Petrology, geochronology, and tectonic implications of c. 500 Ma metamorphic and igneous rocks along the northern margin of the Central Asian Orogen (Olkhon terrane, Lake Baikal, Siberia)

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Abstract: A significant portion of the continental crust of northern Eurasia is thought to have formed during the evolution of the Central Asian Orogenic Belt at the time of accretion of continental terranes and island arcs. Records of this event are well preserved within the Siberian craton–Central Asian Orogenic Belt transition zone in Lake Baikal region, particularly in the Olkhon terrane. Our results establish granulite-facies conditions for peak metamorphism in the Olkhon terrane, and indicate that the granulites were derived from island arc mafic volcanic rocks and back-arc basin sediments. Sensitive high-resolution ion microprobe dating of metamorphic zircons from two mafic granulites yielded 238U/206Pb ages of 507 ± 8 and 498 ± 7 Ma, and magmatic zircons from syntectonic syenite yielded an age of 495 ± 6 Ma. The main metamorphic event occurred at about 500 Ma, and was probably related to collision of the Barguzin microcontinent with the Siberian craton. Ages from 535 to 2750 Ma for detrital zircon cores in early Palaeozoic metasediments of the Olkhon terrane were obtained. Archaean ages of detrital zircons in such metasediments suggest that the Barguzin microcontinent was originally part of the Aldan Province of the Siberian craton that was detached in late Mesoproterozoic, and reattached to the craton during early Palaeozoic collision.

Most of the Earth’s continental crust is generally thought to have been generated during the Archaean and Palaeoproterozoic (e.g. Condie 1990, 2004; Windley 1995). However, substantial crustal material also formed in accretionary orogens during the Neoproterozoic (Reymer & Schubert 1986), the Palaeozoic and the Mesozoic (Samson 1989, 1995; DePaolo et al. 1991; Chen & Jahn 1998; Jahn et al. 2000). A good example of such an orogen is the Central Asian Orogenic Belt (Zonenshain et al. 1990; Hu et al. 2000), which formed mainly during the Palaeozoic by closure of the Palaeoasian ocean (Khain et al. 2003). A significant portion of the continental crust of northern Eurasia formed during the evolution of the Central Asian Orogenic Belt at the time of accretion of continental terranes and island arcs. The early stage of formation of this orogenic belt was partly synchronous with the final phases of the Pan-African orogeny (about 550–600 Ma), which led to Gondwana assembly (Hoffman 1991; Rodgers 1996).

Collision and accretion of several microcontinents and fragments of intra-oceanic complexes of the Palaeoasian ocean onto the Siberian craton margin led to formation of strongly deformed and metamorphosed terranes in the contact zone between the craton and Central Asian Orogenic Belt. A chain of such terranes extends for 1000 km along the southern craton margin, and includes (from east to west) the Olkhon, Slyudianka, Kitoykin and Derba terranes (Fig. 1). These terranes are separated from the Siberian craton by the Primorsky and Main Sayan faults. Records of terrane accretion and collision, as well as the geological histories of the various terranes, are particularly well preserved in metamorphic and igneous complexes within the craton–orogen transition zone in the southern Lake Baikal region of the Central Asian Orogenic Belt, particularly in the Olkhon terrane (Fedorovsky et al. 1995).

All the terranes listed above contain relics of high-grade metamorphic rocks (up to granulite facies). These metamorphic terranes have long been interpreted as early Precambrian rocks of the Siberian craton margin or as relics of ancient basement of the Central Asian Orogenic Belt (Petrova & Levizkii 1984, and references therein). A probable Archaean or early Proterozoic age for these rocks was based on their high metamorphic grade. Subsequent isotopic data demonstrated that high-grade metamorphism occurred in the early Palaeozoic (Bibikova et al. 1990; Salnikova et al. 1998; Donskaya et al. 2000). However, there is no detailed petrological or geochronological coverage of most of the metamorphic terranes, owing mainly to poor exposure or strong alteration of the rocks during low-temperature retrogression.

An exception is the Olkhon terrane in the southern Lake Baikal region, which is well exposed and has been mapped in detail (Fedorovsky et al. 1995). In this paper, we present new petrological, geochemical and isotopic data for the Olkhon terrane. The results have important implications for the history of terrane accretion and orogenic evolution in the northern Central Asian Orogenic Belt, and for the processes responsible for the growth of continental crust in northern Eurasia.

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Geology of the Olkhon terrane

The Olkhon terrane is located along the western shore of Lake Baikal and includes Olkhon Island (Fig. 2). It has been suggested that the Olkhon terrane was generated as a result of accretion and collision of the Barguzin microcontinent with the margin of the Siberian craton (Fedorovsky et al. 1995), which caused strong deformation and metamorphism in the Olkhon rocks and was accompanied by local igneous activity. We have investigated several metamorphic domains and related intrusions within the Olkhon terrane.

The Olkhon terrane is separated from the Siberian craton by a blastomylonite zone that contains tectonic slivers of Palaeozoic granulites and relics of early Proterozoic cratonic basement (Sukhorukov et al. 2005). The Olkhon terrane consists of sedimentary, volcanic and plutonic rocks (Fig. 2), the majority of which have undergone regional Barrovian-type metamorphism ranging from lower amphibolite to granulite facies (Fedorovsky et al. 1995; Rosen & Fedorovsky 2001). Metamorphic zonation corresponds to the andalusite–sillimanite and kyanite–sillimanite transition, and $P$–$T$ conditions of metamorphism increase from SE to NW.

Five main metamorphic zones, increasing in grade toward the craton margin, have been distinguished according to their mineral associations: (1) staurolite–chlorite–muscovite and staurolite–biotite–andalusite–muscovite zones; (2) staurolite–sillimanite–biotite–muscovite zone; (3) biotite–sillimanite–muscovite–orthoclase zone; (4) biotite–sillimanite–orthoclase and cordierite–garnet–orthoclase zones; (5) granulite zone (Rosen & Fedorovsky 2001). The metamorphic rocks can be divided into two main groups (Fig. 2). The first is composed of intra-oceanic (mainly arc-related) volcanic, volcano-sedimentary and sedimentary associations including ultramafic rocks, amphibolite, gneiss, calc-silicate rocks, marble and quartzite. The second group consists of granitoid gneiss and migmatite. Both groups are intruded by granite and pegmatite dykes as well as rare mafic dykes.

Fig. 1. Precambrian microcontinents and early Palaeozoic metamorphic terranes in the Sayan–Baikalian region of southern Siberia. Age of granulite event for the Derba terrane is after Nozhkin et al. (2004), for the Kitoykin terrane after Donskaya et al. (2000), for the Slyudyanka terrane after Salnikova et al. (1998) and for the Olkhon terrane from this study.

Fig. 2. Generalized geology of the Olkhon terrane (after Fedorovsky et al. 1995), showing locations of studied samples.
Detailed structural studies resulted in the recognition of three main phases of synmetamorphic deformation (Rosen & Fedorovsky 2001). The earliest phase is marked by oblique (transpressive) compression, resulting in thrusting and folding with varying dips of axial surfaces from subhorizontal to subvertical. The general direction of thrusting was to the NW (in present coordinates), towards the Siberian craton. The second phase of deformation resulted in the formation of numerous large dome structures. The third phase is particularly well documented in the Olkhon terrane and was responsible for numerous strike-slip (mostly dextral) faults.

The present positions of tectonostratigraphic units within the Olkhon terrane can be interpreted as the result of tectonic interdigitation of rocks from different crustal levels. Dipping and strike-slip faulting events have modified the initial structure so dramatically that reconstructions of the original stratigraphy are not possible. The general style of deformation is similar to that observed in modern accretion–collision belts (Fedorovsky et al. 2005).

Previous speculations on the age and evolution of the Olkhon terrane, based on whole-rock K–Ar and Rh–Sr dating (see Rosen & Fedorovsky 2001, and references therein), can be considered invalid in view of the likelihood of open-system behaviour of the above isotopic systems during metamorphism. There is a single conventional U–Pb multi-grain zircon age of 485 ± 5 Ma for a garnet– sillimanite granulite that was interpreted as age of metamorphic event (Bibikova et al. 1990).

Geological setting, sample descriptions and petrography

Our field and laboratory investigations in the Olkhon terrane focused on metamorphic rocks of the Khada and Khoboy complexes, which contain relicts of granulite and a foliated syntectonic syenite massif (Fig. 2). Additional maps and details of geochemical and geochronological analyses are available online at http://www.geosos.org.uk/SUP18285.

Khoboy granulite

Granulite-facies rocks occur in coastal cliffs at Khoboy Cape and on several neighbouring promontories in the northern part of Olkhon Island. These rocks are separated from amphibolite-grade rocks, which occupy large parts of the island, by a high-temperature basaltymylonite zone. The amphibolite-grade rocks include biotite and amphibole–biotite gneiss, calc-silicate rocks, marble and quartzite. The granulite-facies rocks include clinopyroxene granulite, garnet–orthopyroxene–biotite gneiss, spinel–garnet–biotite gneiss with corundum and sillimanite, and garnet–cordierite–biotite gneiss. Typical mineral assemblages (major phases) are as follows: (1) clinopyroxene granulite: Cpx + Pl + Qtz (Bt as minor phase; ilmenite (Im), apatite (Ap), zircon (Zrn), titanite (Ttn) as accessories; Amp as secondary metamorphic phase replacing Cpx); (2) garnet–orthopyroxene–biotite gneiss: Grt + Opx + Bt + Pl + Qtz (Ilm, Ap, Zrn as accessories); (3) garnet–cordierite–biotite gneiss: Grt + Crd + Bt + Kfs + Pl + Qtz (Sill as minor phase; Ilm, Ap, Zrn, Mnz as accessories; Ms after Sill); (4) spinel–garnet–biotite gneiss: Sp1 + Grt + Bt + Qtz + Kfs + Pl (Sill, Cm as minor phases; Zrn, Mnz, Ap as accessories). Both the amphibolite- and granulite-facies rocks are enriched in secondary (non-metamorphic) graphite. A sample of clinopyroxene granulite (sample 03210) was collected for U–Pb zircon geochronology from a fresh coastal cliff outcrop (Fig. 2) close to Cape Khoboy.

Khada granulite

The high-grade rocks of the eastern Khadarta Cape consist of two-pyroxene biotite- or amphibole-bearing mafic granulite and clinopyroxene–scapolite-bearing granulite. Garnet–biotite plagiogneiss, garnet–orthopyroxene granulite, garnet–two-pyroxene amphibole-bearing granulite, and marble also occur within the high-grade zone. The following mineral assemblages (major phases) were observed: (1) Bt + Cpx + Opx + Pl + Qtz + Ilm (Ap, Zrn as accessories); (2) Hbl + Cpx + Opx + Qtz + Pl (Bt as minor phase; Ilm, Ap, Zrn as accessories); (3) Cpx + Scp + Pl + Qtz (Bt as minor phase; Ttn, Ilm, Ap and Zrn as accessories). Minor replacement of pyroxene by amphibole was observed in some mafic granulites. All Khadarta granulite-facies rocks exhibit granoblastic textures.

Towards the NW, the granulite-facies rocks are replaced by diopside–plagioclase schist (locally garnet-bearing), diopside–quartz–carbonate rock, diopside–calcite schist, diopside-bearing quartzite and marble. These rocks belong to the amphibolite-facies complex. Small lenses of olivine pyroxenite occur among the metamorphic rocks of Khadarta Cape. Numerous non-foliated granitoid veins (trondjhemite, granite) intrude all the metamorphic rocks of this area.

A sample of mafic granulite (sample 03132) was collected for U–Pb zircon geochronology from an outcrop located near the southern end of Cape Khadarta (Fig. 2) and facing Maloe More (Small Sea) Strait.

Foliated quartz syenite–granite, Olkhon Island

A foliated quartz syenite massif, together with granitic rocks, which is over 20 km from SW to NE and 1–1.5 km in width, occurs in the southern part of Olkhon Island. This large linear intrusion cuts gneiss, granite–gneiss and amphibolite, and contains large enclaves of metamorphic rocks. Medium-grained quartz syenite is mainly composed of K-feldspar, quartz, plagioclase and amphibole. Minor phases include epidote and biotite, and apatite, titanite and zircon are accessory minerals. A well-developed foliation in the syenite is defined by the orientation of K-feldspar, amphibole and biotite. Because this foliation is parallel to the main regional foliation in the surrounding gneisses we consider the syenite to be of pre- or syntectonic age. Sample 03240 was collected for U–Pb zircon geochronology from an exposure of foliated quartz syenite located in the southeastern part of Olkhon Island (Fig. 2) within the granite syenite massif. The northern and western parts of the massif are intruded by steeply dipping, NW–SE-trending aplite dykes, up to 10 m wide, of undeformed massive granite. These dykes (more than 100) locally cut the syenites, and some also contain xenoliths of metamorphic rocks (Fedorovsky et al. 2005).

P–T estimates

Polished thin sections were carbon-coated and analysed at the Geological Institute, Siberian Branch, Russian Academy of Sciences in Ulan-Ude on a LEO1430VP microprobe (15 kV accelerating voltage, 5 nA beam current, 3–5 μm beam diameter) and on a MAR-3 electron microprobe (20 kV accelerating voltage, 40 nA beam current, 2–3 μm beam diameter). Analyses were calibrated using natural mineral standards.

Khoboy granulite

We studied mineral reactions and estimated P–T conditions for representative metamorphic rocks of the Khoboy Cape complex, including spinel–garnet–biotite gneiss (Fig. 3a and b), garnet–cordierite–biotite gneiss (Fig. 3c), and garnet–orthopyroxene–biotite gneiss (Fig. 3d).

For these rocks estimates of peak P–T conditions are not possible because of extensive retrogression. The retrograde P–T conditions were calculated using the post-peak metamorphic assemblage Grt(rim) + Opx(corona texture) + Bt(matrix) + Pl(matrix) + Qtz (garnet–orthopyroxene–biotite gneiss, sample
03206), Grt(rim) + Spl(matrix) + Bt(matrix) + Pl(matrix) + Qtz + Sil (spinel–garnet–biotite gneiss, sample 03218), Grt(rim) + Crd(corona texture) + Bt(matrix) + Pl(matrix) + Qtz + Sil (garnet–cordierite–biotite gneiss, sample 03225). Thermobarometry was performed using a multi-equilibrium approach with the self-consistent thermodynamic data of Holland & Powell (1998) and Thermocalc, version 27.5 and various geobarometers (Grt–Opx–Pl–Qtz (Bhattacharya et al. 1991), Grt–Pl–Qtz–Sil (Koziol & Newton 1989), Grt–Crd–Sil–Qtz (Thompson 1984)) and geothermometers (Grt–Opx (Bhattacharya et al. 1991), Grt–Bi (Hoinkes 1986), Grt–Crd (Bhattacharya et al. 1988)). Average P–T results using Thermocalc are broadly similar to those of geobarometry and geothermometry. Temperature estimates using geothermometry vary insignificantly (643–773 °C), whereas pressure estimates based on geobarometry range from 4.5 to 7.9 kbar.

Our P–T estimates together with mineral reactions and textures (Fig. 3a–d) observed in the Khoboy granulites indicate a near-isothermal decompression (ITD) path (Harley 1989).

**Khadarta granulite**

Samples of two-pyroxene biotite- or amphibole-bearing mafic granulite from the eastern part of the Khadarta Cape were selected for detailed studies. Pyroxenes in the mafic granulite are hypersthene (En$_{61-65}$Fs$_{34-38}$Wo$_{1}$) and augite (En$_{40-42}$Fs$_{10-14}$Wo$_{45-48}$), following the classification of Morimoto (1988). Biotite is close to the phlogopite end-member (Guidotti 1984). All analysed biotites have high TiO$_2$ contents (3.39–5.39%). Plagioclase is labradorite to bytownite. Primary amphibole (pargasite in composition according to Leake et al. 1997) is a major mineral phase in the only sample studied (03135). Secondary amphibole (magnesiohornblende in composition according to Leake et al. 1997) replaces pyroxene in all mafic granulite samples.

Pressures and temperatures were calculated using the equilibrium peak-metamorphic assemblage of mafic granulites with the association Cpx–Opx–Bt–Pl–Qtz (samples 03132, 03134 and 03135). Thermobarometry was performed using Thermocalc, version 27.5 (Holland & Powell 1998) and geobarometry (Cpx–
 Whole-rock geochemistry
For whole-rock chemical analyses representative homogenized powders (about 250 g) were used. Samples were dried in an oven overnight and fused into lithium borate glass discs and pressed into powder pellets for major oxide and trace element measurements, respectively. The analyses were carried out by XRF on a S4 PIONEER XRF spectrometer at the Institute of Geochemistry of the Siberian Branch of the Russian Academy of Sciences (SB RAS), Irkutsk.

Additional trace elements, including REE, were determined by inductively coupled plasma mass spectrometry (ICP-MS) using a VG Plasmaquad PQ-2 Plus (VG Elemental) in the Limnological Institute, SB RAS, Irkutsk. Instrumental configuration, operating conditions and the acid dissolution technique used to prepare the samples for ICP-MS analysis were described by Garbe-Schönberg (1993). Calibrations used international standards and international rock standards (BHVO-1, BIR-1, W-2 and RGM-1). Precision and accuracy are 0.5–1.0% for major elements and up to 5% for trace elements and REE (XRF, ICP-MS).

More than 35 fresh samples from the locations mentioned above were collected for whole-rock geochemistry. The degree of element mobility in high-grade metamorphic complexes is controversial (Tran et al., 2003, and references therein). Some workers (Ferry 1983; Roser & Korsch 1986; Passchier et al. 1990) have suggested that element mobility in high-grade metamorphic process could be insignificant. Considering this assumption we used major elements for the whole-rock chemical classification, but as a tool for tectonic settings discrimination we mainly discuss relatively immobile elements (Th, Zr, Nb, Ta, Y, Ti, Cr, Ni, Sc and REE, according to Tran et al., 2003, and references therein).

Khadarta granulate
The two-pyroxene mafic granulites have Nb/Y ratios varying from 0.01 to 0.09 and correspond to basalt according to the criteria of Winchester & Floyd (1977). These rocks are characterized by high Mg-number values up to 70 and high Ni + Cr contents (550–790 ppm) that are close to primary melt compositions. All samples are slightly enriched in REE (total REE 33–76 ppm). They show slight REE fractionation and REE-normalized abundance patterns with (La/Yb)n = 1.5–3.5, (Gd/Yb)n = 1.2–1.5 and Eu/Eu* = 0.83–0.96.

The mafic granulites are characterized by low Nb/Y (0.01–0.09) and Nb/Ce (0.01–0.07) ratios, which are close to those of island-arc basalts (IAB). The two-pyroxene granulites are similar to volcanic arc basalts, according to their Zr–Nb–Y ternary proportions (Meschede 1986) and the Ti–Zr classification of Pearce (1982). The samples demonstrate enrichment in Th, U and light REE (LREE), and relative depletion in high field strength elements (HFSE) and heavy REE (HREE) that distinguishes them from normal (N-MORB) and enriched mid-ocean ridge basalt (E-MORB) (Fig. 4). Such geochemical affinities are typical of IAB and may be interpreted as a possible slab contribution from altered MORB and subducted sediments (Hawkesworth et al. 1993).

Clinopyroxene scapolite-bearing granulites have very high CaO contents (16.4–19.1%), at SiO2 concentrations of c. 52%, and Al2O3 contents from 10.3 to 12.2%. The chemical compositions of these granulites are more typical of sediments than of igneous rocks (Condie 1993). According to the log(Na2O/K2O)−log(SiO2/Al2O3) classification of Pettijohn et al. (1987), the composition of these rocks corresponds to greywacke. The high CaO contents may be explained by calcareous components in the original sediments. The granulites are characterized by immobile element ratios (Ti/Zr 56–62; La/Sc 1.2–1.3) close to values for oceanic arc clastic sediments (Bhatia & Crook 1986).

Foliated quartz syenite–granite association
The foliated felsic igneous rocks making up the granitoid massif in the southern part of Olkhon Island can be classified as quartz syenite and granite according their SiO2–total alkali proportions (Middlemost 1985). Variations in chemical composition of the quartz syenite are not significant: SiO2 61.5–62.9%, K2O + Na2O 8.8–9.6%, P2O5 ≤0.23%. Coexisting granites have high total alkaline contents (up to 8.7%), corresponding to the subalkaline series.

The quartz syenite shows enrichment in incompatible trace elements such as large ion lithophile elements (LILE; Rb 42–68 ppm, Ba 1925–2981 ppm, Sr 1542–1919 ppm) and HFSE (Zr 114–162 ppm, Y 25–38 ppm), with a distinct depletion in Th, Nb, P and Ti. LREE abundances are generally ≥100 ppm. (La/Yb)n varies in the range 9–22. The association of quartz syenites with granites in the same massif classifies them as rocks of the granite syenite series. Subparallel abundance patterns in both rock types and pronounced negative anomalies in Nb–Ta, P and Ti (Fig. 5) confirm this suggestion.

Ion microprobe U–Pb zircon geochronology
Heavy mineral fractions were obtained from each sample using a Wilfley table, Frantz magnetic separator and high-density liquids. A representative selection of zircon crystals was extracted from each sample by hand-picking under a binocular microscope. Crystals, together with appropriate standard minerals, were cast in epoxy mounts, which were then polished

KL–Qtz (Raith et al. 1983) and geothermometry (Cpx–Opx (Powell 1978)). Average results of P–T calculations using Thermocalc are broadly similar to those of geobarometry and geothermometry. Temperature estimates using geothermometry vary from 804 to 877°C, whereas pressure estimates based on geobarometry range from 7.0 to 10.0 kbar. All results indicate granulite-facies metamorphism.
to section the crystals for analysis. Each mount was cleaned thoroughly, and the polished surface was documented with transmitted and reflected light micrographs, then vacuum-coated with a c. 500 nm layer of high-purity gold. To eliminate any residual water from the sampling surface, each mount was pumped down to high vacuum overnight in the sensitive high-resolution ion microprobe (SHRIMP) sample lock prior to analysis.

U–Pb zircon analyses were conducted using SHRIMP II ion microprobes at the Research School of Earth Sciences at the Australian National University, Canberra, and the John de Laeter Centre for Mass Spectrometry at Curtin University of Technology, Perth. U–Pb ratios and U, Th and Pb concentrations were determined relative to zircon standards CZ3 in Perth, and SL13 and FC-1 in Canberra. Analyses of standards were interspersed with those of unknown grains. Measured compositions were corrected for common Pb using non-radiogenic 204Pb. Prior to analysis, each site was cleaned by rastering the primary ion beam over the area for up to 3 min. Subsequently, 204Pb counts for most analyses remained low and constant, and showed no tendency to decrease over the course of a 20 min analysis, suggesting that common Pb in these crystals is mainly inherent to the mineral rather than surface-related. In almost all cases, corrections are sufficiently small to be insensitive to the choice of common Pb composition, and an average crustal composition (Stacey & Kramers 1975) appropriate to the age of the mineral was assumed. Data were processed using SQUID, version 1.02, and Isoplot, version 3.00 (Ludwig 2001, 2003). Weighted mean ages are reported below with 95% confidence intervals; uncertainties on mean 238U/206Pb ages include a contribution arising from calibration against the zircon reference standard.

**Clinopyroxene granulite sample 03210 (Khoboy complex)**

The sample contains both elongate, prismatic zircons, and oval to spherical, multi-faceted zircons. The majority contain cores that exhibit concentric growth zoning, surrounded by featureless rims. A total of 29 analyses (17 rims, 12 cores) were conducted.
on 22 zircons. With two exceptions, the proportion of common $^{208}\text{Pb}$ is less than 1%, with a median of 0.1%; analyses of zircon rims 11.1r and 20.1r indicated 2.4 and 17.8% common $^{206}\text{Pb}$, respectively, and are not considered further. Concentrations of $^{238}\text{U}$ and $^{232}\text{Th}$ in rims vary from 129 to 1314 ppm and 30 to 1080 ppm, respectively, with Th/U ratios between 0.06 and 0.63. One zircon rim (22.1r) has much higher U and Th contents (5400 and 590 ppm), although its Th/U ratio (0.11) is low. Zircon cores have U and Th concentrations of 113–338 ppm and 47–2010 ppm, respectively, and Th/U ratios between 0.23 and 0.84.

Fifteen zircons yield $^{238}\text{U}/^{206}\text{Pb}$ ages, ranging from 459 to 537 Ma, that are dispersed well beyond analytical precision (Fig. 6). Most are concordant, although a few lie slightly below concordia after correction for common Pb. There is a coherent group of seven $^{238}\text{U}/^{206}\text{Pb}$ ages that indicate a weighted mean of 498 ± 7 Ma (MSWD = 1.86). Two analyses, 5.1r and 12.1r, yielded significantly older ages of 519 and 537 Ma, and may indicate incorporation of inherited core material. Six analyses with $^{238}\text{U}/^{206}\text{Pb}$ ages between 484 and 459 Ma are interpreted to reflect Pb loss. The best estimate of the time at which zircon rims were formed is 498 ± 7 Ma, based on the main group of seven analyses.

Analyses of zircon cores are mainly concordant (Fig. 6) and yield apparent ages between 486 and 2753 Ma (ages <1000 Ma are based on $^{238}\text{U}/^{206}\text{Pb}$; those >1000 Ma on $^{207}\text{Pb}/^{206}\text{Pb}$). The four oldest results, between 1658 and 2753 Ma, are slightly to moderately discordant and may be considered minimum ages. Analysis 4.2c yielded a $^{238}\text{U}/^{206}\text{Pb}$ age of 486 Ma and an imprecise $^{207}\text{Pb}/^{206}\text{Pb}$ age of 717 Ma, and probably reflects loss of radiogenic Pb. Nine analyses of cores indicated mainly concordant ages between 535 and 1012 Ma.

**Mafic granulite sample 03132 (Khadarta complex)**

Fourteen analyses were obtained from eight zircons; six zircons were analysed twice. With one exception, the zircons are highly enriched in $^{238}\text{U}$ (1014–4049 ppm), relatively low in $^{232}\text{Th}$ (90–377 ppm), and have Th/U ratios between 0.07 and 0.34. Analysis 2.2 has 232 ppm U and 3.9% common $^{206}\text{Pb}$, is highly discordant and imprecise, and is not considered further. Common Pb for the remaining 13 analyses is very low; values for $f_{\text{Dol}}$ range from 0.07 to 0.14%. The measured compositions are concordant to slightly discordant following correction for common Pb (Fig. 7). Twelve of $^{207}\text{Pb}/^{206}\text{Pb}$ ratios indicate a mean age of 494 ± 11 Ma (MSWD = 0.98). However, $^{238}\text{U}/^{206}\text{Pb}$ ratios are dispersed beyond analytical precision. A coherent group of seven analyses is centred on concordia and yields a mean $^{238}\text{U}/^{206}\text{Pb}$ age of 507 ± 8 Ma (MSWD = 0.89). Two zircons yielded four significantly younger results (1.1, 1.2, 4.1 and 4.2) and probably reflect minor loss of radiogenic Pb. Two reversely discordant analyses (2.1 and 9.2) yielded older $^{238}\text{U}/^{206}\text{Pb}$ ages of 525 and 530 Ma, indicating either some unrecognized inherited material or enhanced sputtering of Pb relative to U from metamict zircon (McLaren et al. 1994), although a lack of correlation between $^{238}\text{U}/^{206}\text{Pb}$ age and U concentration argues against the second explanation. The age of zircons in this sample is taken to be 507 ± 8 Ma, based on the mean $^{238}\text{U}/^{206}\text{Pb}$ age of the concordant group of seven zircons.

**Quartz syenite sample 03240**

Zircons from quartz syenite are elongate, prismatic, brown in colour, and show euhedral concentric zoning. U and Th concentrations are 270–560 ppm and 68–262 ppm, respectively, with Th/U ratios between 0.15 and 0.74. Common Pb is low; the fraction of common $^{206}\text{Pb}$ is less than 1.1% for all analyses, with a mean of 0.3%. All $^{207}\text{Pb}/^{206}\text{Pb}$ ratios agree to within analytical precision and indicate a weighted mean age of 493 ± 11 Ma (MSWD = 0.84). Eight of 10 $^{238}\text{U}/^{206}\text{Pb}$ ratios agree to within analytical precision and yield a weighted mean age of 495 ± 6 Ma (MSWD = 1.45) (Fig. 8). Two slightly younger results (4.1 and 5.1) are consistent with minor loss of radiogenic Pb.

**Sr–Nd isotopic analysis**

Nd and Sr isotopic analyses employed about 100 mg of fine-grained powder (<50 µm) from each sample. The powders were dissolved in a...
mixture of HF and HNO₃, using a three-step microwave approach (Todand et al. 1995). Following dissolution, Sr and REE chemistry was determined by standard ion exchange methods. For Sm–Nd, HDHP columns were used following procedures described by White & Patchett (1984). Isotopic measurements were carried out at the Max-Planck-Institut für Chemie in Mainz using a Finnigan MAT 261 mass spectrometer. The isotope composition of Sr and Nd were measured with multiple collectors operating in static mode. Uncertainties in isotopic ratios are given as 2σ errors of the block mean (10–25 blocks and 10 mass scans for each block). Blanks for Nd and Sr were below 100 pg and are thus not significant. All Nd and Sr measurements were performed as IC measurements. Rb/Sr and Sm/Nd ratios are based on ICP-MS concentration analyses. The εNd(t) values and Nd mean crustal residence ages (TDM) were calculated using the depleted mantle (DM) values of Goldstein & Jacobsen (1988): ¹⁴⁳Nd/¹⁴⁴Nd = 0.513151 and ¹⁴⁷Sm/¹⁴⁴Nd = 0.2136, and chondritic uniform reservoir (CHUR) parameters of Jacobsen & Wasserburg (1984): ¹⁴⁳Nd/¹⁴⁴Nd = 0.512638 and ¹⁴⁷Sm/¹⁴⁴Nd = 0.1967.

Initial ⁸⁷Sr/⁸⁶Sr ratios for samples 03210, 03132 and 03240, calculated for 500 Ma, are between 0.7043 and 0.7078, and reflect the wide range of isotopically primitive to more evolved components that participated in the generation of these samples. This is also apparent from the Nd isotopes. Values of εNd(500 Ma) range from −3.2 to +4.0 and ¹⁴³Nd/¹⁴⁴Nd(500 Ma) between 0.511820 and 0.512143, with present-day ¹⁴³Nd/¹⁴⁴Nd(500 Ma) between 0.51272 and 0.5172, demonstrate the involvement of crustal sources as well as more primitive reservoirs.

The least isotopically evolved sample is Khadarta granulite sample 03132, derived from basalt, which has the highest εNd(500 Ma) Value of +3.0 and low ⁸⁷Sr/⁸⁶Sr ratio (0.7046). This reflects a significant juvenile component, consistent with the inferred basaltic composition of the protolith. However, Nb–Ta and Ti negative anomalies on primitive-mantle normalized patterns (Fig. 4) document a crustal component in source of this basalt that is typical for island-arc basalts. Khoboy granulite sample 03210, of sedimentary origin, has a largely crustal isotopic composition that indicates a significant contribution from crustal material in its genesis. The inferred sedimentary nature of the protolith, the negative εNd(0) and the correspondingly high Nd mean crustal residence age (TDM = 1524 Ma) all suggest significant input of early Proterozoic crustal material. Syenite sample 03240 has intermediate Nd isotopic characteristics (εNd(500 Ma) = −0.3), low ⁸⁷Sr/⁸⁶Sr ratio (0.7043) and Nb–Ta depletion (Fig. 5) that may reflect a mixture mantle and crustal sources; its Nd model age (TDM = 1286 Ma) supports this speculation.

**Discussion**

**Significance of U–Pb zircon ages**

Our new data permit some preliminary conclusions to be made regarding the early history of the Olkhon terrane. We consider the U–Pb zircon ages of 498 ± 7 and 507 ± 8 Ma for the Khoboy and Khadarta granulites to indicate the timing of the granulite-facies metamorphic event in the Olkhon terrane. The age of 495 ± 6 Ma for the quartz syenite is interpreted as the time of its emplacement.

Constraints on the nature and age of terranes involved in the formation of the Central Asian Orogenic Belt are provided by the ages of zircon cores in Khoboy granulite sample 03210. These data document a chronologically heterogeneous source terrane for the precursor sediments deposited in successions of the Olkhon terrane. The sedimentary protolith of the Khoboy granulite, probably a greywacke, has a maximum depositional age of 535 ± 5 Ma, and reflects a continental source(s) dating back to at least 2753 Ma. The zircon core ages (Fig. 6), which have a wide range, form two main groups: Archaean–Palaeoproterozoic and Neoproterozoic. The older group represents input from ancient craton(s), whereas the younger group contains zircons that probably crystallized in Neoproterozoic microcontinents or intra-oceanic complexes.

The oldest zircon core yielded a slightly reversely discordant ⁷⁰⁷Pb/⁷⁰⁶Pb age of 2753 Ma. This is a typical age for granites within the southern part of the Siberian craton (Aldan Province, Larin et al. 2004) and suggests this area as a possible source for the detrital zircons. The next youngest core furnished a normally discordant ⁷⁰⁷Pb/⁷⁰⁶Pb age of 2488 Ma. Rocks of this age are rare in most of the Siberian craton but occur mainly in granitoid complexes of the Aldan Province (Salnikova et al. 1997). The presence of 2.4–2.5 Ga rocks is not unique to the Siberian craton, and igneous and metamorphic activity around the Archaean–Proterozoic boundary is known from several ancient cratons, notably northern China (Windley 1995). Nevertheless, the Siberian craton may be considered a possible source of 2.4–2.5 Ga zircons. Some cores are variably discordant and indicate Palaeoproterozoic minimum ⁷⁰⁷Pb/⁷⁰⁶Pb ages of 1658 and 1794 Ma. Rocks of this age occur within the southern part of the Siberian craton and are represented by small granite massifs in the Angara–Kann block (Bibikova et al. 2001) and by volcano-plutonic associations in the Aldan Province (Ulkab belt, Larin et al. 1997). Thus, the presence in sample 03210 of zircon cores corresponding to ages of igneous events in the Aldan Province makes it likely that the southern part of the Siberian craton is a possible source of Archaean and Palaeoproterozoic detritus in Olkhon metasediment sample 03210.

The younger group of concordant or moderately discordant zircon core ages ranges from 1012 to 535 Ma (Fig. 6). We suggest that these cores were derived from late Mesoproterozoic, Neoproterozoic, and earliest Palaeozoic arc terranes formed within the Palaeoasian ocean. Based on U–Pb zircon ages for ophiolites and arc terranes in southern Siberia, Khain et al. (2003) suggested that the main crust-formation events in the Palaeoasian ocean occurred at 1000–1010, 830, 740–700, 670–640, 570, 540 and 500–490 Ma. We suggest that the younger zircon cores in sample 03210 were introduced into the Olkhon...
sedimentary succession from a marginal arc (active continental margin of the Barguzin microcontinent).

**Evolution of the Olkhon terrane**

Peak $P-T$ conditions of metamorphism in Olkhon granulites are 800–900 °C at moderate pressures of 8–10 kbar. The $P-T$ path of the studied rocks suggests near-isothermal decompression (ITD) following the metamorphic peak. ITD granulites are generally interpreted to have formed in crust thickened by collision, with magmatic additions being an important supplemental heat source (Harley 1989).

At least two scenarios can be considered for evolution of the Olkhon metamorphic terrane that are compatible with these considerations. The first involves successive accretion of oceanic terranes of different ages and compositions (including island-arc and back-arc basin complexes) against the Siberian margin to generate a large crustal unit (the Barguzin microcontinent). In this case, we would expect to find metamorphic rocks younger than the studied granulites (c. 495–500 Ma) within the transition zone between the craton and fold belt, reflecting the processes of collision or accretion of the Barguzin microcontinent to the Siberian craton. However, there are no high-grade rocks or syncollisional igneous complexes in the study area younger than 495–500 Ma. Moreover, other evidence of amalgamation within the interior of the Barguzin microcontinent should be present. However, rocks such as granulite and syncollisional hypersthene-bearing granite (496 ± 3 Ma, U–Pb zircon age, Fedorovsky et al. 2005) occur only locally within the transitional zone near the craton margin and not in the interior of the Barguzin microcontinent.

Alternatively, the Barguzin microcontinent formed elsewhere and drifted in a northwesterly direction (in present coordinates) to collide with the Siberian craton. This scenario, involving early Palaeozoic collision of an assembled Barguzin microcontinental marginal arc and related basins with the Siberian craton, is consistent with the timing and evolution of metamorphism determined for the Olkhon terrane. The chemical characteristics and lithological affinities of the Khadarta (island-arc association) and Khoboy (arc-related basin sequence) complexes support this suggestion. The age of ITD metamorphism (c. 500 Ma) reflects collision of the marginal Barguzin arc with the Siberian craton. Collision was accompanied by syntectonic igneous activity and intrusion of 495 ± 6 Ma quartz syenite (this study) and 496 ± 3 Ma hypersthene-bearing granite (Fedorovsky et al. 2005) into thickened and heated lithosphere of the collisional zone.

This scenario corresponds generally to the Scandinavian-type orogen, which involves crustal thickening and diapirc rise, followed by decompression and melting (Dewey 1988). Subsequent orogenic collapse is marked by intrusion of granite veins, limited in age between 475 and 465 Ma (U–Pb zircon data, Yudin et al. 2005). At about 450–440 Ma, most of the earlier high-grade syn- and post-collisional igneous complexes were subjected to a significant overprint related to the development of large- and small-scale strike-slip faults within the Olkhon terrane. The earlier thrust- and dome-related structures were compressed and folded into large sigmoids (Fedorovsky et al. 2005). Strike-slip displacement along crustal segments was accompanied by formation of blastomylonite (Sukhorukov et al. 2005), which contains relics of c. 500 Ma granulites.

Peak pressure estimates for the Khadarta complex yielded moderate $P$ values not exceeding 10 kbar, and argue against significant lithospheric thickening during the collision event. Peak metamorphic temperatures (up to 880 °C) calculated for peak-granulite formation could therefore not have been achieved through lithospheric thickening alone, and other heat sources must have played a significant role. The c. 500 Ma syenites in the Olkhon terrane may be responsible for advection of additional heat into the collision zone. As shown by Jung et al. (2004), such alkaline rocks may be derived from enriched upper mantle sources (Dawson 1987), from asthenospheric mantle (Fitton 1987), or from interaction of asthenosphere-derived melts with the overlying lithosphere (e.g. Menzies 1987). In all these cases, generation of syenite magma involves a mantle component in the source. Therefore, syenite in the Olkhon terrane may reflect mantle magma transport into thickened lithosphere of the collision zone. Following Jung et al. (2004), we suggest that the c. 500 Ma syenite in the Olkhon terrane could be the result of mixing of mantle-derived silica-undersaturated alkaline magma with a crustal component. The isotopic systematics of quartz syenite sample 03240 (in particular the low negative $e_{Nd(t)}$ value) is compatible with such a genesis.

Suspecting the Aldan Province as a possible source of the older zircons (2753–1658 Ma), and regarding the younger grains (1012–535 Ma) as recording a marginal arc contribution, we develop the following model for the evolution of the Barguzin microcontinent (Figs 9 and 10a–c). The precursor of this block may have been part of the southeastern margin of the Siberian craton (Aldan Province) that was detached from the craton in the early–middle Neoproterozoic (Figs 9 and 10a). The precise time of detachment is unclear, because the southern part of the Aldan Province was strongly reworked in the Mesozoic. However, this time was probably close to 1.10–1.05 Ga and reflects the earliest stage of passive margin evolution along the southern part of the Aldan Province (Gallet et al. 2000; Khudoley et al. 2001). This detached fragment (the Barguzin microcontinent) probably drifted from SE to NW (in present coordinates) across the Palaeoasian ocean (Figs 9 and 10a,b). Subduction of oceanic crust beneath the northwestern margin of the microcontinent led to the generation of a marginal arc (Figs 9 and 10b). The Khadarta mafic granulite may be a relict of this marginal arc. The group of mid-Neoproterozoic zircon ages (Fig. 6) may be

![Fig. 9. Speculative model of movement of the Barguzin microcontinent through the Palaeoasian ocean.](image-url)
somehow related to this marginal arc. This SE-directed subduction (present coordinates) (Fig. 9) probably caused the shrinking of the oceanic crust between the Siberian craton and the Barguzin microcontinent and eventually led to their oblique collision at c. 500 Ma (Figs 9 and 10c).

During oblique collision with the SE part of the Angara–Anabar Province of the Siberian craton at about 500 Ma (Fig. 9), the active margin of the Barguzin microcontinent, the Siberian margin, and the marginal arc complexes and arc-related sedimentary basins were deformed and metamorphosed to granulite facies and now form the Olkhon high-grade terrane (Fig. 10c). Collision was accompanied by emplacement of mantle-derived magmas, which caused melting of thickened lithosphere and intrusion of syntectonic granitoids. Oblique collision also resulted in the development of large-scale strike-slip (mostly dextral) faults.

A similar model was proposed by Kuzmichev et al. (2001) for the Tuva–Mongolian, Dzabkhan, and Central Mongolian microcontinents, and those workers also considered these microcontinental fragments as detached parts of the Aldan shield.

Our results, together with the data of Fedorovsky et al. (1995), Salnikova et al. (1998), Donskaya et al. (2000) and Kuzmichev et al. (2001), provide evidence for an early Palaeozoic collision belt along the southern margin of the Siberian craton. This belt, along Lake Baikal, developed by accretion of arcs and microcontinents relatively early during the evolution of the Central Asian Orogenic Belt and was responsible for substantial continental growth in southern Siberia.

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