagioclases are saussuritized and occur with calcite, phengitic mica and epidote. Pale-green hornblende, actinolite, chlorite and epidote also occur as matrix constituents.

The actinolite + chlorite metabasite is the most common rock type in the HP complex; the hornblende- and

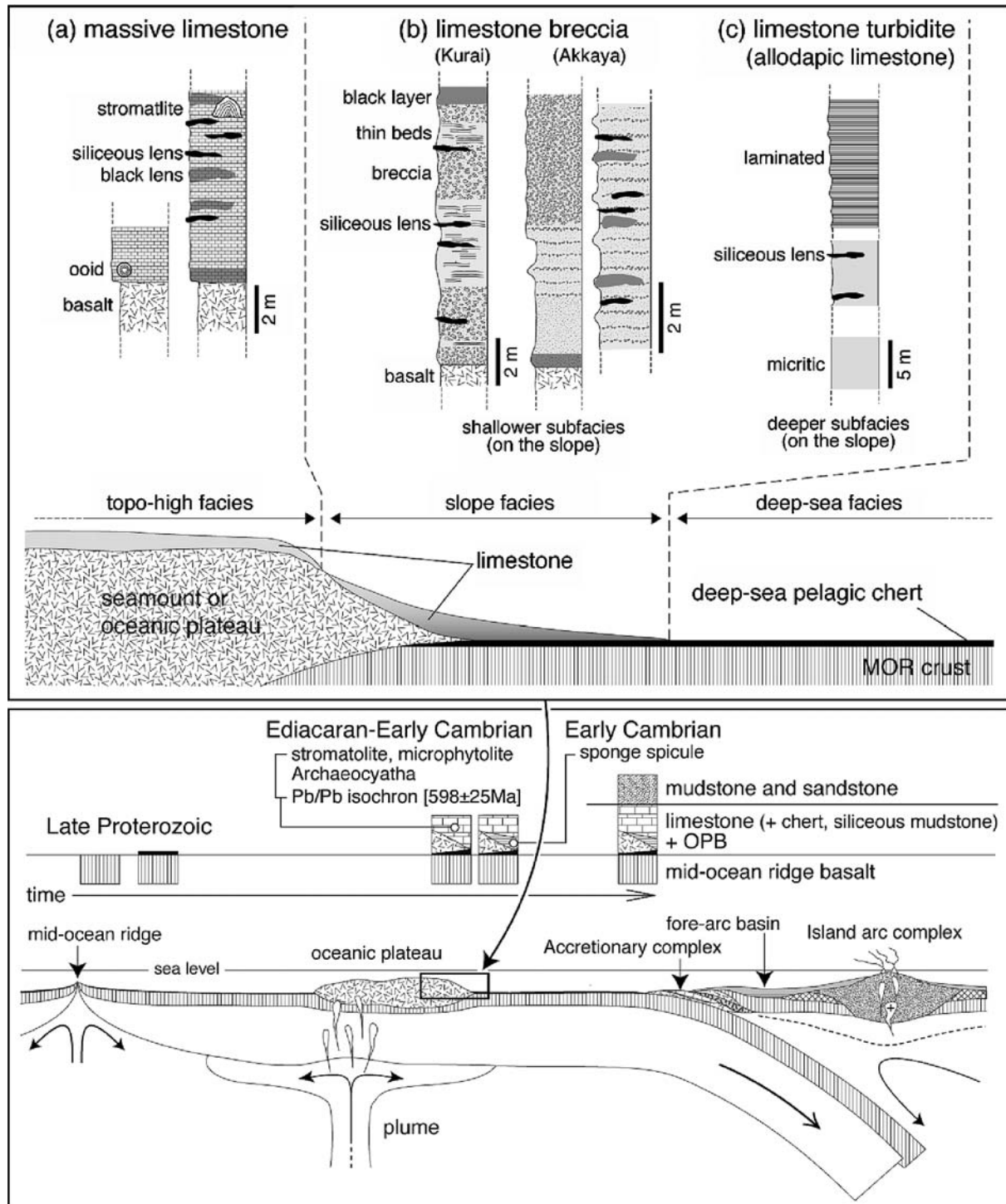


Fig. 7. Three distinct facies of limestone in the Kurai and Chagan-Uzun areas deposited on and around an ancient mid-oceanic plateau and their sedimentary settings, modified from Uchio et al. (2004) (above), and a schematic diagram showing the pre-accretion primary stratigraphy of oceanic sediments, modified from Isozaki et al. (1990) (bottom). MOR, mid-ocean ridge; OPB, oceanic plateau basalt.

garnet-bearing ones are predominant in the eastern Akkaya river area, suggesting an eastward increase of metamorphic grade. In the Kurai area (Fig. 10b), mineral zones of actinolite (ACT), hornblende (HBL), garnet-barroisite (GB) and garnet-hornblende (GH) show a symmetrical pattern with the highest grade in the central part of the complex (Ota et al., 2002).

4.3. Ophiolitic complex

The OP complex consists of three fault-bound blocks tectonically sandwiched between the island arc complex and the HP complex (Fig. 6a and c). The western and central blocks are mainly composed of peridotites and amphibolites, respectively, while the southeastern block

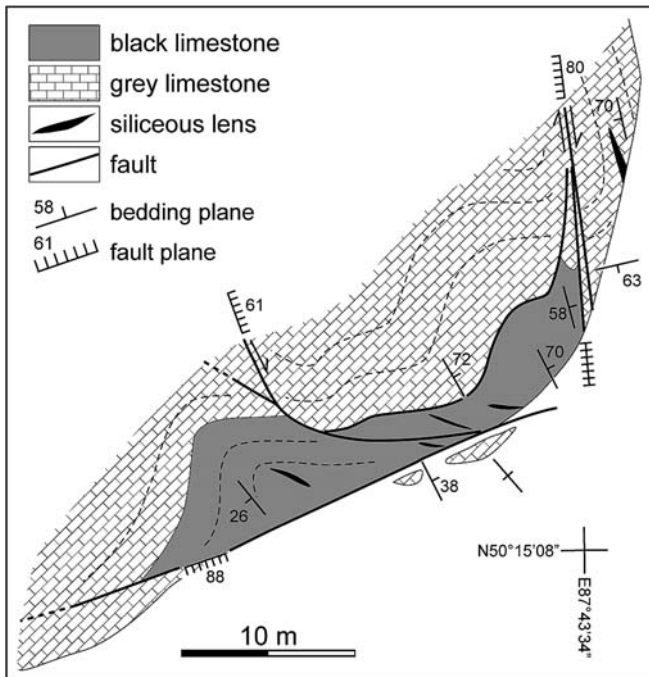


Fig. 8. Geologic sketch map of Baratal valley, near Aktash, southern Gorny Altai region.

consists of basaltic rocks with intercalations of limestone, and peridotite (Figs. 6c and 12). The total thickness of the OP complex is less than 250 m. Thus this unit is a dismembered ophiolite that lacks several parts of the standard definition of an ophiolite, such as sheeted dikes and bedded chert. Peridotites are intruded by dikes of basalt, gabbro

and pyroxenite, which are in places deformed into lenticular masses. Massive amphibolites and basaltic lavas are commonly intruded by basaltic dikes. Buslov and Watana-be (1996) reported a K–Ar age of 523 ± 23 Ma (early Cambrian) for hornblendes separated from the amphibolite.

Peridotites include harzburgite and dunite that are highly deformed and serpentinized near the boundary faults. Primary minerals in the peridotites are olivine, orthopyroxene, spinel and clinopyroxene. Olivines are altered to serpentine with tiny magnetites. Coarse-grained orthopyroxenes have dusty cores with clinopyroxene lamellae, and are often replaced by anthophyllite along their cleavages. Minor clinopyroxenes and red or dark brown spinels occur as granular crystals in the matrix or along grain boundaries.

Gabbros, occurring as lenses and dikes in peridotites, are less than a few meters wide. They are mainly composed of intermediate- or coarse-grained plagioclase and clinopyroxene, but recrystallized to chlorite, prehnite, calcite and quartz along shear planes and in fractures. Plagioclases are partly altered to fine-grained calcite, epidote and mica. Clinopyroxenes are often replaced by brown or green hornblende, actinolite and chlorite. Rodingites, mostly composed of anhedral diopside and epidote, are frequently associated with the gabbro lenses. Pyroxenites commonly occur as dikes, a few meters wide, in the peridotites; they are composed of coarse-grained euhedral orthopyroxenes and rare clinopyroxenes; minor olivines are partly serpentinized.

Amphibolites, composed of green hornblende, plagioclase, and minor titanite, magnetite, epidote and ilmenite, are recrystallized, and exhibit a weak foliation. In some

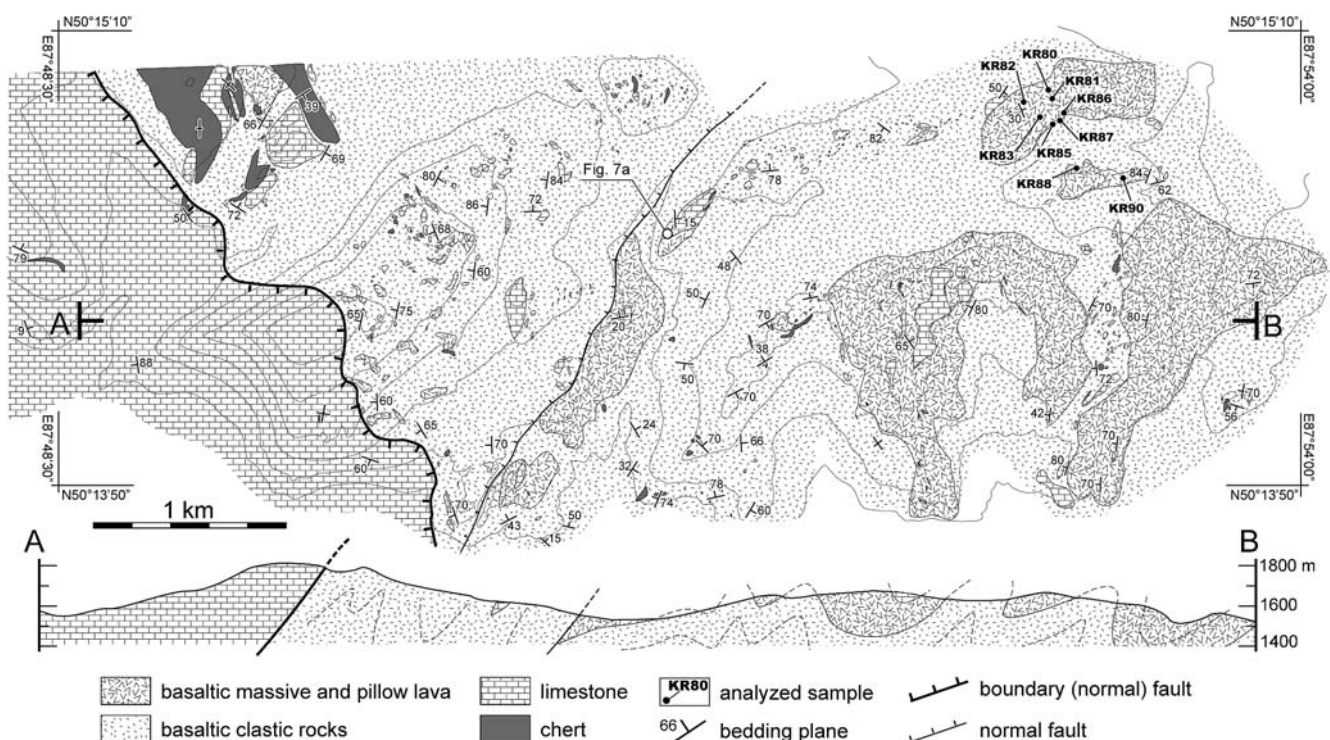


Fig. 9. Geologic map and profile of western Kurai, southern Gorny Altai region. The locality of the columnar section in Fig. 7a is indicated.

amphibolites, shear planes contain chlorite, colorless actinolite, and veinlets of prehnite, calcite and quartz.

Basaltic rocks occur as pillowed or massive lavas and dikes. Most of the lavas are porphyritic with clinopyroxene and plagioclase phenocrysts; some are ophitic. Clinopyroxenes are often fractured and are partly replaced by titanite, epidote, and chlorite. Plagioclases are partly altered to tiny calcite, phengitic mica, titanite, and pumpellyite. The groundmass is composed of plagioclase, clinopyroxene, magnetite, secondary calcite, chlorite, quartz, pumpellyite, and veinlets containing calcite, quartz, albite, prehnite, or epidote. Basaltic dikes in the basaltic lavas and the amphibolites are less altered, and have similar constituents and textures to the basaltic lavas. On the other hand, most dikes in the peridotites are highly recrystallized; clinopyroxenes are replaced by brown or green hornblende and grain boundaries are filled with epidote, chlorite and calcite. Less altered dikes, rarely recognized in the peridotites, are plagioclase-porphyritic; plagioclase, clinopyroxenes, and secondary titanite, chlorite and calcite occupy their intergranular or intersertal groundmass.

The limestones, which are interbedded with basaltic lavas, are grey micrites composed of recrystallized calcite with minor carbonaceous material.

The above petrographic data indicate that basaltic dikes in the lavas were recrystallized under prehnite–pumpellyite facies conditions, while the basaltic dikes, gabbroic dikes and lenses in the peridotites, and the peridotites themselves, were metamorphosed under greenschist to amphibolite facies conditions (Fig. 12). Such low-pressure recrystallization of prehnite–pumpellyite type in the amphibolite facies may have been due to hydrothermal alteration at a mid-ocean ridge. However, the hydrothermal alteration is unlikely to generate amphibolites with a foliation, implying penetrative deformation when the hot ophiolite was emplaced.

On the other hand, the basaltic dikes in the amphibolites, and the plagioclase-porphyritic basalt and the pyroxenite dikes in the peridotites, are less metamorphosed than their host rocks. Such a difference in metamorphic grade suggests that these dikes would have intruded into their host rocks after the low-pressure type metamorphism. In addition, some dikes intruding the peridotites are petrochemically close to rocks of the calc-alkaline island arc series (Buslov et al., 1993, 2002). After low-pressure metamorphism, the OP complex was emplaced into the subduction zone and overprinted by island arc volcanism.

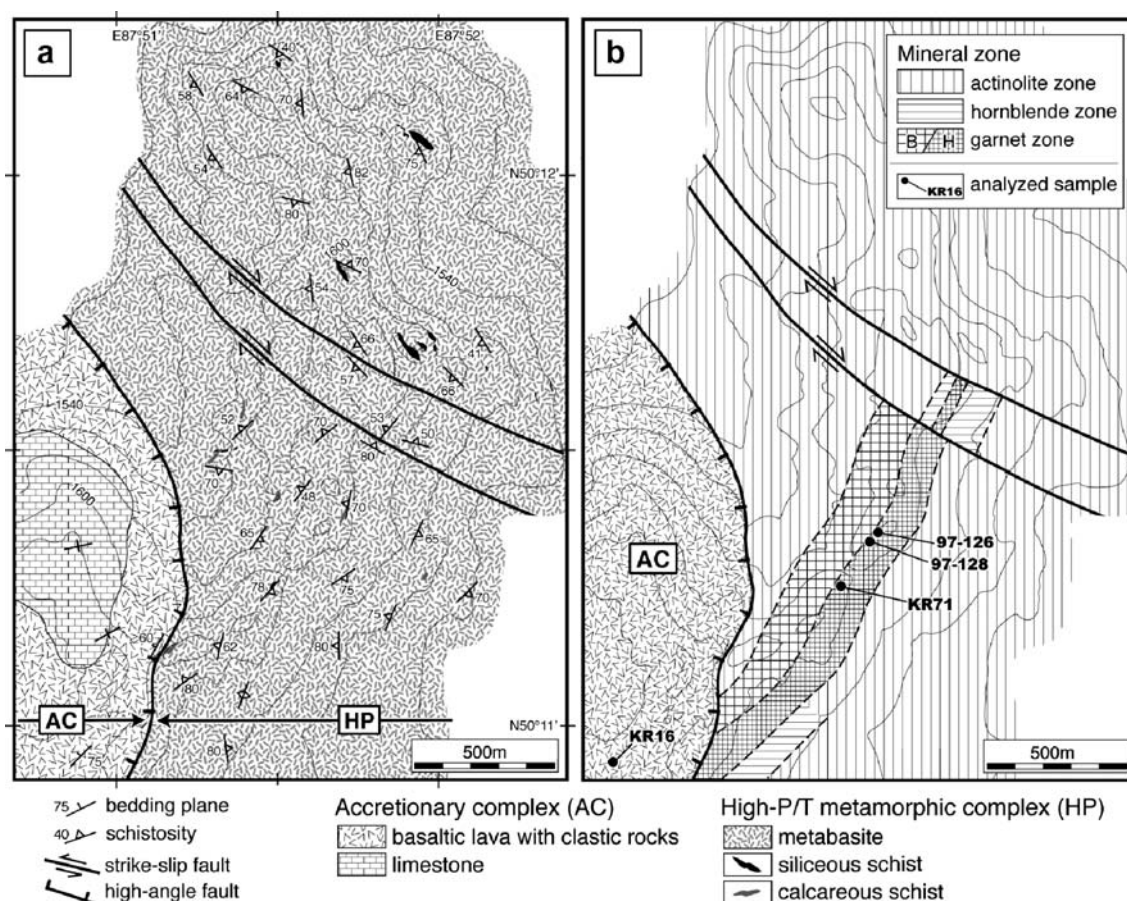


Fig. 10. Geologic map of the southwestern Kurai (a), showing the mineral zones of the high-*P/T* metamorphic complex (b), modified after Ota et al. (2002). Zone B: garnet-barroisite subzone; Zone H: garnet-hornblende subzone.

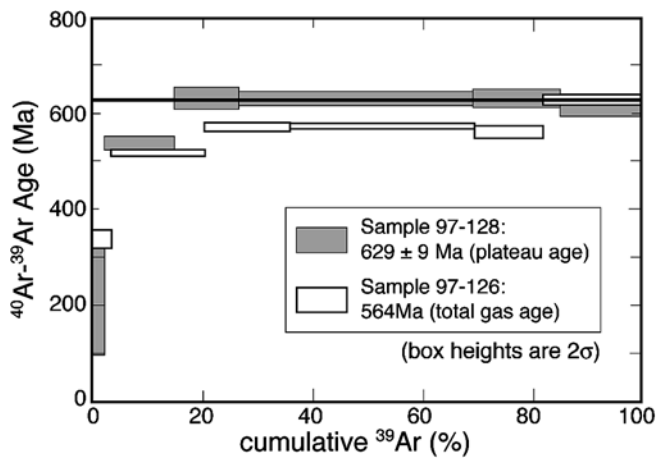


Fig. 11. Results of amphibole Ar–Ar dating for metabasites from the garnet-hornblende subzone in Kurai, southern Gorny Altai (Analyzed by Alexey Travin, the Institute of Geology and Mineralogy, Siberian branch, Russian Academy of Sciences, Novosibirsk). Sample 97-126, massive metabasite preserving the metamorphic peak assemblage; Sample 97-128, highly deformed and retrograded metabasite. A thick horizontal line indicates a plateau age for sample 97-126. See text for further explanations.

4.4. Island arc complex

The island arc complex in the northeast of Chagan-Uzun village is characterized by an imbricated nappe structure (Fig. 6a and b), in which tholeiite–boninite rocks overlie calc-alkaline rocks on low- to moderate-angle faults, or alternatively they occur with calc-alkaline series rocks.

The tholeiite–boninite series rocks in the upper part of the structure include volcanogenic sedimentary rocks, a dike-sill unit, sheeted dikes, and a layered gabbro–pyroxenite (Dobretsov et al., 1992; Buslov et al., 1993; Simonov et al., 1994). The gabbro–pyroxenite is composed of layered gabbro, cumulative clinopyroxenite, wehrlite and serpentine, and is intruded by dikes of quartz–diorite and plagiogranite of the calc-alkaline series (e.g., Buslov et al., 1993). Clinopyroxenes from the clinopyroxenite have been dated at 647 ± 80 Ma (Ponomarchuk et al., 1993; Dobretsov et al., 1995). The dike, sills and sheeted dikes include dolerite, gabbro and boninitic rocks; the occurrence of boninitic rocks within the sheeted dikes suggests their origin in a spreading centre during the formation of a primitive island arc (Dobretsov et al., 1992; Simonov et al., 1994). In the boninitic rocks, porphyritic textures are associated with rare phenocrysts of olivine, clinopyroxene and chromite, but no orthopyroxene has been found. Previous petrochemical studies on boninite series rocks (Simonov et al., 1994; Buslov et al., 2002) indicate that they are comparable with western Pacific boninitic rocks (e.g., Crawford et al., 1989; Hickey-Vargas, 1989).

Calc-alkaline rocks in the lower part of the structure consist of andesitic lava, tuffaceous rocks, calcareous sandstone, limestone, mudstone, sandstone and minor chert. They are juxtaposed by a fault against the OP complex

(Fig. 6a and b); mudstone near the boundary fault is tightly folded.

In the northeast of Akkaya river (Fig. 6c), the sedimentary unit of the island arc complex, associated with the calc-alkaline rocks, occurs between Devonian sedimentary rocks and the AC with strike-slip fault contacts. The sedimentary unit is composed, in ascending order, of calcareous sandstone, calcareous conglomerate, and sandstone. The calcareous sandstone forms interbeds within andesitic lava and tuff, and limestone, and in places alternates with thin mudstone and greywacke layers. The andesitic lava preserves a porphyritic texture, although phenocrysts and matrix constituents have been altered under subgreenschist-facies conditions. A calcareous conglomerate contains fragments of micritic limestone, andesite, and quartzite in a calcareous matrix.

To the southeast of Aktash village (Fig. 5), the sedimentary unit is mainly composed of massive arkose sandstone, interbedded calcareous sandstone and mudstone, conglomerate, and siliceous mudstone (Fig. 13). A massive arkose contains angular lithic clasts of limestone and plagioclase–porphyritic volcanic rocks, together with angular and sub-angular clasts of quartz, feldspars and clinopyroxene, and occasionally shows rhythmic layering (Fig. 13a). The arkose gradually changes upward into calcareous sandstone and mudstone (Fig. 13b). The conglomerate contains subangular to subrounded pebbles and granules of limestone, plagioclase–porphyritic volcanic rocks, andesitic tuff, quartz, feldspars, amphibole and clinopyroxene, together with rare rounded pebbles of serpentinite, pyroxenite, gabbro, amphibolite, quartz–schist, and granitic rocks (Buslov et al., 1993). A siliceous mudstone occurs in the upper horizons where massive arkose gradually changes from green to red siliceous mudstone in an upward-fining sequence (Fig. 13b). Judging from their lithological characteristics, the sedimentary unit of the island arc complex probably accumulated in a fore-arc basin as suggested by Buslov et al. (1998, 2002).

Some sedimentary rocks contain informative fossils for age discrimination; the red-colored siliceous mudstone contains Middle and the Late Cambrian sponge spicules. Limestone clasts in the conglomerate contain Middle Cambrian trilobites and brachiopods, and Early Cambrian archaeocyathids (Buslov et al., 1993, 1998). These ages indicate that the sedimentary unit of the island arc complex was probably deposited in the Middle–Late Cambrian and arc volcanism was active at that time.

5. Petrochemistry of basaltic protoliths in the subduction–accretion complex

In the Gorny-Altai subduction–accretion complex, basaltic rocks with various geological and petrological features are widespread (Buslov et al., 1993, 2002; Utsunomiya et al., 1998). In order to extract igneous petrochemical data from these rocks, we selected less

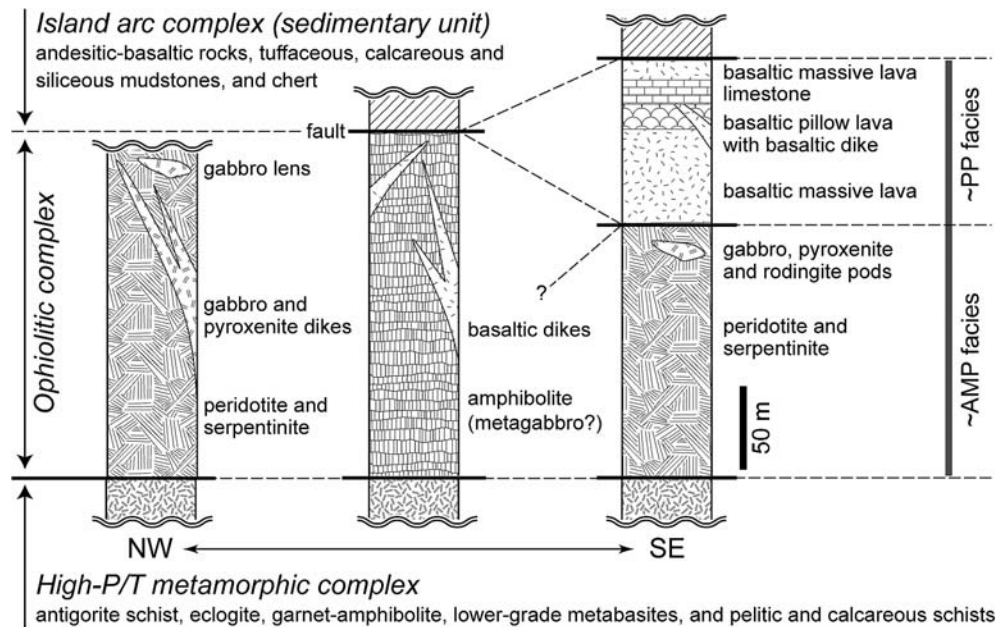


Fig. 12. Stratigraphic columns of the ophiolitic (OP) complex in Chagan-Uzun. The OP complex is composed of three fragments (see also Fig. 6c). Basaltic rocks are recrystallized under prehnite–pumpellyite (PP) facies, and amphibolite (metagabbro) and peridotite under amphibolite (AMP) facies conditions. See text for further details.

deformed and metamorphosed lavas from the AC, the HP and the OP complexes (Figs. 6c, 9 and 10). We then determined major and rare earth element (REE) compositions of relict igneous clinopyroxenes in selected samples. For some of the selected samples, whole-rock major, trace, and rare earth elements were also analyzed. Even in the amphibolite section of the OP complex (Fig. 6c) and the high-grade mineral zone of the HP complex (Fig. 10), some samples are less recrystallized and partly retain their basaltic textures. Thus, their whole-rock compositions were also analyzed. Localities of analyzed samples are shown in Figs. 6c, 9 and 10, and representative analyses of clinopyroxene and whole-rock compositions are listed in Tables 1 and 2.

5.1. Petrography of analyzed samples

Basaltic samples selected from the HP complex are porphyritic and hyalocrystalline. Phenocrysts include clinopyroxene and plagioclase; most of the clinopyroxenes survived metamorphic recrystallization, although they are often fractured and rarely replaced by chlorite along their margins. Plagioclase phenocrysts and groundmass are recrystallized to pumpellyite–actinolite facies minerals.

Lavas from the AC and the OP complex contain porphyritic and ophitic basalts, with very rare aphyric basalts. The porphyritic basalts contain phenocrysts of clinopyroxene and plagioclase. The clinopyroxenes are often fractured and partly replaced by secondary minerals, but preserve fresh cores. Plagioclase phenocrysts and groundmass are commonly replaced by secondary minerals such as calcite, albite, chlorite, phengitic mica, titanite, pumpellyite and opaque minerals, indicating less

than prehnite–pumpellyite facies conditions. In the ophitic basalts, the igneous mineral assemblage is clinopyroxene + plagioclase + magnetite. Clinopyroxenes survived the secondary alteration, but plagioclases were albitized. In some samples, brown or brownish green hornblendes, rimmed by colorless actinolite and chlorite and by irregular-shaped epidote, also occur in the matrix; they probably formed by hydrothermal alteration on the ocean floor. Such highly altered rocks were excluded from samples for whole-rock analysis.

5.2. Major and rare earth element compositions of relict igneous clinopyroxenes

Major elements and REE compositions of relict igneous clinopyroxenes were determined with the electron probe microanalyser (EPMA) and with the secondary ion mass spectrometer (SIMS), Cameca ims 3f, respectively, housed at the Tokyo Institute of Technology. We followed the SIMS analysis procedure of Wang and Yurimoto (1994).

Analyzed clinopyroxenes are augites with Mg-numbers ($Mg\# = 100 \times Mg / (total\ Fe + Mg)$) ranging from 57.9 to 89.3; Al contents of the clinopyroxenes from the HP and the OP complexes show negative and positive correlations against their Mg# and Ti contents, respectively (Fig. 14). Such compositional trends for the clinopyroxenes contrast with those in mid-ocean ridge basalt (MORB), with decreasing Al contents at higher degrees of fractional crystallization.

In terms of chondrite-normalized REE patterns (Fig. 15), the relict clinopyroxenes from the HP complex exhibit convex patterns with slight depletions in both light- and heavy-REEs. Such REE patterns are similar to those

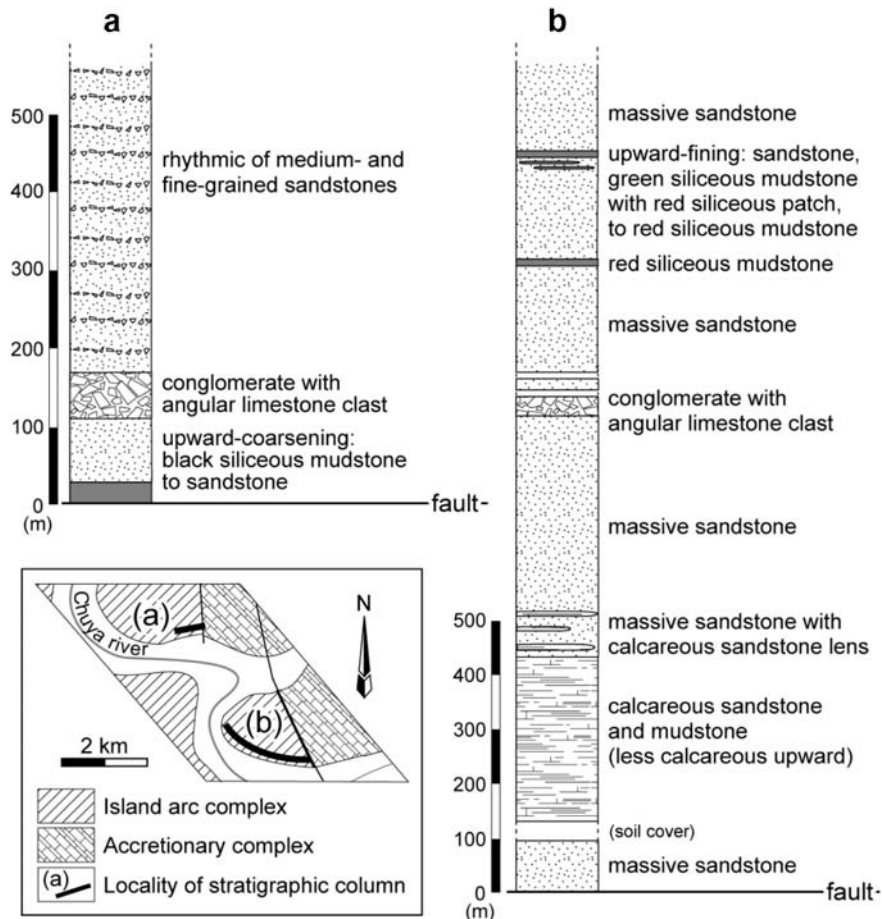


Fig. 13. Stratigraphic columns of sedimentary rocks of the island arc complex bound with the accretionary complex to the southeast of Aktash village. An inset shows a simplified unit map, with localities of columns (a) and (b).

of clinopyroxenes from Hawaiian oceanic island tholeiites (OIT). On the other hand, clinopyroxenes from the AC and the OP complex have REE patterns with high depletion of light-REEs, which are similar to those of MORB clinopyroxenes.

5.3. Major, trace and rare earth element compositions of whole-rocks

Whole-rock compositions were determined using X-ray fluorescence spectroscopy (XRF) at the Tokyo Institute of Technology for major elements, and at the Ocean Research Institute, the University of Tokyo, for trace elements; REEs compositions were obtained with the inductively coupled plasma mass spectrometer (ICPMS) at the Tokyo Institute of Technology. Analytical methods and uncertainties in these determinations are after Goto and Tatsumi (1994) and Hirata et al. (1988) for the XRF and the ICPMS, respectively.

In MgO-variation diagrams, the basaltic rocks from the AC show different trends for some major elements relative to MORB (Fig. 16a). Their Al_2O_3 contents decrease with

increasing MgO contents. TiO_2 and FeO contents of the analyzed rocks are nearly constant or slightly decrease, whereas CaO contents slightly increase with MgO contents. Compositions of samples from the OP and the HP complexes plot broadly on the compositional trend of basaltic rocks from the AC.

With attention to high field strength elements, less mobile as well as REEs, most analyzed rocks in the AC and the OP complex show flat patterns similar to those of N-MORB in primitive-mantle-normalized trace element variation diagrams (Fig. 16b). One sample from the AC (KR16, Table 2) exhibits a pattern moderately enriched in Nb and light-REEs and depleted in Y and heavy-REEs, which is similar to that of oceanic island basalts (OIB). This sample has the highest TiO_2 content and an intermediate MgO value among the analyzed samples. Trace element patterns of most analyzed samples show Nb-depletion, commonly regarded as a typical feature of island arc basalt (IAB). However, the depletion in Nb/La of 0.40–0.61 for the analyzed samples (except for the most depleted sample) is included within a variety of MORBs (Nb/La = 0.4–1.7; e.g., compilation by Lassiter and DePaolo, 1997).

Table 1
Selected analyses of clinopyroxene compositions of the southern Gorny Altai basaltic rocks

Area	Kurai						Chagan-Uzun		
Complex	AC						HP	OP	
Sample No.	KR80	KR82	KR83	KR85	KR88	KR89	CO23	M113	UZ12
<i>Major elements (wt.%)</i>									
SiO ₂	52.6	52.6	48.3	50.5	51.2	51.3	51.7	52.5	49.9
TiO ₂	0.6	0.6	1.4	1.1	0.9	1.0	0.9	0.6	0.8
Al ₂ O ₃	2.7	2.1	3.4	3.1	2.9	2.1	3.3	3.4	5.1
Cr ₂ O ₃	0.4	0.2	b.d.l.	b.d.l.	0.0	0.0	0.6	0.7	0.1
FeO ^a	6.4	7.8	11.4	13.3	9.8	10.1	7.0	5.5	11.6
MnO	0.2	0.3	0.3	0.2	0.3	0.2	0.1	0.1	0.3
MgO	16.3	16.5	13.5	12.2	14.6	13.8	16.3	16.1	13.7
CaO	20.5	19.7	19.2	18.2	18.9	19.1	19.6	21.0	18.7
Na ₂ O	0.3	0.3	0.3	0.3	0.4	0.4	0.3	0.2	0.3
Total	99.9	100.1	97.8	98.9	99.0	98.0	99.8	100.1	100.5
<i>Rare earth elements (ppm)</i>									
La	0.2	0.3	1.2	1.7	0.2	0.4	0.9	1.2	0.3
Ce	1.4	1.7	6.5	9.4	1.3	2.7	4.1	5.5	1.5
Pr	0.3	0.4	1.6	2.3	0.3	0.7	0.8	1.0	0.4
Nd	2.3	2.8	10.0	14.1	2.3	4.7	4.2	5.6	2.4
Sm	1.3	1.8	5.6	8.5	1.5	2.9	1.9	2.4	1.8
Eu	0.5	0.7	1.8	2.7	0.5	0.9	0.7	0.9	0.8
Gd	2.0	2.2	7.6	11.2	1.8	4.1	2.4	2.9	2.3
Tb	0.4	0.5	1.4	2.0	0.4	0.8	0.4	0.5	0.5
Dy	3.1	3.8	10.7	16.2	2.8	7.0	2.7	3.6	4.2
Ho	0.6	0.8	2.2	3.0	0.6	1.5	0.6	0.7	0.8
Er	1.8	2.5	6.6	9.5	1.8	4.5	1.3	1.6	2.8
Tm	0.2	0.3	1.0	1.3	0.3	0.8	0.2	0.3	0.4
Yb	1.8	2.0	6.2	8.6	1.7	4.1	1.4	1.5	2.4
Lu	0.3	0.3	0.9	1.5	0.2	0.7	0.2	0.2	0.4

AC, Accretionary complex; HP, High-*P/T* metamorphic complex; OP, Ophiolite complex.

^a Total iron as ferrous; b.d.l., below detection limit.

6. Discussion

6.1. Origin of basaltic rocks

6.1.1. Basalt variety

In all of the AC, the HP and the OP complexes, the basaltic rocks are closely associated with marine carbonates and lack any coarse-grained terrigenous clastic material. This indicates that these basaltic rocks were primarily derived from the surficial oceanic crust in shallow water-depth above the calcium carbonate compensation depth (CCD) in a mid-oceanic setting. Additionally, most basaltic rocks have a phenocryst assemblage of clinopyroxene ± plagioclase, which is different from that of typical MORB. The negative correlation between Al contents and Mg# for the relict clinopyroxenes from the HP and the OP complexes (Fig. 14b), together with that between the Al₂O₃ and MgO contents for the basaltic rocks from the AC (Fig. 16a), supports the idea that clinopyroxene fractionation predominated during crystallization of these rocks. The compositional trends for other major elements are consistent with clinopyroxene-dominant fractional crystallization. These rocks differ considerably from typical MORB that is characterized by olivine + plagioclase frac-

tional crystallization, and often associated with deep-sea chert. These features indicate that all the basaltic rocks formed in the mutually similar tectonic setting of a shallow mid-ocean.

However, the relict clinopyroxenes and the whole-rock geochemistry of the Gorny Altai basalts are very different for different complexes and samples: The relict clinopyroxenes from the HP complex have heavy-REE-depleted patterns and their light-REEs are less depleted compared than those from the AC and the OP complex (Fig. 15). Such differences cannot be accounted for by fractional crystallization of an identical parental magma, because the relict clinopyroxenes from the HP complex have higher Mg# than those from the other two (Fig. 14a and b). Accordingly, the basaltic rocks in the HP complex are obviously different in origin from those in the AC and OP complexes that have an N-MORB-like signature, and likely evolved from a primary magma similar to that of a Hawaiian OIT. In terms of whole-rock compositions, the AC includes two kinds of basaltic rocks; one with the highest-TiO₂ rock and the remainder that show geochemistry similar to that of OIB and N-MORB, respectively (Fig. 16b). Differences in trace element patterns between the above two kinds of rocks also cannot be explained

Table 2
Selected analyses of whole-rock compositions of the southern Gorny Altai basaltic rocks

Area	Kurai								Chagan-Uzun	
Complex	AC								HP	OP
Sample No.	KR16	KR80	KR81	KR82	KR83	KR86	KR87	KR90	KR71	CU53
<i>Major elements (wt.%)</i>										
SiO ₂	50.1	48.2	47.6	49.2	48.7	49.6	50.8	49.7	47.6	49.0
TiO ₂	3.01	2.03	1.64	2.05	2.42	1.02	2.07	1.68	1.15	2.13
Al ₂ O ₃	16.1	14.2	12.5	14.2	14.4	15.2	16.4	13.8	14.1	14.7
FeO*	11.7	12.1	11.0	12.1	13.1	10.7	10.5	10.6	9.2	12.9
MnO	0.13	0.21	0.19	0.21	0.24	0.22	0.21	0.45	0.18	0.22
MgO	6.15	6.19	8.24	6.05	4.91	5.27	4.35	6.04	7.09	6.70
CaO	10.3	8.73	9.41	8.79	7.47	9.09	8.07	8.30	12.8	8.20
Na ₂ O	3.64	3.55	2.53	3.92	3.88	5.15	5.00	5.27	3.24	4.09
K ₂ O	0.43	0.21	0.48	0.21	1.45	0.03	0.45	0.12	0.07	0.26
P ₂ O ₅	0.36	0.22	0.17	0.23	0.25	0.10	0.25	0.18	0.10	0.22
Total	101.8	95.7	93.8	97.0	96.8	96.3	98.0	96.2	95.6	98.4
<i>Trace elements (ppm)</i>										
Ni	67	41	94	39	14	20	23	42	41	42
Rb	8	7	6	3	10	b.d.l.	4	1	1	4
Sr	214	291	238	189	125	81	361	252	186	155
Y	34	44	37	50	37	19	48	42	21	38
Zr	218	164	123	174	133	43	189	132	71	119
Nb	19	2	2	3	3	b.d.l.	3	2	1	2
Ba	21	77	63	57	160	26	177	98	31	39
La	15.9	5.04	n.a.	6.08	5.40	1.86	6.59	4.51	n.a.	3.36
Ce	37.9	15.8	n.a.	19.0	16.1	5.54	20.4	13.8	n.a.	11.2
Pr	5.22	2.76	n.a.	3.26	2.74	0.99	3.50	2.39	n.a.	2.02
Nd	23.2	14.7	n.a.	17.35	14.6	5.56	18.4	12.8	n.a.	11.1
Sm	6.16	5.14	n.a.	5.73	4.75	2.05	6.05	4.45	n.a.	4.17
Eu	2.70	1.81	n.a.	2.04	1.82	0.85	1.97	1.50	n.a.	1.64
Gd	7.16	7.12	n.a.	7.97	6.48	3.05	8.17	6.10	n.a.	5.93
Tb	1.16	1.27	n.a.	1.44	1.14	0.56	1.47	1.11	n.a.	1.07
Dy	7.36	8.73	n.a.	10.0	7.61	3.93	10.1	7.59	n.a.	7.36
Ho	1.39	1.80	n.a.	2.08	1.53	0.82	2.11	1.59	n.a.	1.53
Er	3.81	5.43	n.a.	6.29	4.46	2.48	6.40	4.76	n.a.	4.55
Tm	0.49	0.78	n.a.	0.91	0.63	0.36	0.94	0.69	n.a.	0.64
Yb	2.86	5.17	n.a.	5.97	4.12	2.38	6.20	4.48	n.a.	4.10
Lu	0.36	0.78	n.a.	0.90	0.62	0.37	0.93	0.68	n.a.	0.60
Pb	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	2	b.d.l.	b.d.l.
Th	2	1	b.d.l.	b.d.l.	b.d.l.	1	1	1	b.d.l.	b.d.l.

n.a., not analyzed. Other abbreviations are the same as those in Table 1.

by fractional crystallization, because the highest-TiO₂ rock has an intermediate MgO content among the basaltic rocks from the AC (Fig. 16a). Consequently, the basaltic rocks from the HP complex and some from the AC could be of OIB origin, which is consistent with their field occurrence and petrography, as mentioned above. In their compilation of accreted material in the Phanerozoic ACs in Japan, Isozaki et al. (1990) established that OIB is the most commonly accreted oceanic basalts in contrast to MORB.

On the other hand, the relict clinopyroxene and the whole-rock compositions of basalts from the OP complex and most of those from the AC exhibit trace element patterns with light-REEs-depletions similar to those of MORB. These patterns are inconsistent with an origin in an oceanic island setting, suggested by the field occurrence and petrography of analyzed rocks.

6.1.2. Oceanic plateau basalt

The complex features of the basaltic rocks from the AC and the OP complex are most similar to those of oceanic plateau basalt (OPB). An oceanic plateau, typical of an oceanic large igneous province (LIP), is a topographic rise in an ocean, and is frequently associated with limestone deposited above the CCD (e.g., Coffin and Eldholm, 1994; Saunders et al., 1996). Oceanic plateaus are mainly composed of aphyric basalts with petrochemistry of N-MORB to E-MORB affinities (Fig. 16c), and most are affected by clinopyroxene fractionation during crystallization (e.g., Mahoney et al., 1993; Tejada et al., 1996). Most basaltic rocks examined in this study are porphyritic, thus different from known OPBs that are mostly composed of aphyric basalt. However, the notion that the majority of OPB are aphyric needs re-evaluation, because the number of OPB samples so far analyzed is still small.

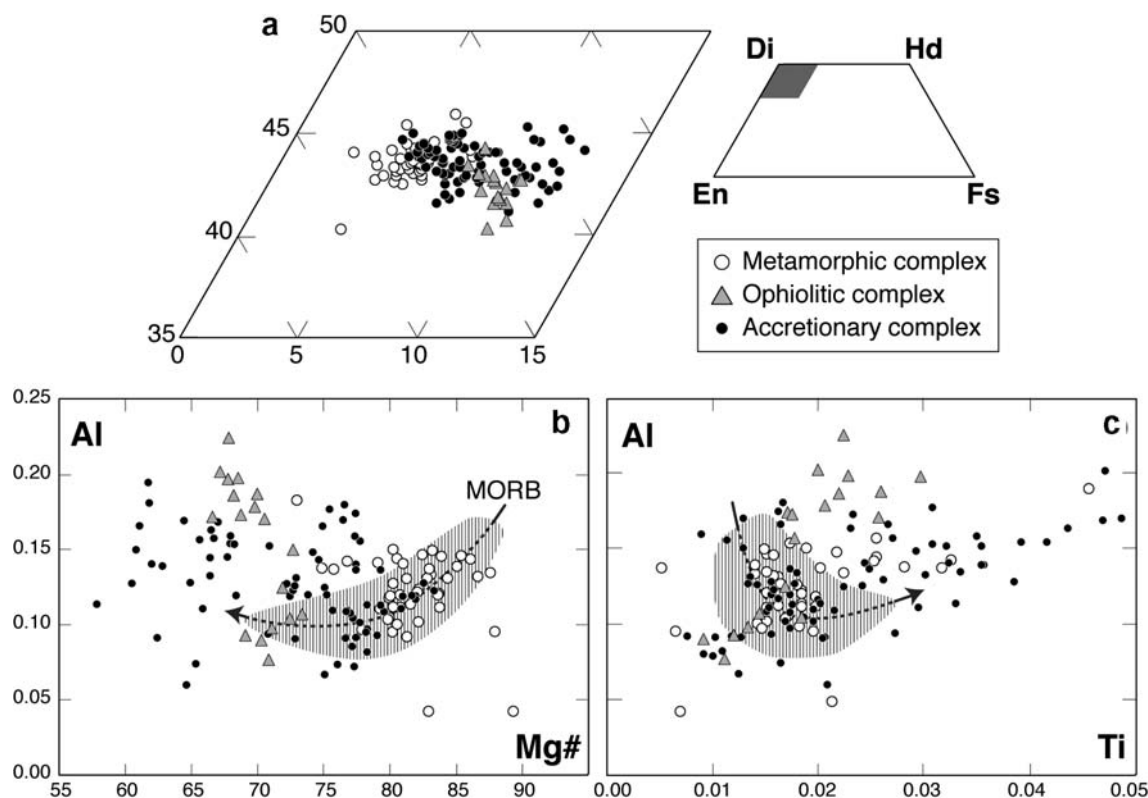


Fig. 14. Compositional varieties of relict igneous clinopyroxenes in basaltic rocks from the southern Gorny Altai. (a) A diopside (Di)–enstatite (En)–ferrosilite (Fs)–hedenbergite (Hd) quadrilateral diagram. (b) Al–Mg# plot. (c) Al–Ti plot. Mg#, $\text{Mg}/(\text{total Fe} + \text{Mg})$. Cation numbers are calculated as six oxygens. Compositional fields of mid-ocean ridge basalts (MORB) and fractionation trends refer to the compilation by Komiya et al. (2002).

In recent years, most models explaining oceanic LIPs with OPBs have invoked mantle plumes, the most frequent model of which is the plume-head (Griffiths and Campbell, 1990). The impingement of a mantle plume at the base of the lithosphere likely raises the temperature and initiates partial melting of depleted oceanic lithosphere. Then lithospheric mantle melts mix with plume-derived melts, generating a spectrum of chemical compositions. It is noteworthy that oceanic LIPs comprise great thicknesses of OPB lava. In order to explain this feature, several studies have proposed that plume heads, consisting principally of peridotite, may contain significant amounts of embedded eclogitic material derived from subducted MORB that once formed in the upper oceanic crust, (e.g., Cordery et al., 1997; Yasuda et al., 1997). Because the MORB solidus is lower in temperature than the fertile peridotite solidus at any upper-mantle depth under dry conditions, the MORB would completely melt in an upwelling mantle plume at temperatures sufficient to cause partial melting of peridotite (Takahashi et al., 1998). The complete melting of embedded MORBs in a mantle plume is more likely to form large quantities of OPB lavas, compared with the amount of lava with MORB-like compositions formed by the partial melting of depleted oceanic lithosphere.

Tejada et al. (2002) calculated incompatible element compositions for batch melts of a composite source containing primitive mantle and recycled (i.e., altered, sub-

ducted and dehydrated) oceanic crust, and showed that primitive-mantle-normalized incompatible element compositions of mixtures are composed of 80–100% melt of recycled MORB and 0.5–20% batch partial melt of the primitive mantle, because the plume peridotite would exhibit similar patterns to those of E-MORB (enriched in Nb and light-REEs) to N-MORB (depleted in Nb and light-REEs); the mixture of lesser amounts (<80%) of the recycled-MORB-derived melts with the 0.5% batch partial melts of the primitive mantle could make the incompatible element patterns more enriched in Nb and light-REEs, and more depleted in Y and heavy-REEs, like those of OIBs. Accordingly, the plume-head model involving the composite source of the plume peridotite and the recycled MORB could contemporaneously generate various magmas with different geochemistry, because the distribution of the recycled MORBs and the degrees of partial melting of the peridotite in the plume head would be variable. Hence, the OPB formation process in the plume-head model can satisfactorily explain the coexistence of the MORB-like and the OIB-like rocks in the AC.

The primary igneous textures of the basaltic rocks in the AC and in the higher-grade metabasites in the HP complex were commonly obliterated. In the HP complex, there are two types of basaltic rocks, the MORB and the OIB, based on their whole-rock geochemistry (Buslov et al., 1993, 2002). However, the basaltic protoliths with MORB-like

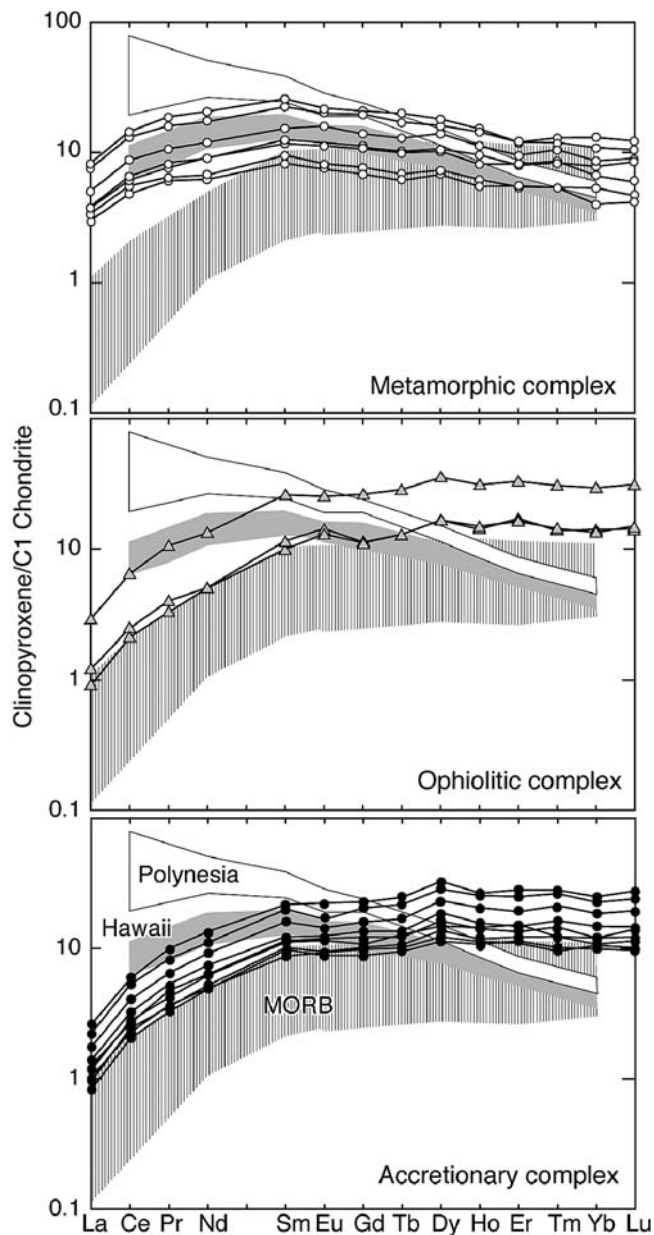


Fig. 15. Chondrite-normalized rare earth element patterns for relict igneous clinopyroxenes in basaltic rocks from the southern Gorny Altai. Chondrite values are after Sun and McDonough (1989). Patterns for clinopyroxenes of oceanic island basalts (Hawaii and Polynesia) and mid-ocean ridge basalts (MORB: ODP Hole 504B) are also shown. Datasets are from Schnetzler and Philpotts (1968) and Jeffries et al. (1995) for Hawaii, from Schnetzler and Philpotts (1968) for Polynesia, and from Dick and Johnson (1995) for Hole 504B.

geochemistry in the AC and the HP complex are different in petrogenesis from common MORB. Although we cannot rule out the possibility that parts of the basaltic protoliths could be of OIB origin, it is most likely that the basaltic protoliths, with their variable geochemical features, in the AC and the HP complex were generated as OPBs at a mid-oceanic and probably off-axis ridge by upwelling of a mantle plume that embedded the recycled MORB. The basaltic protoliths with MORB geochemistry

would have evolved from a composite magma that mixed two kinds of melts, derived from the complete melting of a greater proportion of the entrained recycled-MORB, and from partial melting of the peridotite matrix of the mantle plume. In contrast, those with OIB-like features would have resulted from a composite magma, composed of a melt with a lesser amount of the recycled, completely melted MORB, and a melt that the plume peridotite partially melted to a lesser degree than generating a MORB magma.

As for the basaltic rocks from the OP complex, the petrography and petrochemistry of relict clinopyroxenes and whole-rock compositions have some analogy with the basaltic protoliths of the AC and the HP complex. We thus consider that the basaltic rocks in the OP complex were accreted OPBs, and that they were different in origin from the amphibolite and peridotite of the basement of an island arc; i.e., lower crust and its underlying fore-arc mantle.

6.2. Accretion to a mid-oceanic arc

6.2.1. Subduction–accretion of an oceanic plateau

The AC in the Gorny Altai region consists solely of the *mélange*-type of Isozaki (1997). No coherent-type AC with imbricated thrust sheets of deep-sea chert (e.g., Matsuda and Isozaki, 1991) occurs. The former type is regarded as a product of seamount subduction at a trench that was formed by destroying older ACs and recycling the wasted material accumulating again in a trench (Okamura, 1991). Thus, the predominance of the *mélange*-type AC and the common occurrence of oceanic basalt and limestone in the studied area are mutually concordant as in Phanerozoic ACs.

The *mélange* matrix contains innumerable fragments derived from OPB and its capping limestone that vary in size from boulder to fine-grained mud. In addition to this oceanic material, the *mélange*-type AC contains lithic and mineral fragments, such as andesite, various kinds of schists, and occasionally amphibolite, derived not from a paleo-seamount but probably from a volcanic arc and from the HP and the OP complexes that already exposed on land. Such a land-derived material was mixed with the oceanic materials at an ancient trench. The best modern analogue exists along the Japan Trench off northeastern Japan, where the Daiichi-Kashima and Erimo seamounts are currently entering the active trench (Cadet et al., 1987; Kobayashi et al., 1987). The Permian Akiyoshi AC in southwestern Japan demonstrates a good ancient analogue of such subduction-related collapse of oceanic topographic-highs (e.g., seamount and plateau) and their mixing with terrigenous materials from a fore-arc domain (Kanmera & Nishi, 1983; Isozaki, 1987; Sano and Kanmera, 1991). In their compilation of all accreted oceanic materials in Japan, Isozaki et al. (1990) concluded that subduction of large oceanic topographic-highs usually causes selective accretion solely of the surficial parts of

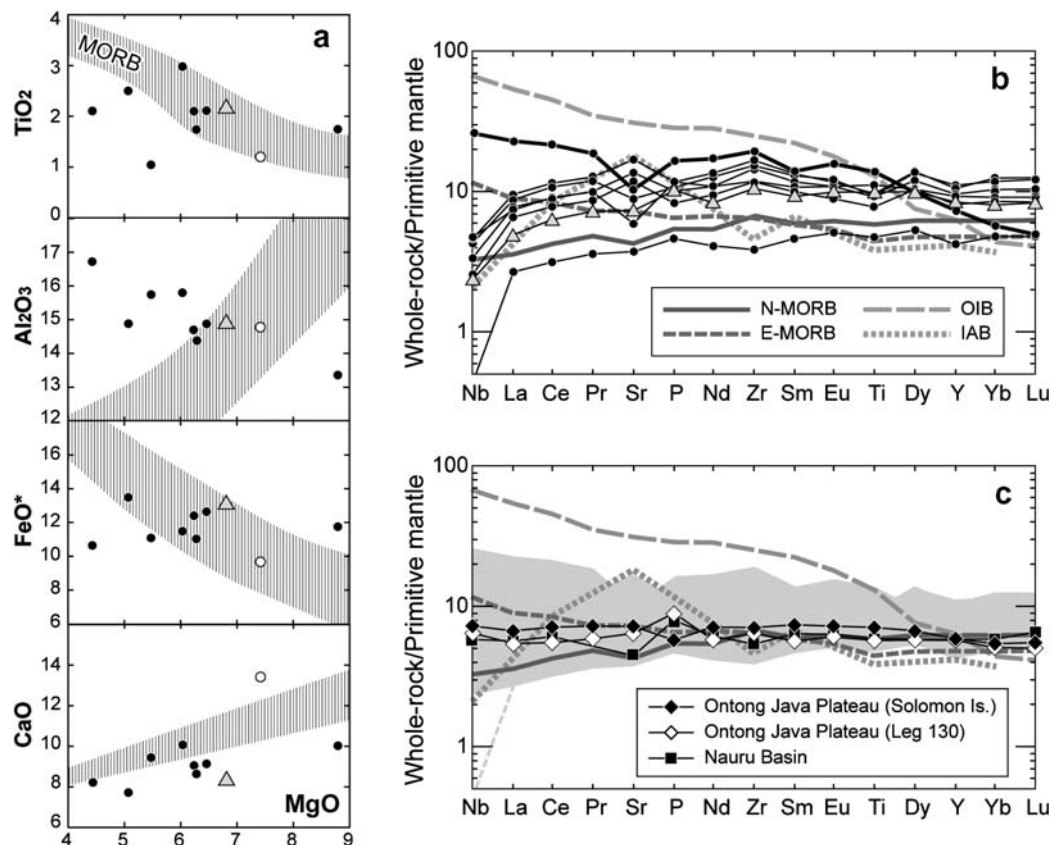


Fig. 16. Whole-rock compositions of basaltic rocks from the southern Gorny Altai. Symbols are the same as in Fig. 14. (a) MgO-variation diagrams (in weight percent) of major elements. Fields of MORB are from the GEOROC database up to 2000 on the web (<http://georoc.mpch-mainz.gwdg.de>), provided by the Max-Planck-Institute. (b) Primitive mantle-normalized immobile trace element patterns. A thick line with solid circles indicates a sample with the highest-TiO₂ content from the accretionary complex. (c) Average values for oceanic plateau basalts from the Ontong Java Plateau (Solomon Islands and ODP Leg 130) and the Nauru Basin, southern Pacific. Datasets are after Tejada et al. (1996) for the Solomon Islands, Mahoney et al. (1993) for the Leg 130, and Castillo et al. (1986) and Saunders (1986) for the Nauru Basin. A shaded field shows the range for basaltic rocks from the southern Gorny Altai. Primitive mantle values are from Sun and McDonough (1989). Reference lines of N- and E-MORBs, and oceanic island basalt (OIB) are from Sun and McDonough (1989), and of island arc basalt (IAB) from McCulloch and Gamble (1991).

the highs and the rest are subducted. The absence of oceanic gabbros in the Cambrian AC in Gorny Altai may be explained likewise.

6.2.2. Serpentinite *mélange*-type AC

The occurrence of serpentinites as the matrix may appear incompatible with the above-mentioned trench setting because ultramafic rocks usually do not crop out in trench environments. However, serpentinite seamounts commonly occur in the fore-arcs of the Izu-Bonin-Mariana intra-oceanic island arcs (e.g., Fryer and Mottl, 1992; Ishii et al., 1992). Such serpentine seamounts are mainly composed of unconsolidated serpentine mud flows, that may have originated as serpentinite diapirs derived from hydrated peridotites of the fore-arc mantle wedge. Another possible source for trench serpentinite is a subducting oceanic plate dissected by a transform fault where hydrated abyssal peridotite may be diapirically exhumed, although their amount may be small. At present, the serpentinite-seamount interpretation appears more realistic. Detrital serpentinite, either from a fore-arc serpentinite seamount or from a transform fault, may have been transported into

or along a trench axis, mixed with other components, then accreted in the inner Cambrian trench.

6.3. Late Proterozoic high-*P/T* metamorphism

The mineral assemblages and chemical compositions of the HP rocks in Chagan-Uzun, Akkaya river, and Kurai areas indicate that the peak metamorphism took place between 300 °C at 0.4 GPa and 660 °C at 2.0 GPa, suggesting a maximum burial depth of about 60 km from the surface, i.e., to the depth of the mantle wedge beneath a fore-arc. Subsequent retrogressive overprint took place during exhumation under greenschist-facies conditions (Ota et al., 2002). The peak *P-T* estimates of the HP rocks in the three studied localities define a single curve in a *P-T* diagram, suggesting that they belong to the same single-phase regional HP complex (Ota et al., 2002). This curve is slightly convex toward the temperature axis on a *P-T* diagram, and is similar to a steady-state *P-T* path in a subduction zone, formed when young, hot lithosphere has subducted, according to numerical calculations (e.g., Peacock, 1996).

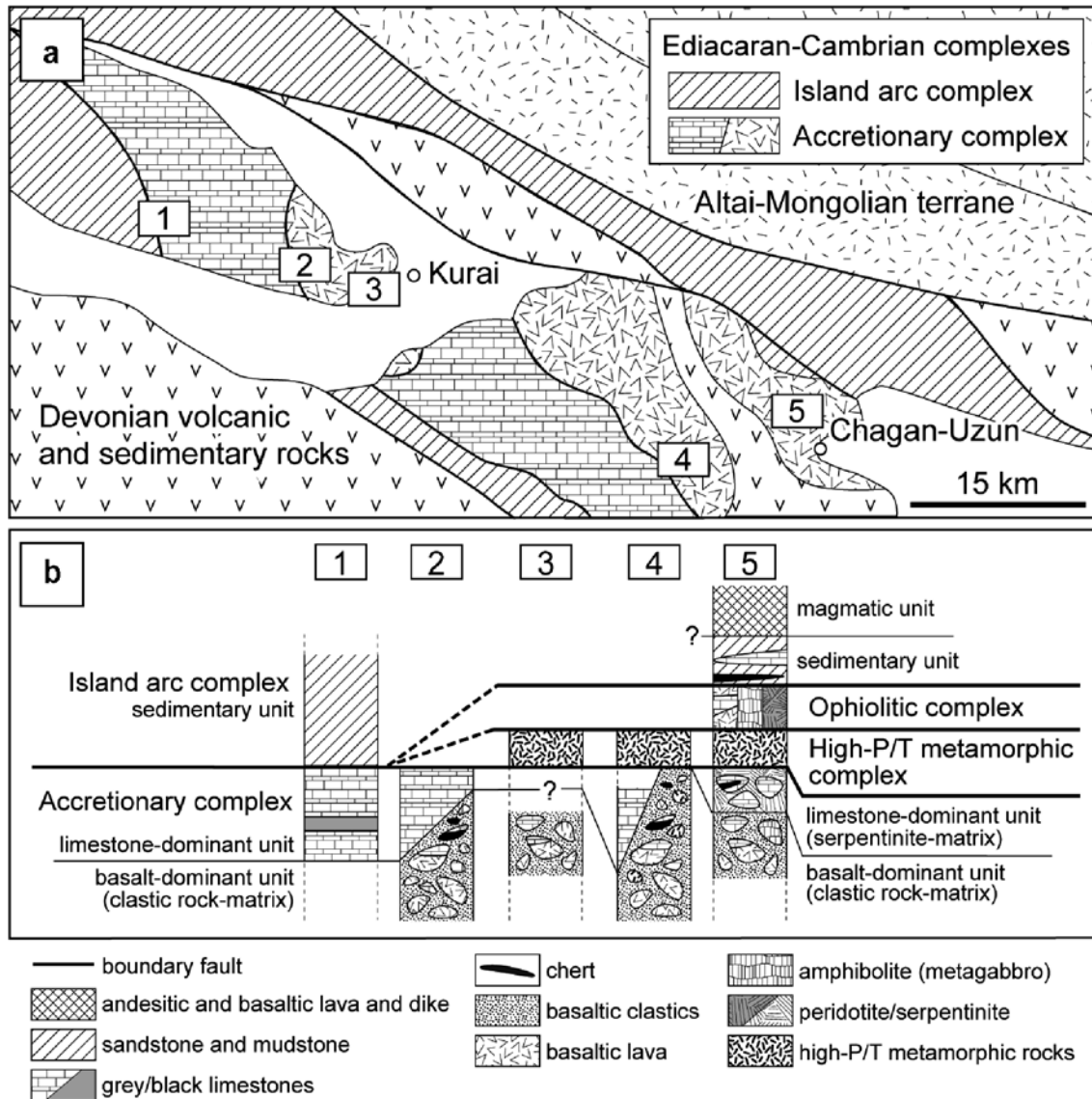


Fig. 17. (a) Index sketch map of the southern Gorny Altai. (b) Schematic diagram showing the order of superposition in the piled nappe structure of the Ediacaran–Cambrian orogenic complexes (not to scale). See the localities of the composite columns 1–5.

The protoliths of the HP rocks were first formed as an AC at a trench. Immediately after that, they were deeply buried by oceanic subduction. Based on the chronological data, the high- P/T metamorphism culminated at ca. 630 Ma (around the Cryogenian/Ediacaran boundary) at a 60 km-deep sub-crustal level, and the complex was exhumed to a shallower depth by ca. 570 Ma (Ediacaran). The above-mentioned P – T path may suggest that the exhumation of the Ediacaran HP complex in Gorny Altai was triggered by subduction of a mid-oceanic ridge accompanying two young hot lithospheres on both sides. After compiling the mode of occurrences of HP belts over the world, Maruyama et al. (1996) generalized the wedge extrusion model for exhumation tectonics of HP rocks induced by ridge-subduction.

It is noteworthy that a mature arc-trench system, with a steep temperature–pressure gradient along the Wadati-Benioff plane that was enough to produce typical high-

P/T assemblages, developed already in the late Cryogenian in the western segment of the CAO (Ota et al., 2002). This subduction system was clearly older, by more than 100 million years, than the system that formed the middle–Late Cambrian AC in both the northern and southern Gorny Altai regions. The protoliths are pre-630 Ma AC formed at the same trench immediately before deep subduction. The 650 Ma (Cryogenian) boninites likewise support the antiquity of the subduction regime. Thus, this convergent margin survived for more than 150 million years, at least from the Cryogenian to the Cambrian.

6.4. Subhorizontal accretionary orogen

The most striking structural feature in the Gorny Altai region, particularly in the south, is the mode of occurrence of all orogenic components within a single fault-bounded

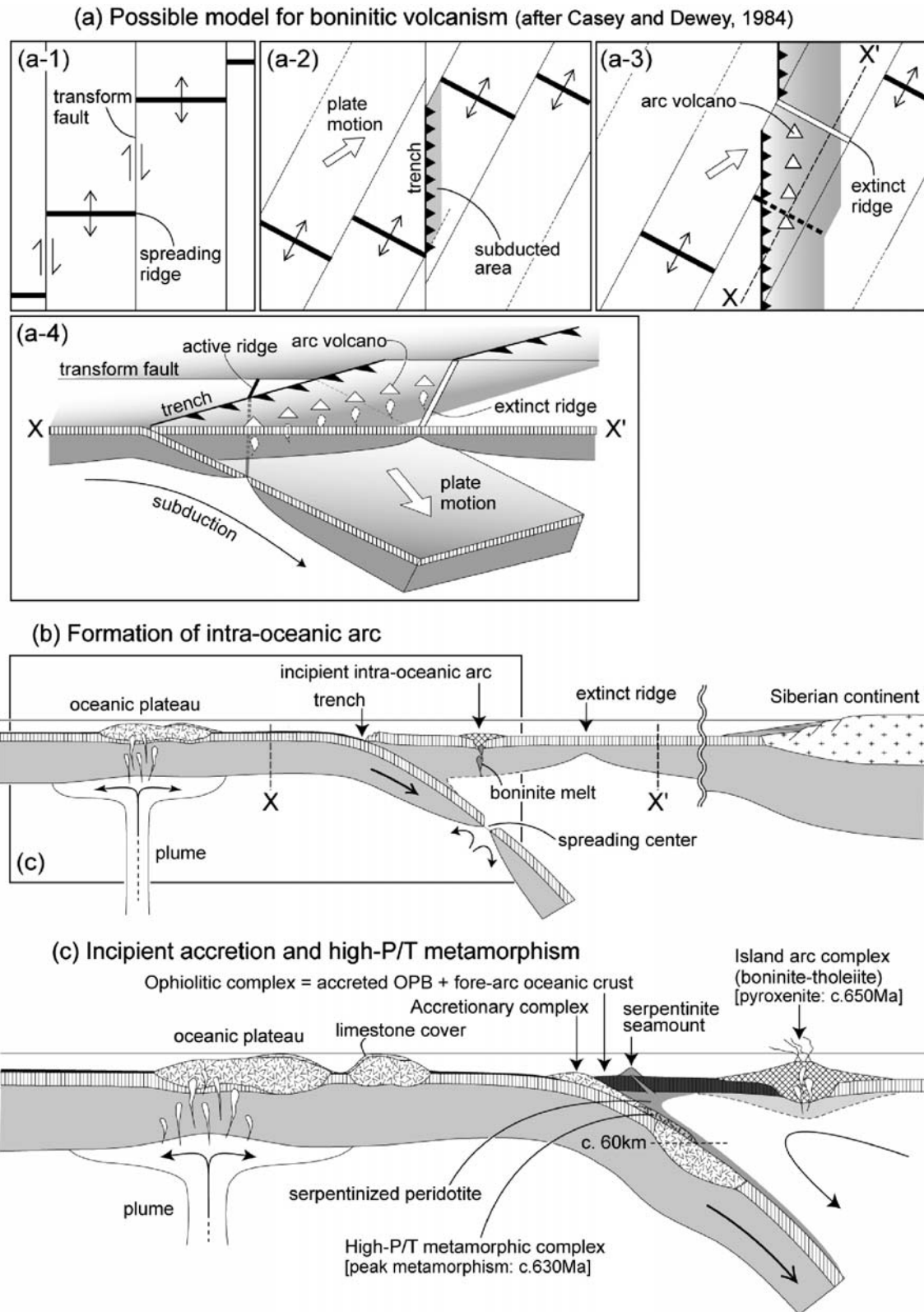


Fig. 18. Cartoons showing the tectonic evolution of the Ediacaran–Cambrian Gorny Altai subduction–accretion complexes. (a) A possible model for boninitic volcanism in the incipient stage of the arc development. (a-1) Normal transform fault and fracture-zone systems. (a-2) Plate re-organization and initiation of young-lithosphere subduction beneath young lithosphere. (a-3) Island arc volcanism, possibly involving boninite genesis. (a-4) A bird-eye view of cross-section X–X' in (a-3). A cross-section of line X–X' is shown in (b). (b) Formation of an intra-oceanic arc. A box outlines the enlarged part in (c). (c) Incipient accretion and high-P/T metamorphism. OPB, oceanic plateau basalt. See text for further details.

subhorizontal nappe (Figs. 2 and 6a–d). The Ediacaran–Cambrian AC plus HP and OP complexes occur as nappes that are all incorporated in a westward vergent, subhorizontal nappe. Although later strike-slip faulting has modified the primary nappe structure throughout the region, the order of structural superposition among the nappes remains constant; the AC nappe, the HP complex nappe, and the OP nappe, in ascending order (as schematically depicted in Fig. 17). This primary structure suggests that horizontal shortening was a main tectonic factor in formation this orogen, and that the order of superposition is critical in reconstructing the primary configuration of the components. The relationship among the three nappes in the northern area appears concordant, although not so clear as in the southern area.

When the 630 Ma HP complex is redefined as a metamorphosed AC that was formed prior to the Cambrian AC, the two nappe units have a structurally downward younging polarity. This downward-younging is consistent with the accretionary processes in both modern trenches and ancient orogens (e.g., Matsuda and Isozaki, 1991; Isozaki, 1996). The westward vergence recognized in the nappe structure indicates that the subduction occurred on the western side of the Gorny Altai intra-oceanic arc.

The occurrence of the thin HP nappe between the non to weakly metamorphosed AC and the OP nappe is particularly important, as this pattern is consistent with the general mode expected of HP rocks (Maruyama et al., 1996). The OP complex tectonically overlying the nappes of the AC and the HP complex probably represents a fore-arc ophiolite derived from an old oceanic plate, beneath that another oceanic plate subducted initially to form an incipient arc. The entire arc complex was originally built upon this OP and its lateral equivalents.

The overall primary nappe structure in the Gorny Altai region is almost identical to that in southwestern Japan that is composed of Late Paleozoic to Cenozoic ACs including the HP equivalents (Isozaki, 1996; Maruyama, 1997) and to that in western North America (Isozaki and Maruyama, 1992; Maruyama et al., 1992). This positively suggests that the Gorny Altai ‘orogen’ was constructed during a Pacific-type (Miyashiro-type) orogeny in the Ediacaran–Cambrian interval.

6.5. Boninite-bearing arc

The orogenic components and their overall structures of the Gorny Altai region are similar to those of the Pacific-type (Miyashiro-type) orogen, as pointed out by Watanabe et al. (1993) and Buslov and Watanabe (1996). However, the region lacks a granitic batholith belt, which is one of the major characteristics of a matured Pacific-type orogen. In addition, the arc-type volcanism with boninites in Gorny Altai requires a specific tectonic setting different from ordinary arc-trench systems. Here we discuss a possible tectonic setting that can explain the various rock types and structure in the Gorny

Altai region with the AC, the HP, the OP, and the arc complexes.

Boninites, long regarded as peculiar island-arc volcanic rocks, are currently considered to have originated from a sub-arc mantle wedge, composed of depleted harzburgite. The wedge harzburgite has usually experienced partial melting at a mid-oceanic ridge, and then overprinted by subduction-zone metasomatism by subducting young, hot oceanic lithosphere. However, to obtain the high temperature required for boninite genesis, it is required that the overlying oceanic lithosphere of the mantle wedge is also very young and hot (Crawford et al., 1989; Tatsumi and Maruyama, 1989; Stern et al., 1991; Pearce et al., 1992).

Pearce et al. (1992) proposed a model to satisfy all these requirements for boninite formation by assuming a transitional tectonic setting from an active ridge-transform system to a new intra-oceanic arc-trench system, as in the Eocene West Pacific (Casey and Dewey, 1984) (Fig. 18). According to this model, the changes of plate motion and the resultant re-organization of plate boundaries enabled the initiation of subduction of a young oceanic lithosphere beneath a young oceanic plate very close to an active ridge, which potentially generated a boninitic melt in the hot mantle wedge (Fig. 18). The boninitic dikes and sills intrude into overlying oceanic crust, and an incipient boninite-bearing island arc develops in an intra-oceanic environment. When the subducted lithosphere reached beneath the active ridge, the overlying ridge was abandoned and subsequent subduction refrigerated the mantle wedge to stop the boninite formation. Consequently, the subducting lithosphere became older and colder, making the arc volcanism shift from boninitic to a normal calc-alkaline.

The arc complex with boninites in the Gorny Altai region probably formed through such transient tectonic processes involving transform and incipient subduction systems. This interpretation appears plausible, because it can also explain other geological features in the Gorny Altai region. In the case of a new subduction system with a brand new oceanic plate subducting, no thick deep-sea sediment such as chert can accumulate on the ocean floor. On the other hand, not much coarse-grained terrigenous clastics are supplied from the incipient intra-oceanic arc, owing to its small size. The limited occurrence of deep-sea chert and paucity of quartzofeldspathic terrigenous clastic in the AC are consistent with this interpretation. The metamorphic facies series of the HP complex are also consistent with the assumption that the subducting oceanic plate was young and hot at the incipient stage of subduction.

6.6. Tectonic synthesis

We synthesize below the tectonic history of the Ediacaran–Cambrian orogenic complexes in the Gorny Altai region within the evolution of the CAOBS and Siberian continental margin.

6.6.1. Birth of the Siberian margin

The tectonic history of the CAO, including the Gorny Altai subduction–accretion complexes, goes back to the breakup of the Mesoproterozoic supercontinent of Rodinia and to the subsequent opening of the Paleo-Asian Ocean and the Paleo-Pacific Ocean (Maruyama, 1994). According to the various tectonic reconstructions (e.g., Hoffman, 1991; Rogers, 1996; Dalziel, 1997; Li and Powell, 2001; Pisarevsky and Natapov, 2003), regardless of minor disagreements among them, the supercontinent Rodinia fragmented likely through several sub-stages of continental rifting during the late Neoproterozoic. A ca. 1020 Ma OP (Khain et al., 2002), the oldest in the CAO, suggests the existence of an oceanic plate; however its tectonic history is not well documented. Later, Siberia became isolated from other major continental blocks, and surrounded by the Paleo-Asian and/or Paleo-Pacific Oceans during the period 680–740 Ma (middle Cryogenian) (Vernikovskiy and Vernikovskaya, 2001; Dobretsov et al., 2003; Khain et al., 2003). During the late Cryogenian, new subduction systems formed in and around the Paleo-Pacific Ocean, triggered by the opening of another ocean on the opposite side of the globe (Maruyama et al., 1997). Likewise, around Siberia, a new subduction developed along its southern margin (relative to the present continental positions) (e.g., Sengör et al., 1993; Dobretsov et al., 1995; Buslov et al., 2001).

6.6.2. Development of an intra-oceanic arc

In a convergent tectonic regime along the margin of the Paleo-Asian Ocean, the Gorny Altai region initially developed as an intra-oceanic island arc by nascent northeastward subduction, probably in the late Cryogenian before 650 Ma. This arc was characterized by boninitic volcanism that formed in a unique transient tectonic setting between an oceanic transform zone and an incipient intra-oceanic subduction. At about 630 Ma, the arc-trench system was subjected to high-*P/T* metamorphism by successive oceanic subduction, and the metamorphosed AC was exhumed around 570 Ma (middle Ediacaran) probably by the subduction of a mid-oceanic ridge. However, the intra-oceanic arc was still small in size, and so did not obtain a large granitic batholith belt.

In the Cambrian, a large oceanic plateau arrived at the Gorny Altai arc-trench system, an AC was formed at a trench incorporating abundant material (OPB and capping limestone) derived from the collapsed oceanic plateau. A major part of the plateau was subducted, whereas its surficial part were accreted to form a *mélange*-type AC. The AC plus the older HP and OP complexes was all incorporated into a sub-horizontal nappe with a westward vergence, and thus a typical Pacific-type (Miyashiro-type) orogen was formed. All these orogenic complexes were covered unconformably by volcanoclastic sediments of an arc affinity. By the middle–late Cambrian, the arc had already evolved into a matured stage characterized by calc-alkaline volcanism. In the Devonian, these Ediacaran–Cambrian complexes in Gorny Altai were

finally intruded by an arc batholith, as the front of the active continental margin migrated oceanward.

6.6.3. Juxtaposition with surrounding terranes

During the Late Paleozoic convergent tectonics proceeded to close the western Paleo-Asian Ocean, juxtaposing the Gorny Altai arc against the neighboring multi-arc systems of the West Sayan terrane to the east and the Altai–Mongolian terrane to the west, and finally to the southern Siberian margin. The mutual collision and rotation of the terranes caused strike-slip dislocations with both dextral and sinistral sense that modified many pre-existing structures of the Ediacaran–Cambrian orogen in the Gorny Altai region.

7. Summary

Our research in the northern and southern Gorny Altai mountains, southern Siberia, has established the following new aspects of the Ediacaran–Cambrian orogenic complex that formed in the western segment of the CAO.

- (1) The Cambrian AC is composed of a *mélange*-type AC with abundant material from oceanic plateau basalt and capping marine carbonates.
- (2) The HP complex underwent peak metamorphism at ca. 630 Ma (late Cryogenian–early Ediacaran, Neoproterozoic), and was exhumed at ca. 570 Ma (Ediacaran).
- (3) Circa-650 Ma (late Cryogenian) boninitic rocks characterized the initial arc volcanism.
- (4) The AC, the HP and the OP complexes formed in an intra-oceanic arc-trench system off Siberia.
- (5) The AC, the HP and the OP complexes occur within a subhorizontal nappe pile with the overall westward vergence; this structure is similar to that of a typical Pacific-type (Miyashiro-type) orogenic belt.
- (6) We summarized the tectonic history of the Ediacaran–Cambrian orogenic complexes from the birth of an intra-oceanic island arc to final amalgamation with the surrounding terranes.

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