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Structural evolution of the Bayanhongor region, west-central Mongolia

Soichi Osozawa [a,*], Gonchig Tsolmon [b], Uondon Majigsuren [b], Jargalan Sereenen [b], Shigeaki Niitsuma [c], Naoyoshi Iwata [c], Terry Pavlis [d], Bor-ming Jahn [e]

[a] Department of Earth Science, Graduate School of Science, Tohoku University, Sendai 980-8578, Japan
[b] Mongolian University of Science and Technology, Ulaanbaatal 46, Mongolia
[c] Department of Earth and Environmental Sciences, Faculty of Science, Yamagata University, Yamagata 990-8560, Japan
[d] Department of Geological Sciences, University of Texis at El Paso, El Paso, TX 79968, USA
[e] Institute of Earth Sciences, Academia Sinica, Nankang, Taipei 11529, Taiwan

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Abstract

Most of previous models suggest that the Central Asia Orogenic Belt grew southward in the Phanerozoic. However, in the Bayanhongor region in west-central Mongolia, volcanic arc, accretionary prism, ophiolite, and passive margin complexes accreted northeastward away from the Baydrag micro-continent, and hence the region constitutes the southwestern part of a crustal-scale syntaxis close to the west. The syntaxis should be original, because presumably reorientation due to strike-slip faulting can be ignored. It is reconfirmed that the Baydrag eventually collided with another micro-continent (the Hangai) to the northeast. A thick sedimentary basin developed along the southern passive margin of the Hangai micro-continent. This region is also characterized by an exhumed metamorphosed accretionary complex and a passive margin complex, which are both bounded by detachment faults as well as basal reverse faults which formed simultaneously as extrusion wedges. This part of the Central Asia Orogenic Belt lacks exhumed crystalline rocks as observed in the Himalayas and other major collisional orogenic belts. In addition, we identified two phases of deformation, which occurred at each phase of zonal accretion as D1 through Cambrian and Devonian, and a synchronous phase of final micro-continental collision of Devonian as D2. The pre-collisional ocean was wide enough to be characterized by a mid-ocean ridge and ocean islands. Two different structural trends of D1 and D2 are observed in accretionary complexes formed to the southwest of the late Cambrian mid-ocean ridge. That is, the relative plate motions on both sides of the mid-ocean ridge were different. Accretionary complexes and passive margin sediments to the northeast of the mid-ocean ridge also experienced two periods of deformation but show the same structural trend. Unmetamorphosed cover sediments on the accretionary prism and on the Hangai micro-continent experienced only the D2 event due to micro-continental collision. These unmetamorphosed sediments form the hanging walls of the detachment faults. Moreover, they were at least partly derived from an active volcanic arc formed at the margin of the Baydrag micro-continent.

Keywords: Bayanhongor; Mongolia; D1; Accretion; D2; Collision; Detachment; Ophiolite

1. Introduction

The Asian continent has undoubtedly grown through several phases of collision, including the on-going collision of India. The collision process is very complex for even the latest collision of India, and involves a time evolution that will continue well into the future. In any collision, prior to complete closure of an inter-continent ocean, there will be episodes of formation of volcanic arcs, accretionary complexes, subduction zone high P/T metamorphism, trench-mid ocean ridge interaction, obduction of ophiolites, deformation of passive margin sediments just front of the colliding continent, etc. Such detachment faulting as the South Tibetan Detachment system (e.g., Hodges et al., 1998) could also be anticipated in a final episode for other collisional suture zones.

* Corresponding author. Tel.: +81 22 217 6619.
E-mail address: osozawa@dges.tohoku.ac.jp (S. Osozawa).

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The Central Asian Orogenic Belt is a very wide Neoproterozoic to early Phanerozoic mobile belt between the Siberian and the Tarim-Sino-Korean continents (Badarch et al., 2002; Fig. 1). The belt has been regarded as a collage made up of micro-continents, island arcs, accretionary complexes, ophiolites, and passive continental margin deposits. Sengor et al. (1993) and Sengor and Natal'in (1996) interpreted the formation of the collage by successive and southward growth of juvenile terranes, accompanied by emplacement of granitic plutons in a single evolving arc. Sengor et al. (1993) concluded that the Altaid Collage (=Central Asian Orogenic Belt) is characterized by the absence of nappe structures as commonly observed in large collisional orogens, such as the Himalayas and Alps. They also suggested that the entire collisional process involved strike-slip stacking, and typically, originally a single major island arc (the Kipchak arc and the presently studied Tuva-Mongolian arc) was duplicated. The importance of strike-slip stacking was also stressed by Buslov et al. (2004). However, probably this view of strike-slip modification has not yet been completely accepted or at the least, remains uncertain.

Badarch et al. (2002) held a different view on the basis of tectonic studies in Mongolia. They considered that the orogenic development involved a large number of closures and subduction of small oceans, obduction of ophiolites, collision of island arcs and micro-continents, and formation of suture zones. However, their data is rather general, and the basic supporting data is not well documented.

In this circumstance, Windley et al. (2007) reviewed the recent literature and proposed a more reasonable tectonic model for accretion of the Central Asian Orogenic Belt. Windley et al. (2007) does not prefer the strike-slip duplication model of the Kipchak and Tuba-Mongolian arcs (Sengor and Natal'in, 2004), but exhibited multiple island-arc accretionary events in the model (Windley et al., 2007, Fig. 9; but the figure is after Sengor and Natal'in, 1996, 2004, and two models are basically similar). In central Mongolia, two micro-continents and their intervened two units composed of ophiolites, accretionary complexes, and island arcs are presented (Windley et al., 2007, Fig. 2; Fig. 1). These mega-units on the order of hundreds of kilometers in length and width are distributed in a U shaped area with a bottom westward, forming syntax, bounded by a suture to the south (Windley et al., 2007, Fig. 2; Fig. 1).

In this study we report our field analysis in the Bayanhongor region, central Mongolia, where an ophiolite belt more than 300 km long is sandwiched between the Baydrag micro-continent in the south and the Hangai micro-continent in the north (Fig. 1).

The "Tarvagatay" continent (Fig. 1) is considered to be a part of the Hangai micro-continent; it may constitute the basement of the Hangai sedimentary basin. To the east, late Silurian to late Devonian radiolarian fossils were found in pelagic chert near UlaanBaatar city (Kashiwagi et al., 2004; Kurihara et al., submitted for publication), which indicates that the Hangai-Khantei ocean (Khangai-Khantey ocean, Sengor and Natal'in, 1996, 2004) back of the Hangai micro-continent was wide and matured enough for deposition of pelagic sediment.

In front of the Baydrag micro-continent, a part of the Tuba-Mongolian block (Windley et al., 2007, Fig. 2), is an ophiolite belt, obducted northeastwards onto the Baydrag micro-continent, and therefore subduction polarity there was also northeast and subduction direction was southwest (present coordinates, Windley et al., 2007, Fig. 2).

Recent field works by Teraoka et al. (1996) and Buchan et al. (2001) in the present Bayanhongor region have already shown that observed northward growth of accretionary complexes was a consequence of southward subduction. The structural analysis of Buchan et al. (2001) also indicated northward reverse faulting for each structural unit and northward "obduction" of the Bayanhongor ophiolite with respect to the passive margin of the Hangai

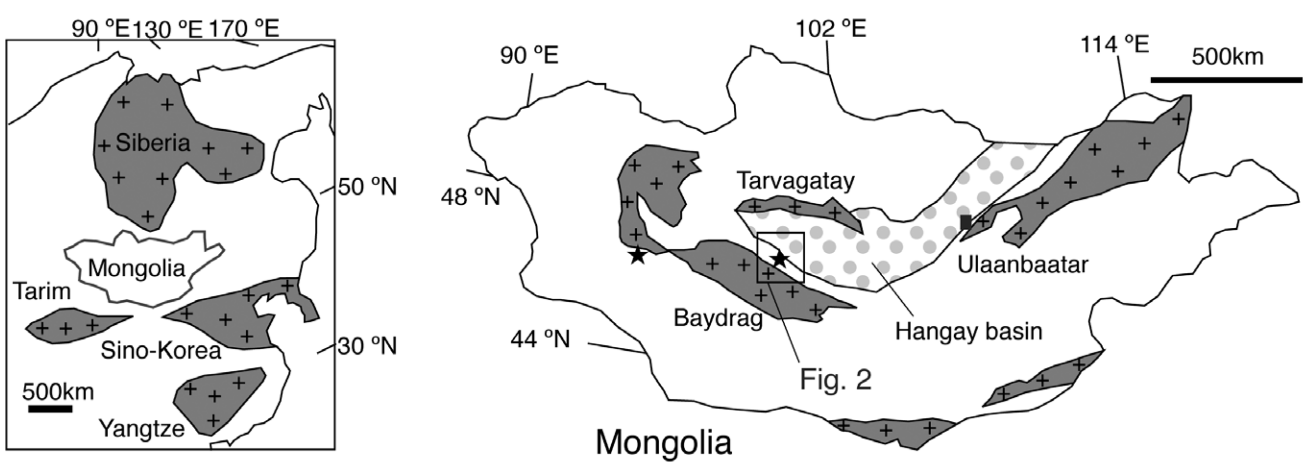

Fig. 1. Index map. Continents and micro-continents are shown. Partly modified after Badarch et al. (2002), and names of micro-continents are in their Fig. 1 referred to as cratonal blocks. The Baydrag micro-continent and its northwestern micro-continent constitute the Tuba-Mongolian block and the syntaxis close to the west. Eastern star: Bayankhongor and western star: Khantaishir and Dariv ophiolites.

micro-continent. We reinvestigated this region in order to obtain data to clarify collision and its precursor accretionary process in more detail, including a full lithological description of accretionary complexes as well as characteristic structural features. Here we present a precise geological observation about the origin of each complex, and a specified structural analysis including subduction polarity, based on which the deformation history should be divided into an earlier accretion (D1) and a later collision phase (D2). We also identified brittle detachment faults labeled D2, for the first time in Mongolia, even though their footwall has metamorphosed accretionary or passive margin lithologies, but not truly crystalline basement rocks. Such detachment faults in accretionary or high P/T metamorphic complexes have recently been recognized in Japan by the first author (Schoonover and Osozawa, 2004; Osozawa and Pavlis, 2007). Finally we reconfirm the structural development model and structural positioning of the Bayanhongor region by Windley et al. (2007).

2. Outline of geology

The structural units of the Bayanhongor region trend NW–SE (Fig. 2) and dip southwest at high angles (Figs. 3 and 4). The dip direction indicates that stratigraphic and structural tops should be southwest in every unit, bounded by D2 faults between each other. According to Buchan et al. (2001), six litho-tectonic units have been proposed from the southwestern Proterozoic Buld Gol unit to the northeastern Paleozoic Dzag unit (Fig. 2). Other than these units, there exists the Baidrag unit of Archaean-Proterozoic gneiss farther to the southwest of the mapped area in Fig. 2, and the Paleozoic Hangai unit in the northeastern part within Fig. 2. The former Baidrag unit contains rock with U–Pb ages of 2646 ± 45, and 1854 ± 5 Ma (Buchan et al., 2001, and references therein), and the latter Hangai unit was not mapped by Buchan et al. (2001).

Seven tectonic units including the Hangai unit and also the Cambrian granite are then mapped in Fig. 2. From SW to NE, these seven units are as follows, but details of lithologies, inferred tectonic setting, ages, structural features, and also structural and stratigraphic relations are provided in next Section 3.

(1) The Neoproterozoic Buld Gol unit, named "Burd Gol melange" by Buchan et al. (2001). The rock assemblage suggests the unit is an accretionary complex (e.g., Sengor and Natal'in, 1996), which probably underwent additional Barrovian-type metamorphism (Buchan et al., 2001).

(2) The next unit to the northeast consists of early Ordovician island-arc volcanics below and fossiliferous sedimentary rocks conformably above the volcanics (Buchan et al., 2001, 2002). Here we name this the Ulaan Sair Formation, as a stratigraphic unit which avoided severe deformation, distinguished from an accretionary complex.

(3) The Delb Khairkhan unit comprises blackschist, greenschist, psammitic schist, and marble. This is an accretionary complex, as named the Delb Khairkhan melange by Buchan et al. (2001), but no age diagnostic fossil and no radiometric ages have been recognized in the unit.

(4) Tectonically below is the Bayanhongor unit which represents an ophiolite. The crystalline age is problematic (Windley et al., 2007) due to the newly proposed Neoproterozoic age (Kovach et al., 2005), but the metamorphic age is known to be early Ordovician (Buchan et al., 2002, and reference therein).

(5) The Haluut Bulag unit was called the Haluut Bulag melange by Buchan et al. (2001), and is also a typical accretionary complex. The unit is composed of a mudstone and sandstone matrix, which contains exotic blocks of basalt, pelagic rocks, and also especially hemipelagic rocks, which is further described in this paper. The pelagic chert contains radiolarians, which is also reported here for the first time. Although the precise age is uncertain, it is important to point out that radiolarian fossils were not known before the Neoproterozoic.

(6) The late Ordovician to Devonian Dzag unit consists of chlorite-rich green psammitic schist only, which has the largest distribution in this region. The monotonous lithology and thickness suggests deposition in the deep-sea passive margin of the Hangai micro-continent (Buchan et al., 2001).

(7) The Hangai unit consists of monotonous green-colored sandstone and mudstone. Occasionally, Devonian chert (Kurimoto et al., 1997) is intercalated. In this paper we interpret the Hangai unit as the sedimentary cover sediments of the Hangai micro-continent.

The contacts between the above tectonic units are southwest dipping brittle faults. These contacts are mostly NE directed reverse faults, but the exceptions are a brittle detachment fault between the island-arc type Ulaan Sair unit above and the metamorphosed Delb Khairkhan accretionary complex below, and also another detachment fault between the Haluut Bulag accretionary complex above and the metamorphosed Dzag unit below (Fig. 3). These brittle faults are both referred in this paper as part of a D2 event, and not distinguished in chronologically different events (cf., Schoonover and Osozawa, 2004; Osozawa and Pavlis, 2007).

The accretionary units furthermore contain another ductile deformation event differentiated from brittle D2 event, and labeled D1. Absolute D1 time is different among these accretionary complexes, but such relative labeling to the post-Silurian D2 is convenient for analyses of accretionary complexes (cf., Osozawa et al., 2004; Schoonover and Osozawa, 2004; Osozawa and Pavlis, 2007). Ductile D1 deformation associates M1 prograde metamorphism, which generates white micas for pelitic rocks and amphiboles for mafic rocks.

Structural data of D2 and D1 are in Fig. 3. The D1 foliations of the Buld Gol and Delb Khairkhan units are overprinted by a D2 fold with oblique rotation axis relative to D1.

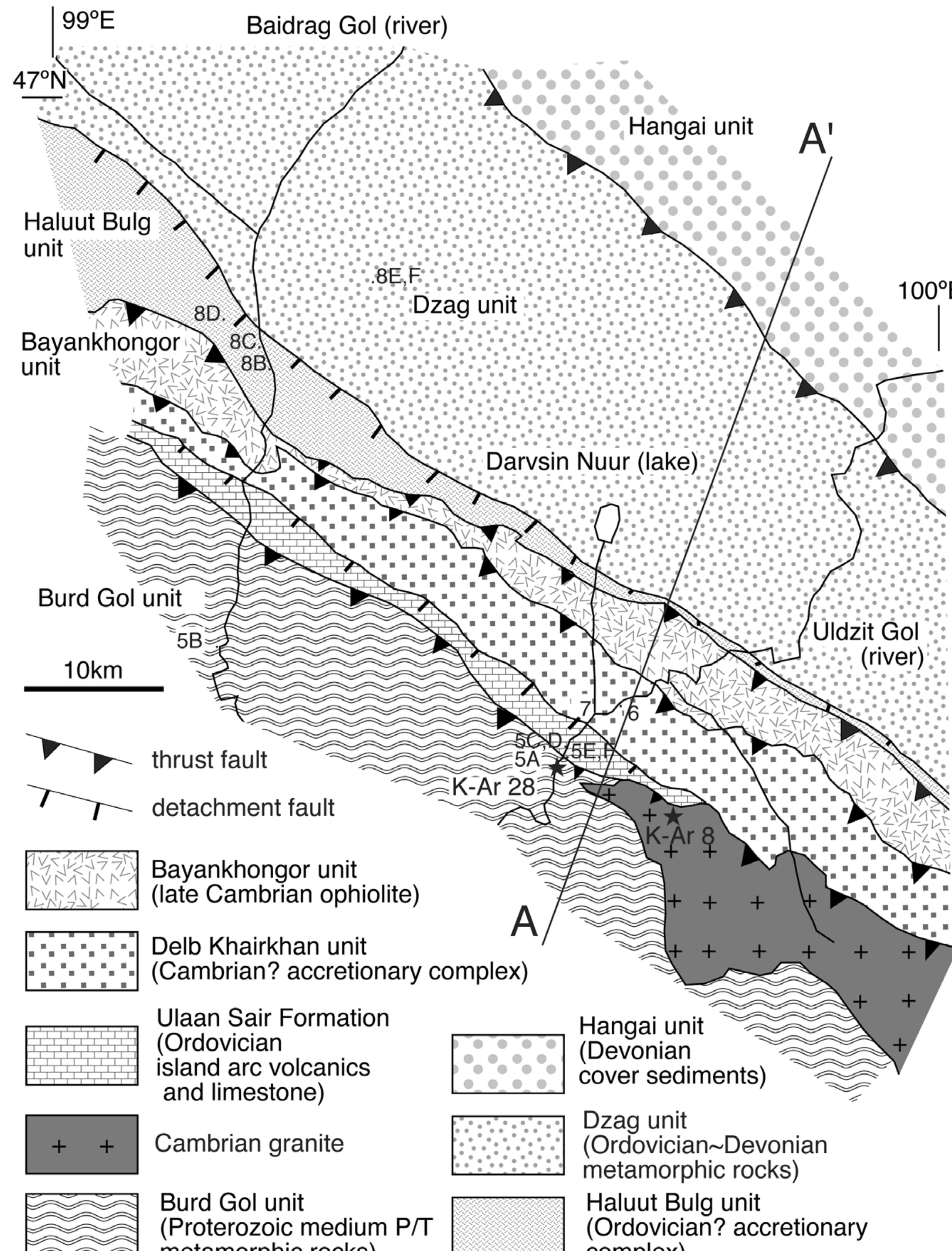

Fig. 2. Geological map of the Bayanhongor region, partly revised after Tomurtogoo et al. (1999). Locations of photo figures are shown (several locations are outside). Sample locations of K–Ar ages are also shown.

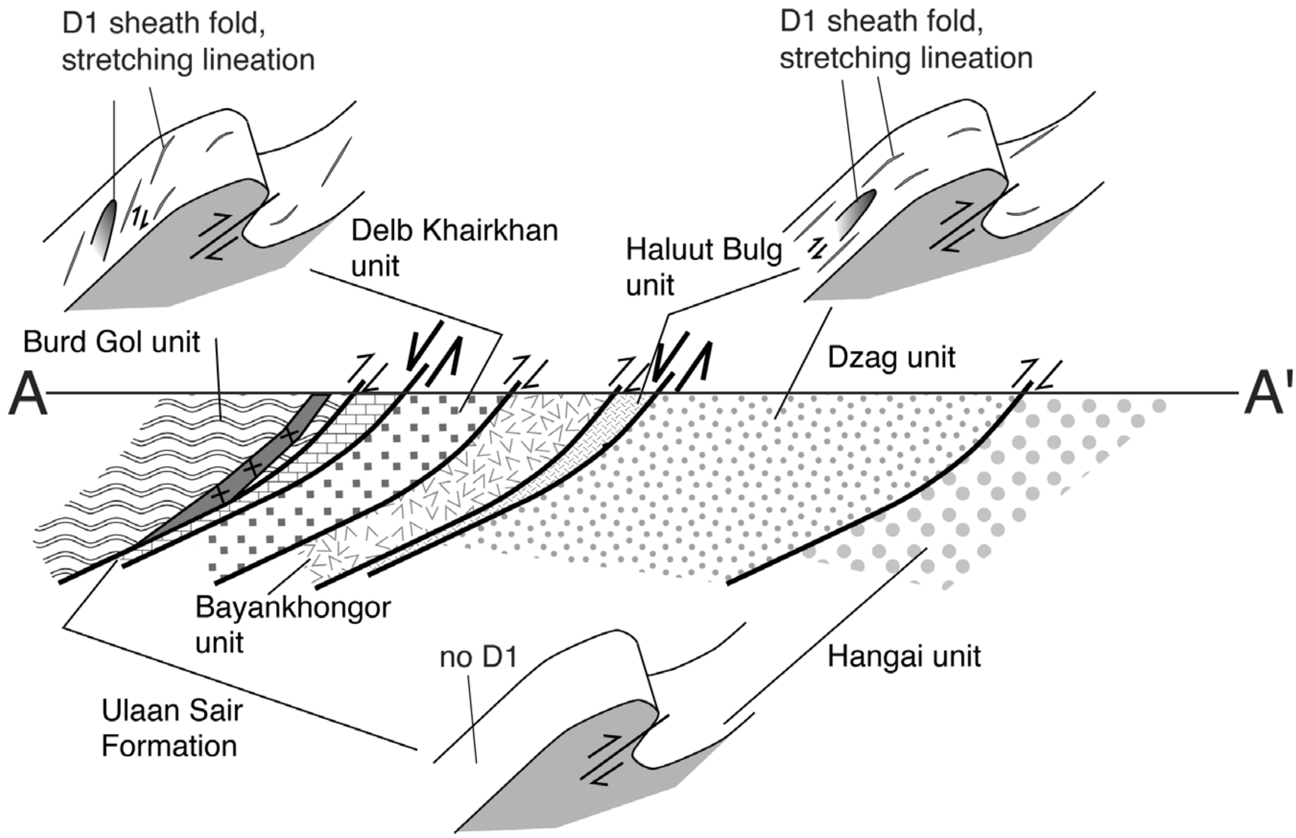

Fig. 3. Idealized NE–SW cross-section of the Bayanhongor region.

3. Lithologies, ages, and structural features

3.1. The Buld Gol unit

The unit comprises pelitic, psammitic, and mafic schist and gneissose rocks. We observed a biotite–calcite–quartz assemblage, whereas Teraoka et al. (1996) reported the occurrence of a hornblende–clinopyroxene–orthopyroxene assemblage, which suggests a granulite-facies metamorphism. Moreover, Buchan et al. (2001) reported biotite–garnet–staurolite and also kyanite–sillimanite assemblages, which indicate Barrovian-type metamorphism.

The psammitic schist with biotite–calcite–quartz assemblage shows a south-southwest plunging stretching lineation, a sheath fold in an outcrop (Fig. 5A), and a top-to-the-north-northeast Riedel shear in a thin section. A north-east-directed D2 asymmetric fold observed in this outcrop bends these D1 structures.

The pelitic schist is a muscovite–biotite–quartz schist. A D1 foliation defined by these M1 metamorphic minerals is crossed by a D2 foliation at an angle of ca. 60° (Fig. 5B). D2 foliations are only dark seams without any newly grown muscovite minerals. A muscovite K–Ar age is 699 ± 35 Ma (Teraoka et al., 1996), which was considered as the minimum age of D1 and M1 of the Buld Gol unit. However, there are discrepancies with a younger U–Pb zircon age of 562 ± 2 Ma for kyanite–sillimanite schist (Kovach et al., 2005, and reference therein) and also ^{40}Ar/^{39}Ar plateau age of 533 ± 3 Ma for biotite from a gneiss (Buchan et al., 2002, and reference therein). Windley et al. (2007) and we ignore the former K–Ar age and prefer the latter biotite age as a minimum age for the metamorphism.

Basalt and dolerite dikes intruded into the D2 anticlinal core of biotite–calcite–quartz schist and elsewhere of the Buld Gol unit. They also cut and post-date the Ulaan Sair Formation, indicating some genetic relation to this island-arc type volcanism (Buchan et al., 2002, and reference therein). Clear chilled margins are observed in these intrusions. No such intrusion is, however, observed in the Delb Khairkhan accretionary complex tectonically below. We could not obtain reliable K–Ar ages for a basaltic dike sample intruded into the Buld Gol schist (Table 1).

K-feldspar granite (Fig. 2) intruding the Buld Gol schist was dated at 539 ± 1, 539 ± 5, 545 ± 2 Ma (^{207}Pb–^{206}Pb zircon evaporation age; Buchan et al., 2002), and 514 ± 10 Ma (^{206}Pb/^{238}U zircon age; Jahn et al., 2004). A granitic sample with the above 545 ± 2 Ma age contains inherited zircons with older ages of 614 ± 6 and 875 ± 3 Ma (Buchan et al., 2002, Fig. 8), and a granite measured by Jahn et al. (2004) also contains many and much older inherited zircons. This Cambrian granite should have a tectonic contact with the younger Ulaan Sair Formation (^{207}Pb–^{206}Pb zircon evaporation age of 474 ± 8 Ma, see below; Buchan et al., 2002) and also the Delb Khairkhan unit, as mapped in Fig. 2. A basalt dike is observed to be emplaced between the Cambrian granite

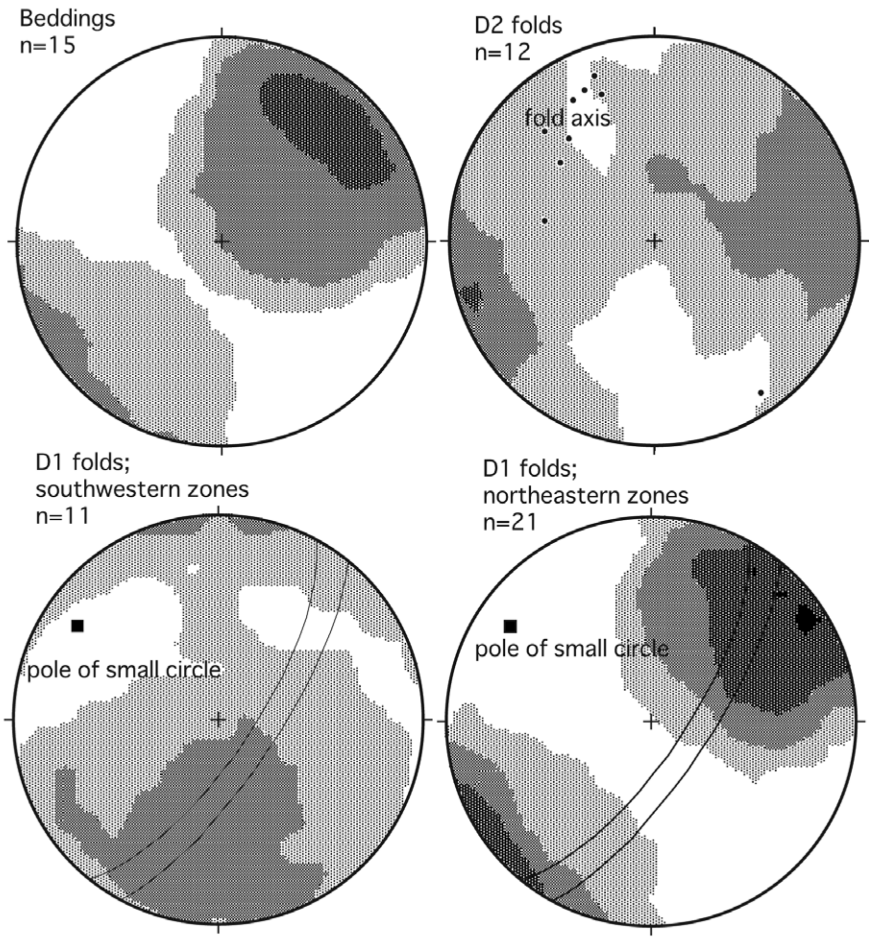

Fig. 4. Structural data projection in the Bayanhongor region.

and Ulaan Sair limestone in an outcrop. We obtained new biotite K–Ar ages for the K-feldspar granite (Table 1), and these 504 ± 22, 509 ± 22, 493 ± 21, and 486 ± 21 Ma represent cooling and hence the exhumation age of the granite.

To the east of the Cambrian older granite younger granites of Permian age (K–Ar age of 268.1 ± 5.6 Ma; Teraoka et al., 2000) are distributed. The granites of the Hangai batholith, out of the mapped area in Fig. 2, have also similar $^{206}Pb/^{238}U$ age of 229.3 ± 5.6 Ma (Jahn et al., 2004). Thus, there are two kinds of granite in the Bayanhongor region.

3.2. The Ulaan Sair Formation

The formation consists of volcaniclastics in the lower part and fossiliferous limestone in the upper part. The lower volcaniclastics are mostly reddish sandstone and conglomerate. This red hematite color may indicate terrestrial sedimentation. The volcanic fragments are basalt to rhyolite, and sometimes contain rounded granitic rocks, quartzites, and limestones. The latter pebble limestones sometimes contain fossils, common to the upper limestone member of the Ulaan Sair Formation. Detrital biotites, likely derived from granitic rocks, are also recognized in the reddish sandstone. According to Buchan et al. (2002), the geochemical characteristics of a rhyolitic dike intruded into the volcaniclastics show an island-arc affinity with clear Nb negative anomaly relative to Th, La, and Ce, in primitive mantle normalized pattern in their Fig. 4. Its $^{207}Pb–^{206}Pb$ evaporation age is 474 ± 8 Ma and 1578 ± 2 Ma, although the latter is from an inherited zircon (Buchan et al., 2002, Fig. 9).

The upper limestone is partly massive and partly bedded. Both types of limestone contain calcareous fossils, such as brachiopods (Fig. 5C), bryozoans, and corals. The fossil assemblages suggest that the limestone is of Carboniferous age (Buchan et al., 2001, and reference therein), but reliable, e.g., foraminifer data are lacking. The bedded limestone is an alternation of calcareous and muddy layers. D2 cleavage without new muscovite crystallization is observed in muddy layers, oblique to bedding at a high dip angle (strike is parallel). This indicates that the strata at this outcrop are not overturned (Fig. 5D). Other than this cleavage, no characteristic deformation was observed in rocks of the Ulaan Sair Formation.

The contact between the Ulaan Sair Formation and the hanging wall (=the Buld Gol schist) is a brittle reverse fault of the D2 phase (Fig. 5E). The reverse fault strikes NNW and dips 60° to WSW. A black gouge, 1.5–2 m thick, occurs between the biotite–calcite–quartz schist and the bedded limestone. The striation on the fault plane is at a high angle plunging to the west-southwest. The footwall

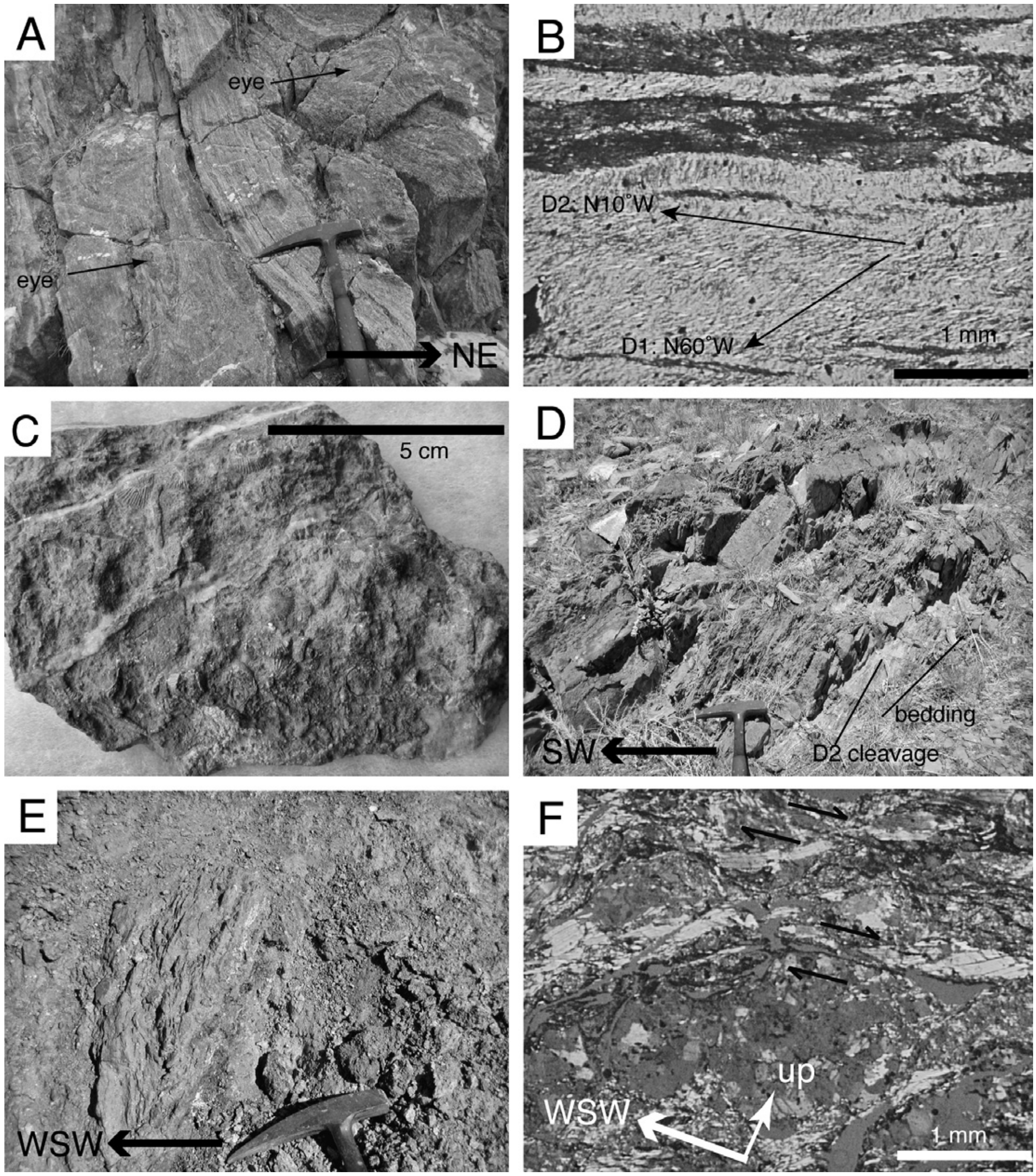

Fig. 5. Structures in rocks of the Burd Gol unit and Ulaan Sair Formation. (A) D1 Sheath fold of biotite–calcite–quartz schist of the Burd Gol schistose unit. Northeastern limb of D2 anticline. (B) D1 (with muscovite crystals) and D2 cleavages of the Burd Gol unit. (C) Brachiopods of the Ulaan Sair limestone. (D) Bedded calcareous sandstone and mudstone of the Ulaan Sair limestone. Cleavage plane (D2) is at high angle to the bedding surface. (E) Reverse fault developed gouge between the Burd Gol schistose unit (left) and Ulaan Sair limestone (right). (F) Microscopic view of above gouge. Biotite of the Burd Gol schistose unit is cut by reverse fault-sense Riedel shears.

Table 1
K–Ar ages of K-feldspar granite and basalt

			K (wt%)	Error	Rad. Ar (10^{-5} cm^3 STP/g)	Error	$^{40}Ar/^{36}Ar$ ratio	Error	Air contamination	Age (Ma)	Error
mo-8-#60-80-1st	Older granite	Biotite	4.52	0.226	10.23	0.081	13695.98	2644.83	2.158	504.96	22.30
mo-8-#60-80-2nd	Older granite	Biotite	4.52	0.226	10.34	0.051	13758.36	545.18	2.148	509.55	22.30
mo-8-#30-60-1st	Older granite	Biotite	5.55	0.278	12.25	0.011	10225.86	1864.66	2.890	493.73	21.96
mo-8-#30-60-2nd	Older granite	Biotite	5.55	0.278	12.05	0.011	5698.81	391.35	5.185	486.55	21.67
mo-28-#60-100-1st	Basalt dike	Whole rock	0.081	0.033	9.19	0.031	355795	13996290	0.083	270.75	100.91
mo-28-#60-100-2nd	Basalt dike	Whole rock	0.081	0.033	8.99	0.032	1322.01	160.73	22.352	265.33	99.09

bedded limestone shows northeast vergent D2 asymmetric folds, associated with axial plane cleavages subparallel to the reverse fault. Microscopic examination revealed that the reverse fault-sense Riedel shears in the foliated fault gouge separate M1 biotite crystals in the biotite–calcite–quartz schist (Fig. 5F).

3.3. The Delb Khairkhan unit

The mafic schist of this complex is an actinolite–hornblende–chlorite–calcite–epidote schist. Some hornblende crystals are found to be rimmed by actinolite, thus we infer the metamorphic facies to be the actinolite greenschist to amphibolite-facies. Limestone is recrystallized and contains no fossils. This is in contrast to the fossiliferous Ulaan Sair limestone. Mafic schist and marble which occur as exotic blocks embedded in the sandy and muddy matrix. Sandstone is also schistose but preserves graded bedding, which is a characteristic of turbidites. Conglomerate shows a schistose and stretched texture. The conglomerate contains well rounded granite clasts.

The pelitic schist is a sericite schist. The D1 foliation is defined by M1 sericite-quartz layers, which are oblique to the D2 foliation. In an outcrop, D1 foliation is N25°W70°SW, and D2 foliation is N80°W75°SW (Fig. 6A). The D1 foliation at the fold limbs is subparallel to bedding defined by sandstone intercalations. The D2 foliation is defined by dark seams without M2 muscovites. Stretching lineation occurs on the D1 foliation surfaces, plunging at a high angle at N65°E80°NE in the above outcrop, and clearly bent by D2 folding (Fig. 6B). Both NNE vergent D1 (Fig. 6C) and NE-vergent D2 asymmetric folds (Fig. 6D) were observed in the same outcrop. Axial planes defined by D1 and D2 foliations are therefore highly oblique to each other.

The northeastern limit of the Ulaan Sair Formation is not a reverse fault as reported by Buchan et al. (2001). The fossiliferous Ulaan Sair Formation containing footwall was detached from the metamorphosed Delb Khairkhan accretionary complex of the hanging wall. The brittle detachment fault bounds a green lapilli tuff to the southwest and marble to the northwest (Fig. 7A). The lapilli tuff is overlain by reddish volcaniclastics, associated with fossiliferous (but age uncertain) limestone pebbles. The detachment fault strikes northwest and dips 70° to the northeast. Some other associated small normal faults in the lapilli tuff dip to the southwest at a high angle. The main detachment fault contains a 10-cm thick reddish foliated gouge (Fig. 7B). Asymmetric minor parasitic folds in this foliated gouge were also observed. Outcrop-scale asymmetric folds involving D2 cleavages were also observed along secondary small faults cutting the lapilli tuff. In thin section of the foliated gouge, a spectacular Riedel shear was observed (Fig. 7C and D). All these asymmetric structures show that the southwestern lapilli tuff slipped down relative to the northeastern marble, constituting a metamorphic core in the Bayanhongor region, although observed faults are at high angle in this outcrop.

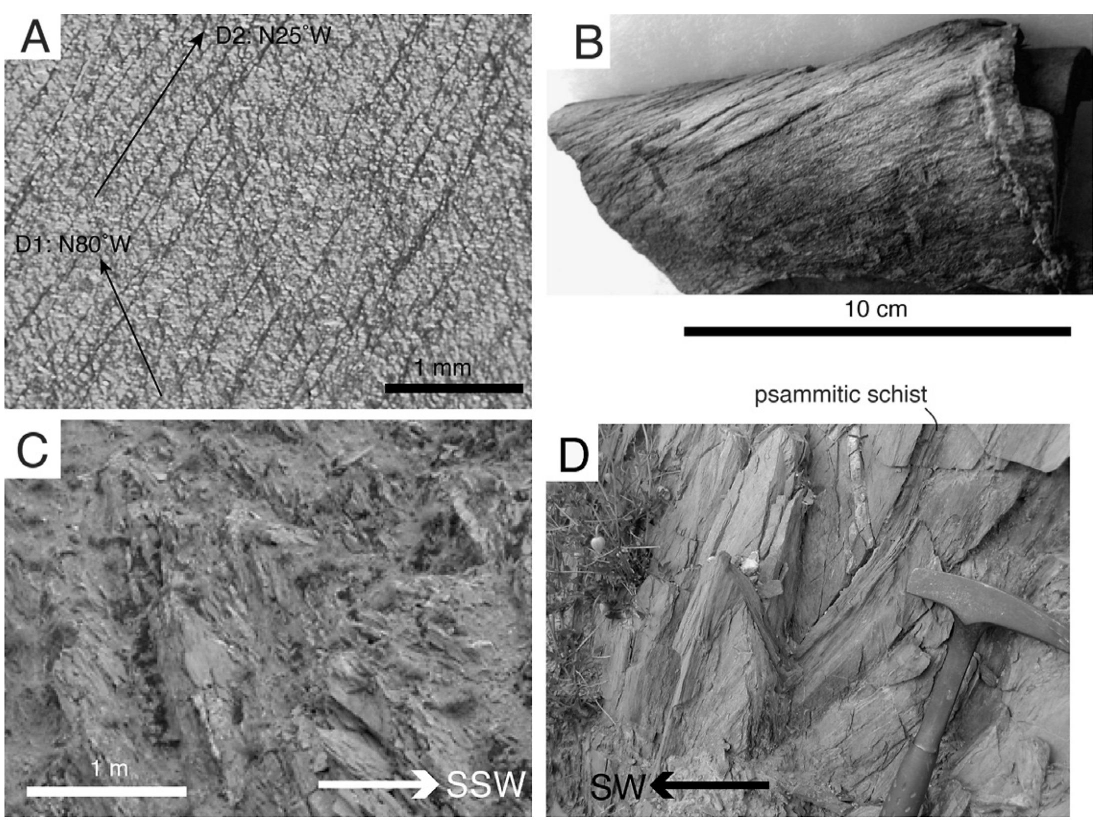

Fig. 6. Structures in rocks of the Delb Khairkhan accretionary complex. (A) D1 (with muscovite) and D2 (without muscovite) pressure-solution cleavages. (B) D1 stretching lineations refolded by D2 fold. (C) N-vergent D1 asymmetric folds with D1 axial planar cleavage. (D) NE-vergent D2 asymmetric folds, with D2 axial planar cleavage.

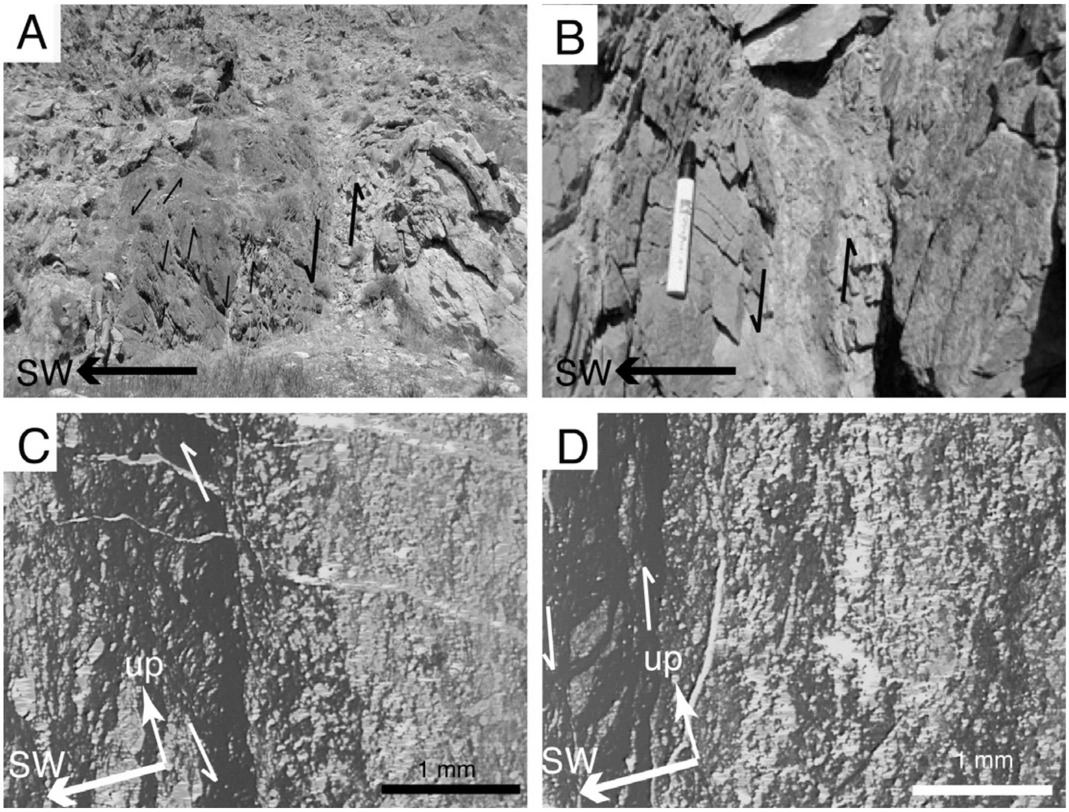

Fig. 7. Detachment fault between the Ulaan Sair volcaniclastics and Delb Khairkhan accretionary complex. (A) Fault between green tuff (left) and white crystalline limestone (right). (B) Reddish, foliated fault gouge. (C and D) Microscopic Riedel shears in the foliated gouge. Normal fault sense of shear.

3.4. The Bayanhongor unit

Pillow lava of the ophiolite consists of vesicular pyroxene basalt. The vesicles are filled with secondary chlorite, epidote, and calcite. We found no hornblende and actinolite. We have not identified any sediment in association with the pillow lava. Consequently, the depositional environment of the overlying sediments of the pillow lava is uncertain. The pillow lava generally dips southwest and is not overturned. However, an exception is found at and near a contact with the underlying Haluut Bulag accretionary complex, where the pillow lava sequence was overturned, forming an asymmetric anticline (Fig. 8A). The observed direct contact is a D2 gouge fault associated with quartz and calcite veins. Also a black gouge, 1-m thick, was observed at the upper contact with the Delb Khairkhan unit.

A sheeted dike complex unit is underlain by pillow lava unit. We could not survey the deep valley of the Uldzit Gol (Fig. 2) and unfortunately did not encounter the gabbro and ultramafic units of the ophiolite.

An Ar–Ar age of 484.5 ± 5.9 Ma was obtained for amphibole forming a mineral lineation within pillow basalt (Buchan et al., 2002, and reference therein; the mineral lineation is comparable to the D1 stretching lineation observed in the Delb Khairkhan unit). Another available age data constraint is a Sm–Nd isochron age of 569 ± 21 Ma, obtained for gabbros (Kepezhinskas et al., 1991). However, data and a much older $^{206}Pb/^{238}U$ age of 665 ± 15 Ma for anorthosite is presented in Kovach et al. (2005). Combined with E-MORB affinity in primitive mantle normalized pattern (Kovach et al., 2005, Fig. 4), the meaning of this age is discussed in Section 4, and was partly discussed in Windley et al. (2007).

3.5. The Haluut Bulag unit

This accretionary complex shows a typical block-in-matrix structure. The exotic blocks, several meters to tens of meters in size, include limestone, basalt (Fig. 8B), bedded radiolarian chert, and varicolored siliceous mudstone.

Some basalt is vesicular, and the vesicles are filled with chlorite. This type of basalt is covered by limestone, which is of sea-mount origin (Xenophontos and Osozawa, 2004). A basaltic dike also occurs within the vesicular basalt. Limestone is crystalline, but occasionally contains some mega-fossils. Basaltic tuff frequently occurs in association with basaltic lava and limestone.

The basaltic tuff and limestone show a D1 cleavage, and further involve NE-vergent D2 asymmetric folds with axial planar cleavages. A first-order asymmetric D2 fold was also observed, with a wavelength of hundreds of meters, involving a basaltic layer of more than 10 m in thickness.

Reddish radiolarian chert is a typical pelagic rock. It sometimes alternates with bedded limestone. Its depositional age cannot be determined due to the poor state of

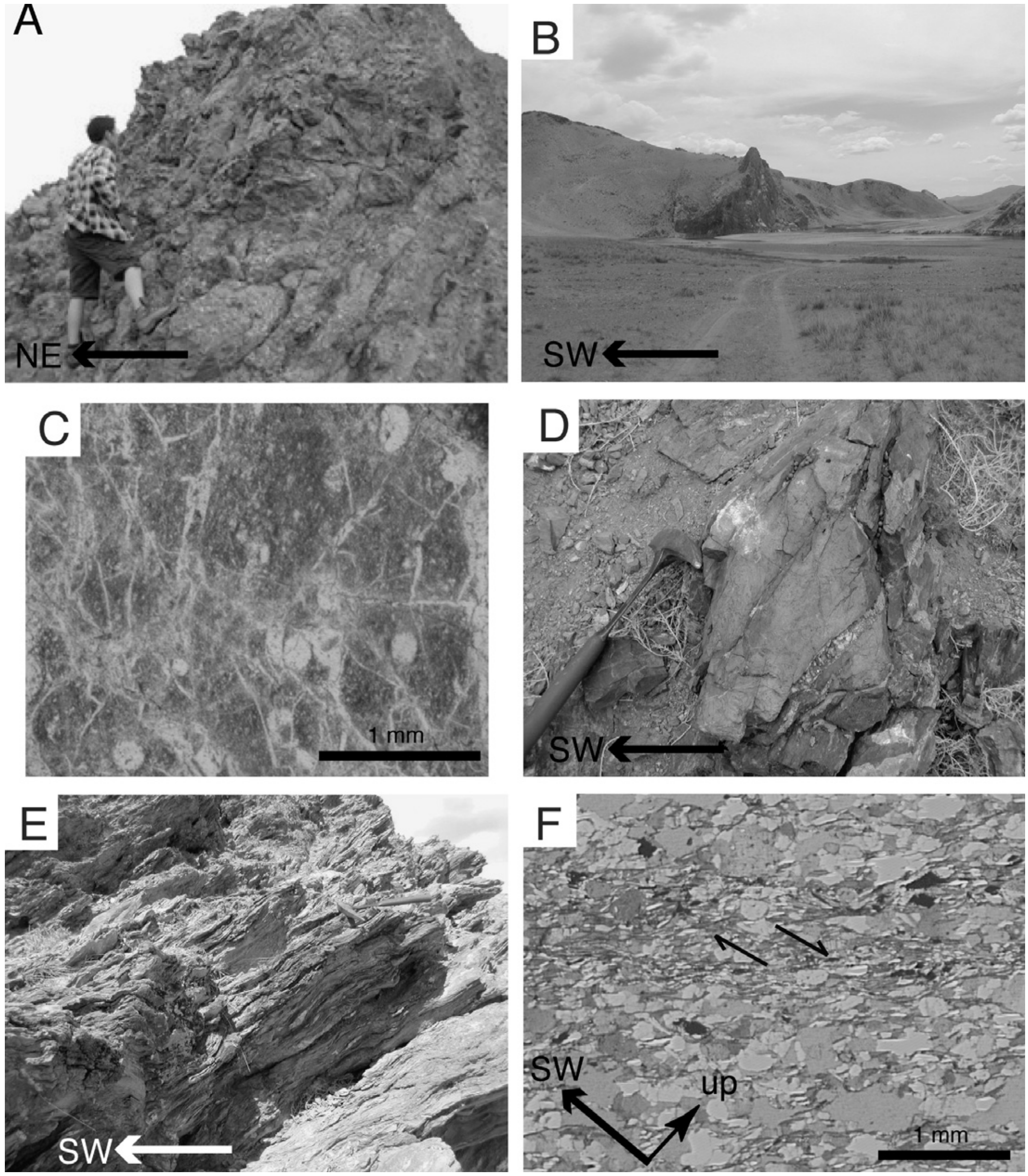

Fig. 8. Structures in rocks of the Bayanhongor, Haluut Bulag, and Dzag units. (A) Pillow lava of the Bayanhongor ophiolite. Top (left) is northeast, to the southwest of mapped area in Fig. 2. (B) Limestone and basalt block of the Haluut Bulag accretionary complex. (C) Radiolaria in chert of the Haluut Bulag accretionary complex, viewed in YZ section cut orthogonal to stretching lineation. (D) D1 axial planar cleavage in sandstone and mudstone of the Haluut Bulag accretionary complex. (E) D2 NE vergent asymmetric folds of the Dzag schistose unit. (F) Microscopic view of chlorite–muscovite–calcite–quartz schist of the Dzag schistose unit. D1 reverse fault sense Riedel shear is shown.

preservation of radiolarians (Fig. 8C). However, the first known occurrence of radiolaria is within the Cambrian, and the basalt age as well as accretion age of the Haluut Bulag unit is thus no older than Cambrian. These radiolaria are constricted to form D1 stretching lineation also in the Haluut Bulag unit. Red, pale green, and black siliceous mudstone also crop out, and these can be attributed to hemipelagic mudstone. Pale green mudstone is originally silicic tuff derived from an island arc. Such silicic tuff is an important component of hemipelagic sediments within accretionary complex (Xenophontos and Osozawa, 2004).

Sandstone corresponds to the terrigenous sediments of the accretionary complex. It rarely contains plant fragments up to 1 cm of two dimensions and the fossils again require that the unit is younger than Ordovician. Alternation of sandstone and mudstone defines a turbidite, and shows asymmetric northeast vergent folds with axial planar D1 cleavage (Fig. 8D). In contrast to the Burd Gol and Delb Khairkhan units, D1 and D2 cleavages in the Haluut Bulag unit and the following Dzag unit have the same strike to the northwest.

3.6. The Dzag unit

The widespread Dzag unit (Fig. 2) is homogeneous in lithology and chemical composition (Jahn et al., 2004). It is a chlorite–muscovite–calcite–quartz schist. The chlorite crystals exhibit a green pleochroism, and show anomalous interference colors under crossed nicols. The macroscopic green color of this psammitic schist is due to the large amount of metamorphic chlorite. Plagioclase crystals are relicts, and a relict crystal of garnet was also observed. The protolith of the Dzag schist is thus such a sandstone associated with some mafic detrital minerals or rocks. The psammitic schist geochemically represents a homogenous and mature crustal rock (Jahn et al., 2004). The Dzag schist has reported K–Ar muscovite metamorphic ages of 395 ± 20, 440 ± 22 (Teraoka et al., 1996), 447 ± 9, and 454 ± 9 Ma (Kurimoto et al., 1998).

A chlorite–muscovite rich layer, which defines the D1 foliation, is truncated by a Riedel shear, with a top-to-the-northeast sense of shear (Fig. 8F). Stretching lineations plunging to the southwest are visible orthogonal to the strike of D1 foliation. The D1 foliation is folded by the NE-vergent D2 asymmetric fold (Fig. 8E). The D2 fold axis generally trends northwest, and the axial plane cleavage dips southwest (Fig. 3). The vergence is northeast, and this D2 cleavage dips at an angle smaller than D1 foliation at the overturned limb of a first-order D2 asymmetric fold.

Unfortunately, no fault contact with the hanging wall Haluut Bulag accretionary complex has been observed. However, near the contact, the D2 cleavage always dips at lower angle than D1 foliation in the Dzag unit, and the fold asymmetry is oppositely southwest. In this case, we explain the SW directed motion as a consequence of brittle detachment faulting and associating mesoscopic parasitic folding. In addition, the metamorphic facies of both units are similar, i.e., greenschist-facies, but the footwall Dzag schist is clearly recrystallized at higher *P–T* conditions than the hanging wall Haluut Bulag accretionary complex, therefore no reverse fault should be expected.

3.7. The Hangai unit

Sedimentary rocks are green sandstone and mudstone (Fig. 9A). The green color is due to detrital chlorite within the sandstone. Other lithic fragments include subrounded quartz, plagioclase, and microcline, derived from granite–gneiss, altered volcanic rocks, epidote schist, and muscovite–chlorite schist, some of which is internally folded (Fig. 9B). Without any age constraints, these sediments, if further metamorphosed, could become the psammitic schist of the Dzag unit. Conversely, the fragments of muscovite–chlorite schist could have been derived as clasts from the Dzag schist.

The sedimentary rocks contain bedded radiolarian chert (Fig. 9C). Although Teraoka et al. (1996) and Kurimoto et al. (1997) considered this chert as exotic blocks included

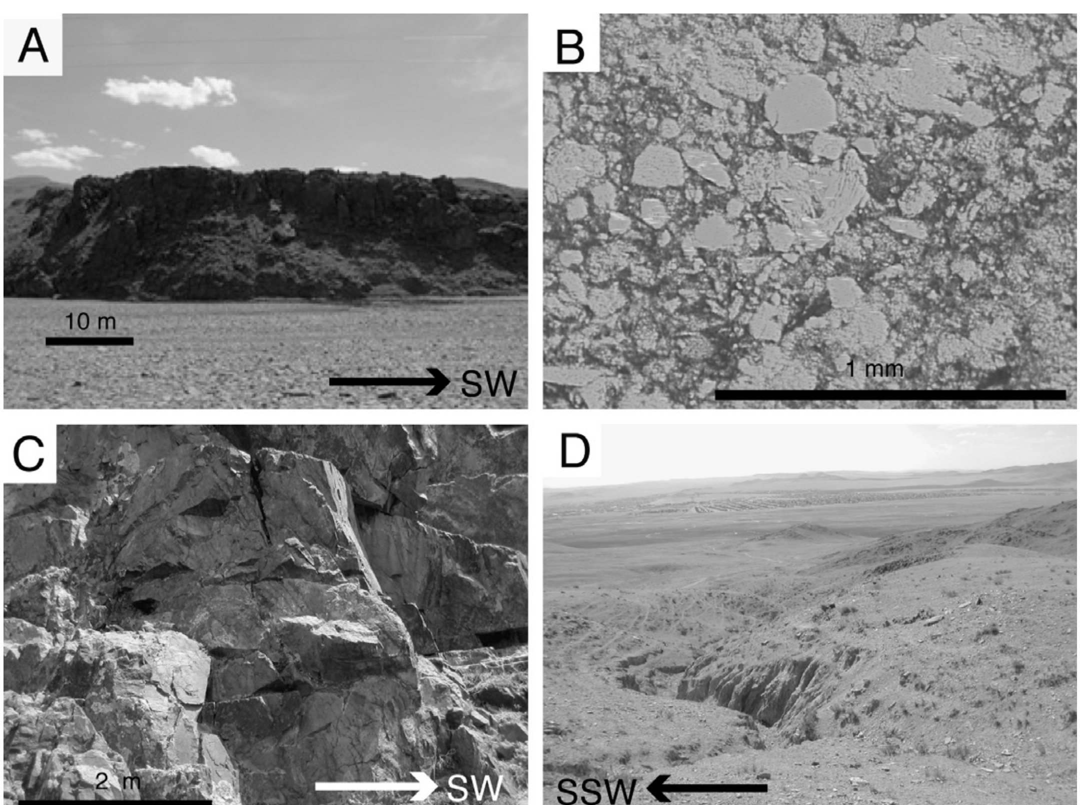

Fig. 9. Structures, rocks, and an upper contact of the Hangai cover sediments. Locations are to the southwest of mapped area in Fig. 2. (A) A NE-vergent asymmetric fold of sandstone of 10-m thick of the Hangai unit. (B) Sandstone of the Hangai unit containing folded muscovite–chlorite schist (center), chlorite, and others. Dark seams are also observed. (C) Radiolarian chert in the Hangai sandstone and mudstone. (D) Reverse fault between the Dzag schistose unit (left) and the Hangai cover sediments (right), background is Bayanhongor, to the southwest of mapped area in Fig. 2.

in an accretionary complex, the chert gradationally changes upward to green mudstone and then downward to green sandstone. Kurimoto et al. (1997) found late Devonian conodonts in the chert. The Hangai sedimentary rocks are thus likely to be younger than the metamorphism of the Dzag unit, and cannot be the source.

Some dark seams are recognized in thin sections and represent the axial planes of asymmetric folds. A folded sandstone was observed in an outcrop, and the asymmetric fold is NE vergent (Fig. 9A). However, in this study area, bedding in the Hangai sedimentary rocks is usually difficult to recognize. Thus, the structure is uncertain in most cases.

We found a D2 reverse fault contact with the hanging wall Dzag unit near Bayanhongor City (Fig. 9D). The Hangai sedimentary rocks do not cover the Dzag schist with an unconformity. The brittle reverse fault is N80°W35SW, and accompanied by a 1-m thick gouge.

4. Discussion

Formation of the Bayanhong region is a consequence of collision between the Baidrag and Hangai micro-continents. Determination of which continent collided relative to the other is important, because it reflects the geotectonic history of the Central Asia Orogenic Belt relative to the nuclear Siberian craton and also the formation process of the major syntaxis in Mongolia (e.g., Windley et al., 2007). As already stressed by Buchan et al. (2001), the Hangai collision relative to the Baidrag is clarified. With a little discrepancy of the Ulaan Sair Formation, formative and prograde metamorphic ages in each unit tend to be younger toward the northeastern Hangai unit, if we consider 562 ± 2 Ma (Kovach et al., 2005, and reference therein) and 533 ± 3 Ma (Buchan et al., 2002, and reference therein) for the Burd Gol unit, 539 ± 1, 539 ± 5, and 545 ± 2 Ma for older granite (Buchan et al., 2002), 474 ± 8 Ma for the Ulaan Sair Formation (Buchan et al., 2002), 484.5 ± 5.9 Ma for the Bayanhongor ophiolite (Buchan et al., 2002, and reference therein), post-Cambrian radiolarian age for the Haluut Bulag unit (this paper), 447 ± 9, and 454 ± 9 Ma for the Dzag unit (Kurimoto et al., 1998), and late Devonian radiolarian age for the Hangai unit (Kurimoto et al., 1997). Furthermore, stratigraphic and D1 structural tops are always to the southwest, except for the overturned fold limb.

At the time of northwestward accretionary growth of each unit, most of these units experienced D1 ductile deformation and M1 prograde metamorphism. The D1 phase is not synchronous, but also younger toward the northeastern unit reflecting the northeastward growth, as partly dated above. On the other hand, D2 gouge faults and associated folds are more brittle, and we infer that they represent the closure of an ocean between micro-continents and collision, followed by exhumation of metamorphic units (including older granite). Ductile D1 is thus only recorded in accretion related units, but brittle D2 is in all units affected by collision. The detachment is included in brittle D2 structure similar to relationships in Japan (Schoonover and Osozawa, 2004; Osozawa and Pavlis, 2007), and the South Tibetan Detachment system (e.g., Burchfiel et al., 1992; Hodges et al., 1998) although in Tibet the early phase also records significant ductile shear (e.g., Carosi et al., 1998).

In the study area, the Buld Gol unit is the oldest metamorphic rock and is Proterozoic in age. We interpret the schist as defining an accretionary complex in view of its lithological characteristics. The Buld Gol schist was first accreted to the core buttress of the Baydrag micro-continent, which formed the upper plate. Although Barrovian-type metamorphism (Buchan et al., 2001) is rare in accretionary complexes, it might have been overprinted in response to the increasing accreted mass, associated partial melting, and consequent Cambrian older granitic intrusive activity.

Both the Delb Khairkhan and Haluut Bulag units are lithologically and structurally suggestive of accretionary complexes, as also pointed out by Buchan et al. (2001). It is important to emphasize that the Central Asia Orogenic Belt contains widespread accretionary complexes (Sengor and Natal'in, 1996). Thus, the Buld Gol, Delb Khairkhan, and Haluut Bulag accretionary complexes formed in front of the Baydrag micro-continent and have not been destroyed; instead, they preserved the record of an ocean between the micro-continents. In particular, the vesicular basalt covered by limestone in the Haluut Bulag unit suggests an origin as a sea-mount (Xenophontos and Osozawa, 2004), which existed in the later closed, relatively wide ocean. Existence of pelagic chert also shows that the ocean was matured enough for development of pelagic sedimentary assemblages.

If the $^{206}Pb/^{238}U$ age of 665 ± 15 Ma by Kovach et al. (2005) were the true age, a time lag of 180 m.y. exists between generation and emplacement (485 Ma) of the Bayanhongor ophiolite. Ophiolites are generally known to be emplaced shortly after their formation (e.g., Nicolas, 1989). The ocean between two micro-continents was probably wide as mentioned above, but this apparent age span of 180 m.y. is longer than the present age of the Pacific Ocean and thus, is probably unreasonable. Radiolarian ages of pelagic chert in the Haluut Bulag accretionary complex also indicates the underlying basalt is not older than Cambrian. When we compare broken zircon grains coated by ferruginous material (Kovach et al., 2005) with clear crystals shown by Buchan et al. (2002), the measured zircons (Kovach et al., 2005) of the Bayanhongor ophiolite suggest they are affected by inheritance. Inherited zircons are not rare in ophiolites and oceanic crust, and are explained by recycling, metasomatism, or assimilation process (Liati et al., 2004).

The probable late Cambrian Bayanhongor ophiolite intervenes between the two accretionary complexes. The footwall is not the Dzag schistose unit which represents a passive margin sediment of the Hangai micro-continent (Buchan et al., 2001), but the Haluut Bulag accretionary complex. Therefore, the Bayanhongor ophiolite has never been directly obducted onto a passive margin, but included in an accretionary complex. Also a metamorphic sole, which normally characterizes Tethyan-type ophiolites (Nicolas, 1989), was not observed. We consider it to represent a subducted but partially accreted mid-ocean ridge or a very young oceanic crust and mantle, as proposed for the Japanese ophiolite in accretionary complexes by Xenophontos and Osozawa (2004) and Osozawa et al. (2004). Although there are no data for the Bayanhongor ophiolite, the Japanese ophiolite is usually directly overlain by hemipelagic mudstone lacking pelagic chert (Osozawa et al., 2004), which shows a close-to-trench spreading of the mid-ocean ridge. If we assume this scenario, an ancient mid-ocean ridge was subducted under the Baydrag micro-continent, emplaced as an ophiolitic accretionary complex in D1, and during D2, obducted onto the Hangai micro-continent (Fig. 10). As in the Tethyan-type ophiolites (e.g., Nicolas, 1989), the Japanese ophiolites contain island-arc type basalts and extremely silicic rocks as well as intra-plate type basalts (Osozawa and Yoshida, 1997; MORB like basalts may be included). Such abnormal composition is also expected for the Bayanhongor ophiolite, and indeed the ophiolite geochemically resembles E-MORB (Kovach et al., 2005).

If the Bayanhongor ophiolite was formed at a spreading center, former accretion of the Delb Khairkhan accretionary complex reflects the plate motion of a frontal and Baidrag side oceanic plate from the mid-ocean ridge, and accretion of the Haluut Bulag unit probably represents the plate motion of Hangai side oceanic plate at the rear of the mid-ocean ridge (Fig. 10 top). Both plate motions were different due to active sea-floor spreading of the mid-ocean ridge (see Fig. 10 top, velocity triangle; Osozawa, 1992), and produced different trends of D1 accretionary structures such as cleavages, although D1 stretching itself is a common event through the units. The motion of the Hangai micro-continent was the same as the oceanic plate of the rear side, and D2, formed at the time of continental collision, had the same trend as D1 in the Haluut Bulag accretionary complex. However, D1 and D2 trends are different in the Delb Khairkhan accretionary complex. This consideration can be applied to the Buld Gol schistose unit, reflecting different plate motions like the former Delb Khairkhan accretionary complex, contrary to reflecting the same motion as the latter Haluut Bulag complex and the Dzag schistose unit.

D1 and D2 thus define deformation phases at the time of accretion and collision, respectively. Folding of consolidated basaltic lava of the Bayanhongor ophiolite and the Haluut Bulag complex was performed as a consequence of the D2 collisional event. The Ulaan Sair Formation and the Hangai cover sediments only record one phase of deformation. These rocks are not metamorphosed and have not experienced accretionary deformation; the deformation phase corresponds to only D2 at the time of continental collision.

The Ulaan Sair Formation is interpreted as a late Ordovician volcanic arc, with a possibility of near-trench volcanic activity (Windley et al., 2007). Volcanic ash in the hemipelagic mudstone of the Haluut Bulag accretionary complex was derived from the Ulaan Sair arc volcanoes.

Tectonically, the mildly metamorphosed Ulaan Sair Formation constitutes the hanging wall of the metamorphosed Delb Khairkhan accretionary complex, bounded by a detachment fault. The Delb Khairkhan complex was exhumed along this detachment fault and basal reverse fault (Figs. 3 and 9), thus the exhumation process is like an extrusion wedge found elsewhere in subduction zones (Ring et al., 1999; Schoonover and Osozawa, 2004; Osozawa and Pavlis, 2007). Probably, the Ulaan Sair rocks were originally deposited on the Delb Khairkhan complex as cover sediments like in a fore-arc basin. Similarly, the Dzag schist, which are accreted and metamorphosed passive margin sediments in front of the Hangai micro-continent, was exhumed as a result of both detachment faulting and reverse faulting, also as another extrusion wedge. Here, both of the exhumed rocks in the extrusion wedge are not metamorphosed crystalline crustal rocks, as typically observed in large collisional orogens, but metamorphosed accretionary and passive margin rocks such as the Japanese examples (Schoonover and Osozawa, 2004; Osozawa and Pavlis, 2007). The pre-final collisional events are thus well preserved in this part of the Central Asia Orogenic Belt.

If the hypothesis is accepted that the Bayanhongor region as a whole may have experienced major strike-slip faulting (Buslov et al., 2004), the southwestward collision of the Hangai continent toward the Baidrag continent may not be primary, and also its present geographic position relative to the Siberian continent may be secondary. However, its preserved internal structure completely lacks a lateral component of shear and reflects the original accretionary history. The other word, crustal-scale syntaxis may be a primal structure only formed by subduction of the Hangai-Hantey ocean (Sengor and Natal'in, 2004) or Mongol-Okhotsk ocean (Badarch et al., 2002).

The Khantaishir and Dariv ophiolites are opposite and southwestern side of the Baidrag continent (Fig. 1). The Pb–U data plots define a discordia line, and the mean $^{207}Pb/^{206}Pb$ ages are 568 ± 4 and 571 ± 4 Ma, respectively (Khain et al., 2003). The Khantaishir ophiolite contains boninitic rocks (Bayarmandal, 2007), and differs from the present Bayanhongor ophiolite with E-MORB affinity (Kovach et al., 2005). These ophiolites and the Bayanhongor ophiolite are therefore not directly correlated. Ocean of the Bayanhongor region existed at least until Devonian time, but the former ophiolites suggest another ocean (the paleo-Asian ocean) existed until the end of Permian time (Xiao et al., 2003). The Khantaishir and Dariv ophiolites obducted onto the northeastward Baidrag micro-continent (present coordinates; Tuva-Mongoria block), although the polarity or subduction direction was later changed northeastward (present coordinates; Windley

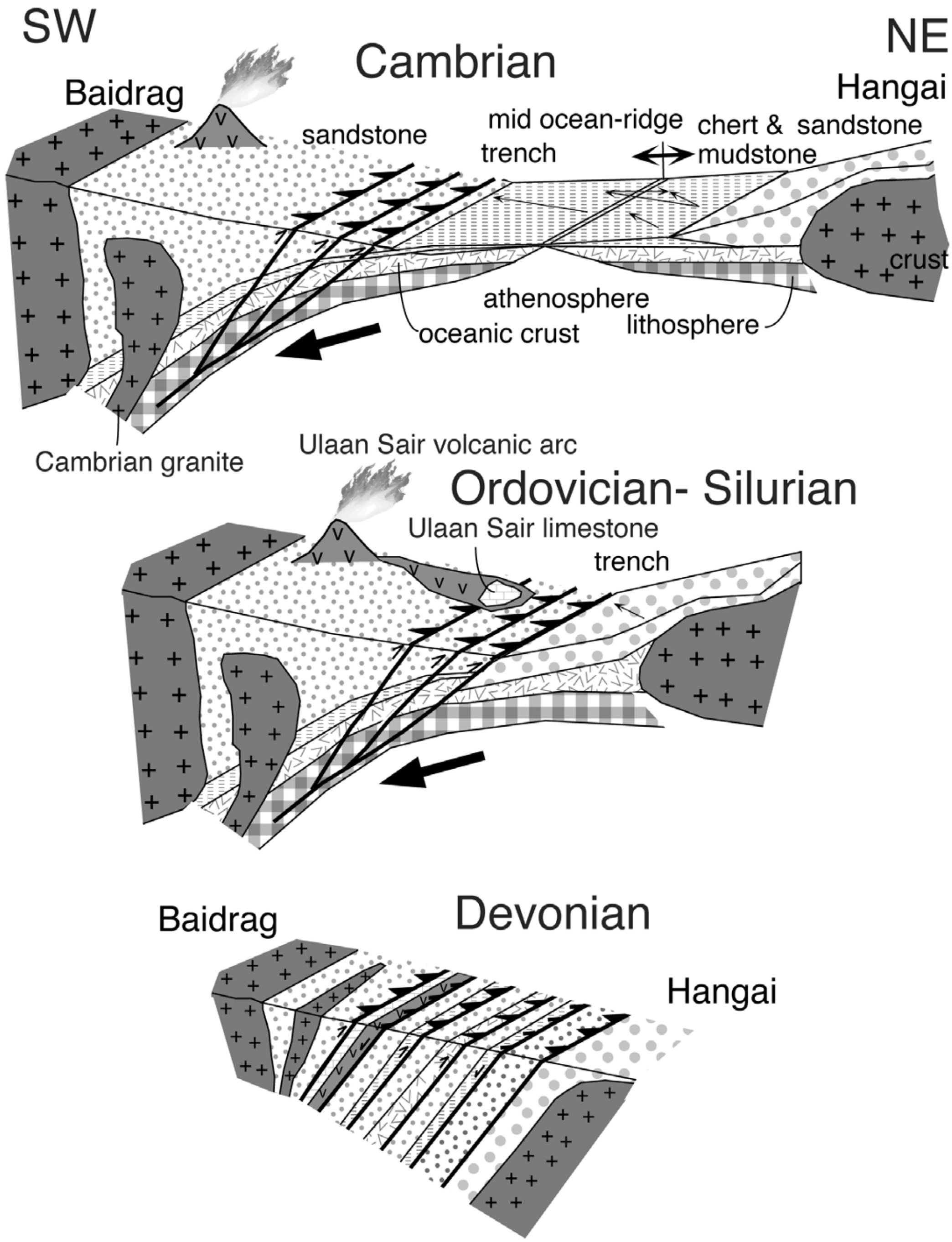

Fig. 10. Accretion and collision process of the Baydrag-Hangai area. Due to sea-floor spreading, oceanic plate motion (a part of arrows) relative to the Baydrag micro-continent is different between the fore and rear plates intervening the mid-ocean ridge, and affects the structural trends. The Ulaan Sair fragments are derived from volcanic arc, deposited in relatively fore-arc area, and finally involved in D2 collisional deformation only. Part of passive margin sediments of the Hangai micro-continent (the Dzag unit) was suffered M1 metamorphism like accretionary complexes. Final collisional event, but labeled D2, is an exhumation for such metamorphic rocks.

et al., 2007). The fate of this subduction zone and paleo-Asian ocean to form the Solonker suture (Windley et al., 2007, Fig. 2) is discussed by Xiao et al. (2003). The leading edge of the Baidrag micro-continent is the Bayanhongor region and then the Hangai micro-continent treated in this paper, and these grew northeastward with southwestward subduction and collision (present coordinates) through early Paleozoic, which was followed by the Triaasic closure of the Hangai-Hantey or Mongol-Okhotsk ocean recorded in eastern Mongolia (Ehiro, 2006).

5. Conclusion

In the Central Asian Orogenic Belt of the Bayanhongor region, west-central Mongolia, accretionary complexes including an ophiolite, a volcanic arc unit, and passive margin units developed widely between the southwestern Baydrag and the northeastern Hangai micro-continents. The belt had grown northeastward, and the Hangai continent collided with the Baydrag continent in the Devonian D2 phase of deformation. Other than sea-mounts, the ocean between the continents had a mid-ocean ridge, which has a genetic link to the Bayanhongor ophiolite. Two sets of oceanic plates were separated by the actively spreading mid-ocean ridge, hence each had different plate motions, and produced different stress fields and ductile D1 structural trends in accretionary and metamorphic complexes. Generally, two periods of deformation, D1 at the time of accretion and another brittle D2 at collision, could be recognized in each complex. In contrast, the cover sediments have recorded only one phase of deformation namely the D2 event associated with the collision. The metamorphosed accretionary complex and also passive margin complex were exhumed to form extrusion wedges. This paper describes the first brittle detachment faults discovered in Mongolia. Strike-slip faulting is completely lacking, and the crustal-scale syntax in western Mongolia could be original to the margin.

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