ELSEVIER

Precambrian Research 000 (2001) 000–000

Precambrian Research

www.elsevier.com/locate/precamres

Neoproterozoic (~ 800 Ma) orogeny in the tuva-mongolia massif (Siberia): island arc–continent collision at the northeast rodinia margin

A.B. Kuzmichev [a,*], E.V. Bibikova [b], D.Z. Zhuravlev [c]

[a] *Institute of the Lithosphere, Russian Academy of Science, Staromonetny Lane, 22, Moscow 109180, Russia*
[b] *Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Science, Kosygina Street 19, Moscow 117975, Russia*
[c] *Institute of Geology of Ore Deposits, Petrography, Mineralogy and Geochemistry (IGEM), Russian Academy of Science, Staromonetny Lane, 35, Moscow 109017, Russia*

Abstract

The Tuva-Mongolia Massif is a composite Precambrian terrane incorporated into the Palaeozoic Sayany-Baikalian belt. Its Neoproterozoic amalgamation history involves early (~ 800 Ma) and late Baikalian (600–550 Ma) orogenic phases. Two palaeogeographic elements are identified in the early Baikalian stage — the Gargan microcontinent and the Dunzhugur oceanic arc. They are represented by the Gargan Glyba (Block) and the island-arc ophiolites overthrusting it. The Gargan Glyba is a two-layer platform comprising an Early Precambrian crystalline basement and a Neoproterozoic passive-margin sedimentary cover. The upper part comprises olistostromes deposited in a foreland basin during the early Baikalian orogeny. The Dunzhugur arc ophiolite form klippen fringing the Gargan Glyba, and shows a comprehensive oceanic-arc ophiolite succession. The Dunzhugur arc faced the microcontinent, as shown by the occurrence of forearc complexes. The arc–continent collision followed a pattern similar to Phanerozoic collisions. When the marginal basin lithosphere was been completely subducted, the microcontinental edge partially underthrusted the arc, and the forearc ophiolite overrode it. Continued convergence caused a break of the arc lithosphere resulting in the uplift of the submerged microcontinental margin with the overthrust forearc ophiolites sliding into the foreland basin. Owing to the lithospheric break, a new subduction zone, inclined beneath the Gargan microcontinent, emerged. Initial melts of the newly formed continental arc are represented by tonalites intruded into the Gargan microcontinent basement and its cover, and into the ophiolite nappe. The tonalite Rb–Sr mineral isochron age is 812 ± 18 Ma, which is similar to a U–Pb zircon age of 785 ± 11 Ma. A period of tonalite magmatism in Meso–Cenozoic orogenic belts is recognized some 1–10 m.y. after the collision. Accordingly, the Dunzhugur island arc–Gargan microcontinent collision is conventionally dated at around 800 Ma. It is highly probable that in the early Neoproterozoic, the Gargan continental block was part of the southern (in modern coordinates) margin of the Siberia craton. It is suggested that a chain of Precambrian massifs represents an elongate block separated from Siberia in the late Neoproterozoic, today. The Tuva-Mongolia Massif is situated in the northwest part of this chain. These events occurred on the NE Neoproterozoic margin of Rodinia, facing the World Ocean.

* Corresponding author.
E-mail address: kuzmich@ilran.ru (A.B. Kuzmichev).

0301-9268/01/$ - see front matter
PII: S0301-9268(01)00183-8

Keywords: Forearc ophiolite; Island arc–continent collision; Noproterozoic; Rdinia; Sberia; Tnalite; Tuva-mongolia massif

1. Introduction

The Tuva-Mongolia Massif (Fig. 1) has been recognized as a Precambrian continental block fringed by Palaeozoic fold belts (Ilyin, 1971). Today, it can be characterized as a composite terrane made up of Early Precambrian continental blocks and Neoproterozoic foldbelts. The Precambrian rocks are unconformably overlain by the Vendian-Cambrian shelf sediments, which permit the massif to be attributed to the Baikalides. The massif was amalgamated in the Vendian, and during the late Vendian-Cambrian was drifting as a microcontinent in the Palaeoasian Ocean. Later, in the early Ordovician, it was incorporated into the Sayany-Baikalian fold belt fringing the Siberia craton (Kuzmichev, 2000).

Previous palaeotectonic reconstructions involved a highly complicated and non-convincing Neoproterozoic pattern, with numerous multiform palaegeographic features: two island arcs, active and passive continental margins and various intra- and back-arc basins (Dobretsov et al., 1986; Belichenko et al., 1988; Dobretsov et al., 1989; Kuzmichev and Buyakaite, 1994; Kheraskova et al., 1995; Khomentovsky and Gibsher, 1996; Kuzmichev and Zhuravlev, 1999and others). Our new data show that these palaeotectonic features existed at two different stages of the Neoproterozoic, which permitted us to model them adequately. This paper focuses on the Early Neoproterozoic (1000–800 Ma) tectonics.

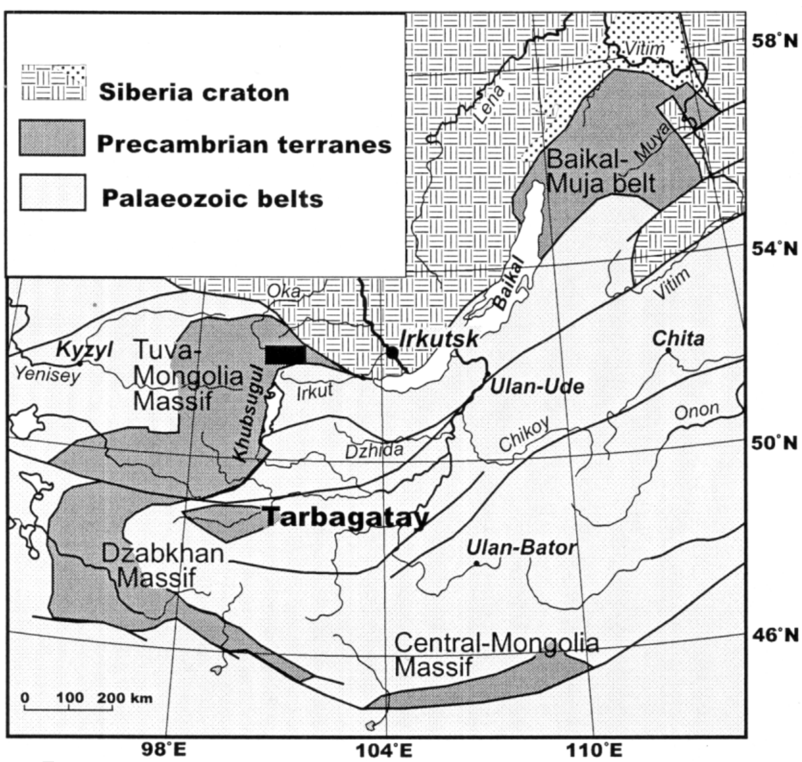

Fig. 1. Precambrian terranes in southern Siberia; filled rectangle shows the position of Fig. 2.

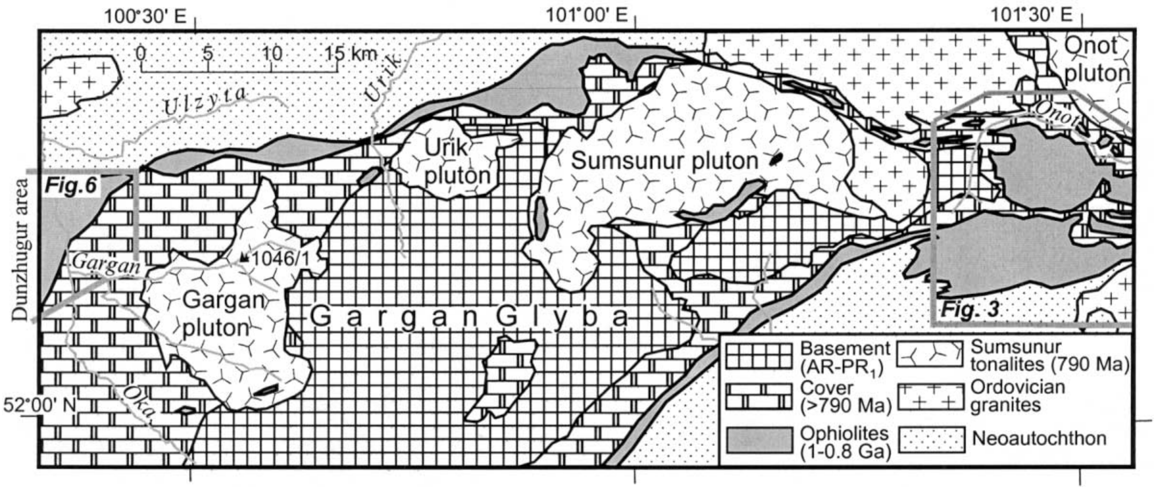

Fig. 2. Sketch map of the Gargan Glyba and surroundings. Based on the State Geologic Mapping and authors' surveys. Locations as in Figs. 3 and 6 are indicated.

We have studied the massif's northeastern part, noted for a prominence of Early Precambrian crystalline basement, known as the Gargan Glyba (glyba — lump, block) (Figs. 1 and 2). Exploration of the area began in the late nineteenth century owing to its ultramafic rocks containing nephrite and asbestos fields (see detailed description in Lodochnikov, 1936, 1941). In the 1950s, gold-bearing veins were discovered in the region, and the area has subsequently been thoroughly investigated. The ultramafic and associated rocks were described as island-arc ophiolites (Dobretsov et al., 1985).

Thus, two Neoproterozoic terranes — a continental block and an island arc have been recognized in the northeastern Tuva-Mongolia Massif. The continental block is conventionally named the Gargan microcontinent, and was presumably part of an Early Neoproterozoic continent. The arc was named the Dunzhugur island arc, as the Dunzhugur Ridge comprises a complete and well-studied ophiolite section (Sklyarov, 1990).

2. Gargan microcontinent

In the present-day structure, the Gargan microcontinent is represented by the Gargan Glyba (Fig. 2) with a classical platform composition: an Archaean–Early Proterozoic crystalline basement is unconformably overlain by the Neoproterozoic shelf deposits.

2.1. Basement

The microcontinent basement is composed of amphibolites, gneisses, and crystalline schists, which typically show plagioclase oikocrysts. Lodochnikov (1941) referred to these as 'pleuraegneisses', 'pleuraeamphibolites', etc. The relict mineral associations — orthopyroxene + biotite + garnet and clinopyroxene + Mg-rich garnet — indicate that these rocks have experienced the granulite facies metamorphism (Avdontsev, 1967). The rocks exhibit at least three stages of superimposed metamorphism (Belichenko et al., 1988; Nikitina, 1963). The Gargan Glyba basement is presumably Archaean, though reliable evidence is missing. The following isotopic data are available: (1) eight K–Ar dates on biotite, muscovite and amphibole obtained in the 1960s, 683–2370 Ma (Avdontsev, 1967; Mitrofanov et al., 1964); (2) a Rb–Sr mineral isochron yielding an age of about 3 Ga (unpublished data by Sklyarov); and (3) U–Pb dating of the Ordovician granites from the Daban-Zhalga 'dome', interpreted as the remobilization effect of an early Precambrian substrate (Khain et al., 1995).

2.2. Cover

The microcontinent cover is built up of the Irkut Formation marbles and the Ilchir Formation shales. Both the formations have been recognized and described as constituting a normal sedimentary succession unconformably overlapping the crystalline basement (Lodochnikov, 1941). Later, these rocks were subdivided into several units with local names. Some of them were considered Palaeozoic and interpreted as allochthons, containing volcanic rocks (for discussion see Kuzmichev, 2001). Our investigations revealed that Lodochnikov's (1941) description is closer to the truth. In the sections that were less deformed in the Palaeozoic, original stratigraphic contacts of the units are preserved, and the metavolcanic rocks are evidently ophiolite–nappe fragments (Figs. 3–5).

The Irkut Formation consists of a thick-bedded dolomitic marble, up to 600 m thick, which contains chert beds, in places altered to quartzite. The formation unconformably covers the crystalline basement, with a 120 m-thick basal conglomerate containing pebbles and boulders of metamorphic rocks (Lodochnikov, 1941; Mitrofanov et al., 1964).

The Irkut Formation was also found northeast of the Gargan Glyba, on the left bank of the Onot River (Figs. 3 and 4) which indicates that these outcrops belong to the Gargan microcontinent. Here the formation overlies the amphibole–biotite gneisses replaced at the top by muscovite

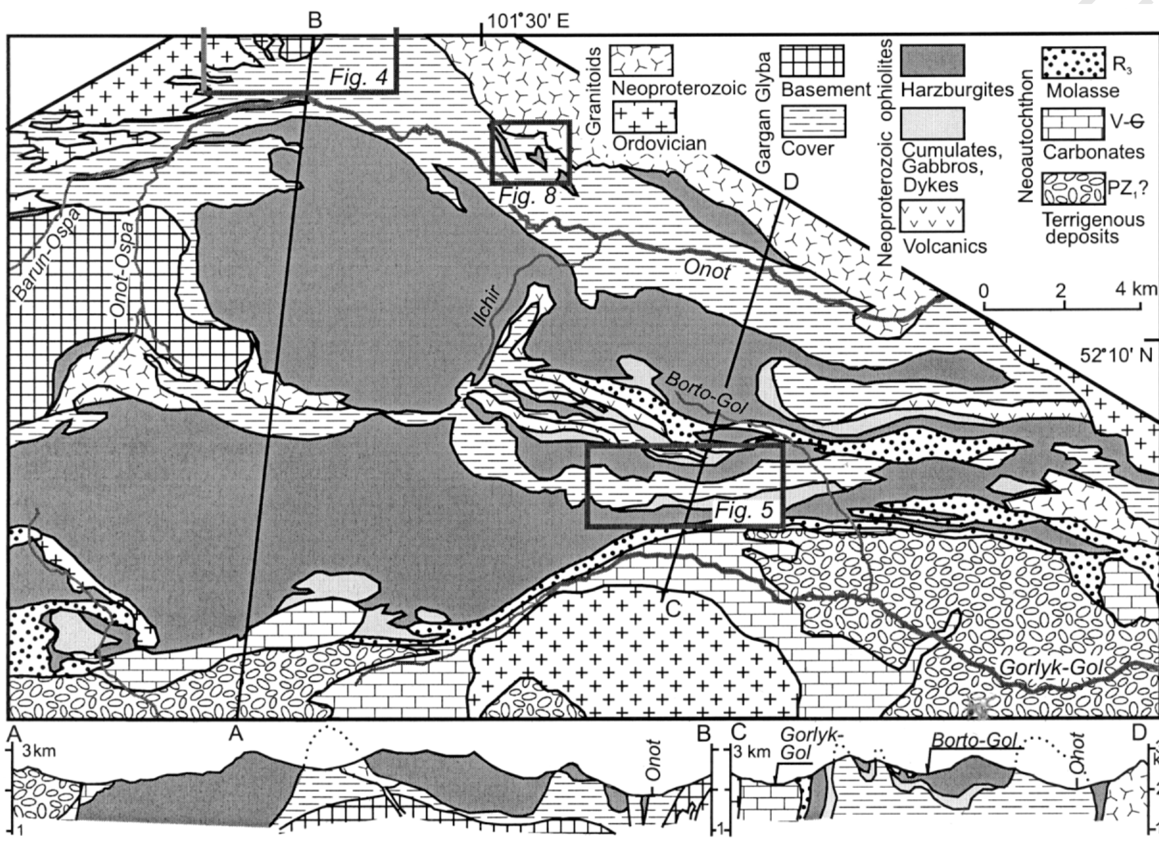

Fig. 3. Ophiolite allochthons in the Onot-Corlyk-Gol area. The map is based on authors' surveys, and data supplemented by Osokin and Khain. Locations as in Figs. 4–6 are indicated by rectangles.

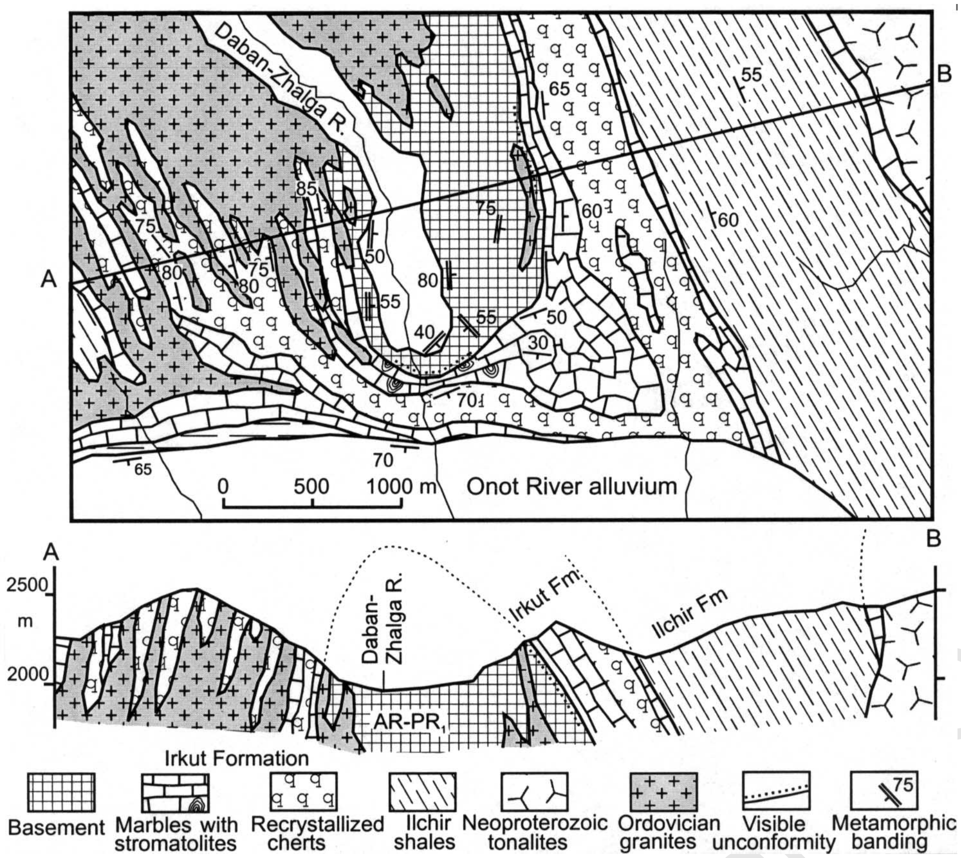

Fig. 4. Gargan Glyba cover on the Onot River north bank. See Fig. 3 for location.

gneisses, which are possibly a product of metamorphism of kaolinized rocks. The basal horizon of the Irkut Formation (about 15 m thick) comprises quartz–pebble conglomerate, muscovite quartzite and quartz–muscovite schist. The last-named contains up to 80% muscovite and is probably the metamorphosed product of an ancient weathering crust. The bulk of the formation, as in the Gargan Glyba, is made up of marble, in places stromatolitic, and recrystallized chert.

The Ilchir Formation conformably overlies the Irkut marbles on the Onot River north bank (Fig. 4). Its concordant lower contact was also described on the northern slope of the Gargan Glyba (Belichenko et al., 1988). Most of the formation is composed of dark gray shale along with sporadic interbeds of fine-grained sandstone and limestone.

The uppermost Ilchir Formation, sporadically exposed between the ophiolite klippen in the region between the Onot and Gorlyk-Gol Rivers (Fig. 3) is noted for the presence of carbonate rocks and olistostromes. These sections correspond to the inner zone of the Gargan microcontinent. Chiefly laminated limestones, varying considerably in thickness, represent the carbonate interbeds. The thick parts are composed of dolomite, possibly because of preferential bioherm dolomitization. Such dolomitic lenses could have easily turned into boudins beneath ophiolite nappes and cannot be confidently distinguished from exotic olistoliths (Fig. 5).

Exotic blocks are represented by massive or brecciated dolomite of various types, including brown dolomitic marble and yellowish dolomite cleaved with quartz veins. Single blocks are a few hundred meters long, forming klippen clearly distinguishable against the Ilchir dark shale. The blocks are accompanied by a breccia–conglomerate, confirming their olistostromal nature. In our opinion, most carbonate olistoliths originate from the Irkut Formation rocks.

The upper horizons of the Ilchir Formation also contain ophiolitoclastic olistostromes. Listwaenite and tremolitite debris (from a few centimeters to several meters long) were found in the shale matrix in the region between the Borto-Gol and Gorlyk-Gol Rivers. The shaly lamination envelops the debris. Serpentinite fragments are included in laminated limestones on the northern bank of the Borto-Gol River. Additionally, the shale contains numerous lenses of serpentinite, listwaenite, gabbro and metabasalt of unclear genesis. These could either be olistoliths or blocks in the clay melange.

Thus, the lower part of the Gargan microcontinental cover is made up of shallow-water dolomite, overlying the crystalline basement. The upper part is mainly composed of black shale. These sedimentary shelf deposits are typical of passive continental margins. The Proterozoic (> 790 Ma) age of the cover is determined from the age of the intruding Sumsunur tonalites (see Section 5). It is possible that the cover is mainly Early Neoproterozoic, as its upper horizons comprise carbonate and ophiolite–clastic olistostromes that were deposited synchronously with the orogeny (~ 800 Ma, see Section 6.1).

3. Dunzhugur island arc

The Dunzhugur island arc is represented by an ophiolite that forms two discontinuous belts fringing the Gargan Glyba. The belts, merging near the Glyba's eastern edge, are assumed fragments of a single, deformed and eroded allochthon, whose klippen also occur on the Glyba. The most representative ophiolite sections are exposed in the Dunzhugur area, to the west of the Gargan Glyba, and the region between the Gorlyk-Gol, Onot, and Sagan-Sair Rivers, to the east of the Glyba (Fig. 2).

The Dunzhugur area shows the most complete ophiolite succession in the Sayany-Baikalian belt, and has been described in detail (Dobretsov et al.,

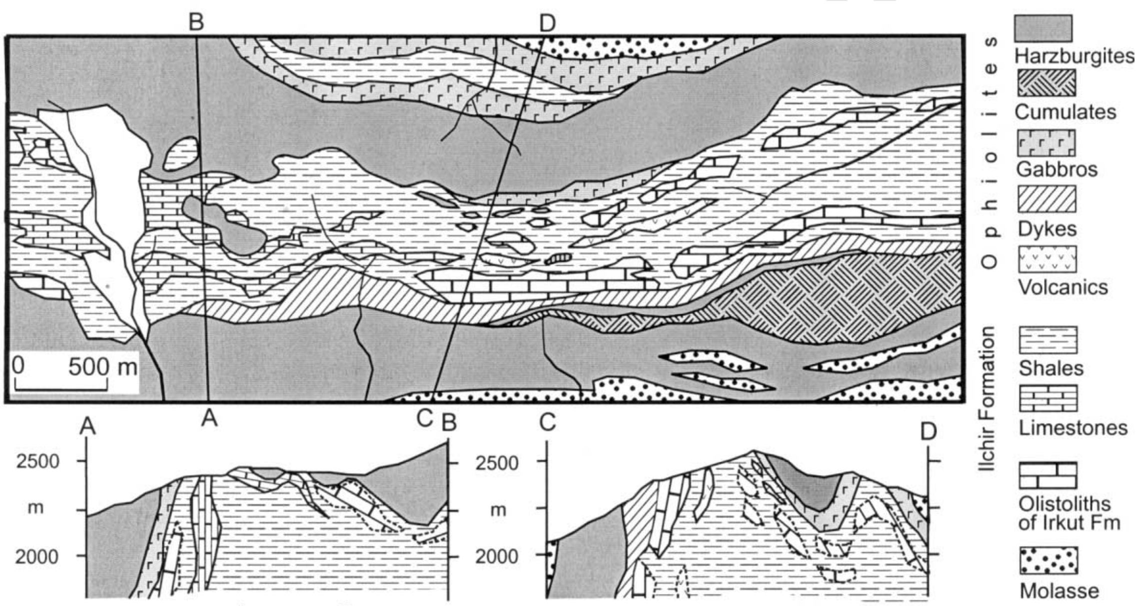

Fig. 5. Sketch map of ophiolite nappes and the underlying Ilchair olistostromes in the Gorlyk-Gol and Borto-Gol watershed. See Fig. 3 for location.

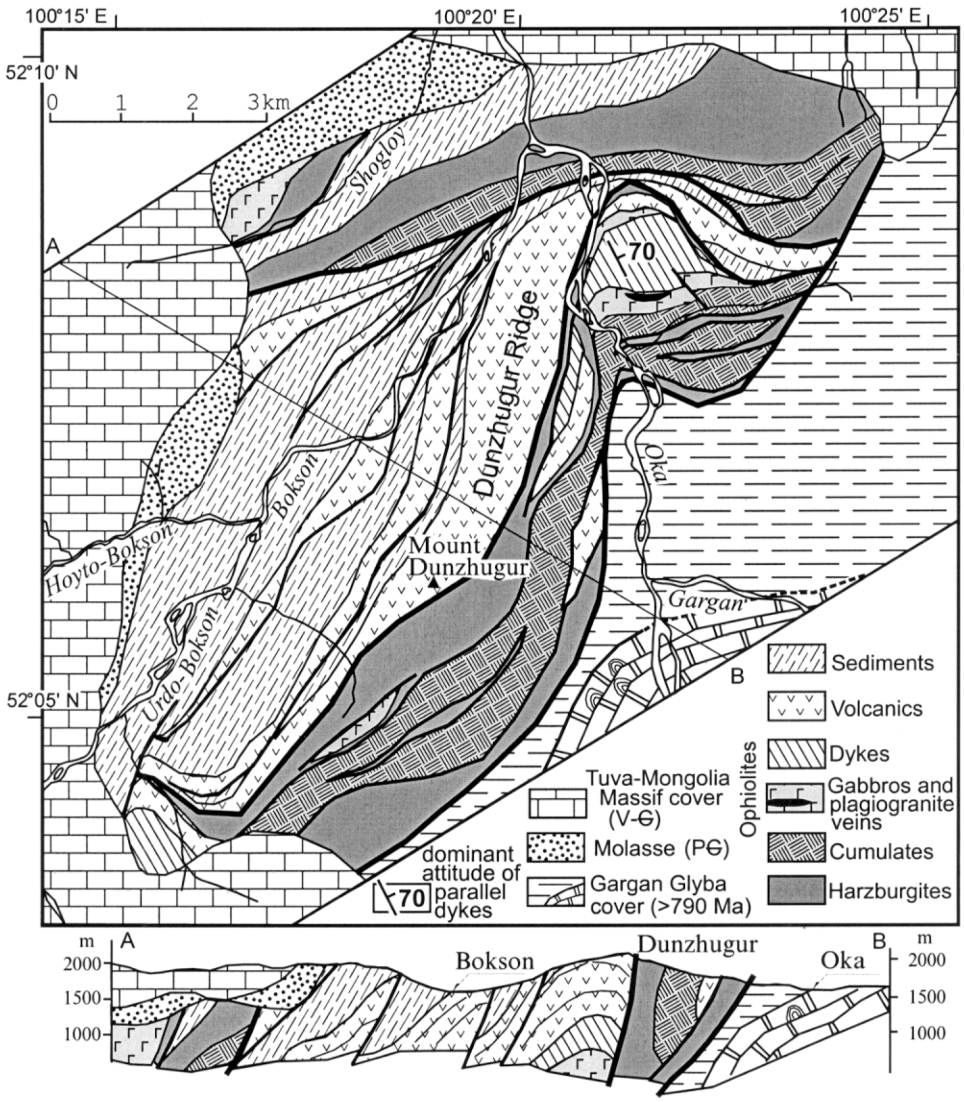

Fig. 6. Ophiolite nappes in the Dunzhugur area. Cenozoic rocks and diabase sills omitted. See Fig. 2 for location.

1985, 1986; Sklyarov, 1990). These authors determined the calc-alkalic and boninitic composition of the dykes and volcanic suites, which indicates an oceanic island-arc environment. Our mapping has shown that the present-day structure of the area is a packet of steeply inclined stacks, where the ophiolites are shuffled without any evident regularity (Fig. 6). Interestingly, the intra-arc sheeted dykes exposed on the eastern bank of the Oka River trend NNW, transverse to the regional strike of the lithological layering (Fig. 6).

The sedimentary rocks associated with the ophiolites are highly fragmented, owing to thrusting and intrusion of numerous post-ophiolitic, tholeiitic dolerite sills. We believe that the lower members of this composite sedimentary succession concordantly overlie the pillow lavas. They are composed of turbidites and hemipelagic cherty shales exposed on the western slopes of the Dunzhugur Ridge and on the Bokson River banks.

The turbidites show lenses of breccia–conglomerate, representing debris flow deposits. The debris composition is identical to the ophiolite volcanic rocks; gabbro and ultramafic fragments are more rare. Monomictic hyaloclastic sandstones and conglomerates consisting of redeposited pillow-lava fragments also occur. Submarine slump folds indicate that the material was transported from NE to SW in modern coordinates. The upper parts of the compound sedimentary section in the Dunzhugur area are noted for the presence of redeposited silicic pyroclastic rocks. They probably represent the late stages of Dunzhugur arc magmatism, and it is unknown whether they are in situ.

Island-arc lavas overlain by turbidities suggest an environment similar to that of recent forearc basins (Bloomer et al., 1995). In the latter, deposits are chiefly represented by turbidities with lenses of unsorted coarse-grained sediment mainly composed of volcanic debris and containing serpentinized ultramafic and coral limestone clasts (Underwood et al., 1995). The presence of mafic sills appears to be a common feature in the intra-oceanic forearc basins (Taylor et al., 1995), and the Dunzhugur rocks agree well with such an interpretation.

East of the Gargan block, the ophiolite allochthon exhibits synforms separated by crest-like anticlines into which the Ilchir shales have been squeezed. The allochthon shows erosional windows and klippen, and is less disturbed than in the Dunzhugur region. It is chiefly composed of serpentinized harzburgite. Layered ultramafic rocks, massive gabbros, sheeted dykes and mafic volcanic rocks are confined to the allochthon foot, and display a reverse sequence (Figs. 3 and 5, Kuzmichev, 2001).

4. Ophiolite emplacement model

The obduction model for the ophiolite can be deduced from their different setting in the marginal and inner parts of the Gargan microcontinent. In the northern and western fringes of Gargan Glyba the ophiolite overlaps different horizons of Glyba's cover, and in places its crystalline basement (Fig. 2). The allochthon's base is highly tectonized. In the inner part, east of the Gargan Glyba, the allochthon lies close to conformity on Ilchir Formation's upper horizons, containing a carbonate olistostrome, which, in places, shows no signs of tectonic effect on the underlying rocks. These relationships can be interpreted as follows: the allochthon overthrust the microcontinent margin, tearing off its platform cover. The cover's fragments were unloaded into the foreland basin, yielding olistostromes on which ophiolites slid during sedimentation. First, their upper portion slid down to form the base for the packet of sheets, thus producing a reverse sequence where ophiolites appeared in inverse stratigraphic order. The suggested model also accounts for the difference in the structure of allochthons in the inner and marginal areas. In the inner site, a slightly disrupted gravitational nappe is reconstructed. In the marginal area, ophiolite stacks are piled up, which can be explained by stacking in front of the bulging edge of the Gargan microcontinent (Fig. 7).

5. The age of ophiolite obduction

Until now, the age of ophiolite obduction was believed to be Vendian or Palaeozoic (for discussion see Kuzmichev, 2001). New dating of the

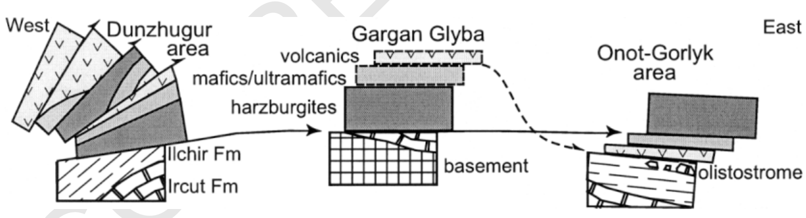

Fig. 7. Ophiolite obduction model. Note difference in the allochthon setting and structure east and west of Gargan Glyba. Most symbols correspond to those in Figs. 5 and 6. For an explanation, see the next.

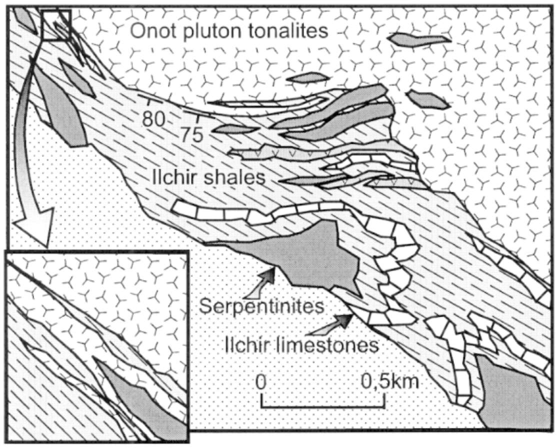

Fig. 8. Onot pluton tonalities intruding ophiolites and rocks of Ilchair formation. Location is Onot River northern bank, 2 km up-stream from the Darmey River entry. See Fig. 3 for location. Symbols are the same as in Fig. 5.

Sumsunur suite tonalites (Fig. 2), which intrude the ophiolite allochthon, show that obduction was Early Neoproterozoic.

Avdontsev (1967) identified the Sumsunur suite and studied the Gargan, Urik, and Sumsunur plutons (Fig. 2) which are made up of tonalite, with occasional trondhjemite, gabbro and diorite. The Onot pluton (Fig. 2) is composed of similar rocks and probably belongs to the same suite. Thus, the suite includes four discordant plutons (each 75–300 km^2) trending along the northern margin of the Gargan Glyba and its eastern extension (Fig. 2). The suite also shows numerous dykes mainly of dacitic composition. The rocks intrude the Gargan Glyba basement and cover as well as the relics of the ophiolite allochthon (Figs. 2 and 8).

The Sumsunur rocks were dated using K–Ar methods in the 1960s as 457, 464, 466, and 484 Ma (Avdontsev, 1967). These dates agreed with the recent dating of the Tuva-Mongolia Massif's common calc-alkaline granites (Khain et al., 1995; Litvintsev and Kalmychkova, 1990) and were considered credible. Our work disproves this viewpoint and establishes a Neoproterozoic age for the Sumsunur suite.

We obtained a Rb–Sr and U–Pb age for the amphibole–biotite tonalites of the Gargan pluton (Fig. 2). The samples were collected on the left bank of the Khoito-Gargan River, 500 m up-stream from the Dunda-Gargan River entry (Fig. 2, sample 1046/1). This is, perhaps, the only exposure in the Gargan pluton that shows fresh massive rocks almost unaffected by alteration. The composition of the dated rock is presented in Tables 1 and 2.

The Rb–Sr isochron is constructed with four points: biotite, amphibole, plagioclase and whole-rock (Table 3, Fig. 9). The isochron parameters are: $T = 812 \pm 19$ Ma; $I_0 = 0.70459 \pm 0.00005$; MSWD = 7.1. By excluding the point for saussuritized plagioclase, we can improve the isochron parameters slightly: $T = 812 \pm 18$ Ma; $I_0 = 0.70458 \pm 0.00003$, MSWD = 2.9. The U–Pb dating was carried out on three zircon fractions (Table 4, Fig. 9). All of them have lost a minor amount of radiogenic Pb. The age determined from the upper intersection with concordia is

Table 1
Chemical composition (in wt%) of sample 1046/1

SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	CaO	MgO	Na_2O	K_2O	P_2O_5	LOI	Total
63.77	0.48	15.96	1.62	3.3	0.09	6.02	1.61	4.15	1.04	0.21	1.04	99.29

Table 2
Contents (in ppm) of trace and rare earth elements in sample 1046/1

Ni	Cu	Zn	Ga	Pb	Rb	Sr	Y	Zr	Ba	La	Ce	Nd	Sm	Eu	Gd	Er	Yb
24	12	60	17	24	21	393	19	94	635	19	35	16	3.8	1.2	3.3	1.5	1.1

Table 3
Rb-Sr isotopic data for sample 1046/1

	Rb (ppm)	Sr (ppm)	^{87}Rb/^{86}Sr	^{87}Sr/^{86}Sr
Whole-rock (WR)	21.4	405	0.1527 ± 0.0008	0.706361 ± 0.000017
Plagioclase (Pl)	4.42	645	0.0199 ± 0.0002	0.704802 ± 0.000017
Amphibole (Am)	7.60	38.6	0.570 ± 0.0008	0.711136 ± 0.000023
Biotite (Bi)	155	15.7	29.7 ± 0.4	1.038059 ± 0.000018

The analyses were performed on a mass spectrometer Finnigan MAT-262 at IGEM, Moscow. Isochron calculation after York (1966).

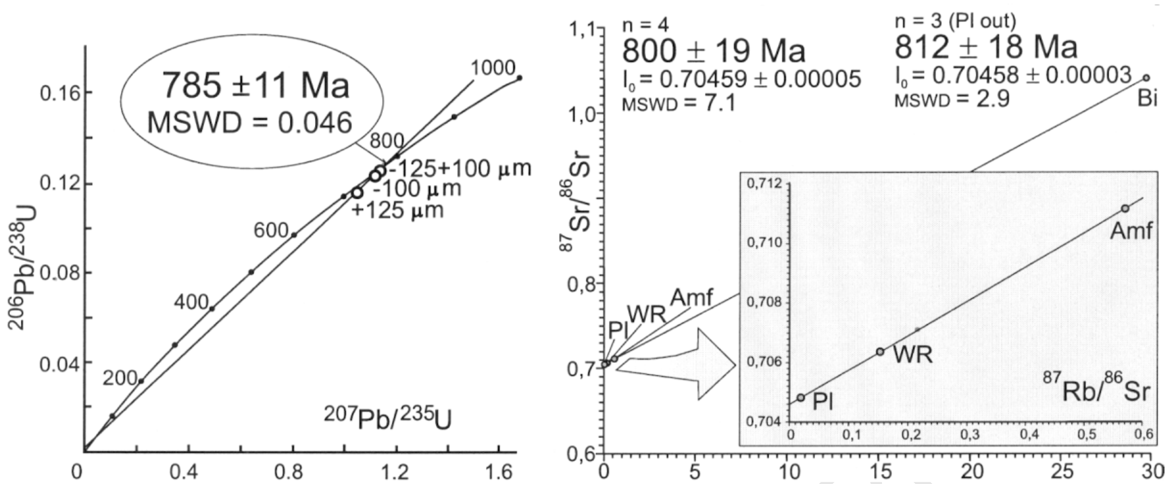

Fig. 9. U–Pb and Rb–Sr diagrams for the Gargan pluton tonalites (sample 1046/1).

Table 4
U–Pb isotopic data for zircons from sample 1046/1

Frac-tion (μm)	Weight g	Contents (ppm)		Isotopic composition of Pb			Isotopic ratio and age, Ma		
		U	Pb	^{206}Pb/^{204}Pb	^{206}Pb/^{207}Pb	^{206}Pb/^{208}Pb	^{206}Pb/^{238}U	^{207}Pb/^{235}U	^{207}Pb/^{206}Pb
+125	0.0029	71.23	9.07	585.5	11.144	5.377	0.1164	1.0472	0.0652
							710	727	782 ± 9
−125	0.0027	63.05	8.45	2400	13.99	5.768	0.1256	1.1351	0.0655
+100							763	770	791 ± 7
−100	0.0026	71.81	9.13	2200	13.908	7.21	0.1237	1.1661	0.0654
							752	761	788 ± 7

Zircons were analyzed by the method outlined by Krogh (1973). U and Pb concentrations were determined by the isotopic dilution using a mixed spike 208Pb+235U. The lead blank was 0.1 ng. Isotopic composition was measured on a mass spectrometer TSN 206A, Cameca Co. at the Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Science. Ages were calculated using the ISOPLOT program of Ludwig (1991). Errors in the U/Pb ratios were 0.5%. Common lead correction was according to Stacey and Kramers (1975) model for 750 Ma.

785 ± 11 Ma, with MSWD = 0.046. The lower intersection (41 ± 187 Ma) is interpreted as a loss of radiogenic Pb in recent times. This result agrees well within the error with the Rb–Sr age. As the U–Pb dating is considered more reliable we interpret the age of the tonalites to be ca. 790 Ma. This is the first Neoproterozoic age available for the Tuva-Mongolia Massif granitoids, and it proves that the Irkut and Ilchir deposits had accumulated, and the ophiolites were obducted, before 790 Ma.

6. Discussion

6.1. Tonalite generation and the age of arc–continent collision

Ophiolite obduction resulted from the Gargan microcontinent–Dunzhugur island arc collision. The younger age limit of this event is determined by dating the cross-cutting Sumsunur-suite tonalites. Theoretically, the collision could have occurred at any earlier stage in Precambrian history. However, it is argued that the arc–continent collision directly the preceded tonalite generation.

Most researchers agree that the trondhjemite–tonalite–dacite rock series (TTD) are typical of continental and oceanic arcs and are subduction-related (Arth, 1979; Barker, 1979; Drummond et al., 1996). The general opinion is that these rocks have formed by the partial melting of a mafic igneous source with little or no fractional crystallization. The two types of TTD recognized are the high-Al (adakites), and the low-Al types (Defont and Drummond, 1990). Geochemical criteria that help to distinguish between them are based on the Al_2O_3, Y, Yb, Sr contents and the Sr/Y, La/Yb, MgO/FeO, $^{87}Sr/^{86}Sr(i)$ ratios, reflecting the mineral composition of the restites. Garnet with high heavy REE and Y concentrations, contained in the restite, is responsible for the peculiar adakite signatures, thought to be generated by the partial melting of a subducted slab at eclogite-facies depths. The low-Al magmas form at shallow depths and are in equilibrium with the amphibole–plagioclase restite, which is interpreted as having a lower-crustal genesis. Experimental data show that when quartz is absent, amphibolite can turn into eclogite at a minimum pressure of 8–10 kbar, corresponding to a depth of 30–35 km (Cloos, 1993 and references therein). Consequently, the chemical composition is not the only criterion sufficient to distinguish lower crustal and slab melts in case of subducton below a thick plate. This is especially true for the Sumsunur-suite rocks, which belong to the intermediate type between the high-Al ('slab') and low-Al ('lower-crustal') TTD series.

Thus, the Sumsunur tonalites could be the product of melting of either the lower continental crust or the subducted slab. The former is unlikely, as this way of melting does not explain the brevity of the Sumsunur magmatic episode, or its specific composition. Partial melting of the mafic lower crust requires the participation of an extra heat source, usually related to magmatic underplating (Barnes et al., 1996; Athernon and Pettford, 1993; Wolge et al., 1996). This must have caused extensive heating of the lower continental crust. As a result, trondhjemite–tonalite intrusions, assumed to be related to underplating, generally enter as part of the diorite–tonalite–granite composite batholiths. Similar Ordovician batholiths are also widespread in the Tuva-Mongolia Massif (Kuzmichev, 2000).

Slab melting was suggested to explain the origin of distinctly Archaean TTD. Similar magmas are also generated in modern supra-subduction environments, though in considerably less volume owing to the greater depth of recent isotherms (Defont and Drummond, 1990; Morris, 1995; Drummond et al., 1996). The calculations show that in recent times, melting can be possible only when a young, warm oceanic plate participates in the subduction (Drummond et al., 1996; Kincaid and Sacks, 1997). The maximum age of the oceanic crust which can still experience melting ranges from 25 (Drummond et al., 1996) to 5 Ma (Peacock et al., 1994) provided there are no additional heat sources. Among other necessary conditions, slow subduction (<3 cm per year) permits convective heating of oceanic crust and additional shear heating (Barnes et al., 1996; Kincaid and Sacks, 1997; Peacock et al., 1994). Other factors favoring melting are: oblique subduction (Yogodzinski et al., 1995; Peacock et al., 1994); low-angle slab inclination ($<27°$, Peacock et al., 1994); and initial subduction (Kincaid and Sacks, 1997; Sajona et al., 1993). Modeling has shown that '…the maximum slab temperatures occur early after subduction initiation and gradually decay with time' (Kincaid and Sacks, 1997, p. 12, 13).

We must bear this in mind when discussing the origin of the Sumsunur tonalites. They are the first suprasubduction melts along the newly formed active margin of the Gargan microcontinent. As shown in Section 6.2, subduction under the Gargan microcontinent was triggered by splitting the Dunzhugur island arc in the area of the highest uprise of isotherms corresponding to the volcanic zone (Fig. 10b). This zone shows the thinnest and weakest lithosphere (Konstantinovskaya, 1999). It includes a hydrated mantle wedge containing some melt (Schmidt and Stefano, 2000). As a result of the arc break, fragments of hot, hydrated, island-arc lithosphere could have been involved in the initial portions of the subducting slab (Fig. 10c). This provides the most favorable melting conditions. It is these fragments that first pass into the mantle and experiences the greatest shear heating. This hypothesis agrees well with the fact that the subduction-related initial TTD magmas were rapidly replaced by calc-alkaline differentiated andesite–dacite–rhyolite compositions. This change can be explained as follows: (1) a decrease in shear stress when subduction had stabilized and the temperature in the slab roof rose to the brittle–plastic transition; (2) depletion and dehydration of the subducted fragment of the island-arc lithosphere; and (3) progressive metasomatism of the overlapping mantle wedge, responsible for the calc-alkaline magmatism.

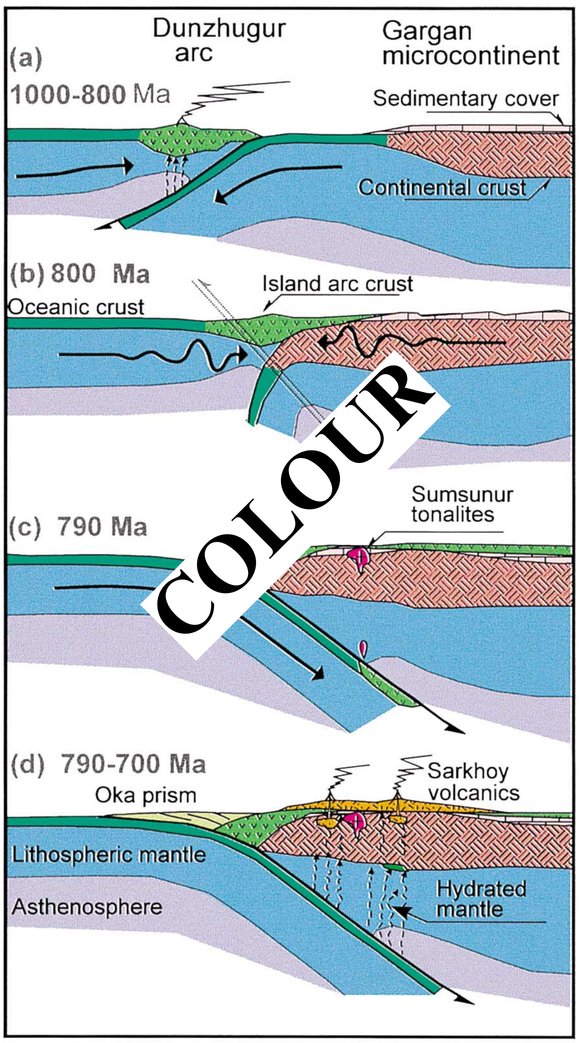

Fig. 10. Model for Dunzhugur arc–Gargan Microcontinent collision. For an explanation, see the text.

The Sumsunur tonalites melted at depths of 35–60 km. Simple calculations indicate that a subsiding plate will reach this depth in several million years after the onset of subduction, even if it moves at a minimal rate of 1 cm per year. Well-studied Meso–Cenozoic examples confirm such estimates. For example, in the Mindanao Islands (Philippines), the slab magmas were generated 1–2 m.y. after the onset of subduction (Sajona et al., 1993); and in the Klamath Mts. (California), tonalite–trondjemite magmatism was manifested some 5–10 m.y. after arc–continent collision and continued for a brief time period (Barnes et al., 1996). These examples assure that such tonalite–trondhjemite magmatic emplacements are generally manifested within 1–10 m.y. after arc–continent collision and the initiation of a new subduction zone. Hence, the early Baikalian orogenesis in the Tuva-Mongolia Massif is conventionally dated at ca. 800 Ma.

6.2. Early Neoproterozoic tectonic events in the tuva-mongolia massif

The Gargan microcontinent passive margin is assumed to have existed since the Mesoprotoerozoic (Nikitina, 1963). The Early Neoproterozoic formed the Dunzugar island arc. The age of the arc is constrained by dating the ophiolitic plagiogranites forming veins in the layered gabbro on the east bank of the Oka River (Fig. 6): the zircon U–Pb coventional and $^{207}Pb/^{206}Pb$ evaporation ages are 1020 Ma (Khain et al., 1999). Such a long-lived volcanic arc (200 Ma) inevitably influenced the crustal composition and thickness. The scarcity of data does not permit this period to be studied in detail, as we can observe only forearc fragments formed mainly at the initial stage of arc evolution. It is possible that the later stages of the island-arc magmatism are represented by felsic volcanics that are observed to be redeposited in the upper part of the forearc sediments in the Dunzhugur area.

The occurrence of forearc complexes indicates that the arc was facing the microcontinent, and the oceanic plate separating them was subducting under the arc (Fig. 10a). The present-day strike of intra-arc sheeted dykes (which develop primarily parallel to the arc trend) suggests that the arc was almost transverse to the Gargan microcontinental margin. An alternative interpretation is that the attitude of the dykes could have been affected by block rotation during oblique collision.

After complete subduction of the marginal basin lithosphere, the edge of the Gargan microcontinent was also partly subducted under the arc. Passive margins, thinned by rifting can subduct as deep as 100 km (Fig. 10b; Cloos, 1993; Malavieille and Chemenda, 1997). Simultaneously, the island-arc lithosphere overrode the continental margin. A similar situation was considered as a most plausible reason for ophiolite obduction (Coleman, 1977). Subsequent events follow a typical pattern as suggested by McKenzie (1969) for the arc–continent collision: a new subduction zone arises, inclined beneath the continent. In recent examples, this zone coincides with the island arc with the thinnest lithosphere (Konstantinovskaya, 1999) (Fig. 10b). The lithosphere's break-off puts an end to pulling the continental margin into the subduction zone. As a result, the submerged microcontinental edge with overthrust forearc ophiolites was torn off its roots and floated upwards. This caused the gravitational sliding of the ophiolite allochthon toward the foreland basin. A split-off part of the heated island-arc lithosphere, probably with active magma chambers, was involved in the initial portion of the slab subducted under the continental margin, which provoked the generation of the Sumsunur tonalites (Fig. 10c). Stabilized subduction resulted in the formation of mainly felsic calc-alkaline magmas represented by volcanics of the Sarkhoi and Darkhat suites (Fig. 10d; Kuzmichev, 1990).

As a result of collision, some 800 Ma ago, the volume of the Gargan microcontinent increased as the mature Dunzhugur arc lithosphere was attached to it and partly obducted. Phanerozoic examples indicate that the island-arc magma production rates estimated for the Aleutians (average for 75 m.y.) and Izu-Ogasawara (average for 47 m.y.) arcs are 60 ± 10 km^3/km per 1 Ma (Holbrook et al., 1999). Evidently, a considerable volume of island-arc crust was generated over 200 m.y. to enlarge the microcontinent. The active continental margin, newly formed in the mid-Neoproterozoic, can be mapped by outcrops of the subaerial Sarkhoi and Darkhat volcanics (Fig. 11).

6.3. Baikalian events in the tuva-mongolia massif as related to rodinia evolution

The Gargan continental block with its granulite basement and Mesoproterozoic cover is presumably a fragment of an old craton. The nearest craton to the Gargan Glyba is Siberia. Along its southern margin, the Baikal-Muja belt has a Neoproterozoic history, congeneric with the Tuva-Mongolia Massif. The belt is known for its Neoproterozoic ophiolites of supra-subduction and MORB-type overthrusting the Siberia cratonic margin (Dobretsov et al., 1992; Tsygankov, 1998). The Sm–Nd age for the MORB ophiolites is 1035 ± 92 Ma (Rytsk et al., 1999). This age corresponds with that of the Dunzhugur ophiolite (Khain et al., 1999). The orogenesis in the Baikal-

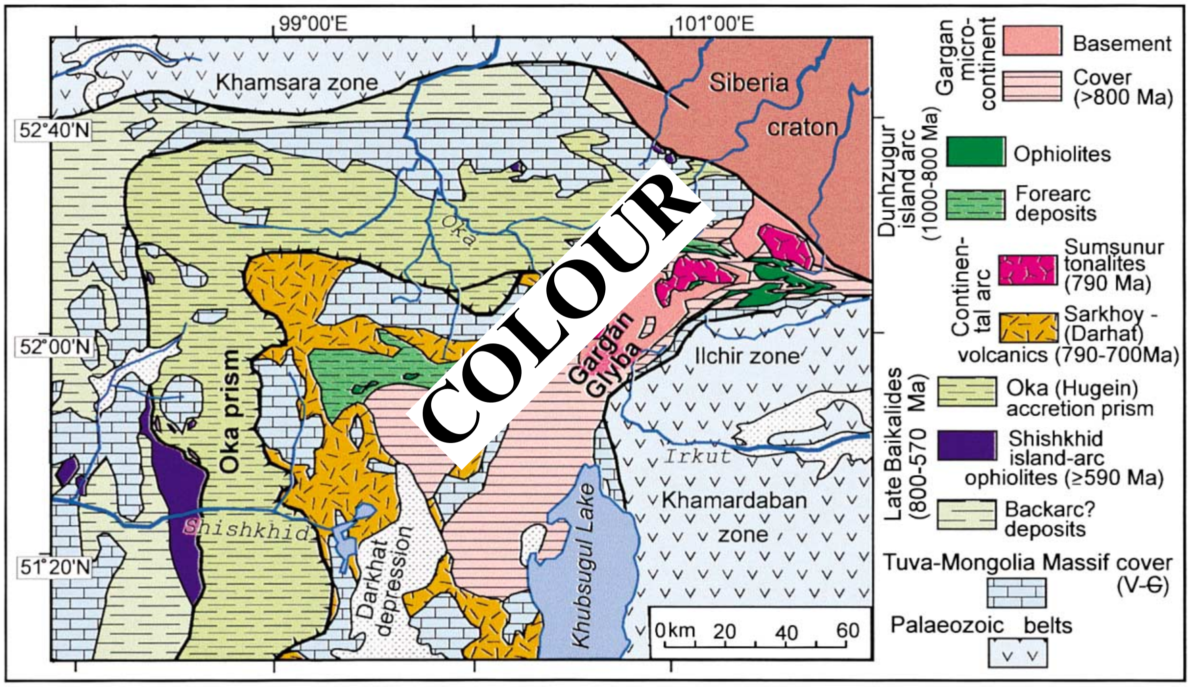

Fig. 11. Geologic zoning of the northern Tuva-Mongolia Massif. Palaeozoic granitoids are omitted. See the toothed line contouring the late Neoproterozoic active margin of the Gargan microcontinent. Compare with Fig. 10d. Late neoproterozoic complexes are not discussed in this paper.

Muja belt and ophiolite obduction on the Siberia craton occurred in the mid-Neoproterozoic: the U–Pb zircon age of syncollisional granites is 815 ± 46 Ma (Rytsk et al., 1999) which is geochronologically indistinguishable from the arc–continent collision in the Tuva-Mongolia Massif. Thus, in the Early Neoproterozoic, the southern margin (in modern coordinates) of Siberia was facing the ocean. This agrees with reconstructions of Rodinia showing an ocean-facing position of the Siberian south margin (Fig. 12a).

This ocean-facing margin probably continued to the South China craton where the Neoproterozoic island-arc ophiolites were found to overthrust the Huanan Block (Li et al., 1997). There are 11 Sm–Nd dates available for ophiolite gabbroids and volcanics, scattered in the range of 1.0 ± Ga, and one SHRIMP zircon age is 968 ± 23 Ma; the age of postcollision granites is 0.9–0.8 Ga (Li et al., 1995; Li, 1999 and references therein).

The above-mentioned data suggest that, within the time interval of 1000–800 Ma, there existed an elongate island-arc system located NE of Rodinia (Fig. 12). This system collided with Rodinia between 0.9 and 0.8 Ga. In the South-China region, this was followed by the continental collision between the Yangtze and Huanan blocks and oceanic basin closure. Meanwhile, the Siberia craton margin was transformed into an active continental margin (Fig. 12c). In the Baikal-Muja belt, continental–arc magmatism was represented by the subaerially differentiated volcanics of the Padrin group (765 ± 50 and 712 ± 40 Ma, Rb–Sr whole rock isochrons, Mitrofanov, 1989).

Similar volcanics also occur in the Tuva-Mongolia Massif, forming the Sarkhoj and Darkhat suites (Fig. 11; 718 ± 30 Ma, Rb–Sr whole rock isochron, Buyakaite et al., 1989). Analogous Neoproterozoic calc-alkaline volcanics (Dzabkhan suite) are also present in the Dzabkhan Massif located to the south (Fig. 1). The Dzabkhan

Massif is also an old continental fragment with the Archaean basement (Kozakov et al., 1997), east of which lies the Central Mongolia Massif. This massif is poorly studied, though its Vendian–Cambrian carbonate cover is known to overlie a metamorphic basement (Osokin, 1999). The last feature is specific to all the three massifs and distinguishes them from the surrounding Palaeozoic belts. Thus, we suggest that the chain of three Precambrian massifs is an elongate fragment of the Siberian craton margin, which primarily filled the gap between the Baikal-Muja and South China Neoproterozoic belts (Fig. 12c). This fragment was most likely detached from the Siberian craton in the Late Neoproterozoic, in the same way as Japan was detached from Asia in the Cenozoic. The unique position of the marginal Baikal-Muja segment, deeply cut into the craton, probably hindered its detachment (Fig. 12d). The Precambrian massifs are now separated from the Siberia craton by Palaeozoic belts [Dzhida zone, Bajanhongor zone, etc. (Fig. 1)] whose most ancient complexes are Vendian ophiolites. This indicates that the craton broke up not later than at the end of the Neoproterozoic. The breakup may have begun earlier as a back-arc extension in the rear of the continental arc.

The hypothesis suggested here appears more realistic than the previous views, based on the assumption that the Tuva-Mongolia, Dzabkhan

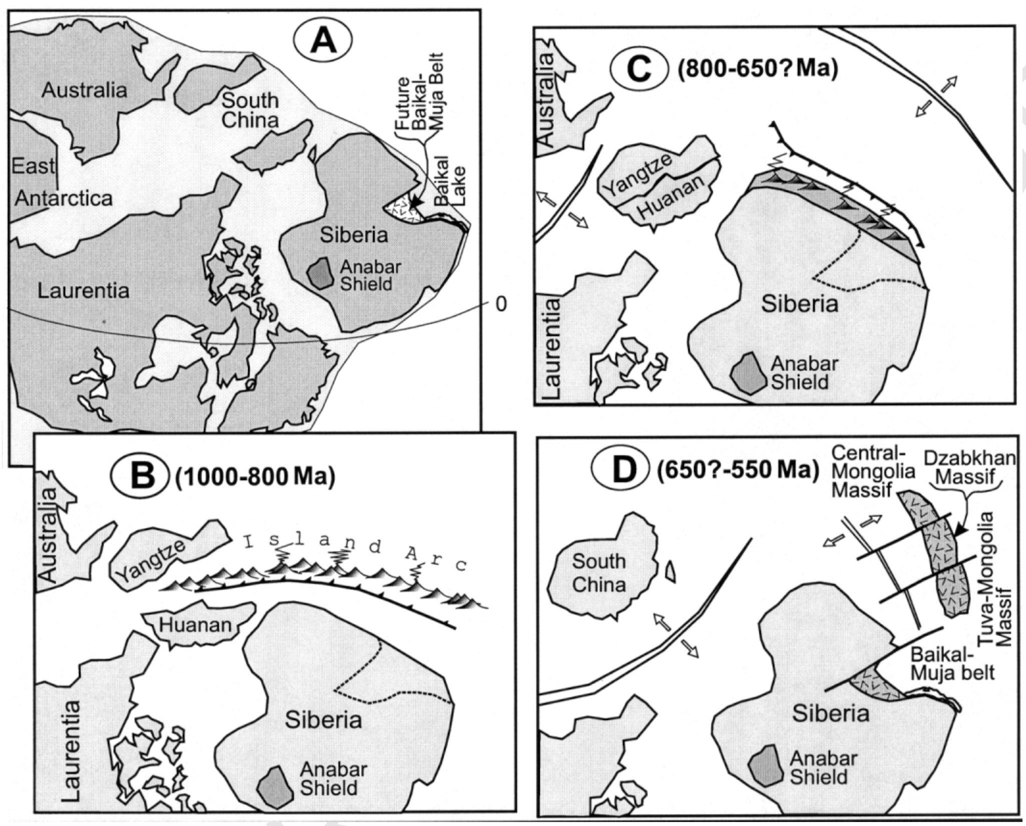

Fig. 12. Speculative model for origin of the Tuva-Mongolia Massif as related to the break-up of Rodinia. (a) Reconstruction of NE Rodinia. Compiled after: Evans et al. (2000), Li (1999), Li et al. (1995), Pelechaty (1996), Unrug (1997). (b) The early Baikalian stage. Dunzhugur island arc is opposite the Siberian craton passive margin. (c) Situation after the island arc–continent collision. Note Siberian craton active margin. Teech on the toothed line indicate direction of subduction. (d) Splitting of the Siberia craton and detachment of elongate fragment yielding Tuva-Mongolia, Dzabkhan, Central-Mongolia Precambrian massivs and the Baikal-Muja belt. (Fig. 12 b–d show relative position of continents without considering their changing physical coordinates).

and Central Mongolia massifs were fragments of the East Gondwanan active margin (Kheraskova et al., 1995).

7. Conclusions

The Early Neoproterozoic features of the northern Tuva-Mongolia Massif include a fragment of a continental passive margin and fragments of an oceanic island arc. About 800 m.y. ago, there was an arc–continent collision and forearc ophiolite obduction on the continental margin. The record of Neoproterozoic orogenic events is incomplete, though according to the available data, the collision resembled Meso–Cenozoic examples. It included the following: (a) blocking of subduction by the continental plate pulled under the forearc lithosphere; (b) reversal of subduction because of an island-arc lithosphere break; and (c) uplift of the subducted continental margin and gravitational sliding of ophiolite nappes into the foreland basin. One of the intriguing features of Neoproterozoic history is the apparently long life span of the oceanic island arc of about 200 m.y. As a result of subduction polarity reversal, an active magmatic belt is formed at the continental margin. The initial magmatism is represented by a tonalite–trondjhemite–dacite series (790 Ma), later replaced by differentiated basalt–andesite–dacite–rhyolite series (718 Ma). It is probable that these events occurred at the southern (in modern coordinates) margin of the Siberian craton. In the Late Neoproterozoic, an elongate block was detached from the craton, which today is preserved as a chain of Precambrian massifs with the Tuva-Mongolia Massif at the northwestern end. This interpretation is in accord with the reconstructions of Rodinia that shows Siberian craton's southern margin facing the Neoproterozoic World Ocean.

Acknowledgements

The study has been financially supported by RFBR project N 98-05-64876. We appreciate the assistance of E. Khain, A. Osokin and E. Mitjukhin, whose help made it possible to carry out the investigations in the Onot-Gorlyk region. Their field maps were very useful in the study of the area, as well as field sketches of the Dunzhugur area kindly provided by E. Sklyarov. We are greatly indebted to Prof. C.McA. Powell who has given considerable time and attention to improve the manuscript and to Dr. Yu.A. Zorin for a constructive review. This paper is a contribution to IGCP 440: Assembly and Breakup of Rodinia.

References

Arth, J.G., 1979. Some trace elements in throndhjemites-their implications to magma genesis and paleotectonic setting. In: Barker, F. (Ed.), Throndhjemites, Dacites and Related rocks. Elsevier, Amsterdam, pp. 123–132.

Athernon, M.R., Pettford, N., 1993. Generation of sodium-rich magmas from newly underplated basaltic crust. Nature 362, 144–146.

Avdontsev, N.A., 1967. The Gargan Block Granitoids. Nauka, Leningrad (in Russian).

Barker, F., 1979. Throndhjemite: definition, environment and hypotheses of origin. In: Barker, F. (Ed.), Throndhjemites, dacites and related rocks. Elsevier, Amsterdam, pp. 1–12.

Barnes, G.C., Petersen, S.W., Kistler, R.W., Murray, R., Kays, M.A., 1996. Sourse and tectonic implications of tonalite-trondhjemite magmatism in the Klamath Mountains. Contrib. Miner. Petrol. 123, 40–60.

Belichenko, V.G., Butov, Yu.P., Boos, R.G., Vratkosksaya, S.V., Dobretsov, N.L., Dolmatov, V.A., Zhmodik, S.M., Konnikov, E.G., Kuz'min, M.I., Medvedev, V.N., Melanina, N.M., Mironov, A.G., Nemirovskaya, N.A., Ogurtsov, A.M., Postnikov, A.A., Roschektaev, P.A., Sklyarov, E.V., Filimonov, A.V., Shafeev, A.A., 1988. In: Dobretsov, N.L., Ignatovich, V.I. (Eds.), Geology and metamorphism of the East Sayany. Nauka, Novosibirsk(in Russian).

Bloomer, S.H., Taylor, B., MacLeod, C.J., Ctern, R.J., Frier, P., Hawkins, J.W., Johnson, L., 1995. Early arc volcanism and the ophiolite problem: a perspective from drilling in the Western Pacific. In: Taylor, B., Natland, J. (Eds.), Active Margins and Marginal basins of the Western Pacific. In: AGU: Geophysical Monograph, vol. 88, pp. 1–30.

Buyakaite, M.I., Kuzmichev, A.B., Sokolov, D.D., 1989. 718 Ma–Rb–Sr isochron of the Sarkhoy suite (Eastern Sayany). Trans. (Doklady) Russ. Acad. Sci. 309, 150–154.

Cloos, M., 1993. Lithospheric buoyancy and collision orogenesis: subduction of oceanic plateaus, continental margins, island arcs, spreading ridges and seamounts. Geol. Soc. Am. Bull. 105, 715–737.

Coleman, R.G., 1977. Ophiolites. Ancient Oceanic Lithosphere? Springer, Berlin.

Defont, M.J., Drummond, M.S., 1990. Derivation of some modern arc magmas by melting of young subducted lithosphere. Nature 347, 662–665.

Dobretsov, N.L., Konnikov, E.G., Medvedev, V.N., Sklyarov, E.V., 1985. The East Sayany ophiolites and olistostromes. In: Dobretsov, N.L. (Ed.), Riphean–Lower Paleozoic Ophiolites of the North Eurasia. Nauka, Novosibirsk, pp. 34–58 (in Russian).

Dobretsov, N.L., Konnikov, E.G., Sklyarov, E.V., Medvedev, V.N., 1986. Marianite-bonninite series in the East Sayany ophiolites. Geol. Geophys. 12, 29–35 (in Russian).

Dobretsov, N.L., Belichenko, V.G., Boos, R.G., Butov, Yu.P., Gordienko, I.V., Zhmodik, S.M., Ignatovich, V.I., Konstantinova, K.K., Kotkin, V.V., Kulikov, Yu.I., Lyuchkin, V.A., Mityukhin, E.N., Nemchinov, V.G., Osokin, P.V., Postnikov, A.A., Rasskazov, S.V., Roschektaev, P.A., Sizykh, N.V., Sklyarov, E.V., Skopintsev, V.G., Shulyak, G.B., 1989. In: Dobretsov, N.L., Ignatovich, V.I. (Eds.), Geology and Ore-bearing Potential of the East Sayany. Nauka, Novosibirsk.

Dobretsov, N.L., Konnikov, E.G., Dobretsov, N.N., 1992. Precambrian ophiolite belts of southern Siberia, Russia and their metallogeny. Precambrian Res. 58, 427–446.

Drummond, M.S., Defant, M.J., Kepezhinskas, P.K., 1996. Petrogenesis of slab-derived trondhjemite-tonalite-dacite/adakite magmas. Trans. R. Soc. Edinb Earth Sci. 87, 205–215.

Evans, D.A.D., Li, Z.X., Kirschvink, J.L., Wingate, M.T.D., 2000. A high-quality mid-Neoproterozoic paleomagnetic pole from South China, with implications for ice ages and the breakup configuration of Rodinia. Precambrian Res. 100, 313–334.

Holbrook, W.S., Lizarralde, D., McGeary, S., Bangs, N., Diebold, J., 1999. Structure and composition of the Aleutian island arc and implications for continental crustal growth. Geology 27, 31–34.

Ilyin, A.V., 1971. The Tuva-Mongolia Massif. Proc. Res. Inst Foreign Geol. 22, 67–71 (in Russian).

Khain, E.V., Neymark, L.A., Amelin, Yu.V., 1995. Isotopic-geochronological study of the granites and granite-gneisses of the Gargan Block the Eastern Sayan Range in the Siberia by Pb–Pb and U–Pb methods on zircons and Sm–Nd method. Trans. (Doklady) Russ. Acad. Sci. 342, 776–780 (in Russian).

Khain, E.V., Bibikova, E.B., Degterev, K.E., Gibsher, A.S., Didenko, A.N., Klochko, A.A., Rytsk, E.Yu., Salnikova, E.B., Fedotova, A.A., 1999. The Paleoasian Ocean in the Neoproterozoic and Early Paleozoic: new isotopic-geochronological data. Geologic Evolution of Proterozoic Pericratonic and Paleoceanic Structures in North Eurasia, Abstract volume. SPb-Moscow, pp. 175–181.

Kheraskova, T.N., Samygin, S.G., Ruzhentsev, S.V., Mossakovsky, A.A., 1995. Late Riphean marginal- continental volcanic belt of East Gondwana. Trans. (Doklady) Russian Acad. Sci. Earth Sci. Sec. 342, 661–664 (in Russian).

Khomentovsky, V.V., Gibsher, A.S., 1996. The Neoproterozoic-lower Cambrian in the northern Gobi-Altay, western Mongolia: regional setting, lithostratigraphy and biostratigraphy. Geol. Mag. 133, 371–390.

Kincaid, C., Sacks, I.S., 1997. Thermal and dynamic evolution of the upper mantle in subduction zones. J. Geophys. Res. 6 (102), 12295–12315.

Konstantinovskaya, E.A., 1999. Geodynamics of island arc–continent collision in the Western Pacific Margin. Geotectonics 33, 353–370.

Kozakov, I.K., Kotov, A.B., Kovach, V.P., Salnikova, E.B., 1997. Crustal growth in the geologic evolution of the Baidarik Block, Central Mongolia: evidence from Sm–Nd isotopic systematics. Petrology 5, 201–207.

Krogh, T., 1973. A low contamination method for hydrothermal decomposition of zircon and extraction of U and Pb for isotopic age determination. Geochim. Cosmochim. Acta 37, 485–494.

Kuzmichev, A.B., 1990. Sarkhoy supersuite in stratotype (stratigraphy, structural setting, age). Late Precambrian and Early Palaeozoic of Siberia. Problems of regional stratigraphy. Novosibirsk, 104–123.

Kuzmichev, A.B., 2000. Palaeozoic granite magmatism in the Tuva-Mongolia Massif Baikalides: tectonic aspects. Geotectonics 34 (in press).

Kuzmichev, A.B., 2001. Early Baikalian tectonic events in the Tuva-Mongolia Massif. Geotectonics 35 (in press).

Kuzmichev, A.B., Buyakaite, M.I., 1994. Rb–Sr data on the volcanics age of Shishkhid ophiolite volcanic belt (West Khubsugul region). Trans. (Doklady) Russ Acad. Sci. Earth Sci. Sect. 335, 340–344 (in Russian).

Kuzmichev, A.B., Zhuravlev, D.Z., 1999. Pre-Vendian age of the Oka Group, Eastern Sayany: Evidence from Sm–Nd dating of sills. Trans. (Doklady) Russ. Acad. Sci. Earth Sci Sect. 365, 173–178.

Li, Z.-X., Zhang, L., Powell, C.McA., 1995. South China in Rodinia: part of the missing link between Australia-East Antarctica and Laurentia? Geology 23, 407–410.

Li, X.-H., Zhao, J.-X., McCulloch, T., Zhou, G.-Q., Xing, F.-M., 1997. Geochemical and Sm–Nd isotopic study of Neoproterozoic ophiolites from southeastern China: petrogenesis and tectonic implications. Precambrian Res. 81, 129–144.

Li, X.-H., 1999. U-Pb zircon ages of granites from the southern margin of the Yangtze Block: timing of Neoproterozoic Jinning Orogeny in SE China and implications for Rodinia Assembly. Precambrian Res. 97, 43–57.

Litvintsev, K.A., Kalmychkova, T.N., 1990. First experience of Rb–Sr dating of the Munku-Sardyk granitoids (East Sayany). Trans. (Doklady) Russ. Acad. Sci. Earth Sci. Sect. 312, 205–208 (in Russian).

Lodochnikov, V.N., 1936. The Ilchir Serpentines and Serpentinites and Related Problems (in Russian).

Lodochnikov, V.N., 1941. Petrology of the Ilchir-Mondy Region. Irkutsk (in Russian).

Ludwig, K.B., 1991. ISOPLOT Program. USA Geological Survey Open-File Report 91.

Malavieille, J., Chemenda, A., 1997. Impact of initial geodynamic setting on structure, ophiolite implacement and tectonic evolution of collisional belts. Ofioliti 22, 3–13.
McKenzie, D.P., 1969. Speculations on the consequences and causes of plate motions. Geophys. J. R. Astronom. Soc. 18, 1–32.
Mitrofanov, F.P., Nikitina, L.P., Buzikov, I.P., Dav, V.N., Avdontsev, N.A., Khil'tova, V.Ya., Krylov, I.N., 1964. Precambrian of the East Sayany. Laboratory of the Precambrian geology, Proceedings, vol. 18. Nauka, Leningrad (in Russian).
Mitrofanov, G.L., 1989. Tectonics and geodynamics of the East Siberia. In: Zonenshain, L.P., Pristavakina, E.I. (Eds.), Geological history of the USSR. Nauka, Moskow.
Morris, P.A., 1995. Slab melting as an explanation of Quaternary volcanism and aseismicity in southwest Japan. Geology 23, 395–398.
Nikitina, L.P., 1963. The Archean of the Gargan Lump. In: Stratigraphy of the USSR. USSR Acad. Sci. II, 116–122 (in Russian).
Osokin, P.V., 1999. Central-Asian phosphorite-bearing province. PhD Thesis, Ulan-Ude.
Peacock, S.M., Rushmer, T., Thompson, A.B., 1994. Partial melting of subducting oceanic crust. Earth Planet. Sci. Lett. 121, 227–244.
Pelechaty, S.M., 1996. Stratigraphic evidence for the Siberia-Laurentia connection and Early Cambrian rifting. Geology 24, 719–722.
Rytsk, E.Yu., Amelin, Yu.V., Krymsky, P.Sh., Shalaev, V.S., Rizvanova, N.G., 1999. On the age of Nyurundukan Formation (Kicherian zone of the Baikal-Muja foldbelt): new U–Pb and Sm–Nd isotopic data. Geologic evolution of Proterozoic Pericratonic and Paleoceanic Structures in North Eurasia. Materials of the Conference. SPb-Moskow, pp. 130–132 (in Russian).
Sajona, F.G., Mauri, R.C., Bellon, H., Cotten, J., Defant, M.J., Pubellier, M., 1993. Initiation of subduction and the generation of slab melts in western and eastern Mindanao, Phillipines. Geology 21, 1007–1010.
Schmidt, M.W., Stefano, P., 2000. Experimentally based water budgets for dehydrating slabs and consequences for arc magma generation. Earth Planet. Sci. Lett. 163, 361–379.
Sklyarov, E.V., 1990. Ophiolites and blueschists of southeast Sayan. First international symposium on geodynamic evolution and main sutures of Central Asia. IGCP Project 283, Guidebook for excursion, Ulan-Ude.
Stacey, J.S., Kramers, J.D., 1975. Approximation of terrestrial lead isotope evolution by a two-stage model. Earth Planet. Sci. Lett. 26, 207–221.
Taylor, R.N., Marlow, M.S., Jonson, L.E., Taylor, B., Sherman, H.B., Mitchell, J.G., 1995. Intrusive volcanic rocks in Western Pacific forearcs. In: Taylor, B., Natland, J. (Eds.), Active Margins and Marginal Basins of the Western Pacific. In: AGU: Geophysics Monograph, vol. 88, pp. 31–43.
Tsygankov, A.A., 1998. Metavolcanites of the Baikal-Muja ophiolite belt: geochemistry and correlation. Geol. Geophys. 39, 19–30 (in Russian).
Underwood, M.B., Ballance, P.F., Clift, P.D., Hiscott, R.N., Marsaglia, K.M., Pickering, K.T., Reid, R.P., 1995. Sedimentation in the forearc basins, trenches and collision zones of the Western Pacific: a summary of results from the Ocean Drilling Program. In: Taylor, B., Natland, J. (Eds.), Active Margins and Marginal Basins of the Western Pacific. In: AGU: Geophysics Monograph, vol. 88, pp. 315–343.
Unrug, R., 1997. Rodinia to Gondwana: the geodynamic map of Gondwana supercontinent assembly. GSA Today 7, 1–6.
B. Wolge and 16 others (Gore-Gambella Geotraverse Team), 1996. Tonalite-trondhjemite-granite genesis by partial melting of newly underplated basaltic crust: an example from the Neoproterozoic Birbir magmatic arc, western Ethiopia. Precambrian Res. 76, 3–14.
Yogodzinski, G.M., Kay, R.W., Volynets, O.N., Koloskov, A.V., Kay, S.M., 1995. Magnesian andesite in the western Aleutian Komandorsky region: Implications for slab melting and processes in the mantle wedge. Geol Soc. Am. Bull. 107, 505–519.
York, D., 1966. Least-squares fitting of a straight line. Can. J. Phys. 44, 1079–1086.