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Geological complexes in the margin of the Siberian Craton as indicators of the evolution of a Neoproterozoic Supercontinent

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Abstract. This paper describes the Late Precambrian geologic complexes from the southern margin of the Siberian Craton, associated with the extension epochs. Analysis of the data available suggests that there were two episodes of intracontinental breakup, which resulted in the opening of the ocean (1300–900 and 850–550 million years ago). The time sequence of the "rift-related volcanic rocks and terrigenous deposits → basic dike swarms → carbonate-terrigenous rocks → ophiolites and island-arc rocks reflects the successive change of geodynamic environments in the marginal part of the craton. The stage of intracontinental rifting was superseded by the stage of advanced rifting which preceded the continent break and the formation of oceanic crust. This period was followed by two phases of oceanic evolution: a passive phase (sedimentary rocks of the passive margins) and an active phase (island arcs, backarc seas, and the like). Several different versions are offered and discussed for the extension processes in the southwestern and southeastern parts of the Siberian Craton is association with the breakup of the Rodinia Continent.

Introduction

The amalgamation of continental blocks into large structural elements of a planetary scale (supercontinents) is among the most interesting features of the Earth's evolution. It was proved that supercontinents had existed since the Early Precambrian [*Dalziel,* 1991; *Hoffman,* 1991; *Moores,* 1991; *Rogers,* 1996]. One of the best documented supercontinents is Rodinia which had originated in the Mesoproterozoic and disintegrated in the Neoproterozoic. A great number of publications appeared in the foreign and a lesser number, in the Russian literature, dealing both with the general reconstructions of the supercontinent [*Dalziel,* 1991; *Hoffman,* 1991; *McMenamin and McMenamin,* 1990; Rogers, 1996, to name but a few] and with its individual cratonic blocks, including the Siberian one [*Condie and Rosen,* 1994; *Sklyarov et al.,* 2000]. These reconstructions, which are fairly similar in many respects, include some ancient cratons which do not find, like in children's mosaic, their own, exactly fixed places. This situation owes its origin, in many respects, to their poor knowledge and to the contradictions of the results obtained for their different segments. The Siberian Craton seems to have the greatest number of its "degrees of freedom", this stemming from a great number of reasons, the most important of them being the poor geochronological knowledge of its indicator metamorphic, magmatic, and sedimentary rocks and also the fact that the resulting data were published mostly in local press. For this reason the aim of this paper is to generalize the indications imprinted in the structure of the southern part of the Siberian Craton, which may provide information on the breakup of the Rodinia Continent in the Neoproterozoic.

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The following rock complexes are usually used to prove the breakup of large cratonic blocks [*Rogers,* 1996]:

1. Swarms of basic dikes in the marginal parts of the cratons, which are known to be the important indications of the early processes of a continent breakup. The presence of these dike swarms does not necessarily mark a continent breakup and the formation of an oceanic space, because rifting can only lead to the formation of intracontinental structural el-

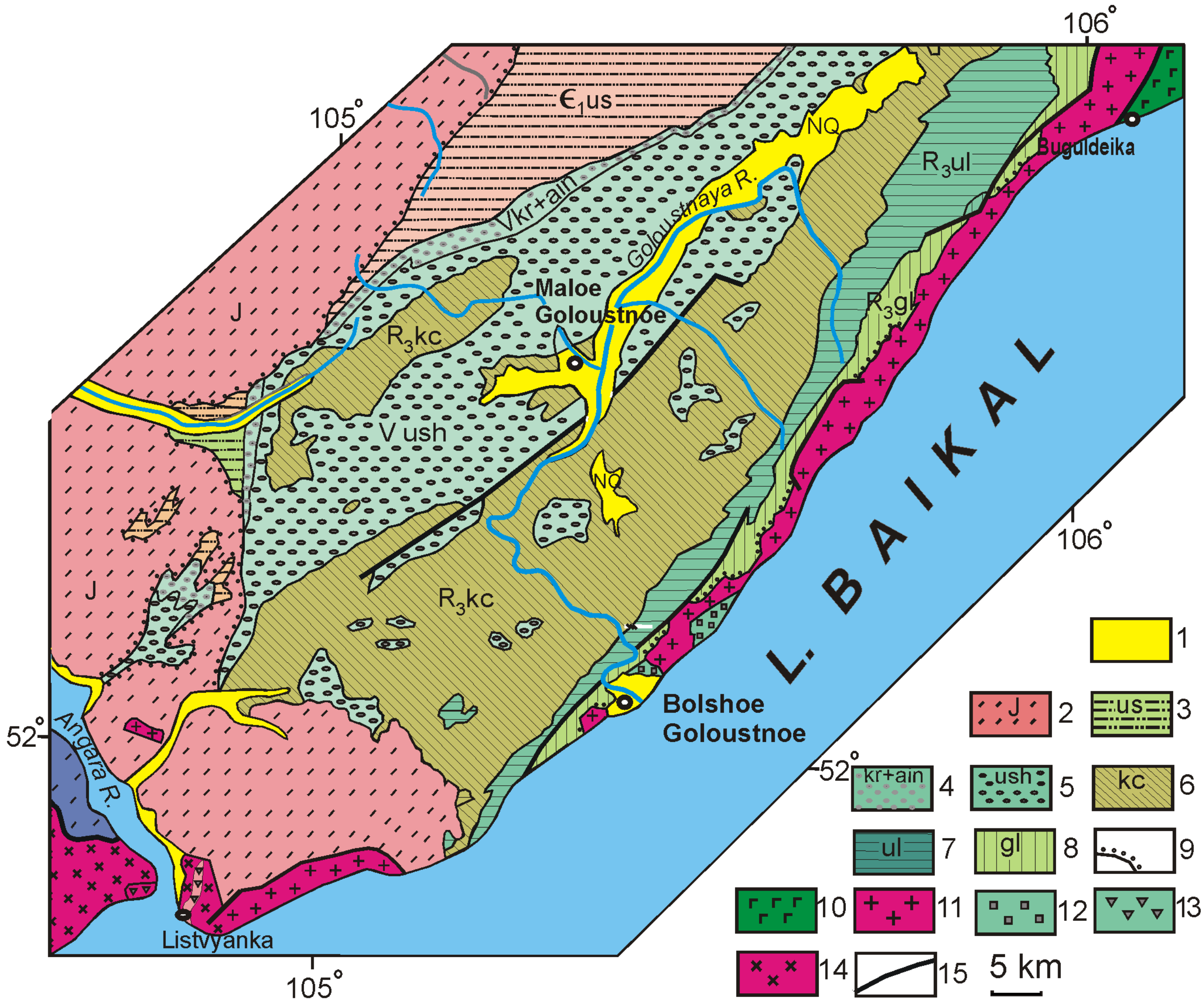

Figure 1. Geological map of the West-Baikal region – a stratotype of the Neoproterozoic sedimentary deposits of the Baikal Group.
1 – Cenozoic sediments; 2 – Jurassic sediments, undifferentiated; 3 – Usolie Formation (Early Cambrian); 4 – Kurtun and Ayanka formations (Vendian); 5 – Ushakov Formation (Vendian); 6–8 – Late Riphean rocks of the Baikal Group: 6 – Kachergat Formation, 7 – Uluntui Formation, 8 – Goloust Formation; 9 – stratigraphic unconformity; 10 – basic volcanics and gabbroids; 11 – Early Proterozoic granites (Primorskii Complex); 12 – gneisses, amphibolites, and migmatites (Archean–Early Proterozoic); 13 – orthoamphibolite and metagabbro of the Listvyanka Complex (Proterozoic); 14 – gneiss, crystalline schist, amphibolite, migmatite, and marble of the Sharyzhalgai Group (Archean); 15 – major fault zones.

ements. The reliable proving of the origin of an oceanic basin calls for the chronological correlation of the processes of basic magmatism with the subsequent sedimentary and magmatic events.

2. Thick carbonate–clastic rock sequences in the marginal parts of the cratons, identified as the sediments of a passive margin, marking a mature stage in the development of Atlantic-type oceans.

3. The oldest ophiolites and associated island-arc formations, found in fold areas adjacent to the cratons. They date a period of active interaction between oceanic and continental plates with the formation of island arcs and back-arc basins.

Added to this list should be volcanogenic clastic rocks characteristic of typical intracontinental rifts, because the breakup process begins with the formation of large intracontinental rifts. Ideally, we can be sure of a continental massif breakup with the formation of a separating oceanic space where we have the following chronologically substantiated sequence: rift-related clastic and volcanic rocks often including volcanic rocks from the destructed continental margins ⇒ basic dike swarms ⇒ thick poorly deformed clastic and carbonate rocks of the newly formed oceanic basin. Different evolution stages of this basin are marked, in their turn, by the ophiolite remnants of the spreading zones, and also by the ophiolites of the back-arc basins, which are associated with the island-arc volcanic, clastic, and intrusive rocks. This sequence of processes in cratonic blocks and in the adjacent regions of foldbelts represents the following succession of geodynamic environments: the origin and evolution of an intracontinental rift system ⇒ an advanced rifting stage, marked by the intrusion of large basalt magma volumes into the upper crust, and the onset of the continental breakup ⇒ the initial stage of the formation of an oceanic space with the absence of an active interaction between the oceanic and the continental plate ⇒ the mature phase of oceanic evolution with an active interaction between the oceanic and the continental plate. We use the term continent breakup here deliberately and do not use the term "supercontinent", because the substantiation of the latter calls for the correlation of geologic events using many present-day cratonic blocks. In this paper we deal only with the southern part of the Siberian Craton.

The southern part of the Siberian Craton includes Neoproterozoic dikes and, locally, thick Riphean clastic–carbonate layers underlain by the rocks of rift origin. Riphean ophiolites and associated island-arc volcanic and clastic rocks have been found in the southwestern segment of the Central Asia foldbelt surrounding the Siberian Craton in the south. Below follows a brief description of the indicator lithologic units of the region, which mark the evolution of the Neoproterozoic Rodinia Supercontinent.

Rift-Related Clastic and Volcanic Rocks

The traces of rifting processes have been most fully reconstructed in the Patoma Upland where they are traced by the rocks of the Medvezha Formation which has been dated Middle Riphean from geological data. The age of the rocks was based only on geologic relations and was not supported by any correct geochronological data. The volcanic and clastic rocks of this formation are represented most fully on the Northwestern slopes of the Chuya and Tonod uplifts where they occur as a NE-trending intermittent band. This formation is distinguished by the accumulation of thick facies-intermittent sandstones and conglomerates interlayered with subaqueous basic volcanic rocks. This formation varies in thickness from a few dozens of meters to 2700 m [*Ivanov et al.,* 1995]. Based on their lithology, most of the sedimentary rocks can be classified as mixtites. Ferriferous quarzites had accumulated in the local stagnation zones of the basin. In terms of their petrochemical composition the volcanic rocks are close to basalts and basaltic andesites with the elevated contents of alkalies. The process of lava flow had been accompanied by the emplacement of basic dikes. The data available on the reconstructed structural elements of the Medvezha epoch suggest the presence of troughs separated by ledges with the Early Precambrian basement [*Ivanov et al.,* 1995]. During the early phases of their development the troughs were roughly N–S oriented (in the present-day coordinates) and had NE trends during the final phases. This rearrangement seems to have been caused by shearing as a reaction to the processes of oblique regional extension which dominated in the region during the period concerned. These rocks covered an extensive area taking into account their potential metamorphic analogs mapped along the SE flank of the Chuya uplift, in the Necher uplift, and in the north of the Baikal–Muya zone.

The next level of the accumulation of rift-related sedimentary and volcanic rocks was mapped in the West Baikal region (west of Lake Baikal). These are the rocks restricted to the base of the Baikal Group and the underlying Late Riphean rocks (Figure 1) [*Mazukabzov et al.,* 2001]. The rocks have a local distribution and are represented by coarse clastic deposits where the clastic material consists of the crushed granitoid and less common metamorphic rocks of the craton's basement. The sedimentary rocks are associated with volcanic rocks represented by tholeiites of normal alkalinity.

The fragmentary distribution of the rift-related rocks does not allow one to trace the zones of Neoproterozoic rifting in the Sayan and West Baikal regions, yet they suggest that the rifts had originated in the southeastern margin of the Angara–Anabar block of the Siberian Craton.

To sum up, the southern marginal part of the Siberian Craton contains rift-related volcanic and sedimentary rocks of two age intervals: Mesoproterozoic and Neoproterozoic. The rift-related rocks have not been dated by any accurate means, yet their geological relations with the dated rocks suggest that these two age intervals agree with the age intervals of 1300–1100 and 850–750 million years.

Dike Swarms in the Marginal Part of the Craton

Diabase and gabbro-diabase dikes, sills, and thin stocks are fairly widely distributed in the marginal part of the craton's basement. There are several fields of the products of basic magmatism, here referred to as dike swarms. The subvolcanics show a close association with sills and lava flows in the zones composed of poorly deformed and poorly metamorphosed rocks. In terms of the geography and relative to some geological structural elements, we mapped the following fields of Proterozoic dike swarms at the southern flank of the Siberian Craton (Figure 2): the North Baikal field and the Sayan field, the latter including the Sharyzhalgai protrusion of the basement.

North Baikal field. A great number of gabbro-diabase dikes and less common sills occur in the Akitka and Baikal ranges which coincide spatially with the North Baikal volcanoplutonic belt. They occur as single bodies and are often grouped into zones and belts. Usually, these are single dikes with distinct salbands, though some of the large bodies show somewhat blurred contacts with the granites. Some dikes have a complex dike-in-dike structure (upper reaches of the

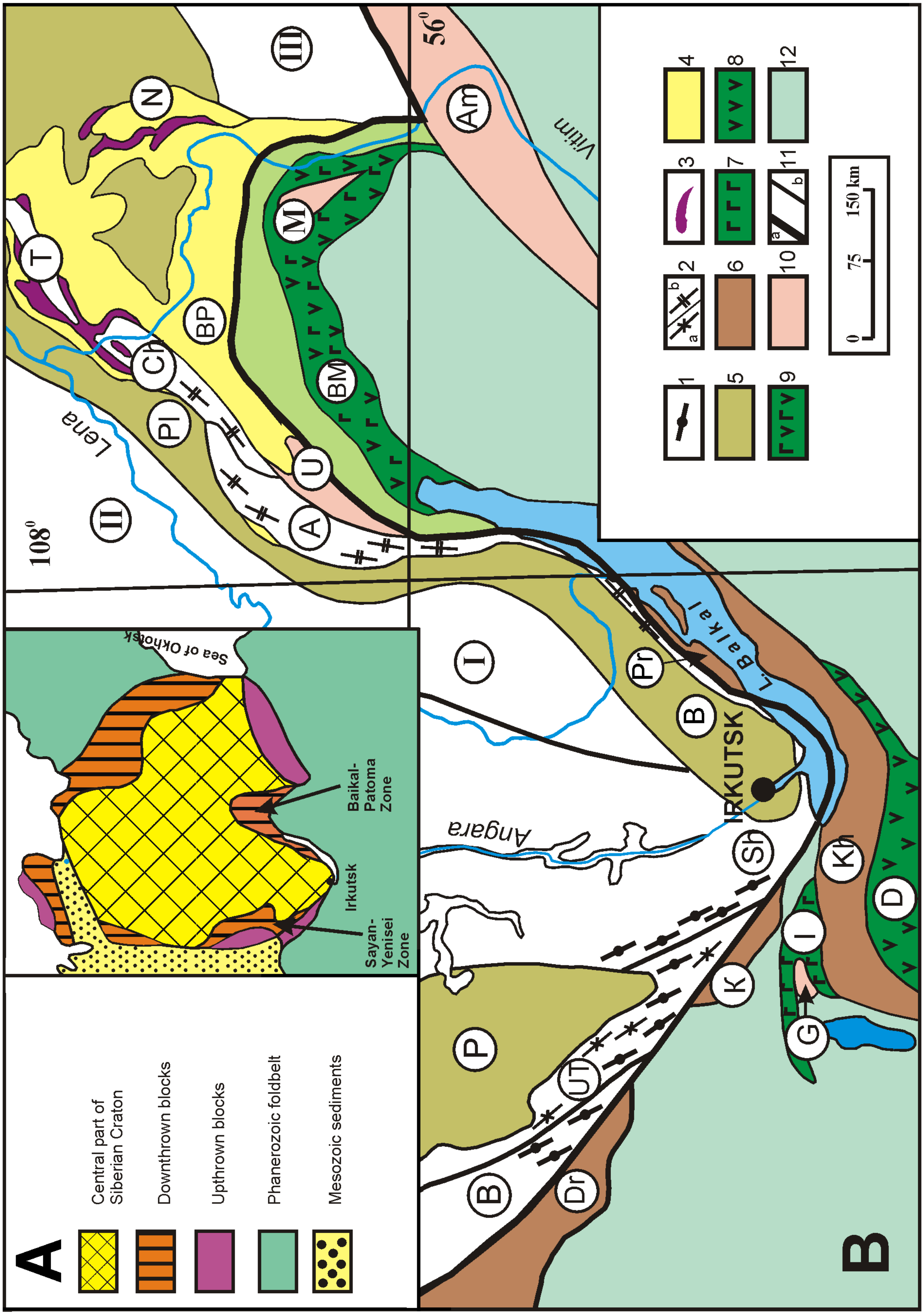
A
Central part of Siberian Craton
Downthrown blocks
Upthrown blocks
Phanerozoic foldbelt
Mesozoic sediments
Sea of Okhotsk
Baikal-Patoma Zone
Irkutsk
Sayan-Yenisei Zone
B
Lena
Angara
Vitim
L. Baikal
IRKUTSK
108°
56°
I
II
III
0 75 150 km

Figure 2. Schematic map showing the positions of the main structural elements of the southern Siberian Craton and the distribution of the geological complexes associated with various processes of the Neoproterozoic Supercontinent evolution: 1 – Late Riphean basite dike swarms (Nersa Complex); 2 – Middle Riphean basite dike swarms: a) Angaul Complex, b) Chaia complex; 3 – basic volcanic rocks (Middle Riphean); 4 – carbonatesclastic deposits of the passive margin (Middle Riphean, Patoma Zone); 5 – Late Riphean carbonate-clastic rocks of the passive margin and transitional to collision-related rocks; denoted by letters are the following collision zones: Lena zone (L), Baikal–Patoma zone (B-P), Baikal zone (B), Sayan Zone (S); 6 – metamorphic rocks supposed to be those of the passive margin and combined tectonically with the fragments of island-arc rocks (Kh – Khamardaban; K – Kitoi; D – Derba); 7–9 – ophiolite belts: 7 – Middle–Late Riphean (1.1–0.8 billion years) (I – Ilchir), 8 – Late Riphean–Vendian (0.7–0.55 billion years) (D – Dzhida), 9 – Middle-Late Riphean and Late Riphean Vendian (BM – Baikal-Muya); 10 – Early Precambrian terranes in a foldbelt (G – Gargan, M – Muya, Am – Amalat); 11a – present-day boundary of the Siberian Craton based on geological and geophysical data without its Phanerozoic sedimentary cover; 11b – various faults and boundaries between the geological complexes; 12 – Sayan–Baikal Foldbelt.
Major structural elements in the margin of the Siberian Craton: A – Akitka volcanoplutonic belt, B – Biryusa Block, N – Necher High, Pr – Primorskaya Zone, T – Tonod High, U – Ukuchikta Block, UT – Urik–Tumanshet Zone, Ch – Chuya High, Sh – Sharyzhalgai basement protrusion;
Major provinces: I – Tunguska, II – Anabar, III – Aldan.
Black rectangles with figures denote the sites of detailed survey and the numbers of the figures given in the text.

Savkina River, Akitka Range). The dikes are as long as 5–8 km and usually have a maximum thickness of 50 m, some being even thicker. They are surrounded by the granitoids of the Irelskii and Primorskii complexes, by the volcanogenic sediments of the Akitka Group, and also by Paleo- and Mesoproterozoic metamorphic rocks. The gabbro-diabase shows a varying range of secondary alteration, the latter being especially high in the dikes of the Baikal Range. The gabbro-diabase of the Akitka Range is almost unaltered and looks fresh in the area bordering the craton, but shows a varying degree of dynamic metamorphism, up to amphibole schist, eastward, closer to the large faults. In terms of their composition the dikes correspond to subalkalic basalt and tholeiite. The dike swarm has been traced southward as far as the Onguren Settlement area (Figure 3).

Based on the fact that the dikes cut the volcaniclastic deposits of the North Baikal Belt and are covered by the carbonate–clastic rocks of the Baikal Group, we place the time of their intrusion in the range of 1900–850 million years. The high secondary alteration of the diabase, often resulting in the complete loss of the primary mineral assemblages, preclude the more exact dating of the dikes.

Sayan field and Sharyzhalgai basement protrusion. The southern flank of the craton includes a few areas of Riphean dike swarms (Figures 4 and 5). Most of the dikes show a steep dip and a northwestern strike, coinciding with the trend of the major faults, and vary in thickness from 20–30 cm to 3–5 m, occasionally being as thick as a few dozens of meters. In some localities the dikes have small dip angles up to a subhorizontal position, resulting from their tendency to adapt themselves to the planar elements of the large folds. In areas of good exposure, some dikes can be traced as far as 1–10 km along the strike. The results of an airborne magnetic survey suggested some dikes to be as long as >15 km [*Sklyarov et al.,* 2000]. Based on the metamorphic grade of the dike rocks and taking into account some geochronological data available, the dike swarms of the region have been classified into three groups of different ages (from the oldest to the youngest), corresponding to the Arban, Angaul, and Nersa complexes, respectively. The rocks of the Arban Complex have the highest metamorphic grade and, according to the geological data, can be dated pre-Riphean. For these reasons they were discarded as the rocks not fitting the subject of this paper.

The rocks of the Angaul Complex occur in the Iya–Urik structure and are represented by dikes and sills of diabase and gabbro-diabase, and also by more rare picrite porphyry for which there are some K-Ar and Rb-Sr dates falling within the age intervals of 1550–1650 and 1100–1300 million years [*Domyshev,* 1976; *Sekerin et al.,* 1999]. The grades of the secondary diabase alteration are fairly high, often with the complete substitution of magmatic minerals. The basic rocks are transitional between tholeiites and subalkalic rocks. A group of locally found high-K basic rocks, classified as a separate series, includes lamproite [*Sekerin et al.,* 1995] and is believed to have been emplaced closely to the time of the intrusion of the above mentioned basic rocks. Based on a single bulk-rock Rb–Sr isochron for the rocks with intensive secondary alterations, these subvolcanic rocks were dated 1268±12 Ma [*Sekerin et al.,* 1995].

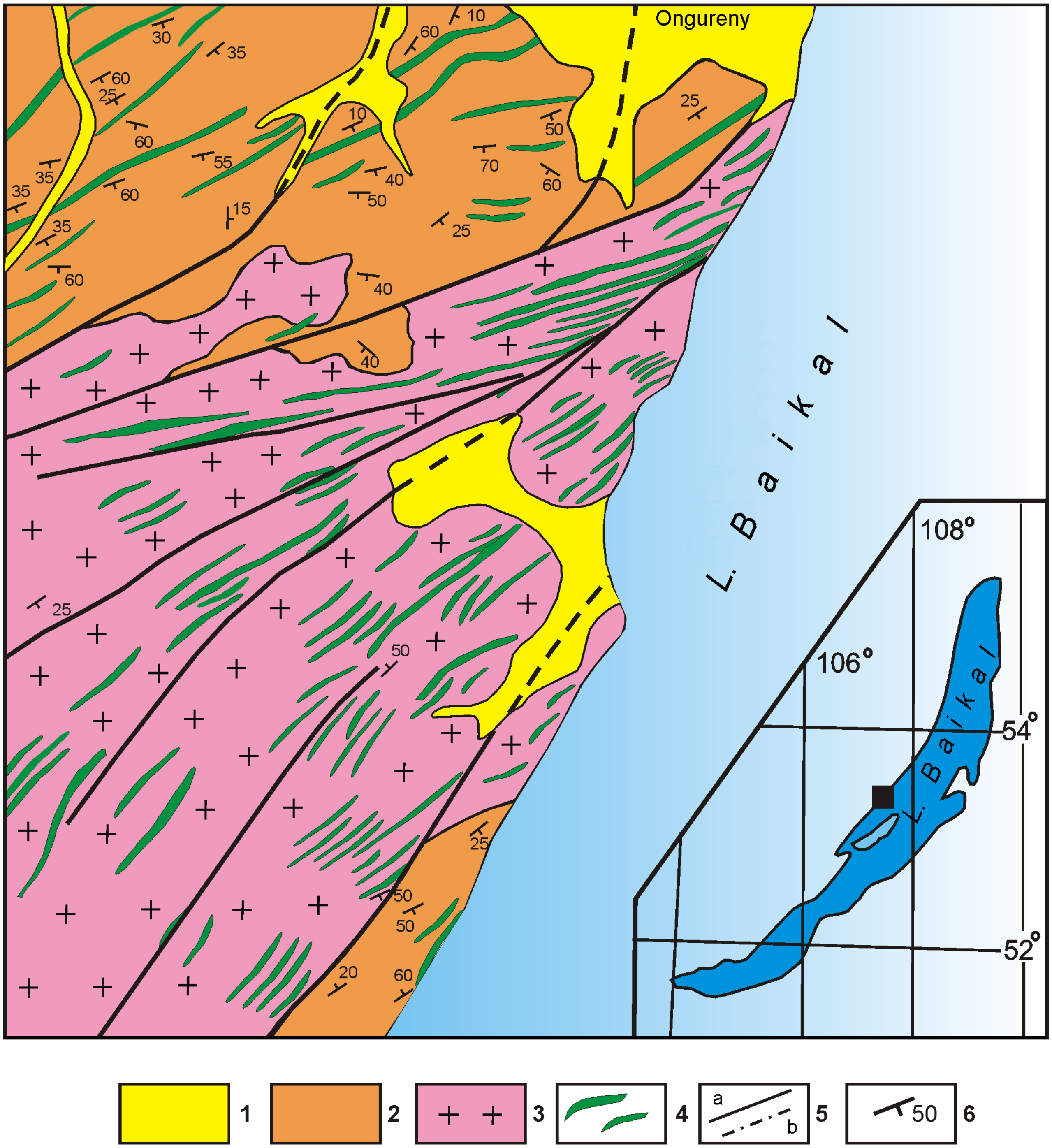

Figure 3. Schematic geological map of the Kocherik area in the Baikal Zone.
1 – Quaternary sediments; 2 – clastic deposits of the Sarma Group (PR_1); 3 – granitoids of the Primorskii Complex (PR_1); 4 – diabase and gabbro-diabase dikes; 5 – faults (a – proved, b – inferred); 6 – dip and strike. The inset map shows the geographical location of the area.

The dike swarms of the younger Nersa Complex, which are often associated with sill rocks, occur primarily in the field of the rocks of the Karagas Group in the Sayan Trough, and also in the Sharyzhalgai basement protrusion, as sills, dikes, and small stocks of an irregular form. The sills separate the members of lithologically different rocks and are as thick as 100 m with a length of >75 km. There are occasional multilevel sills [*Domyshev,* 1976]. The dikes usually have a NW trend (330–340°), although in some areas their dominant strike changes to the NE one. The mapped dikes vary in length from a few hundred of meters to a few kilometers. The thick dikes often show a differential structure [*Domyshev,* 1976]: the margins are composed of microdiabase, and the central parts, of gabbro-diabase.

Most of the dikes are steeply dipping, although some dikes have a gentle dip, being conformable with the gneissic structure of the host rocks.

Most of the dikes are composed of holocrystalline medium- and fine-grained rocks. These are typical diabase and gabbro-diabase with variable ratios between the major rock-forming minerals, the leading of them being clinopyroxene and plagioclase. Less common are olivine and pigeonite. The typical accessory mineral is titanomagnetite; the magnesian rocks contain chromite in the form of fine inclusions in olivine.

Most of the dikes have a tholeiite composition [*Gladkochub et al.,* 2001] and can be classified with the N-MORB type or with the type transitional to E-MORB. Less developed are subalkalic varieties. The geochemical characteristics of the studied gabbro-diabase samples suggest that the initial magmas parental for the subvolcanic rocks of the tholeiite and subalkalic rock series had been generated in an enriched lithospheric source.

The dikes were dated 850–890 Ma by the Ar-Ar method (plagioclase) [*Gladkochub et al.,* 2000]. Later, a bulk-rock Sm–Nd isochron was obtained for these dikes, dating them 750 Ma old. This substantial difference can be explained as follows. The configuration of Ar spectra, having a poorly expressed U-shaped form, can be indicative of an excessive radiogenic argon. In this case the results obtained can be taken only as the lower age limit of the dikes (not older than 850 Ma). Hence, the more realistic value seems to be 750 Ma.

To conclude, by convention there are two age generations of the post-Paleoproterozoic dikes, which were intruded 1100–1300 Ma ago (Angaul Complex) and 750–850 Ma ago (Nersa Complex). Convention mainly concerns the dating of the Angul dikes.

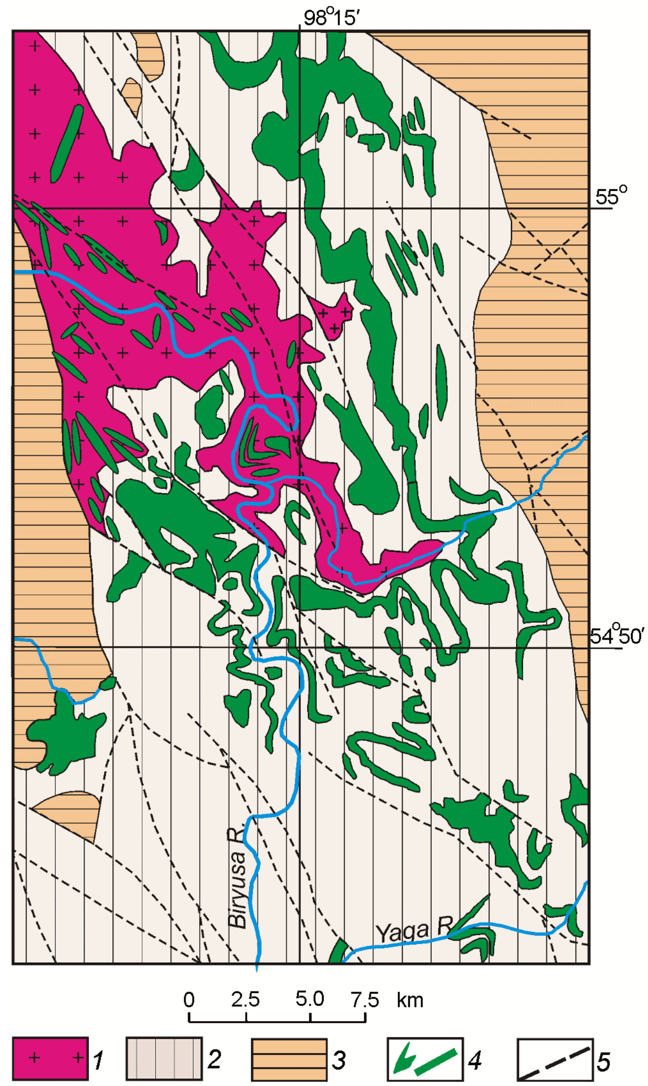

Figure 4. Schematic geological map of the Biryusa R. middle-coarse area (compiled using the field maps of G. A. Belozerov and T. V. Gorbovskaya).
1 – Early Proterozoic granitoids of the Sayan Complex; 2–3 – Late Riphean rocks of the Karagas Group: 2 – Shangulezh and Tagul formations (bottom of the section), 3 – Uda Formation; 4 – Late Riphean Nersa dolerite complex (sills and dikes); 5 – faults.

Poorly Deformed Clastic–Carbonate Rock Sequences

The preserved fragments of the sedimentary clastic–carbonate rock sequences reflect, to some extent, the potential orientation of the sedimentation regions associated with the formation of the passive continental margins in the south of the Siberian Craton. The relations of these deposits with the underlying rocks vary from marginal to cryptic conformable unconformities which were obviously controlled by extension. There are three structural-facies zones that can be outlined in terms of their sedimentation patterns and spatial positions: the Patoma, Baikal, and Sayan zones. The tectonic relicts of these sedimentary sequences can be traced in the Baikal–Muya Zone.

In the Patoma Zone, the Middle–Late Riphean rocks of the Ballaganakh Group (up to 7 km thick) rest conformably on the rocks of the Medvezha Formation in the grabens and with an unconformity elsewhere, where they rest on the Early Proterozoic rocks. This suggests that sediments had accumulated at that time over a significantly larger region, at least, within the limits of the Baikal–Patoma Highland. It has been proved that the lower layers of the sequence had accumulated irregularly, their accumulation being controlled by large fault zones with the formation of clinoforms. A characteristic feature of these deposits is a combination of immature and poorly differentiated clastic sediments (polymictic to oligomictic sandstones and occasional conglomerates and carbonaceous shale) with interlayers of highly aluminiferous shale [*Nemerov and Stanevich,* 2001]. The origin of this deep and extensive sedimentary basin seems to have been associ-

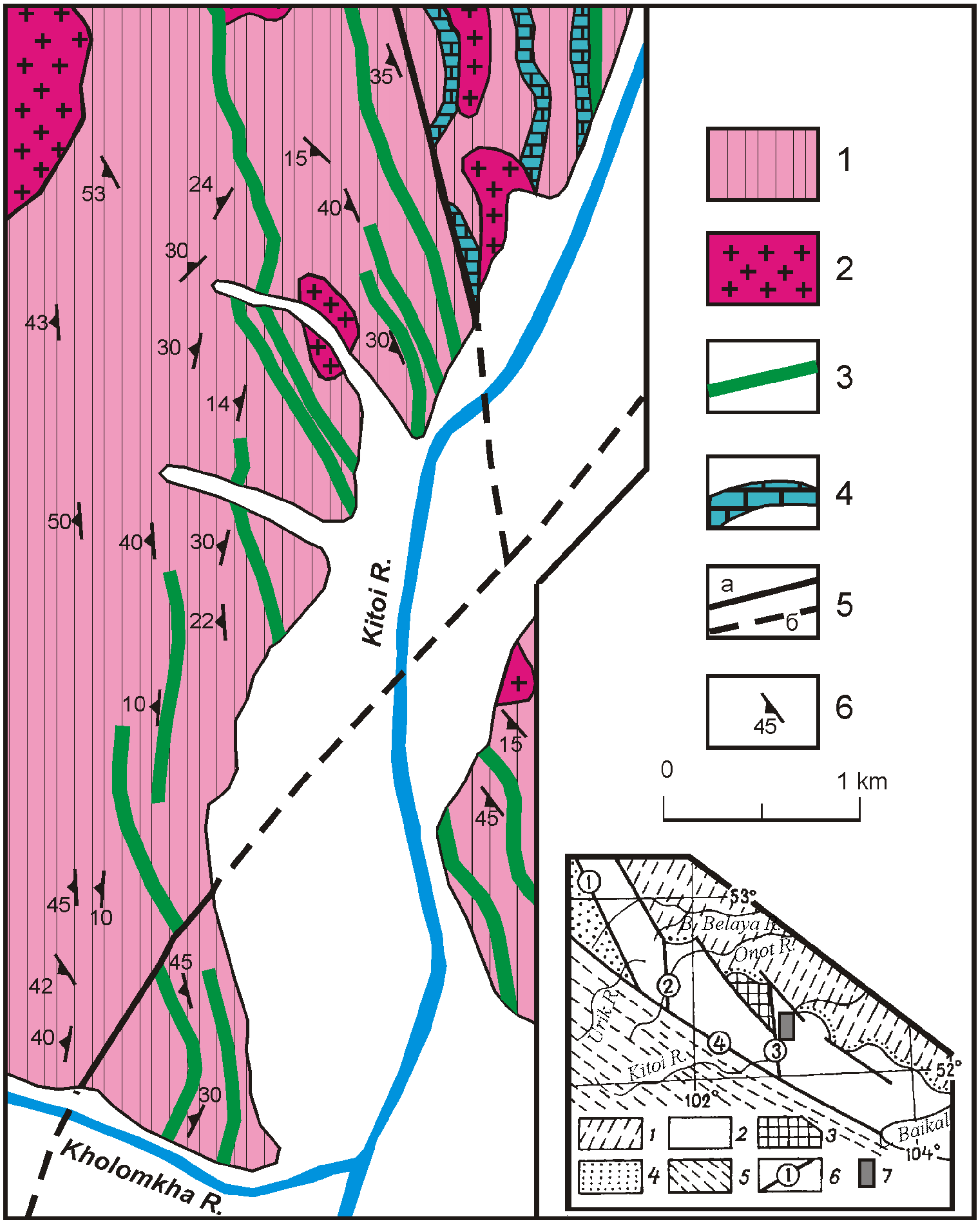

Figure 5. Neoproterozoic gabbro-dolerite dikes in the area of the Kitoi R. middle course.
1 – Early Precambrian metamorphic rocks; 2 – Proterozoic granites; 3 – Neoproterozoic gabbro-dolerite dikes; 4 – marble layers; 5 – faults: (A) observed, (B) inferred; 6 – dip and strike.
The inset is a schematic map of the southern flank of the Siberian Craton and adjacent fold areas.
1 – sedimentary rocks, 2 – Sharyzhalgai basement protrusion, 3 – Onot graben, 4 – Urik–Iya graben, 5 – Sayan–Baikal Foldbelt, 6 – faults: Tagninskii (1), Onot (2), Kitoi (3), Glavnyi Sayan (4); 7) survey area.

ated with the extension and thermal relaxation of the lithosphere after the high thermal effect of the preceding phase. These rocks are overlain by the rocks of the Nygra and Dalnyaya Taiga groups, as thick as 3600 m. Their characteristic feature is the highly variable composition of the clastic and carbonate deposits with the elements of a flyschoid structure. The character of their facies changes suggests that they accumulated in a marginal basin under shallow- to deep-sea conditions. The clastic material was transported from the adjacent areas of the craton and, possibly, from the incipient Baikal–Muya island arc. The rhythmical patterns of the rock sequences suggest the continuation of the slow tectonic subsidence of the region with the origin of undercompensation environments in the central parts of the basin. These phenomena in the evolution of the pericratonic basin were obviously associated with its rearrangement [*Stanevich and Perelyaev,* 1997], the latter being obviously associated with changes in the dynamic conditions of the Baikal–Muya island arc [*Nemerov and Stanevich,* 2001].

The subsequent evolution history of the region included the transformation of the peripheral basin to a foreland basin with collision events that occurred during the final phase of the Baikalian cycle.

The problem of the occurrence of the clastic–carbonate deposits of this age further south and southwest is of great interest and is still a matter of debate. It is known that thick (5–12 km) clastic–carbonate deposits occur along the southern and southwestern boundaries of the craton, within a foldbelt (Figure 2), which seem to have accumulated in the environments of a continental slope and its foot. These metamorphic sediments, known as the Olkhon, Slyudyanka, and Kitoi groups, are zonally metamorphic (to a granulite facies) and have been classified by many geologists, on the basis of their high-grade metamorphism, as Early Proterozoic or even as Late Archean rocks. The results of more recent studies [*Bibikova et al.,* 1990; *Donskaya et al.,* 2000; *Salnikova et al.,* 1998] proved the Early Proterozoic age of the high-grade metamorphism (460–480 Ma). The problem of the age of the parent rocks remained unsolved. However, the rhythmic structure and lateral persistence of the carbonate and clastic–carbonate rock sequences, similar to that of the rocks of the Ballaganakh Group, as well as the analysis of the geologic situation, seem to justify their most probable classification as the sediments of a Riphean passive margin. It should be emphasized that there are no indications of their older age, other than their high metamorphic grade. Since the metamorphism was associated with the processes of the Early Paleozoic collision, we can assume that the fragments of the marginal part of the craton, covered by thick sediments, were involved into the zone of the greatest collision effect.

As to the West Baikal zone, fragments of sedimentary rocks have been preserved only in the proximal zone of the basin, documenting the evolution of the passive margin (Figure 6). The time of their accumulation is dated by the sediments of the Goloustenskaya Formation of the Baikal Group and by the underlying rocks of the Late Riphean Nuganskaya and Khotskaya formations. The passive margin seems to have existed there up to the accumulation of the basal layers of the Kachergatskaya Formation. Most of the clastic material was added during the Prekachergatskoe time from the side of the Siberian Craton and was sufficient for the accumulation of sediments as thick as 2.5 km. The deposits of the Kachergatskaya Formation document the conditions of a foreland basin, which is emphasized by the onset of the addition of a clastic material from the perioceanic region.

The Sayan branch of the Late Riphean passive continental margin was located in the territory of the craton, which was subject to extension not sufficient for rifting. Under these conditions a sedimentary basin was formed (Sayan Trough) where the sediments of the Karagas Group, as thick as 1100–3700 m, accumulated. The basal beds of this group accumulated in the continental and lagoonal environments and are represented by red conglomerates, cross-bedded sandstones and siltstones, and less common arenaceous dolomites. Above follow carbonate-clastic and carbonate deposits. The sedimentary sequence is terminated by the clastic flyshoid and siliceous-clastic-carbonate rocks with a high phosphate content. The sedimentation process was accompanied by the intrusion of the diabase bodies of the Nersa Complex. This distinguishes the Sayan branch of the passive continental margin from the Patoma and Baikal margins. On this basis the Sayan margin can be classified as a volcanic-type passive margin.

The rocks of the Karagas Group are overlain by those of the Oselkovaya Group of Vendian age. The rocks of the latter rest transgressively on the different layers of the Karagas Group and can be classified as the rocks of a foreland trough, the formation of which terminated the history of the Baikal cycle [*Sovetov,* 2001].

To summarize, like in the case of the rifting-related deposits, there are two intervals of clastic-carbonate deposits which can be interpreted as the sediments of a passive margin or as the deposits that accumulated during a period of transition from the passive to the active margin. In the Patoma zone, these deposits are associated, in terms of space and tectonic events, with the rifting-related volcanic and clastic rocks of the Medvezha Formation. Coarse clastic deposits and volcanics were also recorded at the bases of the Baikal and Karagas groups in the Sayan and Baikal zones, respectively. Like in the case of the rift-related volcanic and clastic rocks, there are no correct isotopic datings for these rocks.

Ophiolites and Island-Arc Rocks

Ophiolites and the island-arc rocks associated with them are widely developed in the Central-Asia Foldbelt, adjacent to the Siberian Craton in the south; they have extensive literature (e.g., *Dobretsov,* 1985; *Khain et al.,* 1997, 2001; *Sklyarov et al.,* 1992]. However, no typical ophiolite sequences, characteristic of the early evolution of the oceanic basins, namely, of their opening, have yet been found. The timing of the individual belts and massifs is still a matter of debate, which stems from the difficulty of their correct dating. Nevertheless, the critical analysis of the data available, along with the latest correct datings, suggest two age levels for the ophiolite complexes and island-arc rocks [*Khain et al.,* 2001; *Rytsk et al.,* 2001].

The oldest ophiolites can be dated roughly 1000 mln years and reside most closely to the Siberian Craton (Figure 2). These are the ophiolite fragments from the Baikal–Muya Belt in the north of the Transbaikalian region [*Rytsk et al.,* 2001], the ophiolites of the Ilchira Belt in the southeastern Sayan region (Figures 7 and 8) [*Khain et al.,* 2001], and the fragments of basic and ultrabasic rock bodies of the Arzybei Block in the eastern part of the Eastern Sayan Mountains [*Nozhkin and Turkina,* 2001], which have been proved to be of ophiolite or island-arc origin. They were dated in the age range of 1035–1042 million years. The most reliable dating

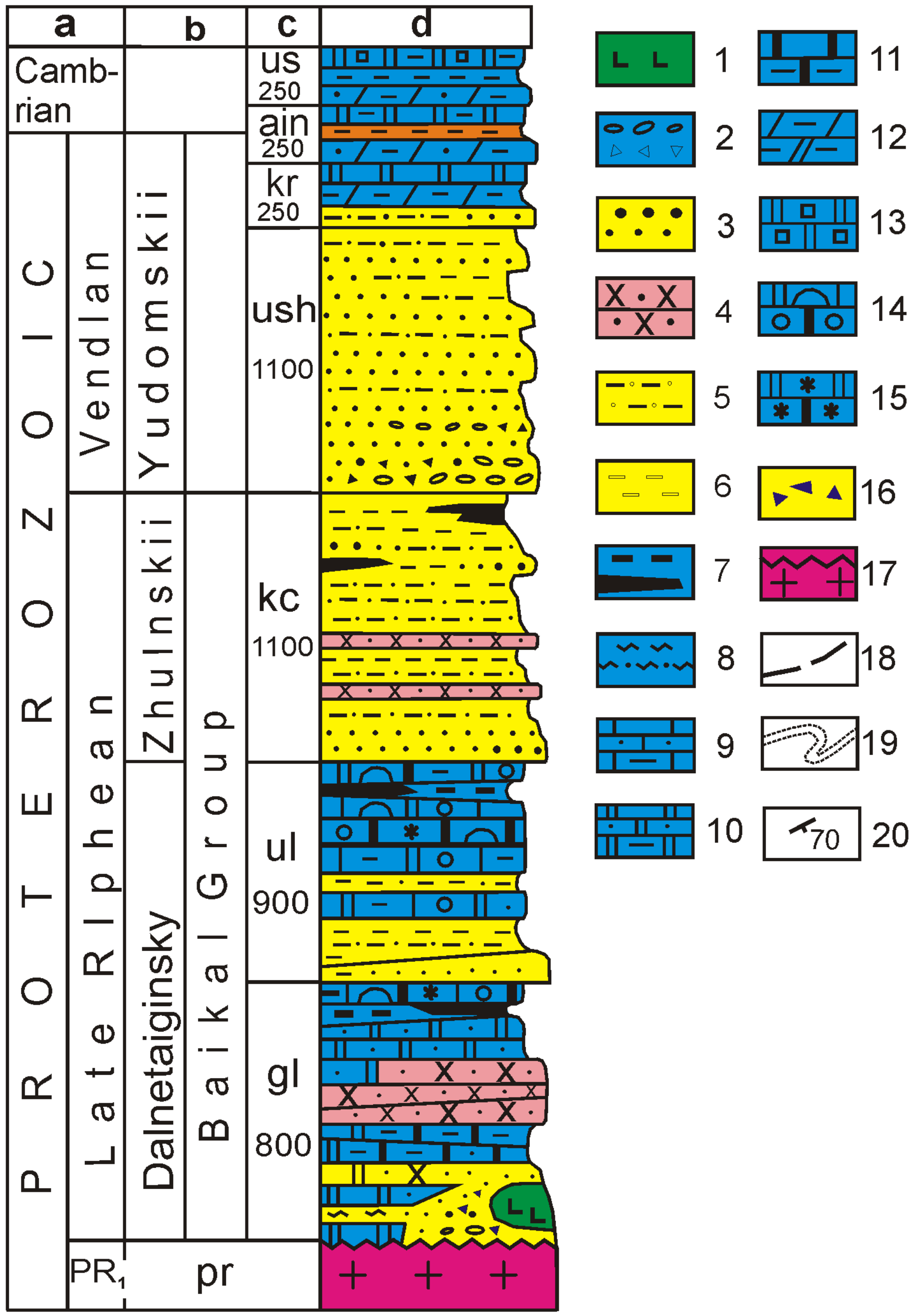

Figure 6. Stratigraphic column of the Neoproterozoic–Early Cambrian rocks.
1 – basic volcanic rocks; 2 – conglomerate and conglobreccia; 3 – polymictic arkose-graywacke gritstone and sandstone; 4 – quartz sandstone; 5 – siltstone; 6 – argillite and silty argillite; 7 – carbon-bearing argillite and silty argillite, 8 – pelitic and aleuropelitic shale; 9 – limestone, including its sandy varieties, and interbedded limestones and silty argillites; 10 – dolomite and sandy dolomite; dolomite and silty argillite interbedding; 11 – dolomitic limestone, calcareous and silty argillite dolomite; 12 – calcareous dolomitic marl; 13 – salt-bearing carbonate; 14 – stromatoliths and microphytoliths; 15 – silicification; 16 – synsedimentation breccia; 17 – unconformity; 18 – fault zone; 19 – structural style; 20 – measured bedding (in degrees).
Stratigraphic column: a) general scale, b) regional units and groups, c) formations and their average thicknesses (in meters), d) lithology; pr is the index for the granites of the Primorsky Complex.

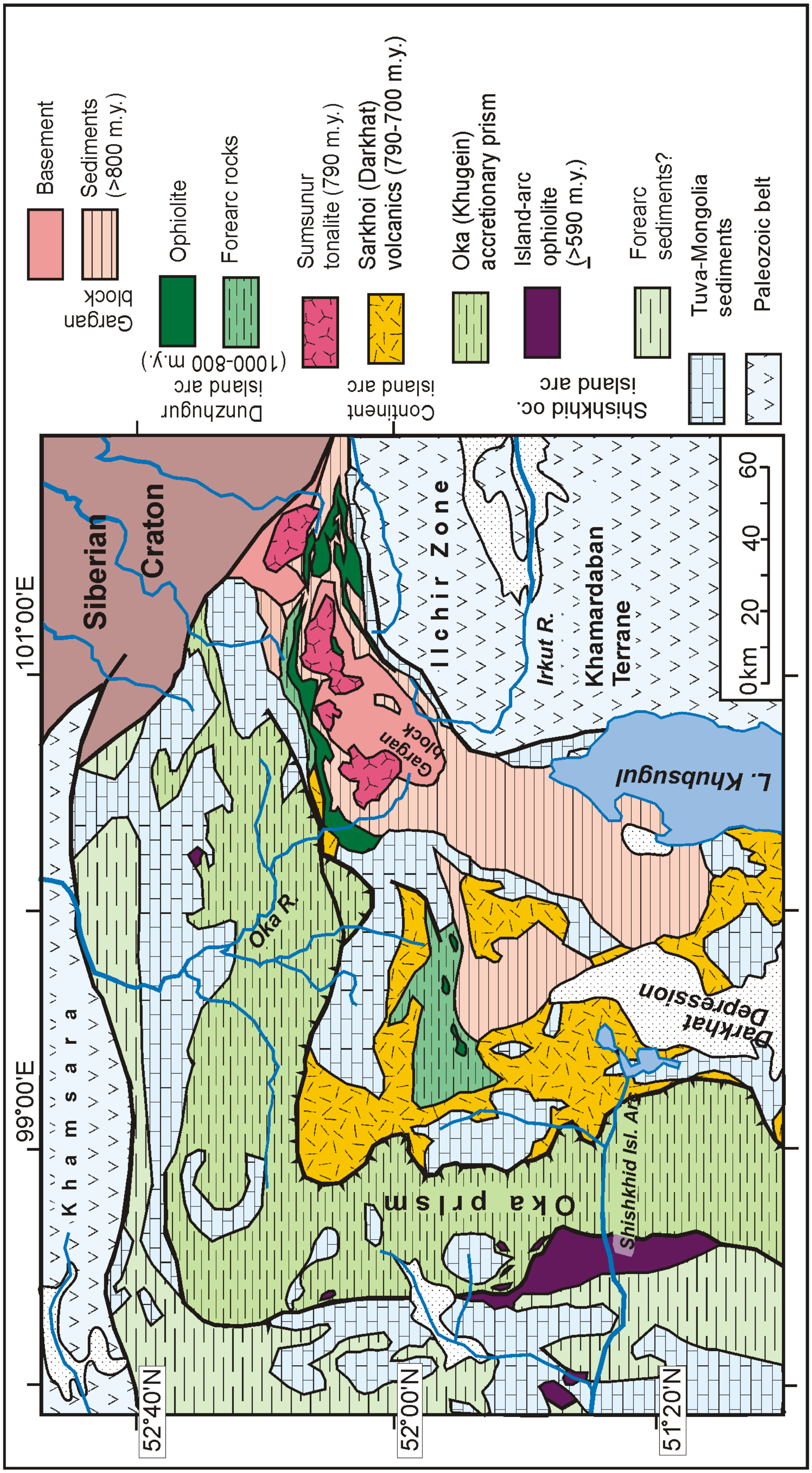

Figure 7. Tectonic map of the Tuva-Mongolia Superterrane (prepared by A. B. Kuzmichev).
Note. Omitted in the map are the Paleozoic granites, as well as the Cenozoic sediments and basalts (except for large depressions).

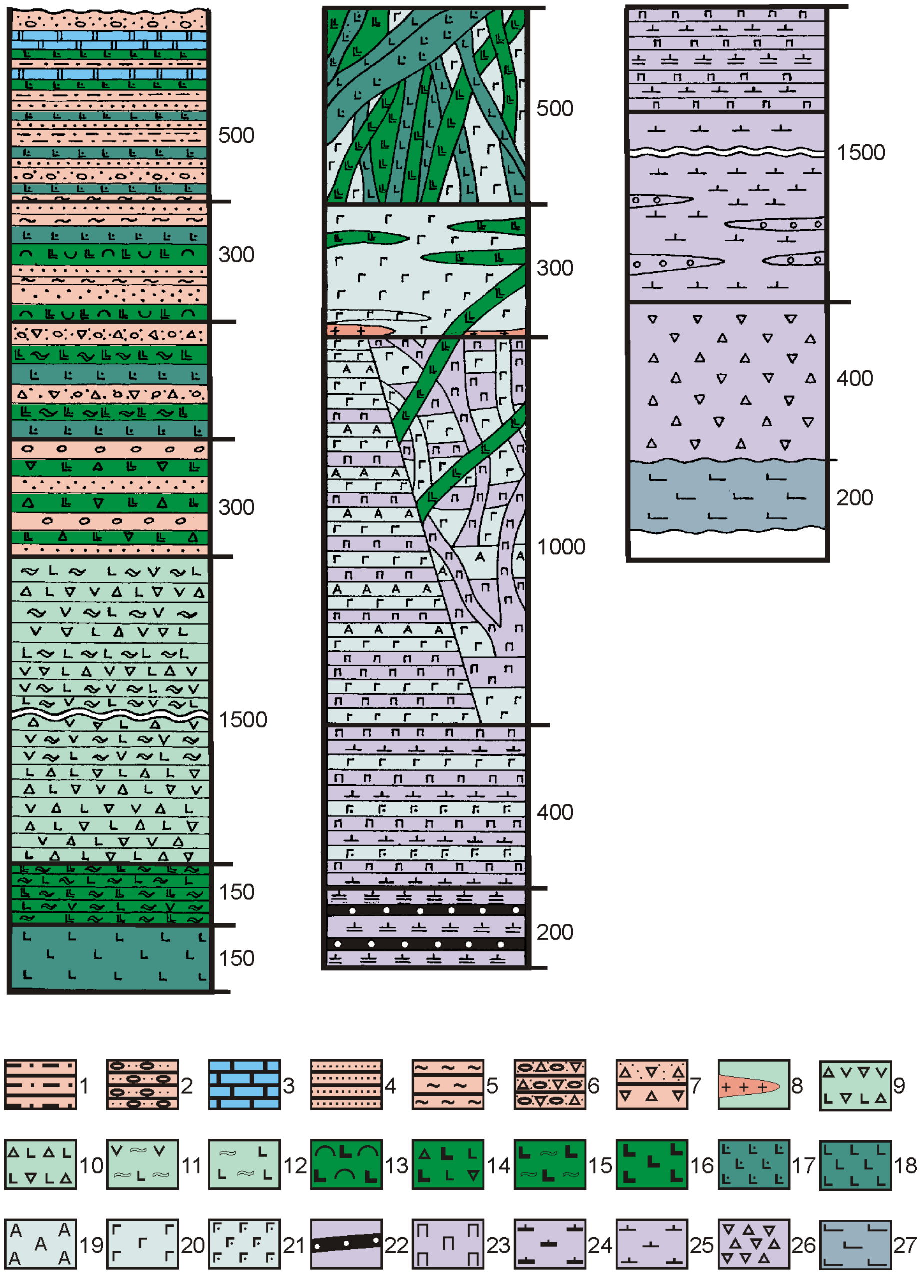

Figure 8. Integrated section of the ophiolite rock association from the Southwest Sayan.
1 – siliceous sediments; 2 – siltstone and conglomerates; 3 – dolomite; 4 – sandstone; 5 – shale; 6 – siltstone with tuff; 7 – sandstone with tuff; 8 – plagiogranite dike; 9 – brecciated basaltic andesite lava; 10 – brecciated basalt lava; 11 – basaltic andesite lava; 12 – basaltic pillow lava; 13 – boninite tuff; 14 – brecciated boninite lava; 15 – boninite lava; 16 – bononite dike; 17 – gabbro-diabase; 18 – massive diabase; 19 – anorthosite; 20 – gabbro; 21 – olivine gabbro; 22 – chromite lenses; 23 – orthopyroxenite; 24 – dunite; 25 – peridotite; 26 – melange zone; 27 – amphibolite.

was derived for the plagiogranites (U–Pb) from the Ilchir ophiolite belt [*Khain et al.,* 2001] with its complete and relatively poorly disturbed rock sequence.

The island-arc formations, mainly intrusive rocks, for which isotopic data are available, yielded a broader range of ages, though most of the valid datings fall within the range of 800–950 Ma [*Nozhkin and Turkina,* 2001; *Rytsk et al.,* 2001, to name but a few]. Naturally, different (younger) ages have been reported. They will be discussed below.

In any case, it is fairly obvious that the oldest ophiolites are of Mesoproterozoic age. It should be noted that in this paper we do not discuss the Paleoproterozoic ophiolites of the Sharyzhalgai basement protrusion, where they are involved into the structure of the metamorphic basement [*Gladkochub et al.,* 2001; *Sklyarov et al.,* 1998]. According to their geochemical characteristics, the ophiolites discussed above suggest their suprasubduction origin in the geodynamic environment of the active continental margin.

The other age-based group of the ophiolites (Figure 2) is more widely distributed in the foldbelt. The ages of the ophiolites and of the island-arc volcanic and intrusive rocks associated with them fall mainly within a range of 700–500 million years [*Khain et al.,* 2001]. As far as the territory discussed is concerned (Figure 2), this group includes the fragments of the ophiolites and associated island arcs of the Baikal–Muya Belt, the Eravninskii island-arc belt, and the ophiolites and island-arc formations of the Dzhida Belt. The rocks of this age are also widely distributed farther westward in the Altai–Sayan region and southward in Mongolia.

To summarize, there are two groups of rocks which originated in the geodynamic environment of active continental margins 1100–800 and 700–500 million years ago. In spite of some doubts concerning the validity of some intermediate values, the ophiolite and island-arc rock complexes of this segment of the Central Asia Foldbelt can be definitely classified into two above-mentioned age groups.

One of the most disputable topics is the position of the ophiolites and associated older island-arc complexes relative to the Siberian Craton, with which these complexes are now in a direct contact. An important point to be emphasized is the fact that the complexes concerned are the constituents of the composite terranes (superterranes), the positions of which relative to the Siberian Craton are problematic for the time when the subterranes were amalgamated into the Central-Asia Foldbelt. At least spatially, the ophiolite and island-arc complexes tend to be associated with two Precambrian blocks: the Gargan Block (Tuva–Mongolian microcontinent) and the Muya Block (Barguzin Microcontinent). Some researchers [*Didenko et al.,* 1994; *Mossakovskii et al.,* 1993] believe that the island arcs and back-arc basins, where the ophiolites and island-arc rocks accumulated, had developed in the course of the active interaction between the oceanic plate and the continental crust of the ancient Gargan and Muya crustal blocks which were located far from the Siberian Craton. As follows from the palinspastic reconstructions proposed by the above authors, the ophiolites and island-arc clastic and volcanic rocks had accumulated in the marginal part of East Gondwana, where the Tuva–Mongolian microcontinent and, possibly, the Muya block had been located. Although the latter seems to have been spatially closer to the Siberian Craton which is believed to have been a constituent of the Rodinia Supercontinent some 1000 million years ago [*Gladkochub et al.,* 2000, 2001; *Sklyarov et al.,* 2000; *Yarmolyuk and Kovalenko,* 2001]. The Muya block is supposed to have been accreted to the southern flank of the Siberian Craton some 570 million years ago [*Rytsk et al.,* 1999]. The accretion of the Tuva-Mongolian microcontinent to the southern flank of the Siberian Craton is believed to have taken place during the Ordovician collision which is marked by the Baikal collision belt [*Donskaya et al.,* 2000]. Hence, it can be assumed that the Mesozoic–Neoproterozoic ophiolites and the associated island-arc rock complexes of the Central Asia Foldbelt were not connected spatially with the margin of the Siberian Craton but had an absolutely different geological history of their own.

Discussion

The disputable and often uncertain ages of some geologic complexes developed at the southern flank of the Siberian Craton and in the adjacent areas of the foldbelt, as well as the ambiguity of their geodynamic interpretation, caused by their late intensive tectonic reworking, explain a great number of different scenarios offered for the geological evolution of the region during the Late Precambrian. Even the authors of one and the same paper disagree with one another concerning many aspects. The discussion of the even most realistic versions would have surpassed the limiting volume of this paper. For this reason we restrict ourselves to the discussion of two potential scenarios based on the existence of two rock sequences which seem to reflect the processes of intracontinental breakup and the subsequent opening and evolution of an oceanic basin.

One group of rocks originated during a time interval of 1300–850 million years. It comprises:

(a) the rift-related volcanic and clastic rocks of the Medvezha Formation and the dike swarms of the Chaiya (in part) and Angaul complexes (1300–1100 Ma), which accumulated under the conditions of the early and advanced phases of intracontinental rifting;

(b) the facially persistent clastic and carbonate deposits of the Ballaganakh Group from the Patoma Zone and the rocks of the Baikal–Muya Zone (1100–900 Ma) which correlate with the former and are interpreted as the fragments of the passive continental margin. Proceeding from the fact that the metamorphic sediments of the Olkhon, Slyudyanka, and Kitoi groups belong to this level, we assume that these rocks, extending throughout the entire southern flank of the craton, can be used as the main stratigraphic indicator of the ocean opening;

(c) the ophiolites and island-arc volcaniclastic and intrusive rocks of the East Sayan (Ilchir) and Baikal–Muya belts developed in the fold area bordering the craton (1100–850 Ma). In contrast to the former two rock complexes developed mainly within the area of the craton, these formations are the constituents of the terranes of the adjacent foldbelt and might have accumulated at a significant distance from the craton.

The rocks of the other age-based group (850–500 Ma) suggest that the craton and the perioceanic blocks moved in different directions. This is expressed in the rather chaotic relations among the fragments of the rift-related, island-arc, and collision-related rock assemblages. These can be classified into the following groups:

(a) the rocks of the obvious rift origin: the dike swarms and the dikes and sills of the Nersa Complex, widespread in the Sayan Zone and in the Sharyzhalgai basement protrusion (800–750 Ma). Also included in this group are the volcanic rocks of the Khota Formation: the typical basalts of destructive continental margins, which underlie the shelf sediments occurring as the basal layers in the Baikal Group;

(b) the clastic–carbonate sediments of the Karagas Group from the Sayan Zone, the basal layers of the Baikal Group (Baikal Zone), and their analogs in the Patoma Zone (850–650 Ma);

(c) the ophiolites and associated island-arc rocks from the Central-Asia Foldbelt (the Dzhida branch and, partially, the Baikal–Muya Zone) dated 700–500 Ma.

It should be emphasized that (1) some of the above-mentioned rock complexes have not been dated exactly, and their ages are still a matter of debate, and (2) the rift-related volcanic and clastic rocks, as well as some dike swarms, might have been formed during intracontinental rifting without any subsequent breakup of the continent. There were many examples of this kind in the geologic history of the Earth. It is only in the case of the well-known time sequence of some associated geologic events that we can assume the transformation of intracontinental extension to a continental breakup with the formation of an oceanic space. Assuming that in spite of these disputable assumptions the proposed sequence of events did exist, two major episodes of large-scale intracontinental extension, which might have ended in the ocean opening, can be inferred. Moreover, the magnitude and role of each extension episode can be treated as diametrically opposed.

Scenario (1) suggests rifting and the opening and evolution of an oceanic basin in the time interval of 1300–900 million years. Thick clastic–carbonate strata, whose fragments are traceable along the entire southern flank of the craton, marked the "Atlantic" phase of the ocean opening. These strata are intimately associated with the preceding rifting-related formations. The ophiolites and the associated island-arc deposits from the adjacent foldbelt regions dated 1100–800 Ma correspond to the "mature" phase of the paleoocean evolution. The collision events dating the closure of the paleobasin or the large-scale processes during the collisions of the terranes with the paleocontinent are imprinted in the granite-gneiss of the Muya Block, dated 825–790 million years [*Rytsk et al.,* 2001], in the tonalite of the Tuva–Mongolia Microcontinent, dated 812–785 million years [*Kuzmichev et al.,* 2001], and in the collision granites from the northwest of the Eastern Sayan, dated 870–860 million years [*Nozhkin and Turkina,* 2001].

The processes that operated during the second stage (800–500 Ma) in the middle of the Neoproterozoic had different trends in the SW (East Sayan) and SE (Baikal–Patoma) segments of the southern flank of the craton. The Baikal–Patoma segment is marked by the development of an extensive back-arc basin bounded in the south by volcanic arcs tapering out in the SW direction. The Eastern Sayan segment shows a wide development of dike swarms and associated rifting-related deposits (Karagas Group), indicative of large-scale rifting. This period of time corresponded to the breakup of Rodinia, as follows from the reconstructions available for some other cratons [*Hoffman,* 1991]. The Vendian–Early Paleozoic ophiolites and island-arc formations, widely developed in the Central Asia Foldbelt, mark the "mature" evolution phase of the Paleoasian ocean, the final closure of which took place during the Ordovician.

In terms of the second scenario, the first rifting event was not accompanied by the opening of any significant newly formed basin. In this case the ophiolites and associated island-arc rocks, as well as the collision products of the respective age, which occur only in the Central Asia Foldbelt, can be interpreted as exotic terranes that were accreted to the Siberian Craton later. It is believed that the craton itself was a constituent of the Rodinia Supercontinent during that time [*Gladkochub et al.,* 2000, 2001; *Sklyarov et al.,* 2000; *Yarmolyuk and Kovalenko,* 2001]. Whereas the ophiolites and island-arc rocks seem to have accumulated away from the margin of the Siberian Craton, in the East Gondwana group of terranes, and were associated with the processes of the origin and evolution of the active continental margins on Precambrian continental massifs, such as the Tuva–Mongolia microcontinent.

The opening and evolution of an oceanic basin (Paleoasian Ocean), documented in the structures of the southern flank of the Siberian Craton, took place in the time interval of 850–650 million years. The intrusion of the dikes of the Nersa swarms into the rocks of the Sharyzhalgai basement protrusion took place during the early phases of extension against the background of a growing dome-shaped uplift of a complex morphology at the expense of the rising asthenospheric mantle and the thinning of the lithosphere. Because of the cooling of the asthenolith or the relaxation of the tensile stresses, the growth of the dome-shaped uplift did not result in the formation of an axial rift valley in its central part.

The breakup of the Rodinia Supercontinent and the opening of the Paleoasian ocean took place as the result of extension along the above mentioned two active paleorift branches. In the East Sayan branch the Neoproterozoic dike swarms and sills were intruded into the sediments of the Karagas Group and into the metamorphic rocks of the basement protrusions.

In the Baikal–Patoma branch of the paleorift, volcanic activity is documented by the lavas of the Khota Formation in the Goloustenskii basement protrusion, which occur at the base of the sedimentary sequences of the Baikal Group. The metavolcanics are the remnants of the rift-related rocks which agree with the environments that had existed prior to the origin of basins in the margin of the Paleoasiatic Ocean [*Sklyarov et al.,* 2001]. In spite of the differences between the two scenarios discussed above, the latter are similar in terms of the Early Vendian molasse sequences bordering the craton. Their properties reflect the conditions of the accretion–collision events which took place at the time of the inferred opening and evolution of the Paleoasian Ocean.

The two scenarios offered in this paper for the geological history of the southern marginal part of the Siberian Craton can be interpreted absolutely differently in terms of the breakup of the Rodinia Supercontinent. According to the first scenario, the formation of the oceanic basin recorded by the rocks of the region occurred during the "assembling" of a supercontinent, the events of the supercontinent breakup being recorded only in the southwestern segment of the southern flank of the craton. This interpretation is in good agreement with numerous palinspastic maps (e.g., [*Rogers,* 1996]), where the segment concerned is believed to have been the marginal part of a supercontinent from the very beginning. The second scenario does not reject the rifting processes with the formation of the oceanic crust, but assumes a kind of their "local" development and smaller size, compared to the events of the later half of the Neoproterozoic, responsible for the global breakup of Rodinia.

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