BEITRÄGE ZUR REGIONALEN GEOLOGIE DER ERDE

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Band 32

The Central Asian Orogenic Belt

Geology, Evolution, Tectonics, and Models

edited by Alfred Kröner

With 109 figures and 2 tables

2015

Borntraeger Science Publishers · Stuttgart

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Front cover: Kröner, A., The Central Asian Orogenic Belt – Present knowledge and comparison with the SW Pacific; p. 1–5; Fig. 1. Geological map of Central Asia showing location of contributions in this volume. Based on Geological Map of Central Asia and Adjacent Areas 1:2500000 (2008), Geological Publishing House, Beijing, China.

ISBN 978-3-443-11033-8

Information on this title: www.borntraeger-cramer.com/9783443110338

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Publisher: Gebr. Borntraeger Verlagsbuchhandlung Johannesstr. 3A, 70176 Stuttgart, Germany mail@borntraeger-cramer.de www.borntraeger-cramer.de

➢ Printed on permanent paper conforming to ISO 9706-1994

Typesetting: DTP + TEXT Eva Burri, Stuttgart Printed in Germany by 00000000 GmbH, Tübingen

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Neoproterozoic accretion of the Tuva-Mongolian massif, one of the Precambrian terranes in the Central Asian Orogenic Belt

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A.B. Kuzmichev

Abstract. The Tuva-Mongolian massif (TMM) is one of the numerous Precambrian terrranes scattered over the Central Asian Orogenic Belt (CAOB). It is likely that most of these terranes are the dispersed fragments of a large continent. This paper provides a schematic overview of the TMM Neoproterozoic tectonics. These data can be compared with those for similar terranes to restore the early history of CAOB evolution. The TMM shows an asymmetric structure with its eastern part representing a Neoarchaean block. Early and late Neoproterozoic arcs were successively accreted to its western margin in the course of two collision events (~800 and ~600 Ma). The first event occurred when the TMM was still attached to the parental continent, whereas during the second stage the TMM was already a microcontinent. The early Neoproterozic Dunzhugur island arc was created not later than 1034 ± 9 Ma existed up to 810 Ma when it collided with the passive margin of the ancient continent. At the end of the early Neoproterozoic (about 850 Ma) the Dunzhugur arc split up, giving birth to the Shshkhid island arc. The mid-Neoproterozoic orogeny involved obduction of the Dunzhugur forearc ophiolite onto the passive margin that produced gravitational nappes in the foreland basin. Subsequently the subduction polarity changed and the subduction channel became inclined beneath the continental margin. This caused Andean-type magmatism on the continental margin to occur that still belonged to the continent. Three tectonic units can be reconstructed in the early late Neoproterozoic. These are the Sarkhoi volcanic belt resting on the above continental margin. The Oka accretionary prism grew outward from it, and the Shishkhid island arc drifted in the Palaeo-Asian ocean. At ~750 Ma an oceanic spreading ridge arrived at the subduction zone and this produced voluminous basaltic magma intrusions into the prism sediments. This event may have triggered back-arc rifting in the rear of the Sarkhoi continental arc. It is likely that, at about the same time, the Shishkhid arc collided with the Sarkhoi margin and generated a triple junction. At least from the middle of late Neoproterosoic the Tuva-Mongolian massif existed as a micrcontinent, separated from its ancestor by a newly formed oceanic basin. By ~600 Ma the Shishkhid arc was completely accreted onto the Sarkhoi margin, and the TMM reached its full size. In the latest Neoproterozoic and Cambrian the Tuva-Mongolian microcontinent drifted through the Palaeo-Asian ocean as a carbonate platform. Collision of the island arc and continental terranes began in late Cambrian and culminated in the Early Ordovician when the TMM and surrounding terranes were squeezed between Siberia and one of the Gondwana-derived continents.

Introduction

General information

The Central Asian Orogenic Belt (CAOB) includes numerous Precambrian terranes dispersed throughout anastomosing intervening Palaeozoic belts (e.g., Şengör and Natal'in 1996, Yakubchuk 2004, Windley et al. 2007, Levashova et al. 2011b). It is likely that most of these terranes are fragments of a large continent and its restoration is uncertain. The Precambrian terranes provide a record of the early geological history of the CAOB that is still poorly understood. This paper concerns the geologic structure and evolution of one such terrane,

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namely the Tuva-Mongolian massif (TMM). It would appear that some CAOB Precambrian terranes show structural patterns similar to that of the TMM. Therefore, this paper provides data for geological correlations with other Precambrian terranes in the CAOB.



Fig. 1. Location of the Tuva-Mongolian and Dzabkhan massifs in the CAOB. The striped eastern portion of the massifs is the ancient core with pre-Neoproterozoic basement. Location of Fig. 2 is outlined.

The Tuva-Mongolian Massif and the related Dzabkhan (or Zavkhan) terrane are located in the Sayan-Baikalian region that fringe the southern part of the Siberian craton (Fig. 1). This area preserves an almost complete record of Neoproterozoic to early Palaeozoic tectonic events and can be regarded as one of the key regions to reconstruct the early phases of CAOB evolution. The paper deals mostly with the northern TMM (Figs. 2, 3), whose Neoproterozoic evolution is reasonably well understood. Some features may be correlated with that of the Dzabkhan terrane where new geochronological data were recently published. A substantial part of the TMM is covered by Palaeozoic rocks and intruded with Palaeozoic granitoids. Both have been removed from the majority of figures in this

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paper to show the Precambrian structures. In spite of the review style of the paper it does not include a complete bibliography and review of previous work. It should be noted that all Neoproterozoic units discussed below were previously thought to be Palaeozoic in age. The complete references lists on each subject discussed in the paper can be found in the cited works.

Outline and structure of the TMM

68

The Tuva-Mongolian Massif is a traditional name, proposed by Ilyin (1971), though most of the "Massif" is built up of sedimentary and volcanic rocks, which underwent low-grade metamorphism. The TMM is predominantly composed of Neoproterozoic rocks that are unconformably covered by Vendian-Cambrian shallow-water carbonate deposits. These were folded and faulted during an Early Ordovician orogeny and no longer look like a platform cover. Yet this carbonate unit is a distinguishing feature of the massif: in the surrounding regions rocks of the same age are represented by an island-arc volcanic suite, ophiolites and shale-sandstone clastic sediments. The boundaries of the TMM terrane differ significantly, depending on the literature cited (e.g., Windley et al. 2007, Badarch



Fig. 2. Early and late Neoproterozoic belts in the Tuva-Mongolian massif. Location of Fig. 3 is outlined.

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et al. 2002, Kovalenko et al. 2004, Kozakov et al. 2011). In this paper all units that were accreted to the ancient core of the TMM by the end of the Precambrian are included in the terrane.

The TMM exhibits an asymmetric structure. Its eastern part comprises a Neoarchaean block (Figs. 1, 3), and early to late Neoproterozoic arcs were successively accreted to its western part in the course of two collision events (~800 and ~600 Ma). Therefore, the Neoproterozoic history of the massif includes two phases: early and late Neoproterozoic. The general trend of this evolution is the successive growth of the continent through the addition of new continental crust due to island arc accretion. During both phases the continental margin faced towards the Palaeo-Asian Ocean and an island arc at some distance. The long duration of each phase (about 200 Ma) leads to a question concerning the life span of oceanic arcs which is discussed below.

The Tuva-Mongolian and Dzabkhan combined terrane shows an arcuate eastern (inner) margin (Fig. 1). It is fringed on the inner side by Vendian-early Palaeozoic belts, i.e. the Dzhida and Ilchir belts on the northern limb (Fig. 2) and Bayankhongor belt in the south (Fig. 1). They include island-arc volcanic rocks, ophiolites and related rocks of latest Neoproterozoic to Cambrian age (e.g., Belichenko et al. 2003, Gordienko 2006, Jian et al. 2010a). The outer, western margin of the massif is also fringed by ocean-related Vendian-Cambrian belts namely the Agardakh in the west (Fig. 2) and the Lake Zone of Mongolia in the southwest (Fig 1) (e.g., Pfänder and Kröner 2004, Yarmolyuk et al. 2011).

The Precambrian units constituting the TMM are discussed below from old to young, namely the pre-Neoproterozoic basement, followed by the early and then late Neoproterozoic belts. Two palaeogeographic features can be restored in the early Neoproterozoic (1000–800 Ma), namely a passive margin of the ancient continent and an oceanic arc (Dunzhugur arc). The late Neoproterozoic segment of the TMM embraces three distinct belts (Fig. 3). The eastern one is the Sarkhoi volcanic belt built upon the ancient cratonic



Fig. 3. The structure of the Northern TMM, the area studied by author. Locations of Figs. 4-6 are outlined.

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terrain. It represents a continental magmatic arc (Kuzmichev and Larionov 2011) and is adjoined in the west to the Oka belt, composed of shales, sandstones, and greenschists corresponding to a late Neoproterozoic accretionary prism (Kuzmichev et al. 2007). The western zone is a discontinuous ophiolite belt that includes the Shishkhid ophiolite as the largest body. Ophiolites and related sedimentary and metasedimentary rocks constitute most of the outer zone of the TMM and in the Neoproterozoic belonged to the Shishkhid island arc system (Kuzmichev et al 2005).

Fragments of an ancient continent in the core of the massif

The ancient continental block in the core of northern TMM is named the Gargan salient (Gargan Glyba, Figs. 3, 4) and is composed of a Neoarchaean crystalline basement unconformably overlain by an early Neoproterozoic carbonate platform cover. The basement is composed of amphibolite, quartzo-felspathic gneiss, and garnet-pyroxene gneiss with plagioclase oikocrysts. These rocks exhibit a retrograde amphibolite-facies mineral assemblage that was superimposed on granulite-facies rocks. Zircons extracted from a biotite-amphibole metatonalitic gneiss were dated with TIMS and yielded a magmatic age of 2727 ± 6 Ma, whereas recrystallized metamorphic zircon was dated at about 2.6 Ga (Anisimova et al. 2009b).



Fig. 4. The Gargan Glyba Archaean salient covered by a carbonate platform and fringed by ophiolite. All three units are intruded by ~800 Ma tonalite. Based on USSR State Geologic Mapping and the authors' surveys. Locations of Figs. 5 and 6 are indicated.

The early Precambrian crystalline rocks are more widespread in the Dzabkhan and Tarbagatai terranes (Fig. 1). In the eastern Dzabkhan massif (Baidarik uplift) metatonalite gneiss and two-pyroxene gneiss revealed three zircon populations, namely ~ 2.8 Ga (protolith age), 2.65–2.5 Ga (granulite-facies metamorphism and ~ 1.8 Ma (amphibolite-facies metamorphism) (Kozakov et al. 2007). These rocks were cut by hyperstene diorite at

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 2364 ± 6 Ma and undeformed isotropic granite at 2308 ± 4 Ma (Kozakov et al. 2007 and references therein). Similar and younger ages were obtained using $^{206}Pb/^{207}Pb$ evaporation and SHRIMP techniques (Demoux et al. 2009a, b). A granite-gneiss in the Tarbagatai terrane was dated as 2219 ± 25 Ma and an anorthosite as 1784 ± 10 Ma (zircon, TIMS, Anisimova et al. 2009a, Kozakov et al. 2011).

The asymmetry of the combined TMM and Dzabkhan terrane implies that it broke off along its eastern side from an ancestral ancient continent. The identification of this continent is a challenging problem: many similar Precambrian fragments are dispersed throughout the CAOB, and at least three competing hypotheses are discussed for their origin. The first postulates that all terranes are fragments of East Gondwana and drifted away from it during the late Neoproterozoic (e.g., Kheraskova et al. 2003, Buslov et al. 2001). This hypothesis is supported by some palaeomagnetic data (Levashova et al. 2011a). The second point of view restores the Precambrian terranes to the southern (recent) margin of the Siberian craton (Kuzmichev 2004, Windley et al. 2007). This view is also supported by palaeomagnetic evidence (Kravchinsky et al. 2010). The third hypothesis reconstructs the Neoproterozoic TMM as a sliver of continental crust attached to the Siberian craton (Sengor and Natal'in 1996, Yakubchuk 2004). The present author believes that the second model is most likely. The known ages for the crystalline basement of the TMM, Dzabkhan and Tarbagatai terranes cannot help in seeking the parental continent. Archaean and Proterozoic rocks with similar ages are known from the South China craton, the Aldan-Stanovic region in the southern part of the Siberian craton and in other places. Furthermore, many of the published zircon ages were obtained using the multigrain TIMS technique, which is hardly applicable for polymetamorphic rocks.

Early Neoproterozoic continental margin deposits

The crystalline basement of Gargan Glyba in the northern TMM is unconformably covered by a platform sedimentary succession (Fig. 4). This cover consists of the Irkut Formation marbles at the bottom and the Ilchir Formation shales above. The Irkut Formation consists of a thickly-bedded dolomitic marble up to 600 m thick, in places stromatolitic, and contains recrystallized chert beds. A basal conglomerate contains pebbles of quartz and metamorphic rocks. The lowermost layers also contain muscovite schist that probably represents a metamorphosed high-alumina weathering product. The Ilchir Formation is composed of dark grey shale along with sporadic interbeds of fine-grained sandstone and limestone. The uppermost Ilchir Formation is exposed between ophiolite klippen in the inner zone of the Gargan block and is noted for the presence of olistostromes with ophiolite and Irkut dolomite exotic clasts up to a few hundred metres long.

The above sedimentary succession is interpreted as a passive-margin shelf deposit. These rocks were intruded at 790 Ma by the Sumsunur tonalite (see below) that defines the upper age limit of sedimentation. It is likely that the sedimentary cover is not older than early Neoproterozoic, because its uppermost horizon is an olistostrome that was deposited synchronously with ophiolite obduction in the course of a mid-Neoproterozoic orogeny.

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Fig. 5. Ophiolite nappes in the Dunzhugur area. Cenozoic rocks and diabase sills omitted. See Fig. 4 for location.

Early Neoproterozoic Dunzhugur oceanic arc

The Dunzhugur island arc is represented by a supra-subduction zone (SSZ) ophiolite that forms two belts fringing the Gargan Glyba (Fig. 4). These belts are reminders of an eroded allochthon, whose klippen also occur on top of Glyba. The most representative ophiolite section is exposed in the Dunzhugur area (west of the Gargan Glyba) (Fig. 5) where a

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complete typical succession from harzburgite to oceanic pillow basalt was mapped and sampled in detail (Sklyarov 1990, Kuzmichev et al. 2001, Khain et al. 2002, Kuzmichev 2004). This area composed of steeply inclined stacked ophiolite slivers, which are shuffled without regularity (Fig. 5). On the opposite, eastern side of the Gargan block the ophiolite allochthon forms synforms separated by crest-like anticlines into which the Ilchir shales have been squeezed (Fig. 6). The allochthon shows erosional windows and klippen and is much less disturbed than in the Dunzhugur region. It is chiefly composed of serpentinized harzburgite. Layered ultramafic rocks, massive gabbro, sheeted dykes and basalts are confined to the allochthon foot and display a reverse sequence (Kuzmichev 2001). The presence of low-Ti basalt and boninite among the ophiolitic sheeted dykes and volcanic rocks was revealed by Sklyarov et al. (1990). These features indicate a supra-subduction setting for the ophiolite suite.



Fig. 6. Sketch map of ophiolite nappes and underlying Ilchir olistostrome in the Onot-Gorlyk-Gol region.

The sedimentary rocks associated with the ophiolite form the upper parts of thrust sheets in the Duzhugur area. The lower member of this sedimentary succession conformably rests upon the pillow basalt. It is represented by turbidites with lenses of clastic rocks composed of basaltic debris and occasional blocks of gabbro and ultramafic rocks. The upper part of the sedimentary section is composed of sandstone and conglomerate with volcanic clasts including andesite, dacite, rhyolite and their plutonic counterparts. They correspond to the later phases of Dunzhugur arc magmatism.

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The Dunzhugur ophiolite succession and especially the sedimentary rocks at its top are densely intruded by diabase sills and dykes with MORB affinity. This igneous suite evidentially indicates extension in the island arc that can be attributed to intra-arc rifting that is discussed in the last section.

The age of the Dunzhugur arc is constrained by dating ophiolitic plagiogranite that forms veins in the layered gabbro on the eastern bank of the Oka River (Fig. 5). The zircon U–Pb conventional and ²⁰⁷Pb/²⁰⁶Pb evaporation age is 1020 Ma (Khain et al. 2002). Keeping in mind the boninitic composition of the ophiolite volcanic rocks, this age probably correspond to the incipient phase of oceanic arc evolution (Shervais 2001, Dilek and Furnes 2011). Subsequently such ophiolite successions form a basement to the forearc zone adjoining the trench. Such a setting is applicable to the Dunzhugur ophiolite and overlying sediments. The later phases of island-arc magmatism are represented by felsic volcanic rocks redeposited in forearc sediments of the Dunzhugur area. As shown in the next section, collision of the Dunzhugur arc and Gargan continental margin occurred at about 810 Ma. Such a long-lived volcanic arc (ca. 200 Ma) inevitably influenced the crustal composition and thickness. Some portions of the inferred mature island arc are masked by Vendian-Cambrian sediments or are overlapped by the Oka thrust (see below). However, it must be admitted that such large newly-formed crustal terrain as can be expected for a 200 Ma old island arc is not documented from field relationships in the TMM.

In order to constrain the time-span of island arc volcanism we dated zircons separated from volcanic sandstone that belong to the upper part of the sedimentary succession associated with the ophiolite. All crystals are near-euhedral and show morphologies, oscillatory CL images, Th/U ratios and U contents that are typical for zircon from felsic igneous rocks (Kuzmichev and Larionov 2013). Twelve zircons were analyzed using a SHRIMP II instrument. The age interval calculated from ${}^{206}Pb/{}^{238}U$ ratios is 910 ± 27 to 1048 ± 12 Ma (1 σ). The five most ancient grains make up a concordant cluster with a mean age of 1034 ± 9 Ma age (2 σ). One more grain that shows a similar 207 Pb/ 206 Pb age $(1037 \pm 17 \text{ Ma}, 1 \sigma)$ can also be included in the cluster. Therefore, mature island-arc volcanism appears to be slightly older than the inferred incipient phase of arc formation because, at 1034 Ma, the Dunzhugur arc already produced felsic volcanic rocks. Analytical points for other grains are dispersed along the concordia curve toward younger ages. The youngest grain yielded a discordant analysis with a 207 Pb/ 206 Pb age of 910 ± 27 Ma. These results show that the arc indeed existed for a long time. A metamorphosed counterpart of the Dunzhugur silicic volcanic rocks was found in the northern Dzabkhan massif where rhyolite and dacite transformed into gneiss yielded a U-Pb zircon age 856 \pm 2 Ma (Kozakov et al. 2012).

Mid-Neoproterozoic orogeny: collision of the island arc and continent

A forearc setting for the Dunzhugur ophiolite indicates that the arc was facing the continent and the oceanic plate separating them was subducting beneath the arc. After complete subduction of the oceanic lithosphere of the marginal basin, the Gargan continental margin was also partly subducted beneath the arc. Passive margins, thinned by rifting, can subduct as deep as 100 km (e.g., Cloos 1993). Simultaneously, the island arc lithosphere overrode the continental margin. A similar situation was considered as a most plausible reason for ophiolite obduction (Coleman 1977).

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The obduction model for the Dunzhugur ophiolite can be deduced from different patterns of nappes in the marginal and inner parts of the Gargan block. On the northern and western margins of Gargan Glyba the ophiolite overlaps different horizons of Glyba's cover, and in places its crystalline basement (Figs. 4, 5), and the allochthon base is strongly disrupted. In the inner, eastern part the allochthon rests almost conformably on the Ilchir Formation upper horizons, containing a carbonate olistostrome (Fig. 6). The obduction in places shows no sign of strain affecting the underlying rocks. These relationships can be interpreted as follows: the allochthon has overthrust the continental margin, tearing off its platform cover. The cover's fragments were unloaded into the foreland basin, yielding olistostromes on which the ophiolites glided during sedimentation. First, their upper portion slid down to form the base of the packet of sheets, thus producing a reverse sequence where ophiolite appears in inverse stratigraphic order. This model also accounts for the difference in the structure of the allochthon in the inner and marginal areas. In the inner



Fig. 7. The ophiolite obduction model. Note difference in the allochthon setting and structure east and west of Gargan Glyba. Most symbols correspond to Figs. 5 and 6. a) Initial stage of obduction. The allochthon was overthrust on the continental margin, tearing off its platform cover, whose fragments were unloading into the foreland basin, yielding olistostrome. b) Uplift of the Gargan margin crowned with ophiolite nappe whose sheets began sliding into the basin above the olistostrome. First, the ophiolite upper horizons slid down to form the base of the packet of sheets, thus producing a reverse tectonic sequence. c) The bulk of ophiolite nappe is sliding down to the east, while in the west ophiolite thrust sheets were stacked in front of the bulging edge of the Gargan block. d) Same situation after erosion.

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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 rising edge of the continent (Fig. 7).

This model is in accordance with a typical scenario suggested by McKenzie (1969) for arc–continent collision. After jamming of the original subduction zone by the edge of a continent, a new subduction zone arises, inclined beneath the continent. The break in the lithosphere puts an end to pulling the continental margin into the subduction zone. As a result, the submerged continental margin with the overthrust forearc ophiolites was torn off its root and floated upwards. This caused the gravitational sliding of the ophiolite allochthon towards 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 beneath the continental margin, provoking generation of the first melts in a newly formed continental arc (see next section).

The age of these melts (~790 Ma), which form the tonalite plutons intruding the ophiolite nappes on Gargan Glyba, was used to constrain the age of obduction (Kuzmichev et al. 2001). It is obvious that collision occurred some time earlier in the northern TMM where syncollisional granitoids are not exposed. However, new age data have recently been obtained in the southern region. In the northern Dzabkhan massif (Songin block) a synkinematic granite has a zircon age of 810 ± 2 Ma (Kozakov et al. 2009b). In the Tarbagatai terrane (see Fig. 1) a similar granite was found with a zircon age of 809 ± 4 Ma (Kozakov et al. 2009a). These ages are close to the expected age of the collision, and it is therefore reasonable to interpret the above granitoids as syncollisional intrusions. Therefore, the age of the Dunzhugur arc and continent collision is probably around 810 Ma.

As a result of collision, the volume of the former continent increased as the mature arc lithosphere was attached to it and partly obducted. The island-arc magma production rate estimated for the Aleutians (average for 75 Ma) and Izu-Ogasawara (average for 47 Ma) is $60 \pm 10 \text{ km}^3/\text{km}$ per 1 Ma (Holbrook et al. 1999). The Dunzhugur arc could have produced an enormous volume of sialic crust during its 200 Ma life time to enlarge the continent, if the above calculation is applicable to the Neoproterozoic situation.

Late Neoproterozoic Sarkhoi continental arc

General geology

76

The Sarkhoi volcanic suite comprises a notable unit in East Sayan and Mongolia (Sarkhoi, Darkhat and Dzabkhan Groups) due to their specific appearance: they are mostly variegated rocks, which can easily be recognized and mapped. This volcanic belt, formed on the active continental margin, is traceable for almost 1500 km and is a major unit in the TMM and Dzabkhan terranes. The geodynamic setting and age of the Sarkhoi volcanic suite are uncertain, and some authors still suggest that they are Ordovician in age and form a tectonic nappe (e.g., Fedotova and Khain 2002).

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Fig. 8. Geologic map of the Sarkhoi volcanic suite in the stratotype area (see Fig. 3 for location). Legend: 1 – Quaternary; 2 – Neogene–Quaternary intraplate basalt; 3 – Ediacaran-Cambrian carbonate platform; 4 – variegated clastic rocks at the base of the carbonate sucession resting unconformably on the Sarkhoi volcanic rocks (wavy line); 5–8 Sarkhoi Group: 5, ignimbrite; 6, greenstone; 7, granophyre; 8, sandstone; 9 – Dibi turbidite; 10 – Ordovician(?) granite; 11 – dykes and other small intrusions; 12 – faults; 13 – structural lines.



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In the northern TMM the Sarkhoi Group rocks lie on a carbonate turbidite, which is somehow related to the Dunzhugur sediments (see above), thrust upon the continent. The roof is the Vendian–Cambrian Tuva-Mongolian carbonate cover. The lower horizons of this cover are composed of variegated clastic rocks that rest with an angular unconformity on volcanic rocks and grade into the overlying carbonates. The Sarkhoi Group is composed of three lithological units, named as sandstone, greenstone and ignimbrite units (Fig. 8). The sandstone unit consists of variegated polymictic or volcanic coarse-grained sandstone and conglomerate. The latter contains pebbles of the underlying sedimentary rocks, quartz, jasper, andesite and dacite. This unit was deposited in a fluvial and deltaic setting. The thickness is up to 2 km.

The greenstone unit consists of olivine basalt, andesite and more abundant dacite. Volcanic rocks occur as massive lava flows, block lava, clastic tuff and rare pillow lava. The rocks are green and greenish-grey or, occasionally, violet-grey. At the top of the section there are flows of megaporphyritic andesite with large (1 cm) plagioclase laths passing into block tuffs and rhyolite flows. The thickness of this unit in the Zabit River basin reaches 550 m. It gradually decreases to the west, and the lower part is replaced by greenish sandstone and conglomerate whereas the upper part is replaced by reddish rocks indistinguishable from those of ignimbrite unit.

The ignimbrite unit is mainly composed of welded and non-welded, usually reddish, rhyolitic ignimbrite and ash. Two other main components of the section are dark-lilac tuff and volcanic gritstone. Typical ignimbrite is a dark coloured rock in the welded lower part and lighter coloured in the upper part. Usually it bears scarce phenocrysts of feldspar and rare quartz in some varietes. Xenogenic clasts of andesite, felsic lava and sandstone are common. Some ignimbrite flows show deep narrow pockets filled with crystals and xenogenic clasts. According to Branney and Kokelaar (2002), such pockets are outgassing tubes because fine material is removed from them with a gas jet. The unit also contains andesite lapilli and pisolite tuff. The latter formed as a result of the sticking of rain drops with volcanic ash in a terrestrial setting. Cross-bedded gritstone consisting of white (plagioclase) and red (glass) grains is also a typical rock deposited by temporary water flows. Another typical facies is poorly sorted coarsely layered dark lilac mudstone representing mudflows. Locally rhyolite, dacite and rare andesite flows were observed. The roof of the ignimbrite unit is washed out. Its visible thickness in the Zabit River basin reaches 1.2 km.

Near the Zabit River source the lower part of ignimbrite unit is transformed into an unusual rock called granophyre in Fig. 8. These are massive non-layered red coloured aphyric or porphyritic (alkali feldspar phenocrysts) jasper-like rocks, sometimes with an indistinct fluidization. Some varieties show a granophyric groundmass. The red jasperoid rock, saturated with opal and iron hydroxides, is a usual product of the hydrothermal alteration of ash tuffs.

Age of the Sarkhoi volcanic rocks

Eleven zircons extracted from a welded rhyolitic ignimbrite were analysed on a SHRIMP II ion microprobe. All analyses showed a Neoproterozoic age of zircon crystallization, and eight of these form a concordant cluster with a mean age of 782 ± 7 Ma (2 σ), which is interpreted as the time of ignimbrite eruption (Kuzmichev and Larionov 2011). The other three analyses were ignored for different reasons. The presence of volcanic clasts in the sandstone unit indicates that volcanism began earlier than the dated sample.

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The related Dzabhan volcanic rocks show a great similarity with the Sarkhoi suite. They are mafic volcanic rocks in the lower part and dacites and rhyolites (mainly Naignimbrites) in the upper part. The reported age for the felsic volcanic rocks is close to the above estimate. Six of thirteen analyzed zircons (LA-ICP-MS) from the upper rhyolite yelded a concordant cluster with an age of 774 ± 4 Ma. For the second sample from the lower part of the section analytical data on five zircon crystals were reported (Levashova et al. 2010, 2011b). Three of these form a concordant cluster with an age of 803 ± 8 Ma. Therefore, the age of Sarkhoi and Dzabkhan volcanic suites is in a range of 805-770 Ma.

Origin of magmas and geodynamic setting

The greenstone unit consists of volcanic rocks that form a continuous series from basalt to rhyolite, with andesite and dacite being predominant. This series probably is the direct product of supra-subduction zone magmatism, which includes differentiation and assimilation of crustal material. These products partly inherited low Ti contents and high Th/Nb and Ce/Nb ratios from the primary magma. The generation of huge masses of rhyolitic ignimbrites of rather homogeneous composition most likely proceeded with melting of the crustal source caused by the intrusion of basaltic magma (see discussion in Kuzmichev and Larionov 2011). This source may have been tonalitic gneisses of the early Precambrian basement, which determined the Na-enrichment of the melts. A crustal source is confirmed by isotopic data. A dacite of the greenstone unit shows a whole-rock $\varepsilon_{Nd(t)}$ value of -1.4 (calculated for 800 Ma) and $t_{DM} = 1623$ Ma; an andesite of the same unit has an $\varepsilon_{Nd(t)}$ value of -5.5 (800 Ma) and $t_{DM} = 2018$ Ma (Vescheva et al. 2008). Independent isotopic analysis of three felsic volcanic samples obtained by V.P. Kovach (personal communication) confirm the strongly negative $\epsilon_{Nd(t)}$ values and Paleoproterozoic model ages for the source. The Sarkhoi volcanic suite is similar to volcanic associations of Andean-type continental arcs which are characterized by a predominance of terrestrial felsic ignimbrites associated with subordinate andesites (e.g., Mamani et al. 2010). Similar volcanic suites are widespread in the Central volcanic zone of the Andes, in Mexico, and in the western USA.

Plutonic rocks comagmatic with the Sarkhoi volcanic rocks

The eruption of large masses of silicic magma resulted from drainage of a magma chamber localized at some depth beneath the ignimbrite caldera. According to general concepts, the volume of non-erupted magma solidified as a tabular intrusion is significantly larger than the volume of erupted magma (e.g., Bachmann and Bergantz 2008b). Therefore, we can suggest the presence of comagmatic granite or diorite-granite batholiths beneath the volcanic units of the Tuva–Mongolian and Dzabhan massifs. The well-known Sumsunur granitoid suite shows a similar age (785 ± 11 Ma) and composition (Kuzmichev 2004). The suite includes several plutons that intrude the Gargan Glyba basement, its early Neoproterozoic cover and the obducted Dunzhugur ophiolite (Fig. 4). The suite is composed of tonalite with some trondhjemite, diorite and gabbro. It reveals a supra-subduction zone setting and a sodic character (Kuzmichev et al. 2001). Some geochemical and mineralogical differences between comagmatic plutonic and volcanic rocks are predictable and were reported elsewhere. According to Bachmann et al. (2007), these differences are mainly due to the fact that the volcanic rocks represent a melt squeezed from a crystal-rich porridge-

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like mass, which then crystallized as plutonic rocks. The comagmatic character of the Sumsunur and Sarkhoi rocks is also confirmed also by their similar Nd and Sr isotopic patterns (Kuzmichev and Larionov 2011).

Recent geochronological studies revealed granitic massifs coeval with the Sarkhoi (Dzabhan) volcanic suite in the Dzabkhan and Tarbagatai terranes. A Neoproterozoic gabbro-diorite-granodiorite-granite pluton in the northern Dzabkhan massif (the Songa uplift) yielded a zircon age of 790 ± 3 Ma (Yarmolyuk et al. 2009) that is identical to that of the Sumsunur suite. Subalkalic porphyritic granitoids were also discovered within the Tarbagatai uplift (Fig. 1), and a U-Pb zircon age of 774 ± 3 Ma was determined (Kirnozova et al. 2009). The number and volume of such granitoids is probably underestimated. Further studies may reveal other granite batholiths that are comagmatic with volcanic rocks in Precambrian massifs of southern Siberia and Mongolia. In contrast to the Sarkhoi ignimbrites with their characteristic appearance, the Neoproterozoic granitoids look like common Palaeozoic rocks.

Oka accretionary prism

The Oka belt is composed of clastic sedimentary rocks and greenschists and extends for about 600 km (Fig. 2). It is bordered by thrust zone dipping east or north (Fig. 3). These are distinct mappable faults in the northern segment and can be traced to the south. The age and tectonic setting of the Oka belt were controversial for a long time. Kuzmichev et al. (2007) suggested that the belt formed in the late Neoproterozoic as an accretionary prism. The imbricated structure, occurrence of oceanic rocks and blueschists and some other features of the Oka belt are typical for modern accretionary prisms.

Structure

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The studied northern portion of the Oka belt contains both strongly deformed mélange zones and more coherent blocks. The least altered rocks are exposed in the Yakhoshop and Dayalyk river basins (Fig. 9). In this area, the Oka belt consists of numerous tectonic slivers separated by sublatitudinal high-angle faults dipping north. Some features indicate that this fault system was originally inclined towards the south and had a "seaward" vergence, typical of modern accretionary prisms. Today the sedimentary strata are steeply inclined to the northwest, which is nearly parallel to the faults, and shows an overturned position. It is very likely that in the Neoproterozoic these strata were in a normal position and were inclined towards the southeast as well as the thrust faults. The orientation of drag folds is inconsistent with a recent southward vergence and suggests original northward thrusting.

Rotation of the thrust planes from flat to steep most likely occurred during early Palaeozoic deformation and did not require significant rotation since the recent thrust planes in the Oka prism dip at angles of 70–85°. Moreover, on the banks of the Hugein River in northern Mongolia (Fig. 2), the lower portion of the Oka belt preserved its original seaward vergence, which is discordant to the sole thrust of the belt.

In the Tustuk River region the majority of sedimentary rocks show disturbed layering that occurs either at microscopic, hand specimen, or outcrop scale. This produced a specific rock appearance, namely the fragments rarely look like slates or slabs, but usually occur in the form of irregular blocks due to an irregular inner structure. This was caused

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Neoproterozoic accretion of the Tuva-Mongolian massif



Fig. 9. Mapped region of the northern Oka prism in the northern Tustuk area (see Fig. 3). The map does not display the imbricated structure of the Oka prism. Due to poor exposure only the main boundaries between different lithologies can be mapped.

by layer-parallel extension and shearing, following layer-parallel contraction, which can be attributed to underthrusting. Layer-parallel extension is exemplified by stretched layers. Boudinage of sandstone layers is common in the Oka clastic rocks. Flattening is most evident in conglomerates where some pebbles were stretched to an enormous degree. Cleavage is oriented at a high angle to bedding and indicates compression directed almost opposite to primary flattening. In Recent and Mesozoic accretionary complexes such wedge shortening, resulting in seaward thrusting and imbrication accompanied by cleavage, was attributed to tectonic underplating (e.g., Hashimoto and Kimura 1999).

Sample and Moore (1987) noted that the above structural features cannot be attributed solely to accretionary wedges but may be expected in any type of thrust belt. However, on

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both sides of the Oka belt there are terrains that include similar lithologies but demonstrate different structures. The Dunzhugur ophiolite described before is overlain by a thick and coherent turbidite sequence that preserves all primary sedimentary textures and does not show the structural features inherent to the Oka belt rocks except for slaty cleavage in the upper portion. On the other side of the Oka belt there occurs the Shishkhid ophiolite that is overlain by the Kharaberin Formation including turbidite and mudstone deposits accumulated in the back-arc setting (see below). These sediments, though folded, were substantially less deformed than the Oka belt. Both Neoproterozoic terrains adjacent to the Oka belt do not show any imbrication. These rocks neither demonstrate early extension nor late stage shortening. All three terranes experienced the same events of deformation at the end of the late Neoproterozoic and at the beginning of the Ordovician (Kuzmichev 2004). Moreover, the Dunzhugur unit underwent one more phase of thrusting in the mid-Neoproterozoic. This implies that in spite of superposed collision events in the region, the Oka belt preserved its specific structural style inherited from the accretionary phase of prism formation.

Trench deposits

The Oka belt is mainly composed of shale-sandstone intercalations. These rocks have usually lost their primary depositional features, though coherent flysch packages occur to the west and southwest of the area shown in Fig. 9. The prevalent sediments in the northern Oka belt are laminated or massive mudstone alternating with siltstone and fine-grained sandstone. Sandstone layers locally show basal erosion and normal grading, indicating a turbidite origin. Soft sediment deformation is common. These packages occasionally include massive sandstone layers up to several metres thick that may represent amalgamated strata. The northern portion of the Oka belt also contains volcanoclastic rocks that were originally mapped by the author as part of the Oka prism but are now interpreted as tectonic slivers of the Shishkhid arc (see below). There is also an olistostrome horizon that probably underlies the isolated stratigraphic member (Kuzmichev et al. 2007).

Oceanic rocks incorporated into the prism

Ocean floor basalt consists of lenses and tectonic slivers that can be found in any part of the Oka belt, though these rocks are more abundant in the Oka inner zone adjacent to the continent. The northern limb of the Oka belt contains rare metabasalt layers that are confined to the soles of thrust planes and are altered into actinolite-chlorite-epidote schist. Southwards their thickness gradually increases. The best occurrence of oceanic basalts (about 1000 m), including both MORB- and OIB-like varieties, is the Hugein River basin (Fig. 2) (Sklyarov et al. 1996). The rocks are generally altered into epidote-chlorite-albite schist although pillow basalt and hyaloclastic breccias are occasionally recognizable.

The metabasalts show a homogenous chemical composition and are classified as tholeiitic basalts with moderate titanium $(1.2-1.5\% \text{ TiO}_2)$ and iron (12-14% FeO) contents, low alumina and low potassium. These features as well as REE patterns and Ti-V and Ti-Zr relations indicate N-MORB signatures. Several kilometres northwest of the Hugein River a picritic basalt is exposed, showing high concentrations of titanium (up to 4%), niobium (up to 83 ppm), and zirconium (up to 340 ppm). Such incompatible element enrichment may indicate an ocean island setting.

82

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Metagabbro, serpentinite and metapyroxenite make up lenses and blocks, enclosed witrhin a greenschist matrix. Such mélange was encountered in the northern part of the Oka belt in areas located to the south and southwest of the area shown in Fig. 9 (see Fig 3), in association with high-pressure metamorphic rocks. These exotic rocks along with basalt form a specific dismembered ophiolitic association that represents oceanic lithosphere.

Oceanic sediments are found in the northern Tustuk River basin. On the map (Fig. 9) these are shown by a general symbol for coloured rocks, which embrace variegated claystone, jasper and associated limestone. Thinly laminated coloured (mostly reddish) and sometimes black argillite may represent a pelagic ooze (Kuzmichev et al. 2007). Jasper is associated with metabasalt and forms brightly coloured layers, up to several metres thick.

Blueschist

A narrow zone, including high pressure-low temperature (HP-LT) metamorphic rocks, was traced along the inner part of the Oka belt (Sklyarov and Postnikov 1990, Sklyarov et al. 1996). It comprises greenschist-facies rocks, mainly metabasites, that contain occasional relicts of HP-LT minerals. Such rocks were found in numerous locations on the area shown in Fig. 3. Most of the zone is composed of actinolite-chlorite-epidote-albite greenschist, formed after basalt and gabbro, and some phengite-chlorite-quartz-albite schist. Winchite is a common mineral, and crossite was found as rare relicts fringed by winchite or actinolite. The Na₂O content reaches 5.6% in the core and decreases towards the rim. White micas are characterized by a high celadonite component (Si = 3.3-3.4 pfu), both in metabasic and associated metapelitic schists. The fact that crossite was only found in rocks of favourable chemical composition as well as the absence of lawsonite and glaucophane tentatively suggests the metamorphic conditions as intermediate between glaucophane-schist and greenschist-facies with the pressure ranging approximately from 5 to 7 kbar. The temperature of metamorphism was about 380-450 °C, judging from the mineral associations and an absence of stilpnomelane.

Tholeiitic intrusions and age of the Oka prism

The northern segment of the Oka belt was intruded by basaltic magma and contains numerous diabase and gabbro-diabase sills and dykes (Fig. 9). These are several metres to several tens of metres wide. In the northern Oka belt the sills are least altered and locally preserve initial intrusive contacts with the Oka shales. Southwards alteration increases and some bodies are completely altered into greenschists that can hardly be distinguished from oceanic metabasalts.

Some intrusions are differentiated into trondhjemite compositions. The siliceous endmember is strongly enriched in rare-earth elements (REE), Zr and other incompatible elements. Differentiation affected the bulk concentration of REE, but has had minor effects on the REE patterns. The latter shows depletion in light REE, typical of N-MORB magmas. A Sm-Nd mineral isochron for gabbro-diabase reveals a crystallization age of 736 ± 43 Ma and $\epsilon_{Nd(t)}$ as high as $+7.8 \pm 0.5$, indicating a strongly depleted mantle source. A felsic member of this intrusion was dated by three zircon multigrain fractions (TIMS technique) at $753 \pm$ 16 Ma (Kuzmicher et al. 2007). This is the upper age limit for the Oka Group host rocks. It is suggested that the host sediments are not much older and are also late Neoproterozoic in age. Most sills are shallow-level igneous bodies that intruded poorly lithified sediments.

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Oka belt summary and similarities with the Shimanto belt of Japan

The matrix of the Neoproterozoic Oka belt is composed of turbidite that contains bodies of exotic rocks. There are distinct structural and compositional features that identify the belt as a fossil accretionary prism. These are: (1) an imbricated structure with a presumably seaward vergence; (2) slivers of MORB- and OIB-like oceanic basalts, and (3) mélange with blueschists.

Apart from the above features, the northern Oka belt demonstrates some uncommon features including tholeiitic intrusions, felsic volcanic material and a carbonate olistostrome. All these features are also inherent to the northern Shimanto belt in Shikoku Island and Kii Peninsula, Japan (e.g., Hibbard and Karig 1990b, Taira et al. 1992). It is possible that tholeiitic sills were emplaced during oceanic ridge subduction. The occurrence of diabase dykes and sills in the Oligocene and Miocene portions of the Shimanto belt was caused by the Shikoku spreading ridge arrival into the subduction zone (e.g., Kimura et al. 2005). Such mode of ocean ridge subduction is not unique (e.g., Sisson et al. 2003). It is therefore likely that mafic igneous bodies in the Oka accretionary prism are of the same origin as in the Shimanto belt. An oceanic spreading centre arrived beneath the Oka accretionary prism in the late Neoproterozoic and produced N-MORB-like melts at 753 \pm 16 Ma ago. Such intrusions occurred only in the northern limb of the Oka belt, which may indicate a near-orthogonal position of the ocean ridge against the convergent margin.

The prism piled up in front of the Sarkhoi continental arc through the second half of the Neoproterozoic though its exact lifetime is unknown. Sediment accumulation in accretionary wedges is one of the major geodynamic mechanisms operating at convergent margins. A huge volume of sialic material has been accumulated in the Oka prism. This reveals that the mechanism was as intensive in the Neoproterozoic as it is at present. The size of the Oka belt is comparable with the adjacent arc terranes even in its present, tightly compressed structure (Fig. 2). The Oka belt is possibly an exceptional example of a well-preserved Precambrian accretionary prism.

Shishkhid island arc

General overview

84

The late Neoproterozoic Shishkhid arc, similar to the Dunzhugur arc, is mainly represented by an SSZ ophiolite. The Shishkhid ophiolite corresponds to the uppermost oceanic lithosphere, comprising depleted mantle tectonites (~6 km thick) and oceanic crust of basaltic composition with andesite at the top (~7.5 km thick). The Shishkhid area shows a meridional structural trend and eastern vergence (Fig. 10). The ophiolite was thrust eastward onto the Oka belt and is underlain by a mélange zone (up to 1 km thick) comprising serpentinite lenses intercalated with shales of the Oka Formation. In the west, the ophiolite is covered by a 3 km-thick sedimentary sequence showing progressive subsidence of the volcanic edifice after cessation of volcanism. This sequence is unconformably overlain by Ediacaran-Cambrian platform sediments (Fig. 10). Therefore, the Shishkhid island-arc ophiolite was thrust upon the Oka prism before the end of the Neoproterozoic.

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Neoproterozoic accretion of the Tuva-Mongolian massif

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Fig. 10. Geological map of the Shishkhid ophiolite and surrounding units (see Fig. 3).

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Shishkhid arc stratifiication

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The ophiolite comprises (from bottom to top): residual ultramafic rocks (\sim 6 km), layered and isotropic gabbro (\sim 4.5 km), sheeted dykes (up to 0.5 km), a bimodal assemblage of basalt and rhyolite (up to 0.7 km), as well as andesite pyroclastic rocks at the top (\sim 2 km) (Kuzmichev et al. 2005). In brief, the composition of the Shishkhid ophiolite is as follows (from bottom to top).

The ultramafic unit is dominated by harzburgite containing 10-15% enstatite and about 1% chromite. Dunite is also abundant, forming lenses and layers. Cpx-bearing harzburgite is rare. Graphitic harzburgite was also observed. The uppermost part of the ultramafic unit is a cumulate sequence (0 to 200 m thick) consisting of harzburgite, lherzolite, dunite, wehrlite and pyroxenite. Ultramafic rocks are completely or partly altered to antigorite serpentinite.

The lower gabbro is composed of cumulates and some isotropic gabbro. Its lower boundary is a distinct thrust zone corresponding to the MOHO, whereas the upper boundary is a vague transition to non-layered gabbro lacking ultramafic rocks. Most of the unit is composed of eucritic (meta)gabbro whose composition changes from anorthosite to pyroxenite end members. Ti- and Fe-rich cumulates were recorded in places. Minor ultramafic cumulates consist of meta-dunite, wehrlite, and clinopyroxenite. The unit also contains cataclasites, mylonites, and small bodies of plagiogranite.

The upper gabbro is dominated by homogeneous isotropic, eucritic gabbro. The rocks locally turn into gabbro breccia. Gradual transitions to fine-grained diabase-like varieties were also observed. There are also diabase dykes cutting the upper part of the unit.

Sheeted dykes are recognized only in the rare good outcrops where typical one-sided chilled set of dykes were observed. The unit decreases in thickness to the north, where part of it is composed of non-parallel dykes.

The lower volcanic unit comprises a bimodal basalt-rhyolite assemblage. Basalt or basaltic andesite are represented by pillowed or massive lava flows. Rhyolite and rare dacite occur as flows or breccias. Phyric varietes contain plagioclase, quartz and rare hornblende phenocrysts. In places rhyolite fragments occur within basaltic hyaloclastite, filling the space between basaltic pillows. Some rhyolite breccia also contains occasional basaltic fragments and pillows.

The upper volcanic unit is composed of andesite, basaltic andesite, and pyroclastic rocks. The latter show chaotically-oriented angular fragments 1 to 25 cm in size. They were mainly erupted in a submarine environment, though above the explosive compensation level, which is generally defined for andesitic magma at a water depth of 500 m or less (Stix 1991). The least altered phyric andesite contains crystals (up to 50% by volume) of greenish-brown hornblende and zoned andesine. Thin flows of basaltic lava were observed within the lower part of the unit.

The sedimentary section overlying the volcanic rocks (Kharaberin Formation) is divided into four units (Fig. 10). The lowermost unit is about 300 m thick and composed of polymictic sandstone, siltstone, as well as sandy and stromatolitic dolomite. The unit contains conglomerate, composed predominantly of porphyry fragments derived from the underlying volcanic rocks and comagmatic intrusions. Some conglomerates show a high abundance of basalt and andesite fragments.

The second unit is about 600 m thick and consists of dark-coloured limestone, and dolostone. Its middle portion contains unsorted greywacke. The third unit is 1000 to 1200 m thick and is mainly composed of rhythmically interbedded (5–20 cm) dark grey shale and sandstone. Locally the rocks exhibit slump folds and graded bedding. There are also

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unsorted non-layered greywacke and pebble conglomerate, up to 50 m thick. The unit also contains syndepositional volcanic material concentrated in tephra layers that consists of pumice, andesite, and felsite fragments. The upper unit is represented by laminated grey limestone with an exposed thickness of 800–1000 m.

The Kharaberin Formation is a transgressive sedimentary sequence. The lower horizons include storm-broken layers deposited in a shallow-water environment near a volcanic edifice. The third unit is mainly composed of relatively deep-water turbidites. The presence of tephra indicates synchronous volcanic activity and a temporal shift of the volcanic front towards the west (in present coordinates) and away from the depositional area. The fourth unit is composed of carbonate ooze and was deposited far away from the source of volcanic or terrigenous clastic material.

Geodynamic model for the Shishkhid ophiolite

High alumina andesite at the top of the igneous section evidently indicates an island arc setting for the Shiskhid ophiolite. Some geochemical features of the basaltic members also imply the supra-subduction setting. These are the low Fe/Mg ratios in the majority of basaltic rocks, classifying them as calc-alkaline. High abundances of fluid-mobile elements as well as a distinct negative Nb-anomaly, high Th, and high La/Nb ratios for most samples suggest subduction-modified mantle sources. Some volcanic rocks show no Nb-anomaly and these were presumably derived from a different mantle source. The felsic magma of the bimodal volcanic unit was produced by partial melting of an amphibole-plagioclase metagabbro protolith, which is the most likely source based on the REE patterns and Nd isotopic composition (see Kuzmichev et al. 2005 for discussion).

The Nd isotopic composition was analyzed for 10 whole-rock samples representing all crustal units of the Shishkhid ophiolite. The data show variable initial ε_{Nd} ranging from +6.9 to 0. This may be due to non-uniform overprinting of a depleted mantle by melts from subducted sediment or isotopic heterogeneity of the upper mantle source, or both. A comparison of geochemical and isotopic features suggests that both options were involved in melt production (see Kuzmichev et al. 2005 for discussion). Two different mantle reservoirs were probably involved in the genesis of the Shishkhid igneous rock assemblage. The main reservoir was depleted sub-arc mantle, metasomatized by subduction-related fluids and slab-derived melts. The second mantle source was possibly upwelling asthenosphere in a back-arc(?) environment. The latter source may account in particular for samples showing only poorly developed subduction-related signatures such as negligible Nb-anomalies.

Taking into account the unusual composition of the volcanic members of Shishkid ophiolite assemblage, an extensional back arc setting is suggested, similar to the "backarc knolls zone" of the Izu-Bonin island arc system. The latter is a region of crustal extension encompassing a 100 km-wide zone of generally shallow bathymetry between the frontal chain of large stratovolcanoes and the Shikoku oceanic basin (Hochstaedter et al. 2001 and references therein). In such a setting the mantle source was influenced by subduction-slab fluids, whereas some magma portions may have originated from fresh upwelling asthenosphere after stretching the subduction-modified upper mantle. Both magma types filled the extensional gaps in the arc crust and formed magma chambers with sheeted dykes at their roofs. Continued extension probably caused the Shishkhid ophiolite to move away from the active volcanic zone. Magmatism continued for some time to the west of the ophiolite which implies that the frontal part of the Shihkhid arc

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was facing westwards (in modern coordinates), whereas its eastern, rear part, including the ophiolite section, was facing the back-arc oceanic basin.

The Shishkhid ophiolite belt

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The thrust zone underlying the Shishkhid ophiolite is a distinct suture demarcating two terranes that were separated by an oceanic basin in the late Neoproterozoic. The suture becomes obscured to the NE and SW of the Shishkhid area due to metamorphism and deformation. Its presumed SW continuation is marked by a chain of serpentinite bodies and associated basalts trending through eastern and central Sangilen (Fig. 2). It is conceivable that the Shishkhid arc farther extends into western Sangilen composed of metamorphic rocks which conforms to detrital zircon ages (see below).

The NE continuation of the Shishkhid ophiolite belt is manifested by a chain of fragmented ophiolites along the northern limb of the Tuva-Mongolia Massif (Figs. 2, 3). These rocks are mostly composed of serpentinite, though several major bodies also contain gabbro, basalt, and minor felsic intrusions. Basaltic members are characterized by low contents in TiO₂ and FeO and exhibit calc-alkaline affinities. They probably belonged to the Shishkhid island-arc system in late Neoproterozoic times. It is suggested that the NE tip of the arc protruded into the Oka prism in its northen part and delivered volcanoclastics rocks to the trench deposits in the Tustuk region (see Fig. 9).

Age of Shishkhid arc

The age of volcanic members of the Shishkhid arc is constrained by U-Pb SHRIMP zircon dating of rhyolite of the bimodal unit. Seven zircons were analyzed and produced concordant data with a mean 206 Pb/ 238 U age of 800.0 ± 2.6 Ma (Kuzmichev et al. 2005). Hence, the Shiskhid arc already existed in the mid-Neoproterozoic.

To constrain its life span two samples of the volcanoclastic deposits were dated at the northeastern end of the arc where it is attached to the Oka prism. The first was a volcanic breccia consisting of andesite and dacite fragments. Ten zircons yielded $^{206}Pb/^{238}U$ ages ranging from 801 to 833 Ma with a mean of 813 ± 7 Ma (2σ). The second sample was a volcanic conglomerate containing rounded pebbles represented by qartz-porphyry, plagiocalse-porphyry, aphyric felsite and granophyre. The $^{206}Pb/^{238}U$ ages of the 12 zircons range between 770 ± 16 and 819 ± 13 Ma (Kuzmichev and Larionov 2013), and the data define two clusters: 819 ± 17 Ma (3 grains), which is the same as the age of the first sample, and 775 ± 8 Ma (2σ , 9 crystals). The data show that Shishkhid arc volcanism began somewhat earlier than Oka prism generation.

Dating of zircons from tephra and volcanic sandstones of the Kharaberin Fm. would be more reliable to determine the duration of the arc. Up to now age data are only avaliable for suggested metamorphosed counterparts of the Kharaberin clastic deposits located in western Sangilen. 58 detrital zircons were analyzed on SHRIMP II, 19 grains from polymictic metasandstone, 9 from metashale and 30 from arkosic metasandstone (Kozakov et al. 2005). Almost all zircons have Neoproterozoic ages in the range of 900–660 Ma using the ²⁰⁶Pb/²³⁸U ratios, three grains are Palaeo- and Mesoproterozoic, and one metamorphic rim yielded an age of granulite metamorphism (474 ± 22 Ma). 16 zircons of the first sample yielded a concordant cluster with a mean at 767 ± 15 Ma (2 σ , whereas all grains but one from the third sample define a cluster at 809 ± 17 Ma (Kozakov et al.

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2005). On the basis of these data the recorded age interval of arc magmatism is in the range of 810–750 Ma for the majority of zircons.

Late Neoproterozoic orogeny and Vendian-Cambrian Tuva-Mongolian microcontinent

After the oceanic lithosphere of the basin dividing the Sarkhoi and Shishkhid terranes had been completely subducted, the terranes collided, and the upper lithosphere of the rear part of the Shishkhid arc was thrust upon the Oka accretionary prism. Collision-related granitoids in the Tuva-Mongolian Massif have not yet been identified, therefore, the age of collision has been inferred from geological evidence. An Ediacaran (late Vendian) to middle Cambrian carbonate platform sequence covers most of the Tuva-Mongolian Massif, and its basal horizons are composed of red clastic sediments filling grabens (Kuzmichev 2004). The lowermost carbonate rocks are Ediacaran, suggesting an age of ~600 Ma for the collision event that may have been related to the late Baikalian orogenic phase which is widely manifested in southern Siberia (Kuzmichev 2004) though it may be somewhat older.

After the Shishkhid arc was accreted to the Sarchoi continental terrane, the Tuva-Mongolian microcontinent evolved as a carbonate platform fringed by a passive margin. The origin of the passive margin on the Shishkhid side of the microcontinent needs to be explained because, prior to being accreted, the Shishkhid arc was fringed by a convergent boundary on the oceanic side. Based on an analogy with northern Taiwan, the collisional orogen could have been rifted off, and the front active zone may have drifted away. In Taiwan the rift gave rise to the Okinawa back-arc basin as the Ryukyu arc moved away from the continent (Teng 1996). Some features in the Shishkhid area are compatible with such a scenario (Kuzmichev et al. 2005).

Synopsis of the Neoproterozoic and early Palaeozoic evolution of the TMM

The author suggests that during most of the Neoproterozoic the TMM represented a marginal region of a large old (Siberian?) continent. In the early Neoproterozoic this margin was a shallow-marine shelf area (Fig. 11a). At the same time the Dunzugur arc that originated not later than 1034 ± 9 Ma drifted in the surrounding ocean. The arc faced the continent, and subduction occurred beneath it (Figs. 11a, 12a). It is likely that such palaeogeography did not change for 200 million years. At the end of this time interval the Shishkhid arc was created by splitting off from the former Dunzhugur arc (Fig. 11b).

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Fig. 11. Summary of the Neoproterozoic evolution of the TMM. The time windows for palaeogeographic cartoons are indicated. a) Early Neoproterozoic: The Dunzhugur island arc faces the passive continental margin of an old continent. b) Splitting up of the Dunzhugur arc and birth of the Shishkhid arc. c) Mid-Neoproterozoic orogeny: collision of the Dunzhugur arc with the continent; forearc SSZ ophiolite obduction. d) Main units of the late Neoproterozoic palaeogeography: Sarkhoi supra-subduction Andean-type volcanic belt, Oka accretionary prism, and Shishkhid arc looking with its rear side to the continent. e) Ocean spreading ridge subduction beneath Oka prism and back-arc rifting in the rear of the Sarkhoi continental arc. f) Tuva-Mongolian micricontinent as a Japan-style continental arc facing the back of the Shishkhid arc. The latter was attached to the active continental margin creating a triple junction. g) Collision of Shishkhid arc and core of the Tuva-Mongolian microcontinent. Birth of the Agardakh arc by rifting of the orogen. h) Completely constructed Tuva Mongolian microcontinent drifting through the Palaeo-Asian ocean.

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The two arcs faced in opposite directions and co-existed for at least 50 Ma (Fig. 12b). This evolutionary model can explain the narrowness of the exposed Dunzhugur arc that hardly matches the long time of island arc evolution. It may also explain the reason for dense basaltic intrusions in the arc upper horizons.

Collision of the Dunzhugur arc with the continental margin may have occurred in the middle Neoproterozoic at ~810 Ma (Figs. 11c, 12c). The forearc lithospere was obducted onto the passive margin and produced gravitational nappes in the foreland basin. From



Fig. 12. Main evolutionary phases of the TMM as displayed in Fig. 11 but in section view.

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this time on, the subduction occurred beneath the continental margin, and an Andean-type volcanic belt began to evolve.

The palaeogeography at the beginning of the late Neoproterozoic included three tectonic units, namely the Sarkhoi volcanic belt resting on the continental margin, the Oka accretionary prism growing in the front of active continental margin, and the Shishkhid island arc (Fig. 11d). The Sarkhoi belt at present can be traced up to 1500 km, and it was probably much longer originally. Such disposition did not last long. At ~750 Ma an oceanic spreading ridge approached the subduction zone (Figs. 11e, 12d) and produced voluminous basaltic magma intrusions into the prism sediments. Back-arc rifting in the rear of the Sarkhoi continental arc was probably triggered by this event (Fig. 11e). As a result, the massif was cut-off from its parent continent. At about the same time or somewhat later the Shishkhid arc most likely collided with the Sarkhoi margin and generated the next triple junction (Fig. 11f). The life span of all the above three units is not well constrained. At least since the middle late Neoproterosoic the Tuva-Mongolian massif existed as a micrcontinent separated from its ancestor by a newly formed oceanic basin (Fig. 12e).

By ~600 Ma the Shishkhid arc was completely accreted to the Sarkhoi margin and partly thrust upon the Oka prism (Fig. 11g). The age of collision is unknown. Grabens filled with terrestrial clastic deposits presumably formed at around 600 Ma but maybe somewhat earlier. They are interpreted as a result of collisional collapse of the late Neoproterozoic orogen. These clastic rocks are overlain by carbonate deposits that constitute a platform cover on the entire Tuva-Mongolian massif. The latter drifted across the Palaeo-Asian ocean as a microcontinent fringed by passive margins during the terminal Neoproterozoic and most of the Cambrian (Fig. 11h). The palaeogeography of the surrounding ocean during this period of time became much more complicated. Numerous island arcs originated and evolved on both sides of the microcontinent. The well-known Agardakh arc presumably orinated by rifting of the margnal part of the Shishkhid collisional orogen (Fig. 11g).

The region was finally amalgamated by numerous collisions of island arcs and continental terranes. The process began in the late Cambrian and culminated in the Early Ordovician when the entire mass was accreted to the southwestern (recent coordinates) Siberian Platform. This was the principal event that formed the main structural style of the region. In the present erosional level about half of the TMM surface exposes Ordovician post-collisional granites and related plutonic rocks. The cause of this event is not yet ascertained. In the author's opinion such large-scale processes imply continental collision. The TMM and surrounding latest Neoproterozoic and Cambrian island arcs were squeezed between Siberia and one of the Gondwana continents.

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289

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307

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