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The Sources of Paleoproterozoic Collisional Granitoids (Sharyzhalgai Uplift, Southwestern Siberian Craton): from Lithospheric Mantle to Upper Crust

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Abstract—The paper presents the geochemical and isotope characteristics of rocks and the U–Pb age and Lu–Hf isotope composition of zircons from three plutons of Paleoproterozoic granitoids in the Sharyzhalgai uplift (southwestern Siberian craton). The age of granitoids of the Toisuk (1838 ± 6 and 1827 ± 9 Ma), Nizhnii Kitoi (1846 ± 7 Ma), and Malaya Belaya (1863 ± 16 Ma) plutons corresponds to the Late Paleoproterozoic collision stage and is correlated with the time of mafic magmatism. The studied rocks have a wide range of silica contents. The Toisuk pluton is composed of a range of rocks from monzodiorites to granodiorites (granosyenites) and granites; the Nizhnii Kitoi pluton, of granodiorites and granites; and the Malaya Belaya pluton, of leucogranites. The rocks of the three plutons are highly ferroan, enriched in LREE, Th, and HFSE, and correspond in composition to *A*-type granitoids. A characteristic feature of melanocratic granitoids of the Toisuk and Nizhnii Kitoi plutons is extremely high contents of Ba: 4080–1500 ppm and 1560–990 ppm, respectively. Based on analysis of experimental data on the melting of various substrates and the results of numerical simulation, it is assumed that monzodiorite–granodiorites of the Toisuk pluton and granodiorites of the Nizhnii Kitoi pluton resulted from the differentiation/melting of a mafic source similar in Ba and Sr contents to intraplate continental basalts. The isotope compositions of zircon and melanocratic granitoids of the Toisuk (ε_{Hf} from –6.0 to –10.7 and ε_{Nd} from –5.3 to –10.2) and Nizhnii Kitoi (ε_{Hf} from –5.0 to –8.1 and ε_{Nd} = –4.0 and –5.1) plutons argue for the generation of their mafic sources from the enriched lithospheric mantle formed as a result of Neoarchean subduction processes. Vein granites of the Toisuk pluton and leucogranites of the Malaya Belaya pluton formed through the melting of quartz–feldspar (granodiorite) substrate. The contrasting isotope parameters of the Toisuk vein granites (ε_{Hf} from –6.7 to –10.1, zircons, and ε_{Nd} = –5.5, rock) and Malaya Belaya leucogranites (ε_{Hf} from 2.9 to 5.9, zircons, and ε_{Nd} from +0.7 to –1.9, rocks) indicate melting of the Archean and Paleoproterozoic crust, respectively. The more radiogenic Hf isotope composition of zircons from vein granites as compared with rocks of the Archean crust of the Irkut terrane is evident of the contribution of juvenile material to the granite formation.

Keywords: collisional granitoids, zircon, Lu–Hf isotope composition, mantle and crustal sources, Paleoproterozoic, southwestern Siberian craton

INTRODUCTION

Granitoids serve as markers of the formation and evolution of continental crust. In collision orogens, most granites derived from intracrustal melting, which resulted in vertical differentiation of the crust without a significant volume increase and a change in bulk chemical composition. On the contrary, granite formation involving mantle melts not only as the sources of heat initiating melting but also as the sources of material leads to a change in the composition of continental crust and its growth at the expense of juvenile-material influx. Thus, the genesis of granitoids determines different styles of crustal evolution.

The compositional diversity of collisional granitoids arises from (1) different compositions of their sources, (2) varying melting conditions, and (3) interaction between mafic and felsic magmas. The presence of mafic plutons spatially associated with granites suggests a direct contribution of mantle-derived magmas. At the same time, it is still unclear whether these magmas and their derivates can form significant mass portions of granitoid plutons in collision belts.

The Nd and Hf isotope compositions of rocks and zircons are markers of granitoid genesis. Ancient crustal and mantle-related (depleted) sources of granites differ strongly in composition and isotope characteristics. In contrast to the rocks of the ancient continental crust, juvenile crustal sources are similar in isotope parameters to the depleted mantle-derived melts. Along with the involvement of mafic melts from the depleted mantle in the granite formation, mafic

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magmas in terranes with thick lithosphere might derive from the subcontinental mantle. In ancient cratons, the latter exhibits the isotope characteristics of the enriched mantle as a result of its metasomatism. Thus, at least four different sources can be involved in granite formation: ancient and juvenile crustal sources and depleted and enriched mantle ones. Therefore, to identify the sources, it is necessary to invoke additional data on the major- and trace-element composition of rocks.

In the southwestern Siberian Platform, the origin of Paleoproterozoic granitoids was associated with collision processes during the amalgamation of Archean terranes into the structure of the Siberian craton and, globally, with the formation of the Paleoproterozoic supercontinent. In the Sharyzhalgai, Biryusa, and Angara–Kan uplifts of the basement (Fig. 1*A*), granitoids formed in the narrow time interval 1.88–1.84 Ga (Donskaya et al., 2002, 2005; Turkina et al., 2006; Nozhkin et al., 2009). By structural position, these granitoids are postorogenic rocks, which suggests their formation under postcollisional extentional setting. The relationship of the Paleoproterozoic granite formation in the southwest of the Siberian craton with the collision of Archean crustal blocks determined their specific isotope composition. The available data show that most of granitoids are characterized by a model Nd age of ≥ 2.5 Ga and negative ε_{Nd} values, which probably reflects the predominant contribution of long-lived crustal sources (Donskaya et al., 2005, 2014; Turkina et al., 2006; Nozhkin et al., 2009). At the same time, the wide range of ε_{Nd} values (–13 to –3) for the Paleoproterozoic granites is evident of the involvement of sources with different crustal prehistory in the melting. We present data on the geochemistry of rocks and the U–Pb age and Hf isotope composition of zircon from Paleoproterozoic granitoids of three plutons in the Sharyzhalgai uplift in order to assess the nature of crustal and mantle sources and their role in the granite formation.

METHODS

The major and trace elements in rocks were determined by XRF analysis on a ARL–9900 XL X-ray spectrometer and by ICP MS on an ELEMENT-II (Finnigan Mat) high-precision mass spectrometer with a U-5000AT+ ultrasonic spray at the Analytical Center of the Institute of Geology and Mineralogy, Novosibirsk. The detection limits of REE and HFSE range from 0.005 to 0.1 ppm. The average analysis accuracy was 2–7 rel.%.

U–Pb zircon dating was carried out on a SHRIMP-II ion microprobe at the Center of Isotopic Research of the Russian Research Geological Institute, St. Petersburg, following the standard technique (Williams, 1998). To choose analytical sites (spots), we used optical (in transmitted and reflected light) and cathodoluminescent (CL) images reflecting the internal structure and zoning of zircons. The intensity of a primary molecular-oxygen beam was 4 nA, and the spot (crater) was 25 μm in diameter and 2 μm in depth. The obtained data were processed using the SQUID program (Ludwig, 2000). The U/Pb ratios were normalized to a value of 0.0668 attributed to the TEMORA standard zircon with an age of 416.75 Ma. The errors of single analyses (isotope ratios and ages) are given at the 1s level, and the errors of calculation of the concordant ages and intercepts with concordia are given at the 2s level. The concordia plots were constructed using the ISOPLOT/ET program (Ludwig, 1999).

The Lu–Hf isotope composition of zircon was determined by ICP MS with laser ablation, using a COMPex-102 193 nm ArF laser, a DUV-193 ablation system, and a ThermoFinnigan Neptune multicollector mass spectrometer with inductively coupled plasma. The analyses were carried out at the Center of Isotopic Studies of the Russian Research Geological Institute, St. Petersburg, following the technique described by Griffin et al. (2000). For the correction of mass discrimination, one normalizing ratio (^{178}Hf/^{177}Hf) was used. Isotopes ^{172}Yb and ^{175}Lu were used for the correction of the interference of ^{176}Lu and ^{176}Yb on ^{176}Hf. Analysis of isotope composition was performed at the same spots as U–Pb SIMS dating, but the crater diameter was ~50 μm, and the crater depth was 20–40 μm. During the measurement period, the average values of ^{176}Hf/^{177}Hf for the zircon standards were 0.282680 ± 23 (TEMORA, n = 10), 0.282497 ± 16 (Mud Tank, n = 6), and 0.281994 ± 20 (GJ-1, n = 6). Zircons from the Malaya Belaya pluton were studied by LA–ICP-MS at the Analytical Centre of James Cook University (Townsville, Australia), using a GeoLas 193 nm ArF excimer laser ablation system and a Thermo Scientific Neptune multicollector mass spectrometer. The analysis procedure is described in detail by Kemp et al. (2009). On the data processing, the decay constant ^{176}Lu = 1.867×10^{-11} years^{-1} (Söderlund et al., 2004) was adopted. For calculation of $_{Hf}$ values, the chondrite values were used: ^{176}Lu/^{177}Hf = 0.0332 and ^{176}Hf/^{177}Hf = 0.282772. The model Hf age (T^{c}_{Hf}(DM)) was calculated relative to the depleted mantle (DM) with parameters ^{176}Lu/^{177}Hf = 0.0336 and ^{176}Hf/^{177}Hf = 0.28325 (Bouvier et al., 2008) by projecting the initial ^{176}Hf/^{177}Hf value for zircon to the DM line and using the average value for crust (^{176}Lu/^{177}Hf = 0.015).

The Sm and Nd contents and isotope ratios were determined on a Finnigan MAT-262 (RPQ) seven-channel mass spectrometer in the static mode, following the procedure described by Bayanova (2004), at the Geological Institute (Apatity). The laboratory contamination was 0.06 ng for Sm and 0.3 ng for Nd. The accuracy of determination was ±0.2% (2σ) for Sm and Nd, ±0.2% (2σ) for ^{147}Sm/^{144}Nd, and ±0.003% (2σ) for ^{143}Nd/^{144}Nd. The measured ^{143}Nd/^{144}Nd values were normalized to ^{148}Nd/^{144}Nd = 0.251578 corresponding to ^{146}Nd/^{144}Nd = 0.7219 and were reduced to ^{143}Nd/^{144}Nd = 0.511860 in the La Jolla Nd standard. The quality of measurements was monitored by measuring the isotope ratios in the La Jolla standard: For the period of study, the weighted average values of ^{143}Nd/^{144}Nd in it

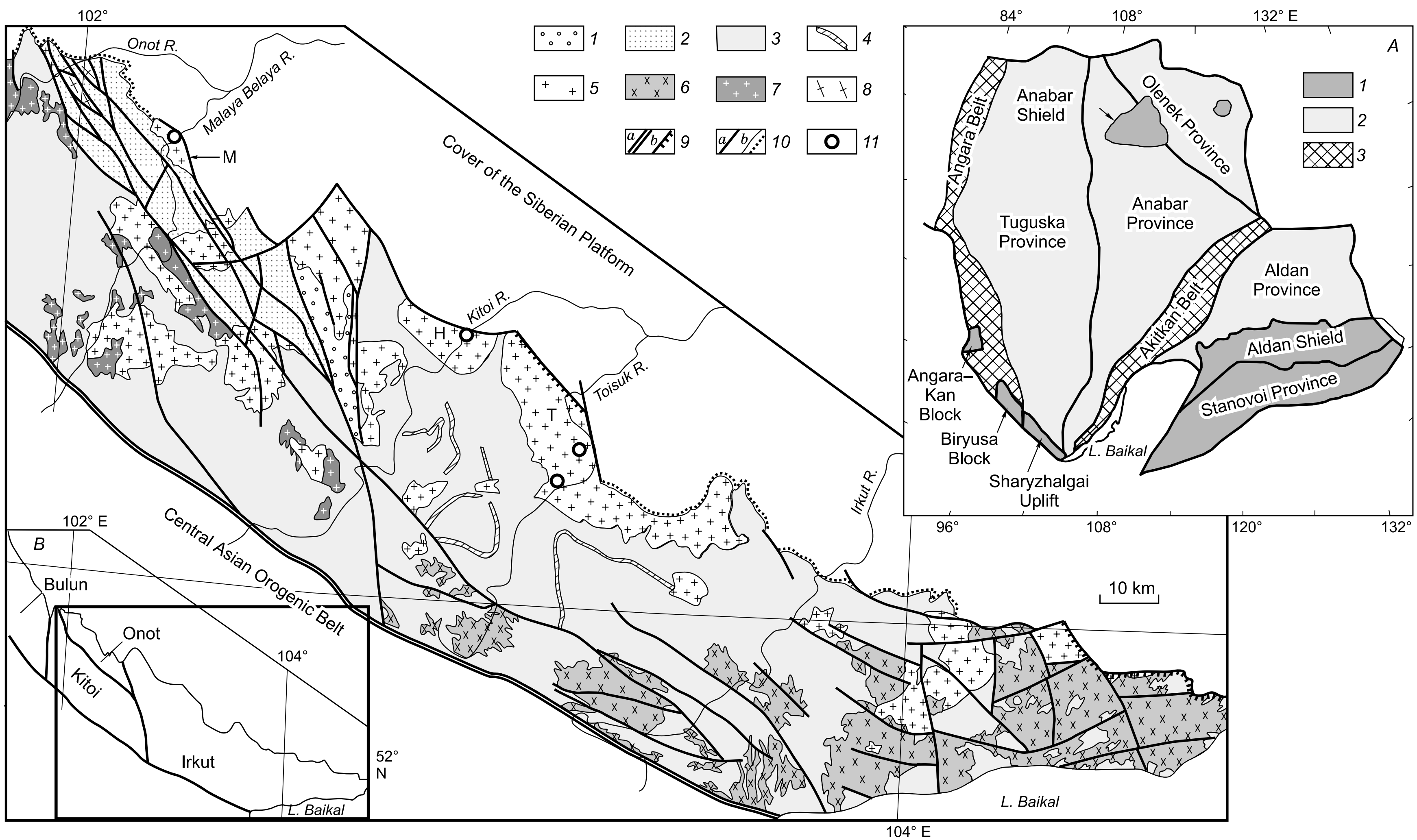

Fig. 1. Schematic geological map of the Sharyzhalgai uplift. *1*, Lower Proterozoic deposits; *2*, volcanometasedimentary deposits of the Onot greenstone belt; *3*, granulite–gneiss complexes of the Irkut and Kitoi terranes; *4*, marking beds; *5*, Paleoproterozoic granitoids; *6*, undivided Archean–Paleoproterozoic granitoids; *7*, Archean granitoids; *8*, Paleoarchean plagiogneisses and plagiogranitoids of the TTG complex; *9*, tectonic boundaries: *a*, faults, *b*, thrusts; *10*, geologic boundaries (*a*), unconformity (*b*); *11*, localities of sampling for isotope-geochronological studies. Plutons: T, Toisuk; N, Nizhnii Kitoi; M, Malaya Belaya. Inset *A*: Major tectonic elements of the Siberian craton: *1*, basement uplifts, *2*, buried basement, *3*, Paleoproterozoic orogenic belts. Inset *B*: Scheme of terranes of the Sharyzhalgai uplift, with the outlined part of the geological scheme in Fig. 1.

(n = 15) were 0.511833 ± 15 (2s). The one-stage model age T_{Nd}(DM) was calculated relative to the depleted mantle (DM) ($^{147}Sm/^{144}Nd$ = 0.2136 and $^{143}Nd/^{144}Nd$ = 0.51315), and the ε_{Nd} values were determined relative to CHUR ($^{147}Sm/^{144}Nd$ = 0.1967 and $^{143}Nd/^{144}Nd$ = 0.512638).

THE GEOLOGIC LOCATION AND COMPOSITION OF GRANITOID PLUTONS

The Sharyzhalgai uplift is an exposed basement in the southwestern Siberian Platform. It extends in the northwestern direction for ~350 km from the shore of Lake Baikal to the Oka River (Fig. 1*A*). From the northwest to southeast, four terranes are distinguished in its structure: Bulun and Onot granite–greenstone and Kitoi and Irkut granulite–gneiss ones; their borders are regional faults of NW and N–S strikes (Fig. 1*B*). The Onot and Bulun terranes are formed by Paleoarchean (3.4–3.3 Ga) plagiogneisses and plagiogranitoids of tonalite–trondhjemite–granodiorite composition (TTG complex) (Bibikova et al., 2006; Turkina et al., 2009) and metasedimentary-volcanic deposits of greenstone belts (GB) (Turkina and Nozhkin, 2008). Based on dating of zircons from metagraywackes of the Urik GB and of detrital zircons from metapelites of the Onot greenstone belt, it is suggested that these belts formed in the Mesoarchean (~2.8 Ga) and Neoarchean (~2.7 Ga), respectively (Turkina et al., 2014a,b).

The Kitoi and Irkut terranes forming most of the Sharyzhalgai uplift are composed of similar rock associations metamorphosed to the granulite facies. These associations include gneisses and granites of intermediate and felsic compositions, metabasites, garnet–biotite and cordierite- and sillimanite-bearing high-alumina gneisses, marbles, and calciphyres (Nozhkin and Turkina, 1993; Gladkochub et al., 2005; Poller et al., 2005). Crustal growth within the Irkut terrane began in the Paleoarchean. The Paleoarchean (~3.4 Ga) crust relics are biotite–orthopyroxene and two-pyroxene granulites of intermediate composition that underwent high-temperature metamorphism at ~3.0 Ga ((Poller et al., 2005; Turkina et al., 2011). The model Nd age (T^{c}_{Nd}(DM) = 2.9–3.3 Ga) of predominant intermediate to felsic Neoarchean granulites and the model Hf age of their magmatic zircons (T^{c}_{Hf}(DM) = 3.0–3.3 Ga) indicate a wide distribution of the Paleoarchean crust at the deep level (Turkina, 2010; Turkina et al., 2012). The magmatic protoliths of mafic and intermediate–felsic granulites predominant in the Irkut and Kitoi terranes formed in the subduction setting; their age is 2.6–2.7 Ga (Poller et al., 2005; Turkina et al., 2012). Two stages of metamorphism were established from the dates for zircons from these granulites: 2.55–2.6 and 1.85–1.86 Ga, which are correlated with the time of the granite formation (2.53–2.54 and 1.85–1.86 Ga) (Gladkochub et al., 2005; Poller et al., 2005; Sal'nikova et al., 2007; Turkina et al., 2012). There are two metasedimentary units within the granulite–gneiss terranes (Turkina and Sukhorukov, 2015). One unit is located in the Kitoi terrane and in the northwest of the Irkut terrane and is composed of cordierite- and sillimanite-bearing and garnet–biotite gneisses. Dating of their detrital zircons yielded the age of the sedimentary protoliths of 2.74–2.70 Ga. The protoliths underwent metamorphism at the Neoarchean–Paleoproterozoic boundary and in the Late Paleoproterozoic (Turkina et al., 2017). The other unit is composed of garnet–biotite, orthopyroxene–biotite, and high-alumina gneisses and calciphyres and occurs in interdomal zones in the southeast of the Irkut terrane. The ages of the youngest detrital cores of zircon from the paragneisses (1.95–2.00 Ga) and its metamorphic rims (~1.85 Ga) constrain the time of sedimentation to the interval 1.85–1.95 Ga (Turkina et al., 2010).

Paleoproterozoic granitoids form a belt of intrusions along the northeastern boundary of the Sharyzhalgai uplift (Fig. 1).

The Toisuk pluton is located in the northwest of the Irkut terrane and is the largest intrusion, 84 km^2 in area (Fig. 1). The host rocks are Archean felsic granulites, gneiss-granites, and garnet–biotite and high-alumina paragneisses, often migmatized. The Vendian–Cambrian sediments overlap the northeastern contact of the intrusion. The intrusion is composed mostly of coarse-grained weakly porphyritic biotite–amphibole and amphibole–biotite granodiorites and granosyenites. These rocks often contain small (2–3 cm) and larger (up to 5–8 cm) melanocratic inclusions enriched in amphibole and biotite. Melanocratic pyroxene–biotite–amphibole monzodiorites form bodies up to a few meters in visible size and are similar in composition to the melanocratic inclusions. The vein facies is represented by biotite granites, leucogranites, and, less often, granodiorites, which form a few meters thick cutting bodies.

Granodiorites and granosyenites of the major phase contain large microcline–perthite grains (20–25%) and have a medium-grained groundmass; their equigranular varieties are scarcer. Biotite (10–15%) prevails among dark-colored minerals, the content of amphibole is 1–10%, and single clinopyroxene grains are represented by diopside. The porphyritic granodiorites and granosyenites are characterized by a nest-like distribution of dark-colored minerals in assemblage with apatite, ilmenite, magnetite, and sphene; the content of accessory minerals is lower than that in the monzodiorites. Amphibole is represented by edenite and hastingsite; ferroan biotite is rich in TiO_2 (3.1–4.8 wt.%). Plagioclase is weakly zoned (An_{31-38}). The secondary minerals are secondary amphibole, epidote, and chlorite. Monzodiorites from large melanocratic inclusions contain porphyroblasts of K-feldspar (microcline) with plagioclase inclusions. Net-like clusters of amphibole and biotite (20–25%) are accompanied by coarse grains of apatite, ilmenite, magnetite, sphene, zircon, and, rarely, magnetite; the content of accessory minerals reaches 2–3%. Clinopyroxene (augite) occurs only as relics in amphibole. Amphibole corresponds in composition to edenite and hastingsite, and ferroan biotite is rich in TiO_2 (2.9–3.8 wt.%). The groundmass of the monzodiorites is composed of plagioclase, microcline, and quartz; the content of the latter does not exceed 5%. Plagioclase of the monzodiorites is strongly zoned, from $An_{40–42}$ in the core to $An_{29–36}$ in the rim. Biotite granites and leucogranites from vein bodies have a medium equigranular texture. The content of ferroan biotite (TiO_2 = 2.6–2.8 wt.%) is no more than 10%; granodiorites bear single amphibole grains. K-feldspar and plagioclase ($An_{26–33}$) are present in the rocks in approximately equal proportions. The main accessory minerals of the granites are apatite, zircon, and magnetite.

The Nizhnii Kitoi pluton is a small intrusion located to the northwest from the Toisuk pluton, in the northwest of the Irkut terrane (Fig. 1). It is formed by coarse- to medium-grained porphyritic amphibole–biotite granodiorites and granites. Granodiorites and granites are composed of microcline, plagioclase, and quartz; the total content of amphibole and biotite is 10–15%, with the latter prevailing. The accessory minerals are apatite, zircon, sphene, ilmenite, and magnetite. At the southwestern exocontact of the pluton, thin (up to a few meters) vein bodies of fine-grained biotite granites are widespread among the host Archean gneiss-granites.

The Malaya Belaya pluton is located in the northeast of the Onot terrane at the boundary with a trough made by metavolcanosedimentary deposits of the Paleoproterozoic Subluk Group (Fig. 1). The pluton has tectonic contacts, which is determined by its position in the zone of the regional fault bounding the Sharyzhalgai uplift. The pluton is hosted by the rocks of the upper section of the Onot greenstone belt: garnet–muscovite–biotite and biotite (± amphibole) schists with interbeds of amphibolites metamorphosed up to the amphibolite facies. In composition the predominant muscovite-biotite garnet schists correspond to terrigenous rocks from aleuropelites to quartz-rich sandstones. The Malaya Belaya intrusion is composed of leucocratic biotite- and amphibole-bearing granites. All rocks are gneissic and have thin-banded textures. The total content of dark-colored minerals—high-Fe biotite (TiO_2 = 2.5–3.2 wt.%) and amphibole (hastingsite)—does not exceed 10%; microcline prevails among feldspars. The accessory minerals are abundant zircon and sphene and scarcer apatite, allanite, and fluorite.

RESULTS OF U–Pb ZIRCON DATING

In **the Toisuk pluton,** zircons from clinopyroxene–biotite–amphibole monzodiorite (sample 63-15) and vein biotite leucogranite (sample 47-15) were dated. In monzodiorite, zircons are long-prismatic crystals 200–400 μm in length, with the length/width ratio (*K*) of 4–2. The cathodoluminescence (CL) images of most crystals show a clear oscillatory zoning; the inner parts of the grains are sometimes unzoned and dark in CL (Fig. 2*A*). In contrast to the zoned crystals (U = 88–376 ppm, Th = 121–248 ppm, Th/U = 0.7–1.5), the grain cores dark in CL are enriched in U (2244–2751 ppm) and Th (934–1453 ppm), but their Th/U ratio (0.43–0.55) (Table 1) is typical of magmatic zircons. The concordant age was calculated by 13 points corresponding to zircons with different U contents and is 1837 ± 6 Ma (MSWD = 0.53) (Fig. 2*A*). In the biotite granite, prismatic crystals of magmatic zircon (100–300 μm, *K* = 3–2) show a clear oscillatory zoning in CL images (Fig. 2*B*). There are also zircon grains consisting of large cores with clear or weak zoning and thin dark and light rims. The magmatic zircons (U = 124–559 ppm, Th = 107–227 ppm, Th/U = 0.42–1.0) and inherited cores (U = 169–315 ppm, Th = 89–297 ppm, Th/U = 0.29–1.3) show U and Th contents and Th/U ratio typical of magmatic zircons. The concordant age of five grains of magmatic zircons is 1827 ± 9 Ma (MSWD = 0.06) (Fig. 2*B*) and does not differ (within the error limits) from that of zircon from monzodiorite. The zircon cores have $^{207}Pb/^{206}Pb$ ages of 2.40–2.55 and 2.80–2.86 Ga (Table 1), which are close to the age of zircons from Neoarchean granites (2.55 Ga) (Turkina et al., 2012) and of the youngest detrital zircons from Archean paragneisses of the Irkut terrane (Turkina et al., 2017).

Zircons from amphibole–biotite granodiorite of **the Nizhnii Kitoi pluton** (sample 43-15) are prismatic crystals 200–400 μm in length, with *K* = 4–2 (Fig. 2*C*). In CL images, these crystals show a clear oscillatory zoning, but their interparts are dark in CL and show a “blurred” zoning. Most grains have low contents of U (114–458 ppm) and Th (68–304 ppm) and a Th/U ratio typical of magmatic zircon (0.48–1.0) (Table 1). The concordant age calculated by 12 points is 1846 ± 7 Ma (MSWD = 0.03) (Fig. 2*C*).

Zircons from leucogranite of **the Malaya Belaya pluton** (sample 43-95) are usually dark in CL and show a clearer zoning only in the rims (Fig. 2*D*). The contents of U (493–652 ppm) and Th (225–346 ppm) and the Th/U ratios (0.47–0.61) correspond to those of magmatic zircons (Table 1). Three grains are enriched in U (1169–7863 ppm) and have highly discordant U/Pb isotope ratios. The age of six zircon grains at the upper intersept of discordia with concordia is 1859 ± 28 Ma (MSWD = 0.81) (Fig. 2*D*). For the entire set of grains, the upper intercept of discordia with concordia is 1863 ± 16 Ma (MSWD = 0.52), this value is taken as the zircon age of the Malaya Belaya pluton.

GEOCHEMISTRY AND Nd ISOTOPE COMPOSITION OF GRANITOIDS

Rocks of **the Toisuk pluton** are characterized by a wide range of SiO_2 contents and are divided into three discrete groups: monzodiorites (55–57%), granodiorite–granosyenites (62–68%,) and granite–leucogranites (71–74%) (Table 2, Fig. 3*A*). They belong to subalkalic series and are predominantly calc-alkalic in terms of the MALI (Na_2O + K_2O–CaO) values (Fig. 3*C*) according to the classification by Frost et al. (2001). The rocks vary from metaluminous to weakly peraluminous, with ASI increasing from monzodiorites (0.83–1.04) to granodiorite–granosyenites (0.91–1.07)

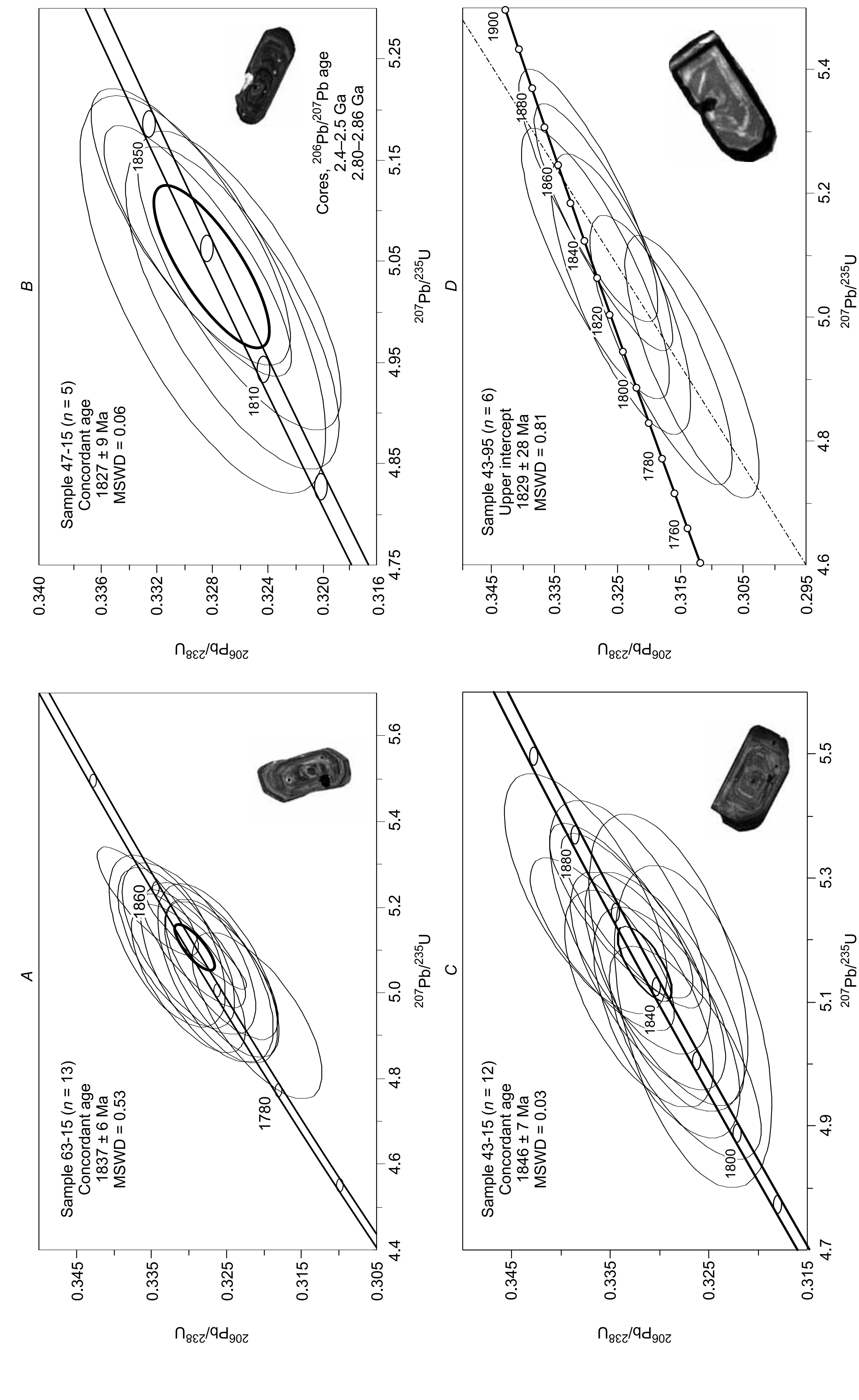

Fig. 2. Concordia diagrams for zircon from Paleoproterozoic granitoids. Toisuk pluton: *A*, monzodiorite, *B*, granite; *C*, Nizhnii Kitoi pluton; *D*, Malaya Belaya pluton.

Table 1. U–Pb isotope data for zircons from Paleoproterozoic granitoids

Spot	% $^{206}Pb_c$	U	Th	$^{232}Th/^{238}U$	$^{206}Pb^*$, ppm	Age, Ma		*D*, %	$^{238}U/^{206}Pb^*$	± %	$^{207}Pb^*/^{206}Pb^*$	± %	$^{207}Pb^*/^{235}U$	± %	$^{206}Pb^*/^{238}U$	± %	Rho
		ppm				$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$										
Monzodiorite, Toisuk pluton (sample 63–15)																	
1	0.19	188	149	0.81	53.6	1839 ± 15	1809 ± 18	–2	3.028	0.9	0.1106	1.0	5.033	1.4	0.3301	0.9	0.683
2	0.14	376	248	0.68	107	1836 ± 14	1838 ± 12	0	3.035	0.9	0.1124	0.7	5.104	1.1	0.3294	0.9	0.792
3	0.51	237	193	0.84	68.2	1853 ± 15	1825 ± 23	–1	3.001	1.0	0.1116	1.3	5.121	1.6	0.3329	1.0	0.601
4	0.12	2751	1453	0.55	794	1865 ± 13	1843 ± 10	–1	2.979	0.8	0.1127	0.6	5.214	1.0	0.3355	0.8	0.833
5	0.17	2244	934	0.43	636	1834 ± 13	1846 ± 6	1	3.038	0.8	0.1129	0.3	5.120	0.9	0.3290	0.8	0.932
6	0.06	180	135	0.77	50.5	1817 ± 15	1834 ± 20	1	3.071	1.0	0.1121	1.1	5.034	1.5	0.3256	1.0	0.664
7	2.68	272	206	0.78	80.3	1851 ± 17	1806 ± 73	–2	2.990	1.1	0.1104	4.0	5.060	4.2	0.3326	1.1	0.258
8	0.12	118	101	0.88	33.4	1839 ± 17	1835 ± 21	0	3.029	1.1	0.1122	1.2	5.103	1.6	0.3300	1.1	0.671
9	0.13	182	196	1.11	51.1	1820 ± 16	1835 ± 24	1	3.064	1.0	0.1122	1.3	5.045	1.7	0.3262	1.0	0.614
10	0.16	141	121	0.88	39	1794 ± 17	1829 ± 20	2	3.116	1.1	0.1118	1.1	4.947	1.6	0.3209	1.1	0.702
11	0.13	142	131	0.95	40.6	1847 ± 15	1826 ± 20	–1	3.014	1.0	0.1116	1.1	5.105	1.5	0.3317	1.0	0.661
12	0.12	88	125	1.47	24.7	1827 ± 17	1828 ± 25	0	3.052	1.1	0.1118	1.4	5.048	1.7	0.3276	1.1	0.619
13	0.25	221	175	0.82	61.9	1817 ± 15	1841 ± 17	1	3.070	0.9	0.1125	0.9	5.052	1.3	0.3256	0.9	0.711
Granite, Toisuk pluton (sample 47–15)																	
1.1	0.07	301	202	0.69	146	2882 ± 21	2864 ± 7	–1	1.774	0.9	0.2047	0.4	15.91	1.0	0.5636	0.9	0.901
2.1	0.07	340	216	0.66	96.4	1835 ± 14	1831 ± 12	0	3.035	0.9	0.1119	0.7	5.084	1.1	0.3294	0.9	0.791
3.1	0.06	201	134	0.69	56.5	1821 ± 15	1819 ± 16	0	3.063	1.0	0.1112	0.9	5.004	1.3	0.3264	1.0	0.732
4.1	0.08	124	107	0.89	35	1831 ± 18	1812 ± 21	–1	3.044	1.1	0.1108	1.1	5.017	1.6	0.3285	1.1	0.708
5.1	0.10	559	227	0.42	158	1833 ± 13	1825 ± 10	0	3.04	0.8	0.1116	0.6	5.060	1.0	0.3289	0.8	0.838
6.1	0.09	281	167	0.61	78.7	1819 ± 15	1831 ± 14	1	3.068	0.9	0.1120	0.8	5.031	1.2	0.3259	0.9	0.757
7.1	0.22	169	123	0.75	73.8	2643 ± 21	2807 ± 11	6	1.971	1.0	0.1976	0.7	13.81	1.2	0.5069	1.0	0.827
8.1	0.28	178	172	1.00	51.7	1874 ± 16	1866 ± 22	0	2.963	1.0	0.1141	1.2	5.308	1.6	0.3373	1.0	0.640
9.1	0.04	230	297	1.33	95.9	2552 ± 20	2549 ± 10	0	2.059	0.9	0.1691	0.6	11.32	1.1	0.4856	0.9	0.850
9.2	0.10	315	89	0.29	120	2359 ± 22	2428 ± 9	3	2.263	1.1	0.1574	0.5	9.590	1.3	0.4418	1.1	0.904
Granodiorite, Nizhnii Kitoi pluton (sample 43–15)																	
1.1	0.19	136	79	0.60	39	1853 ± 17	1859 ± 22	0	3.001	1.0	0.1137	1.2	5.221	1.6	0.3330	1.0	0.648
2.1	1.34	165	158	0.99	43.6	1708 ± 15	1778 ± 37	4	3.288	1.0	0.1087	2.0	4.550	2.3	0.3033	1.0	0.455
3.1	0.27	151	78	0.54	42.4	1822 ± 16	1838 ± 33	1	3.060	1.0	0.1124	1.8	5.060	2.1	0.3266	1.0	0.499
4.1	8.44	433	304	0.72	128	1738 ± 15	1822 ± 98	5	3.171	1.0	0.1114	5.4	4.750	5.5	0.3094	1.0	0.175
4.2	0.75	119	68	0.59	34	1834 ± 17	1861 ± 29	1	3.034	1.0	0.1138	1.6	5.163	1.9	0.3291	1.0	0.541
5.1	0.06	395	273	0.72	113	1857 ± 14	1864 ± 12	0	2.995	0.9	0.1140	0.7	5.247	1.1	0.3339	0.9	0.804
5.2	0.13	167	78	0.48	47.5	1844 ± 15	1839 ± 18	0	3.018	1.0	0.1124	1.0	5.134	1.4	0.3312	1.0	0.697
6.1	0.04	283	156	0.57	81.7	1867 ± 14	1834 ± 13	–2	2.977	0.9	0.1122	0.7	5.193	1.1	0.3358	0.9	0.783
6.2	0.14	114	69	0.63	32.2	1836 ± 17	1844 ± 19	0	3.034	1.0	0.1127	1.1	5.121	1.5	0.3295	1.0	0.705
7.1	0.14	140	87	0.64	40.5	1875 ± 16	1854 ± 20	–1	2.962	1.0	0.1133	1.1	5.274	1.5	0.3375	1.0	0.678
8.1	0.47	213	130	0.63	61.2	1849 ± 15	1824 ± 20	–1	3.008	0.9	0.1115	1.1	5.105	1.4	0.3321	0.9	0.645
9.1	0.08	175	139	0.82	49.3	1826 ± 15	1819 ± 18	0	3.054	1.0	0.1112	1.0	5.018	1.4	0.3274	1.0	0.705
9.2	0.65	248	167	0.70	70.7	1835 ± 15	1826 ± 21	–1	3.033	0.9	0.1116	1.1	5.067	1.5	0.3293	0.9	0.628
10.1	0.03	458	214	0.48	131	1852 ± 14	1864 ± 11	1	3.004	0.9	0.1140	0.6	5.231	1.1	0.3328	0.9	0.836
Leucogranite, Malaya Belaya pluton (sample 43–95)																	
1	0.02	587	339	0.60	165	1825 ± 22	1865 ± 14	2	3.056	1.4	0.1140	0.8	5.145	1.6	0.3272	1.4	0.864
2	6.33	7153	502	0.07	318	303 ± 4	946 ± 87	212	20.64	1.3	0.0706	4.3	0.469	4.5	0.0482	1.3	0.296
3	0.11	586	299	0.53	158	1756 ± 21	1863 ± 20	6	3.193	1.4	0.1139	1.1	4.92	1.8	0.3131	1.4	0.789

(*continued on next page*)

Table 1 (*continued*)

Spot	% $^{206}Pb_c$	U, ppm	Th, ppm	$^{232}Th/^{238}U$	$^{206}Pb^*$, ppm	Age, Ma: $^{206}Pb/^{238}U$	Age, Ma: $^{207}Pb/^{206}Pb$	*D*, %	$^{238}U/^{206}Pb^*$	± %	$^{207}Pb^*/^{206}Pb^*$	± %	$^{207}Pb^*/^{235}U$	± %	$^{206}Pb^*/^{238}U$	± %	Rho
4	0.94	648	295	0.47	148	1502 ± 19	1811 ± 29	21	3.803	1.4	0.1107	1.6	4.006	2.1	0.2625	1.4	0.657
5	0.08	652	346	0.55	185	1836 ± 22	1870 ± 15	2	3.034	1.4	0.1144	0.84	5.196	1.6	0.3296	1.4	0.852
6	0.09	504	239	0.49	142	1831 ± 23	1837 ± 17	0	3.044	1.5	0.1123	0.97	5.086	1.8	0.3285	1.5	0.835
7	0.13	549	325	0.61	150	1780 ± 23	1845 ± 19	4	3.143	1.5	0.1128	1.1	4.947	1.8	0.3181	1.5	0.809
8	3.76	1169	669	0.59	171	975 ± 13	1582 ± 70	62	6.084	1.4	0.0978	3.8	2.201	4	0.1633	1.4	0.354
9	0.23	493	225	0.47	138	1811 ± 22	1842 ± 22	2	3.081	1.4	0.1126	1.2	5.037	1.8	0.3245	1.4	0.763
10	1.58	7863	1060	0.14	464	422 ± 5	600 ± 45	42	14.79	1.3	0.0599	2.1	0.558	2.5	0.0676	1.3	0.534

Note. The errors are given at the 1σ level. Pbc and Pb* indicate the common and radiogenic portions of lead, respectively. Correction for common lead was made using the measured ^{204}Pb. The error of TEMORA standard calibration is 0.36%. *D*, discordance (%) calculated by the equation $D = 100 \times [(^{206}Pb/^{207}Pb\ age/^{206}Pb/^{238}U\ age) - 1]$. Rho, error correlation between $^{207}Pb^*/^{235}U$ and $^{206}Pb^*/^{238}U$ values.

and granites (1.00–1.09). All granitoids are highly ferroan: f (FeO*/(FeO* + MgO)) = 0.77–0.91; the f value increases with the SiO_2 content (Fig. 3*B*). Monzodiorites are highly enriched in TiO_2 (1.5–2.0 wt.%) and P_2O_5 (0.68–0.86 wt.%), whose contents slightly increase with the SiO_2 content, whereas the contents of TiO_2 and P_2O_5 decrease in the series of granodiorites, granosyenites, and granites. A characteristic feature of the rocks of major phase and monzodiorites is extremely high contents of Ba (4100–1200 ppm) and enrichment in HFSE (Zr, Hf, Nb, Ta, and Y), especially Zr (Table 2). All rocks, from monzodiorites to granodiorites, are divided into two groups according to the contents of REE and trace elements (Fig. 4*A*, *B*). The rocks of the first group (I) are enriched in LREE ($(La/Yb)_n$ = 31–48), Ba

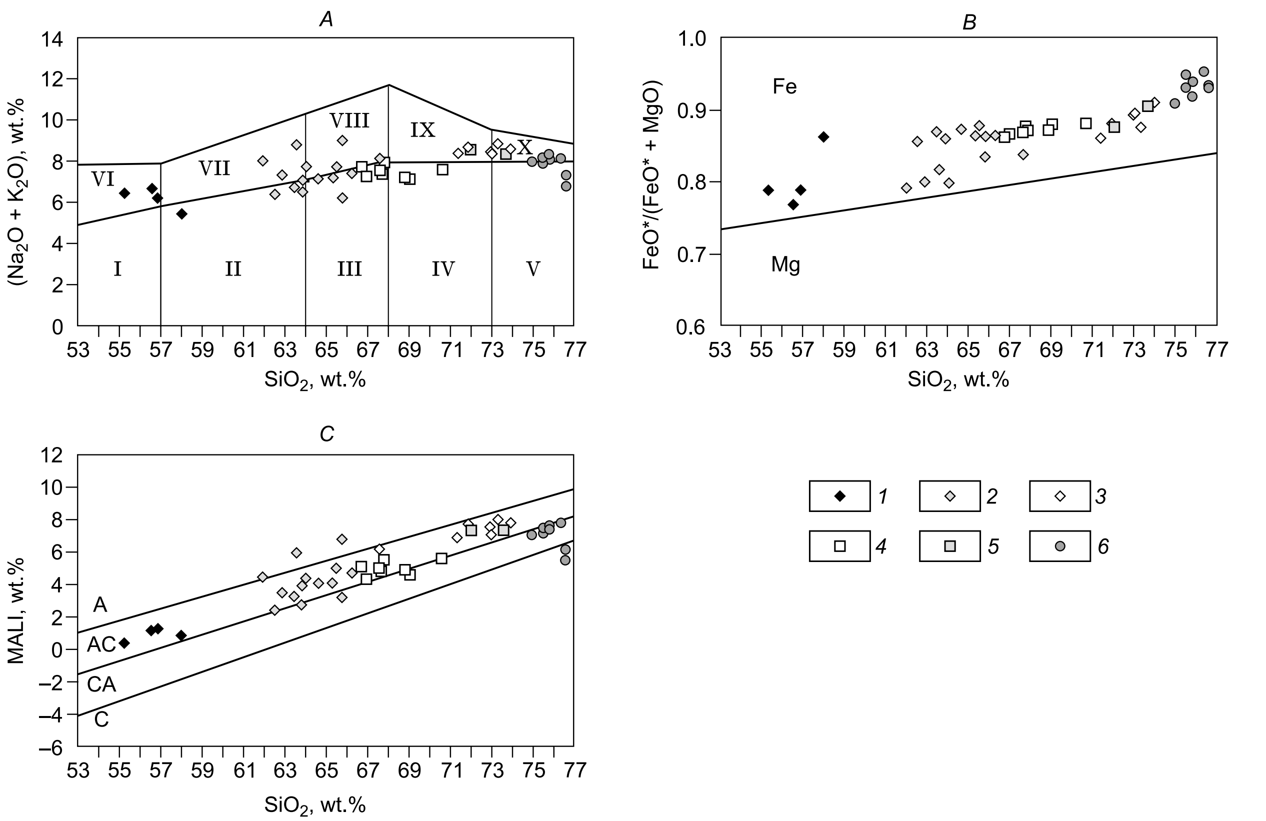

Fig. 3. SiO_2–(Na_2O + K_2O) (*A*), SiO_2–FeO*/(FeO* + MgO) (*B*), and SiO_2–MALI (*C*) diagrams for Paleoproterozoic granitoids. *1–3*, Toisuk pluton: *1*, monzodiorites, *2*, granodiorite–granosyenites, *3*, vein granites; *4–5*, Nizhnii Kitoi pluton: *4*, granodiorite–granites, *5*, vein granites; *6*, Malaya Belaya pluton. *A*, Fields of: I, diorites; II, quartz diorites; III, granodiorites; IV, granites; V, leucogranites; VI, monzodiorites; VII, quartz monzodiorites; VIII, quartz syenites; IX, subalkalic granites; X, subalkalic leucogranites; *B*, fields of: Fe, ferroan and Mg, magnesian granitoids (Frost et al., 2001); *C*, fields of: C, calcic; AC, alkali-calcic; CA, calc-alkalic; A, alkali granitoids (Frost et al., 2001).

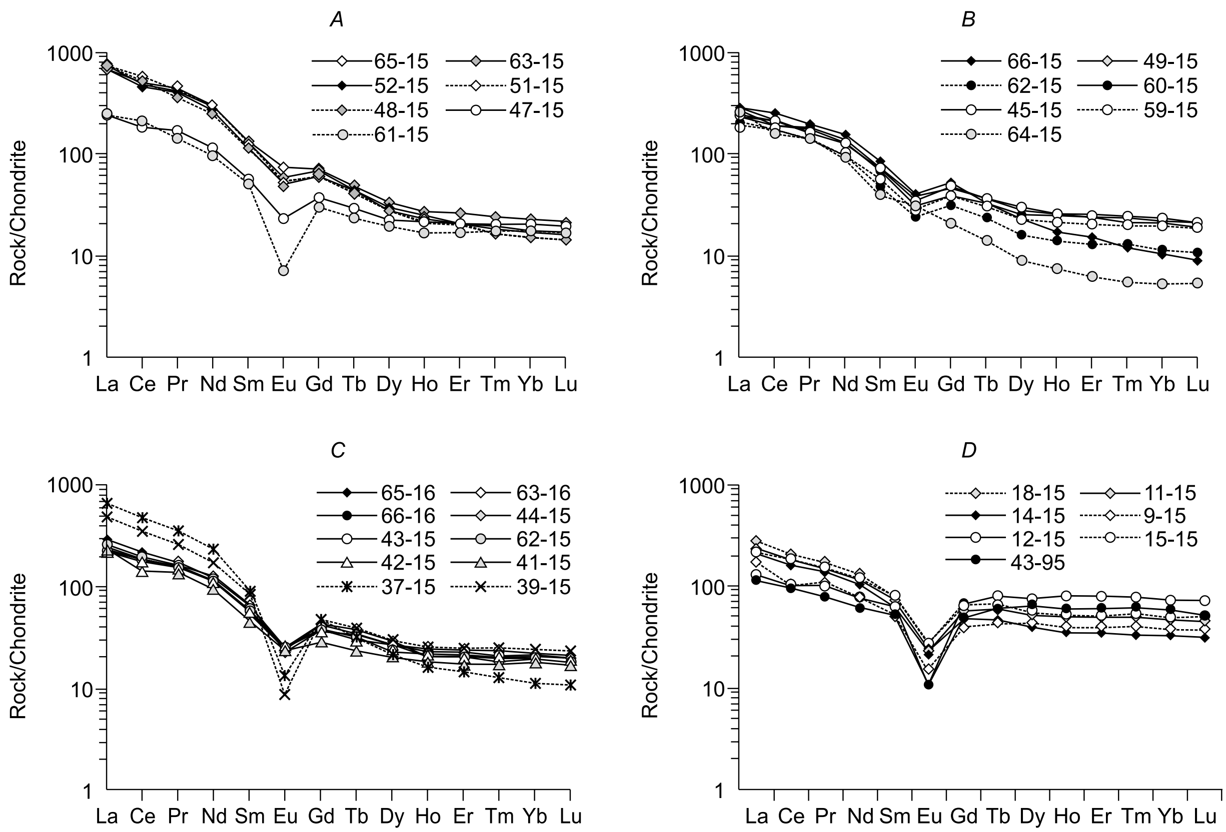

Fig. 4. REE patterns for Paleoproterozoic granitoids. Plutons: *A*, *B*, Toisuk: *A*, group I (monzodiorites, quartz monzonites, and granodiorites) and vein granites, *B*, group II (mopnzodiorite and granodiorites); *C*, Nizhnii Kitoi (dotted lines mark vein granites); *D*, Malaya Belaya. Sample numbers follow Table 2.

(2900–4100 ppm), Sr (1550–796 ppm), and, to a lesser extent, Zr (316–745 ppm) and Th (16–38 ppm). The rocks of the second group (II) are depleted in LREE ($(La/Yb)_n$ = 9–27), Ba (1200–2100 ppm), and Sr (270–738 ppm) relative to the group I rocks, having similar contents of Zr (354–645 ppm) and Th (12–22 ppm). One sample from this group has low contents of HREE and lacks a Eu anomaly (Fig. 4*B*). The REE patterns of the rocks of both groups show a weak negative Eu anomaly (Eu/Eu* = 0.73–0.56). Vein granitoids are diverse in REE composition, varying from highly HREE-depleted rocks with a high $(La/Yb)_n$ ratio (36) and a weak negative Eu anomaly (Eu/Eu* = 0.74) to moderately fractionated ones ($(La/Yb)_n$ = 12–14) with a clear negative Eu anomaly (Eu/Eu* = 0.18–0.50) (Fig. 4*A*). The contents of incompatible trace elements in the vein granitoids are lower as compared with the rocks of major phase. The multi-element patterns of the monzodiorites and granodiorites of major phase show strong negative Ti and Nb anomalies; in addition, the granodiorites show negative Sr and P anomalies (Fig. 5*A*, *B*). Granites demonstrate similar anomalies on multielement patterns (Fig. 5*A*).

The rocks of the Toisuk pluton are characterized by negative ε_{Nd} values; monzodiorites and granodiorites of group I have the lowest values (–10.2 to –7.2). The depleted rocks of group II and vein granites show noticeably higher ε_{Nd} values (–4.6 to –5.5) (Table 3, Fig. 6*A*).

Rocks of **the Nizhnii Kitoi pluton** have a narrower range of SiO_2 contents (66.7–69.0 wt.%). These are metaluminous subalkalic granodiorites and granites (Fig. 3*A*). According to the MALI ($Na_2O + K_2O–CaO$) values, the granitoids are of calc-alkalic type (Fig. 3*C*). The vein facies corresponds in composition to subalkalic granites. All granitoids are highly ferroan (f = 0.82–0.91) (Fig. 3*B*) and moderately enriched in TiO_2 (0.6–0.82%) and P_2O_5 (0.2–0.28%). The major-phase rocks are characterized by a high Ba content (1560–990 ppm), decreasing in vein granites (400–700 ppm). In contents of REE, HFSE, Th, Rb, and Ba the granodiorites and granites are comparable with the group II rocks of the Toisuk pluton but are poorer in Sr (204–331 ppm). The REE patterns of the granitoids are moderately fractionated ($(La/Yb)_n$ = 10–14) and show a weak negative Eu anomaly (Eu/Eu* = 0.49–0.63) (Fig. 4*C*). The vein granites and leucogranites are enriched in LREE and show high $(La/Yb)_n$ values (20–60) and a strong negative Eu anomaly (Eu/Eu* = 0.13–0.19). Multielement patterns of granitoids of major phase demonstrate strong minima of Nb and Ti and weaker negative Sr and P anomalies (Fig. 5*C*). Two samples (granodiorite and granite) from the Nizhnii Kitoi pluton are characterized by ε_{Nd} = –4.0 and –5.1 and T_{Nd}(DM) = 2.6 and 2.7 Ga (Table 3). Their points lie above the field of the isotope composition of the Archean crust of the Irkut and Kitoi terranes (Fig. 6*A*).

Table 2. Chemical composition of Paleoproterozoic granitoids of the Sharyzhalgai uplift

Component	65-15	63-15	52-15	51-15	48-15	66-15	49-15	62-15	45-15	60-15	59-15	64-15	47-15	61-15	65-16	63-16
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
SiO_2, wt.%	55.29	56.54	61.96	62.86	64.04	56.84	62.52	63.57	63.84	63.88	65.31	65.81	73.01	73.93	66.73	66.96
TiO_2	1.61	1.75	1.10	1.11	0.91	1.45	1.39	0.94	1.35	1.20	1.07	0.56	0.20	0.19	0.81	0.82
Al_2O_3	17.03	15.21	15.13	14.84	14.57	17.43	13.92	16.30	13.41	13.58	13.69	16.23	13.21	13.00	13.88	13.85
Fe_2O_3*	9.20	9.94	6.89	7.10	6.36	9.29	9.25	5.41	9.03	8.21	7.48	4.09	2.30	2.33	5.80	6.07
MnO	0.13	0.12	0.08	0.08	0.08	0.09	0.12	0.05	0.12	0.11	0.11	0.03	0.02	0.03	0.08	0.08
MgO	2.22	2.68	1.63	1.59	1.44	2.25	1.40	1.08	1.33	1.21	1.06	0.73	0.24	0.20	0.84	0.84
CaO	6.08	5.53	3.62	3.85	3.33	4.91	4.00	2.87	3.75	3.16	3.14	2.18	1.34	0.78	2.60	3.00
Na_2O	3.38	3.14	2.88	3.13	3.08	3.48	2.62	3.09	2.69	2.64	2.68	2.99	2.54	3.12	2.83	2.93
K_2O	3.09	3.53	5.17	4.17	4.63	2.72	3.78	5.72	3.79	4.45	4.54	6.03	5.85	5.47	4.91	4.33
P_2O_5	0.67	0.68	0.39	0.44	0.36	0.68	0.51	0.39	0.49	0.43	0.42	0.20	0.05	0.05	0.26	0.28
LOI	0.39	0.52	0.47	0.57	0.50	0.60	0.40	0.64	0.30	0.70	0.50	0.78	0.36	0.47	0.28	0.29
Total	99.65	100.1	99.84	100.2	99.68	99.97	100.2	100.4	100.3	99.81	100.22	100.10	99.24	99.65	99.22	99.63
Th, ppm	16.2	25.0	27.0	38.0	38	12.1	13.9	16.6	17.1	15.2	13.2	21.0	22	51	29	17.4
U	1.9	2.1	2.9	4.8	3.3	1.87	1.34	2.1	2.3	2.6	2.2	1.6	3.0	2.4	4.4	3.4
Rb	79	91	131	105	150	168	93	198	132	189	128	157	183	297	193	170
Ba	4085	3133	4009	3062	2 908	1 343	1624	2093	1 549	1504	1665	3310	1 186	318	1 272	987
Sr	1548	1165	900	853	796	738	343	296	282	254	270	516	206	64	226	221
La	209	215	239	240	230	87	73	64	75	85	56	81	76	79	90	69
Ce	378	424	387	461	423	203	154	135	171	153	144	127	153	171	173	148
Pr	49.1	54.3	50.4	54.8	45	23	22.4	17.6	19.9	21.4	17.0	16.9	20	17.7	21	19.6
Nd	175	180	167	182	151	91	81	56	77	75	61.2	55.2	68	57	72	75
Sm	25.3	26.0	22.4	24.1	22	16.4	14.0	9.4	14.1	13.5	11.1	7.6	10.8	9.9	12.3	12.9
Eu	5.3	4.2	3.7	4.0	3.7	3.0	2.8	1.8	2.5	2.2	2.1	2.3	1.7	0.52	1.8	1.8
Gd	18.3	18.1	15.7	16.5	16.2	13.2	11.9	8.0	12.5	10.2	9.8	5.4	9.7	7.7	10.5	11.3
Tb	2.2	2.3	2.0	2.0	1.90	1.61	1.7	1.1	1.70	1.6	1.4	0.7	1.39	1.10	1.6	1.8
Dy	9.4	10.8	9.0	9.2	8.9	7.4	9.0	5.1	9.6	8.2	7.5	2.8	7.3	6.2	8.6	9.2
Ho	1.8	1.9	1.7	1.7	1.5	1.2	1.8	1.0	1.8	1.8	1.5	0.5	1.6	1.2	1.73	1.84
Er	4.4	5.5	4.1	4.4	4.2	3.1	5.0	2.7	5.1	4.9	4.3	1.3	4.3	3.6	4.8	5.1
Tm	0.60	0.78	0.63	0.63	0.54	0.40	0.69	0.40	0.77	0.75	0.66	0.18	0.66	0.57	0.68	0.76
Yb	3.70	4.72	3.49	3.66	3.3	2.2	4.29	2.33	4.8	4.60	4.05	1.10	4.3	3.7	4.4	4.5
Lu	0.54	0.68	0.50	0.51	0.47	0.29	0.63	0.34	0.66	0.66	0.60	0.17	0.63	0.54	0.63	0.65
Zr	316	745	570	528	505	472	388	645	401	354	354	465	302	186	333	321
Hf	7.6	17.1	13.4	12.5	11.5	9.7	9.5	15.5	9.5	9.1	8.8	11.3	8.0	5.7	9.0	8.9
Ta	0.99	1.28	0.82	1.02	1.04	0.81	1.07	0.85	1.27	1.33	1.08	0.31	1.34	1.61	1.26	1.10
Nb	20	25	18	17.6	17.7	18.4	19.1	15	21	21	17.6	7.5	17.0	15.6	17.3	16.3
Y	46	56	45	46	44	38	47	30	52	46	42	14	46	36	46	51
Cr	39	75	65	50	39	39	30	47	33	37	37	52	56	39	13.8	10.4
Ni	12.7	14.3	12.5	14.4	6.1	12.7	10.0	11.5	8.7	11.0	10.0	10.6	10.6	6.1	0.48	0.45
$(La/Yb)_n$	38.1	30.6	46.0	44.2	47.6	27.0	11.5	18.5	10.6	12.5	9.4	49.5	11.9	14.4	13.8	10.4
Eu/Eu*	0.73	0.57	0.57	0.57	0.58	0.59	0.64	0.63	0.56	0.56	0.60	1.0	0.50	0.18	0.48	0.45
T, °C	754	825	833	827	828	817	798	863	801	797	799	844	811	769	801	795

(*continued on the next page*)

Component	66-16	44-15	43-15	62-16	42-15	41-15	37-15	39-15	18-15	11-15	14-15	9-15	12-15	15-15	16-15	43-95
	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32
SiO_2, wt.%	67.60	67.72	67.83	68.82	69.03	70.63	72.0	73.6	74.97	75.46	75.49	75.80	75.79	76.59	76.60	76.36
TiO_2	0.81	0.79	0.79	0.81	0.74	0.60	0.52	0.32	0.20	0.19	0.16	0.13	0.08	0.22	0.28	0.112
Al_2O_3	13.47	13.19	13.48	12.64	12.83	13.18	13.2	13.1	12.06	12.07	11.75	11.99	12.46	11.59	11.29	11.87
Fe_2O_3*	5.91	6.10	5.98	5.69	5.68	4.63	3.29	2.81	3.08	2.90	2.41	2.92	2.07	2.53	3.04	2.29
MnO	0.08	0.08	0.08	0.08	0.07	0.06	0.03	0.04	0.03	0.04	0.03	0.02	0.03	0.03	0.03	0.03
MgO	0.80	0.76	0.80	0.75	0.70	0.56	0.42	0.26	0.28	0.19	0.12	0.17	0.16	0.16	0.20	0.1
CaO	2.54	2.58	2.45	2.36	2.51	2.03	1.24	0.97	0.88	0.75	0.70	0.45	0.91	1.16	1.28	0.39
Na_2O	2.73	2.72	2.71	2.57	2.76	2.66	2.50	2.89	2.96	2.90	2.87	2.88	3.39	3.32	3.05	2.76
K_2O	4.80	4.68	5.25	4.64	4.37	4.98	6.05	5.45	4.99	5.00	5.29	5.21	4.93	3.99	3.74	5.4
P_2O_5	0.27	0.27	0.27	0.26	0.25	0.20	0.09	0.06	0.02	0.03	0.04	0.02	0.02	0.02	0.02	0.03
LOI	0.18	0.62	0.43	0.35	0.40	0.22	0.66	0.55	0.35	0.50	0.26	0.49	0.00	0.13	0.27	0.58
Total	99.40	99.70	100.29	99.17	99.55	99.96	100.1	100.2	99.93	100.17	99.24	100.18	99.95	99.88	99.94	99.92
Th, ppm	28	22.0	23.0	42	29.0	19.6	66.0	72.0	27	42	24	43	52	30	35	60
U	2.8	2.9	3.5	1.9	3	3.8	3.2	7.9	3.8	6.3	3.4	6.4	12.6	3.9	3.9	7.4
Rb	186	182	202	69	181	215	268	358	176	194	200	202	191	114	–	221
Ba	1 233	1175	1561	368	1184	1433	720	429	702	699	844	507	270	815	–	215
Sr	211	204	219	331	215	208	97	67	39	39	43	28	24	60	–	16
La	80	75	68	114	69	70	207	147	86.2	73	64.8	53	39.7	67.9	–	37
Ce	160	152	145	194	140	114	390	287	165.5	147	132.9	81	82.6	153.8	–	81
Pr	19.3	19.9	18.7	19.4	18.2	16.7	44	32	20.9	18.3	17.1	13.1	12.3	18.5	–	9.8
Nd	69	68	65	56	67	56	139	105	80.2	69	63.4	46	46.4	72.3	–	37
Sm	11.3	10.7	11.7	6.6	10.7	9.0	17.4	16.1	15.8	14.3	12.0	10.1	12.6	15.8	–	10.3
Eu	1.7	1.7	1.8	1.1	1.9	1.8	0.97	0.62	1.83	1.72	1.66	1.13	0.82	2.05	–	0.79
Gd	10.4	9.6	9.8	5.3	9.2	7.4	12.1	12.1	17.1	15.1	12.3	10.3	17.6	17.0	–	12.8
Tb	1.6	1.4	1.5	0.63	1.4	1.1	1.51	1.8	3.1	2.8	2.2	2.1	3.8	3.1	–	2.9
Dy	8.3	7.3	8.4	2.8	7.4	6.4	6.8	9.9	17.9	17.4	12.7	13.9	24.5	18.6	–	20.3
Ho	1.52	1.57	1.62	0.57	1.49	1.31	1.18	1.83	3.7	3.6	2.6	2.8	5.8	3.8	–	4.3
Er	4.4	4.3	4.7	1.46	4.2	3.7	3.1	5.2	10.8	10.5	7.4	8.4	17.0	10.8	–	12.8
Tm	0.64	0.65	0.66	0.21	0.63	0.57	0.40	0.81	1.7	1.63	1.1	1.32	2.5	1.7	–	2.02
Yb	3.9	4.3	4.3	1.40	4.0	3.7	2.3	5.0	10.3	10.0	7.0	8.0	15.7	10.6	–	12.3
Lu	0.57	0.62	0.63	0.21	0.59	0.54	0.34	0.73	1.6	1.47	1.0	1.24	2.4	1.6	–	1.64
Zr	330	299	324	402	311	290	495	290	243	270	162	147	155	303	–	150
Hf	8.8	8.00	8.50	11.3	8.50	8.00	14.7	9.7	9.5	9.8	6.4	7.8	9.6	11.6	–	8.8
Ta	1.18	1.3	1.2	0.32	1.2	1.1	1.0	2.6	2.0	1.93	1.3	1.48	2.0	2.4	–	3.4
Nb	16.2	16.9	16	5.0	15.4	13.5	19.4	27	24	26	18.1	26	28	27	–	33
Y	45	45	48	15.8	42	40	34	55	111	97	77	68	187	113	–	107
Cr	13.7	11.9	10.7	55.1	11.8	–	–	–	–	–	–	–	–	–	–	–
Ni	0.48	0.50	0.50	0.54	0.56	–	–	–	–	–	–	–	–	–	–	–
$(La/Yb)_n$	13.7	11.9	10.7	55.1	11.8	12.6	60.9	20.0	5.7	4.9	6.3	4.5	1.7	4.3	–	2.0
Eu/Eu*	0.48	0.50	0.50	0.54	0.56	0.63	0.19	0.13	0.34	0.36	0.41	0.34	0.17	0.38	–	0.21
T, °C	801	791	797	822	797	799	861	808	793	806	757	755	750	811	–	757

Note. 1–11, Toisuk pluton: group I: 1, 2, monzodiorites, 3–5, quartz monzonites and granodiorites; group II: 6, monzodiorite, 7–11, granodiorites; 12–14, vein granites; 15–24, Nizhnii Kitoi pluton (15–20, granodiorites and granites, 21–24, vein leucogranites); 25–32, Malaya Belaya pluton, leucogranites. Fe_2O_3*, total iron. Temperature was determined based on Zr saturation, after Watson and Harrison (1983). Dash, no data.

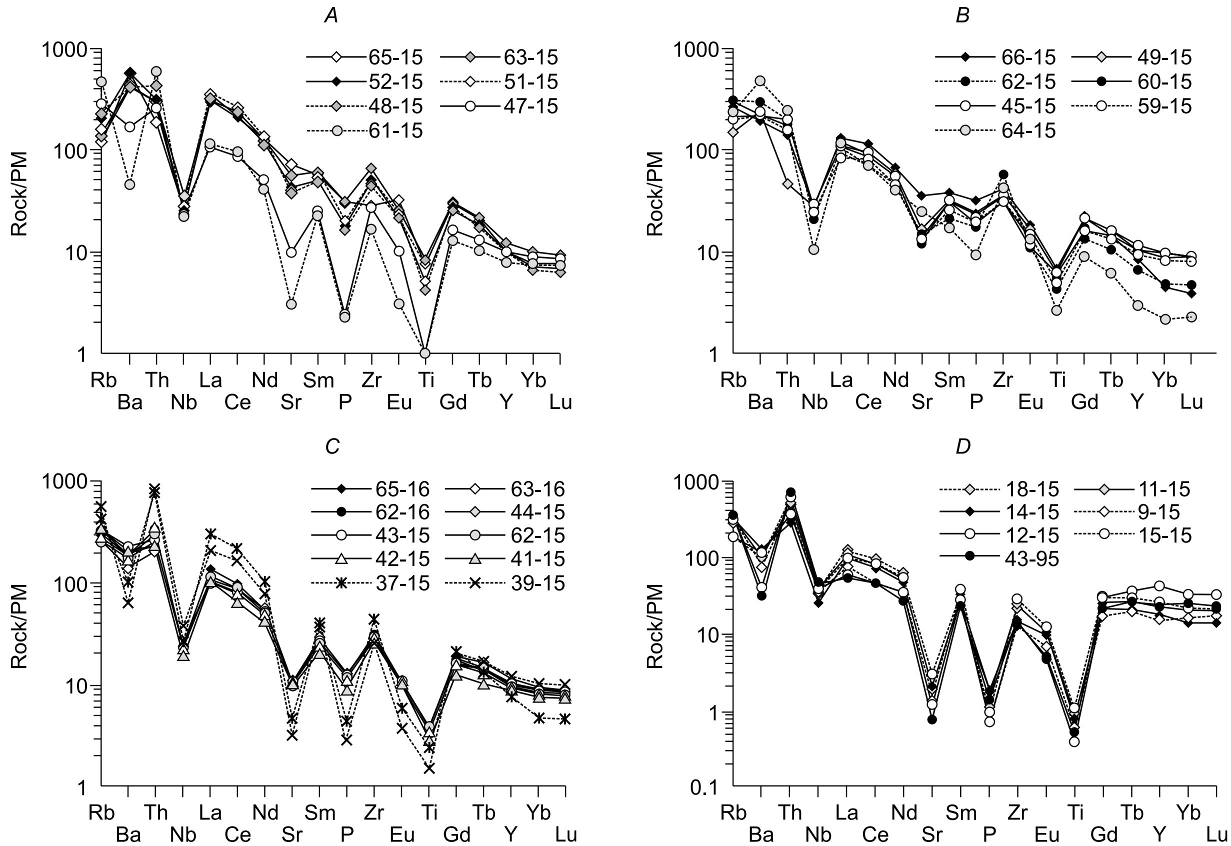

Fig. 5. Multielement patterns of Paleoproterozoic granitoids. Plutons: *A*, *B*, Toisuk: *A*, group I (monzodiorites, quartz monzonites, and granodiorites) and vein granites, *B*, group II (monzodiorite and granodiorites of major phase); *C*, Nizhnii Kitoi (dash lines mark vein granites); *D*, Malaya Belaya. Sample numbers follow Table 2.

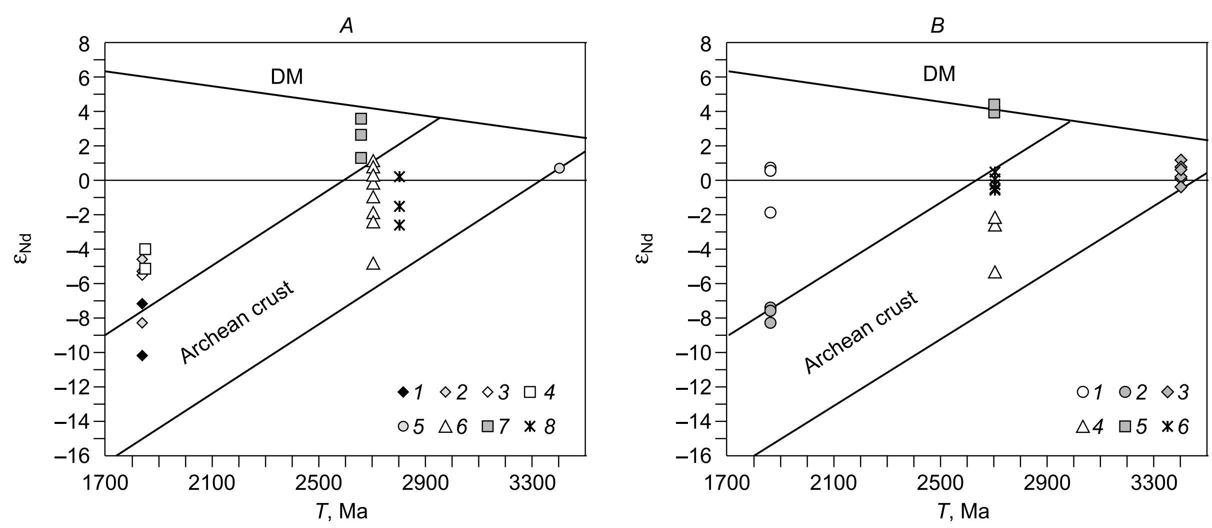

Fig. 6. T–ε_{Nd} diagrams for Paleoproterozoic granitoids and the host rocks. *A*, Irkut and Kitoi terranes: *1–3*, Toisuk pluton: *1*, monzodiorites, *2*, granodiorites, *3*, granite; *4*, Nizhnii Kitoi pluton; *5*, Paleoarchean granulite; *6*, *7*, Neoarchean granulites: *6*, intermediate–felsic, *7*, mafic; *8*, Archean metasedimentary gneisses; *B*, Onot terrane: *1*, Malaya Belaya pluton; *2*, Shumikha pluton; *3*, plagiogneisses of TTG complex; *4*, felsic gneisses; *5*, amphibolites; *6*, garnet–staurolite schists. Data for the host rocks are after Turkina et al. (2006, 2012, 2014b), Turkina and Nozhkin (2008), and Turkina and Sukhorukov (2015).

Table 3. Sm–Nd isotope data for Paleoproterozoic granitoids

No.	Sample	*T*, Ma	Nd	Sm	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	T_{Nd}(DM), Ma	$\varepsilon_{Nd}(T)$
			ppm					
1	65-15	1840	171.3	26.8	0.9438	0.510879±5	2885	–10.2
2	63-15	1840	177.4	26.4	0.0898	0.510978±13	2659	–7.2
3	51-15	1840	183.5	25.7	0.0847	0.510862±13	2691	–8.2
4	62-15	1840	58.3	10.0	0.1039	0.511246±10	2632	–5.3
5	64-15	1840	53.3	7.6	0.0859	0.511063±8	2480	–4.6
6	47-15	1840	39.5	5.7	0.0868	0.511028±16	2539	–5.5
7	43-15	1850	74.0	12.9	0.1051	0.511318±7	2561	–4.0
8	41-15	1850	56.8	9.9	0.1053	0.511264±7	2640	–5.1
9	14-15	1860	61.4	12.8	0.1259	0.511808±12	2293*	0.7
10	16-15	1860	108.3	24.6	0.1371	0.511937±11	2306*	0.6
11	12-15	1860	44.2	12.5	0.1711	0.512227±9	2508*	–1.9

Note. *T*, age of granitoids. 1–6, Toisuk pluton: 1, 2, monzodiorites, 3–5, granodiorites, 6, granite; 7, 8, Nizhnii Kitoi pluton; 9–11, Malaya Belaya pluton.
* Two-stage model Nd age.

The rocks of **the Malaya Belaya pluton** are leucogranites (SiO_2 = 75.0–76.6 wt.%) with the contents of alkalies ($K_2O + Na_2O$ = 6.8–8.3 wt.%) falling on the boundary of subalkalic rocks (Fig. 3*A*). The granites are weakly peraluminous (ASI = 0.98–1.07) and, in contrast to rocks of the other two plutons, are of alkali-calcic type according to the MALI values (Fig. 3*C*). A specific feature of the leucogranites is extremely high *f* values (FeO*/(FeO* + MgO) = 0.91–0.95) (Fig. 3*B*). The granites have low contents of Ba (840–215 ppm) and extremely low contents of Sr (16–60 ppm). Compared with the rocks of the Toisuk and Nizhnii Kitoi plutons, the leucogranites are enriched in Th (27–60 ppm), U (3.4–12.6 ppm), Nb (18–33 ppm), and Y (77–187 ppm) and are depleted in Zr (150–303 ppm). These rocks show weakly fractionated REE patterns ($(La/Yb)_n$ = 1.7–6.3) with enrichment in both LREE and HREE, and a strong negative Eu anomaly (Eu/Eu* = 0.17–0.41) (Fig. 4*D*). Their multielement patterns demonstrate strong negative Sr, P, and Ti and weak negative Nb and Ba anomalies (Fig. 5*D*).

The leucogranites of the Malaya Belaya pluton are characterized by ε_{Nd} varying from +0.7 to –1.9 and T_{Nd}(DM-2st) = 2.3–2.5 Ga (the two-stage model age is calculated, because the rocks have high $^{147}Sm/^{144}Nd$ values, 0.13–0.17 (Table 3, Fig. 6*B*)). The Nd isotope composition of the granites is in strong contrast to that of Archean crustal rocks of the Onot terrane, whereas the coeval granodiorites and granites of the Shumikha pluton of the same terrane (ε_{Nd} from –7.4 to –8.3) are similar in isotope composition to the Archean crust.

The temperatures of formation of the granitoids were estimated using a zirconium saturation thermometer (Watson and Harrison, 1983) and a phosphorus saturation thermometer (Harrison and Watson, 1984) as well as a Ti-in-Zr thermometer (Ferry and Watson, 2007). The presence of ilmenite in the rocks permits us to accept αTiO_2 = 0.7. The zirconium saturation temperatures of major-phase rocks and monzodiorites of the Toisuk pluton and granitoids of the Nizhnii Kitoi pluton are 786–897 and 823–855 ºC, respectively (Table 2). Highly fractionated leucogranites of the Malaya Belaya pluton are characterized by lower saturation temperatures, 782–844 ºC. According to Miller et al. (2003), the calculation yields the minimum temperature of melt formation for granitoids lacking inherited zircon cores. Close temperatures were obtained using a Ti-in-Zr thermometer for zircons from monzodiorite of the Toisuk pluton (826–905 °C, Ti = 16–31 ppm) and granodiorite of the Nizhnii Kitoi pluton (818–886 ºC; Ti = 14–26 ppm). Magmatic zircons from vein granite of the Toisuk pluton are characterized by lower temperatures (776–801 ºC; Ti = 10–12 ppm). Since these granites contain inherited Archean zircon cores, the Zr saturation temperature calculated for them (802–844 °C) is probably close to the melt temperature. The maximum crystallization temperatures are established using a phosphorus saturation thermometer. They decrease from 933–1024 °C (monzodiorites and rocks of major phase of the Toisuk pluton) and 973–993 ºC (granodiorite-granites of the Nizhnii Kitoi pluton) to 863–916 and 782–844 ºC (granites and leucogranites, respectively, of the Toisuk and Malaya Belaya plutons). The high temperatures obtained for the most melanocratic rocks are correlated with the early crystallization of apatite present as inclusions in plagioclase and amphibole. These results are consistent with the experimental data showing that melts with FeO* + MgO > 5% formed at 900–1000 °C during the melting of crustal sialic and mafic substrates (Vielzeuf and Holloway, 1988; Beard and Lofgren, 1991; Rapp and Watson, 1995; Sisson et al., 2005; Bogaerts et al., 2006).

Lu–Hf ISOTOPE COMPOSITION OF ZIRCON

In **the Toisuk pluton** we studied zircons from the rock end-members: monzodiorite and leucogranite. Zircons from monzodiorite are characterized by ε_{Hf} from –6.0 to –10.7 and model age T^{c}_{Hf}(DM) = 2.9–3.1 Ga (Table 4, Fig. 7*A*).

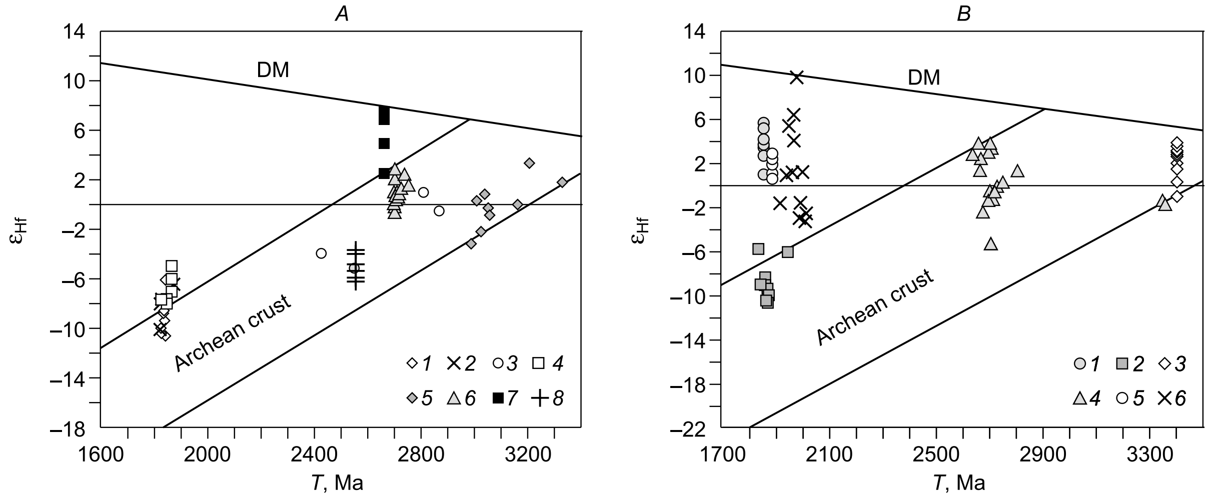

Fig. 7. *T*–ε_{Hf} diagrams for zircons from Paleoproterozoic granitoids and the host rocks. *A*, Irkut terrane: *1–3*, Toisuk pluton: *1*, monzodiorite, *2*, *3*, granite: magmatic (*2*) and inherited (*3*) zircon; *4*, granodiorite of the Nizhnii Kitoi pluton; *5*, Paleoarchean granulite, *6*, *7*, Neoarchean granulites: *6*, intermediate–felsic, *7*, mafic; *8*, Neoarchean granites. *5–8*, data after Turkina et al. (2012) and unpublished data of O.M. Turkina; *B*, Onot terrane: *1*, leucogranite of the Malaya Belaya pluton; *2*, granite of the Shumikha pluton (Turkina and Kapitonov, 2017); *3*, plagiogneisses of TTG complex (Turkina et al., 2013); *4*, garnet–staurolite schists (Turkina et al., 2014b); data for comparison: *5*, zircons from leucogranite of the Tagul pluton (Turkina and Priyatkina, 2017); *6*, detrital zircons from Paleoproterozoic metasedimentary gneisses of the Irkut terrane (Turkina et al., 2016).

Magmatic zircons from granite are similar in ε_{Hf} values (–6.7 to –10.1) and T^{c}_{Hf}(DM) = 2.9–3.1 Ga to those from monzodiorite. The inherited zircon cores with ages of ~2.5 and 2.8 Ga, ε_{Hf} = –5.3 to +1.0, and T^{c}_{Hf}(DM) = 3.2–3.4 Ga correspond to the Archean crust of the Irkut terrane, whose isotope parameters were determined by studying zircons from Paleo–Neoarchean intermediate–felsic granulites and Neoarchean granites. Zircons from granodiorite of the Nizhnii Kitoi pluton lack ancient cores and are characterized by ε_{Hf} = –5.0 to –8.1 and T^{c}_{Hf}(DM) = 2.8–3.0 Ga (Table 4, Fig. 7*A*). These zircons are similar in isotope parameters to those from rocks of the Toisuk pluton.

Zircons from leucogranites of **the Malaya Belaya pluton** are characterized by positive values of ε_{Hf} (5.9 to 2.4), and their model age T^{c}_{Hf}(DM) is 2.1–2.3 Ga and differs strongly from most of Paleoproterozoic granitoids of the Sharyzhalgai uplift (Table 4, Fig. 7*B*). According to Hf isotope composition, zircons from leucogranite differ strongly from zircons (ε_{Hf} = –8.4 to –10.6) from coeval granites of the Shumikha pluton (Turkina and Kapitonov, 2017) and from the host Archean rocks of the Onot terrane (Turkina et al., 2013, 2014b).

DISCUSSION

The age and geochemical types of Paleoproterozoic granitoids of the Sharyzhalgai uplift. The new U–Pb dates for zircons from rocks of the Toisuk (1838 ± 6 Ma and 1827 ± 9 Ma), Nizhnii Kitoi (1846 ± 7 Ma), and Malaya Belaya (1863 ± 16 Ma) plutons agree with the earlier obtained dates for granitoids of the Alar (1.85 Ga, Bulun terrane) and Shumikha (1.85 Ga, Onot terrane) plutons (Turkina and Kapitonov, 2017). Garnet-bearing vein granites, diatexites, and migmatites of the Irkut terrane formed at the same stage (1854 ± 11 Ma) (Turkina and Sukhorukov, 2017a). A similar age (~1.87 Ga) was established for charnockites in the east of the Irkut terrane (Aftalion et al., 1991) and abundant vein granitoids in the Kitoi and Irkut terranes (Poller et al., 2005; Sal'nikova et al., 2007). The formation of granitoids in the Kitoi and Irkut terranes was closely related to granulite metamorphism (1.88–1.85 Ga) (Poller et al., 2005; Turkina et al., 2010, 2012, 2017). Close ages of metamorphism (1.88 Ga) (Turkina and Nozhkin, 2008) and granitoid intrusions (1.85–1.86 Ga) were also established in the Onot terrane. Thus, all granitoids of the Sharyzhalgai uplift fall in the earlier determined time interval of granite formation (1.87–1.84 Ga) in the southwestern Siberian craton as a whole (Turkina et al., 2006). The subsynchronous metamorphism and granite formation indicates the connection of these events with the same collision process.

The formation of collisional granitoids was accompanied by basic magmatism. Granitoids of the Toisuk and Nizhnii Kitoi plutons are spatially associated with gabbro-dolerite dikes (1864 ± 4 Ma) intruding Archean gneiss-granites, migmatized gneisses, and Neoarchean granites (Gladkochub et al., 2013). In composition the gabbro-dolerites correspond to highly magnesian basalt; they have high contents of K_2O (2.3–2.5 wt.%), P_2O_5 (0.74–0.77 wt.%), LREE (La = 116–120 ppm), Sr (900–990 ppm), and Th (7–12 ppm), which indicates their formation from an enriched mantle source. The Malyi Zadoi intrusion located in the west of the Irkut terrane is composed of rocks from plagioperidotites to gabbronorites (1863 ± 1 Ma) (Mekhonoshin et al., 2016). The

Table 4. Lu–Hf isotope composition of zircons from Paleoproterozoic granitoids

Grain	T, Ma	$\frac{^{176}Lu}{^{177}Hf}$	$\frac{^{176}Hf}{^{177}Hf}$	±σ	ε_{Hf}	±2σ	T^{c}_{Hf}, Ma
Monzodiorite, Toisuk pluton (sample 63-15)							
2-2	1838	0.0004480	0.281363	0.000036	–9.4	1.3	3067
4-1	1834	0.0003683	0.281388	0.000030	–8.5	1.1	3009
4-1a	1834	0.0003561	0.281401	0.000044	–8.0	1.5	2980
5-1	1835	0.0004248	0.281458	0.000042	–6.0	1.5	2860
6-1	1835	0.0004981	0.281404	0.000023	–8.1	0.8	2984
7-1	1829	0.0003547	0.281384	0.000029	–8.7	1.0	3019
8-1	1826	0.0003427	0.281340	0.000027	–10.4	1.0	3116
8-1a	1826	0.0004299	0.281351	0.000027	–10.1	0.9	3099
9-1	1828	0.0005104	0.281349	0.000029	–10.2	1.0	3108
10-1	1841	0.0003968	0.281323	0.000030	–10.7	1.1	3147
Granite, Toisuk pluton (sample 47-15)							
1-1c	2864	0.0009370	0.280978	0.000033	–0.5	1.2	3327
2-1	1831	0.0003786	0.281406	0.000041	–7.9	1.5	2972
3-1	1819	0.0003486	0.281353	0.000045	–10.1	1.6	3093
5-1	1825	0.0004120	0.281403	0.000029	–8.2	1.0	2986
6-1	1831	0.0004447	0.281412	0.000044	–7.8	1.5	2965
7-1c	2807	0.0006772	0.281040	0.000045	1.0	1.6	3196
8-1	1866	0.0005113	0.281425	0.000030	–6.7	1.1	2921
9-1c	2549	0.0012070	0.281058	0.000040	–5.3	1.4	3370
9-2c	2428	0.0004541	0.281135	0.000046	–4.0	1.6	3200
Granodiorite, Nizhnii Kitoi pluton (sample 43-15)							
1-1	1859	0.0006001	0.281436	0.000020	–6.5	0.7	2908
4-2	1861	0.0004204	0.281443	0.000041	–6.0	1.5	2876
5-1	1864	0.0007264	0.281481	0.000049	–5.0	1.7	2817
6-2	1844	0.0003751	0.281394	0.000052	–8.1	1.8	2990
7-1	1854	0.0005464	0.281407	0.000051	–7.6	1.8	2971
8-1	1824	0.0004509	0.281417	0.000041	–7.8	1.5	2958
10-1	1864	0.0004161	0.281417	0.000041	–6.8	1.4	2931
Leucogranite, Malaya Belaya pluton (sample 43-95)							
1.1	1863*	0.0016025	0.281769	0.000008	4.1	0.3	2256
3.1	1863	0.0014456	0.281730	0.000009	3.0	0.4	2329
4.1	1863	0.0013974	0.281726	0.000009	2.9	0.3	2334
6.1	1863	0.0013011	0.281724	0.000008	2.9	0.3	2331
7.1	1863	0.0016533	0.281765	0.000008	3.9	0.3	2270
8.1	1863	0.0016728	0.281759	0.000007	3.7	0.3	2285
9.1	1863	0.0018702	0.281811	0.000009	5.3	0.2	2184
10.1	1863	0.0073889	0.282024	0.000016	5.9	0.3	2147

Note. Spot numbers follow Table 1. T, $^{207}Pb/^{206}Pb$ age. T^{c}_{Hf}, model Hf age. Grains 4.1 and 4.1a and grains 8.1 and 8.1a were analyzed in two spots, c, zircon core.
*Age accepted for zircons from the Malaya Belaya pluton.

enriched mantle source of these rocks is also evidenced by high contents of LREE (La = 8–19 ppm), Rb, Ba (up to 300 ppm), and Th.

Despite their varying composition, granitoids of the Toisuk, Nizhnii Kitoi, and Malaya Belaya plutons have a number of similar features. They are highly ferroan and are enriched in HFSE and LREE, that is, they have all geochemical characteristics typical of *A*-type granitoids (Whalen et al., 1987; Frost and Frost, 2011). Amphibole–biotite granodiorites and granites of the Shumikha pluton in the Onot terrane are also *A*-type granitoids (Donskaya et al., 2002), whereas granites and leucogranites of the Alar pluton of the Bulun terrane are similar in composition to highly differentiated *I*-type granites (Turkina and Kapitonov, 2017). Amphibole- and pyroxene-bearing low- and high-K charnockites of the southwestern Irkut terrane correspond to magnesian granitoids (f = 0.65–0.76), have low contents of Nb (4–17 ppm) and Y (3–18 ppm), and widely varying contents of Zr (119–550 ppm) and LREE (Ce = 50–190 ppm), which indicates their similarity to *I*-type granites (Turkina and Sukhorukov, 2017b). Garnet-bearing vein granites and diatexites of the Irkut terrane are typical *S*-type granitoids (Turkina and Sukhorukov, 2017a). They are characterized by a peraluminous composition (A/CNK = 1.1–1.4), wide variations in f values (0.65–0.84), and low contents of HFSE. Thus, the Paleoproterozoic granitoids of the Sharyzhalgai uplift are of three geochemical types, with Fe-rich *A*-type granites prevailing. The same three geochemical types were earlier established for nearly coeval Paleoproterozoic granitoids of the Biryusa terrane (1.87–1.86 Ga) (Donskaya et al., 2014). The diversity of granitoids formed in a narrow time interval seems to be a characteristic feature of the Paleoproterozoic collisional orogen in the southwestern Siberian craton.

Genesis of Paleoproterozoic granitoids and the sources of melts. *A*-type granites are assumed to form from contrasting crustal and mantle sources, including mantle-related mafic rocks, which can generate *A*-type granites through partial melting or intense fractional crystallization (Turner et al., 1992; Vander Auwera et al., 2003; Li et al., 2007; Shellnutt and Zhou, 2007), as well as metaigneous rocks of tonalitic to granodioritic composition (Patiño Douce, 1997; Frost and Frost, 2011). These granites can also result from the mixing of crustal and mantle melts (Yang et al., 2006).

A-type granitoids vary in composition from metaluminous to weakly peraluminous and from alkali-calcic to calc-alkalic and alkalic. Based on the composition of these ferroan granitoids and the experimental data, two main models of their genesis are proposed: (1) differentiation of mafic magmas along with crustal contamination, resulting in calc-alkalic and alkalic *A*-type granitoids, and (2) melting of crustal quartz–feldspar substrates, leading to the generation of alkali-calcic leucocratic granites (Frost and Frost, 2011). Based on this information, we consider the probable sources of Paleoproterozoic granitoids of the Sharyzhalgai uplift.

The Malaya Belaya pluton. The high contents of SiO_2 and low contents of femic components in granites of the Malaya Belaya pluton testify to their formation from melt that underwent intense fractional crystallization. Amphibole, plagioclase, Fe–Ti oxides, and apatite were major fractionating phases, as follows from the deep negative Sr, P, and Ti anomalies on the multielement patterns and from the low contents of Fe, Mg, and Ca. The alkali-calcic type of the granites implies two possible models of their formation: melting of crustal quartz–feldspar material or extreme differentiation of tholeiite-basaltic melt. The latter model is supported by the high f values of leucogranites (0.91–0.95), which is similar to those of granophyres of the Skiergaard intrusion (f > 0.9) (Wager and Brown, 1967). At the same time, the Malaya Belaya pluton lacks low-SiO_2 rocks and is not associated with mafic rocks, which contradicts the second model. According to Frost and Frost (2011), most of alkali-calcic metaluminous and weakly peraluminous granites rich in SiO_2 (>70 wt.%) result from the melting of crustal quartz–feldspar material (Fig. 8), which is confirmed by the experimental data (Patiño Douce, 1997; Bogaerts et al., 2006). Leucogranites of the Malaya Belaya pluton are most similar in indicator petrochemical features to melts produced at the low degree of melting of granodiorites at 4 kbar (Bogaerts et al., 2006) (Fig. 9). As follows from the isotope parameters of zircon (ε_{Hf} = 5.9 to 2.4) and rocks (ε_{Nd} = +0.7 to –1.9), the leucogranites formed from the Paleoproterozoic juvenile crust but not the Archean rocks of the Onot terrane (Fig. 7*b*).

The Toisuk pluton. Monzodiorites and rocks of major phase are poorer in SiO_2 and richer in femic components (FeO* + MgO = 7.2–11.6%) than melts from quartz–feldspar sources (tonalites and granodiorites) (SiO_2 = 64–80 wt.% and FeO* + MgO < 5 wt.%) (Singh and Johannes, 1996; Skjerlie and Johnston, 1996; Bogaerts et al., 2006) (Fig. 9). Such low-SiO_2 sources melts can be produced from mafic substrates at 3–8 kbar and >900 ºC (Beard and Lofgren, 1991; Rapp and Watson, 1995; Sisson et al., 2005). In the diagram summarizing the experimental data on the melting of various sources (Laurent et al., 2014), the Toisuk pluton rocks (except for vein granites) fall in the field of melts from mafic substrates and are localized along the boundary between low- and high-K mafic-source rocks (Fig. 8). The contribution of mantle-related melts to the formation of the Toisuk pluton is directly evidenced by monzodiorites (SiO_2 = 55–57 wt.%) enriched in TiO_2 and P_2O_5, which per-

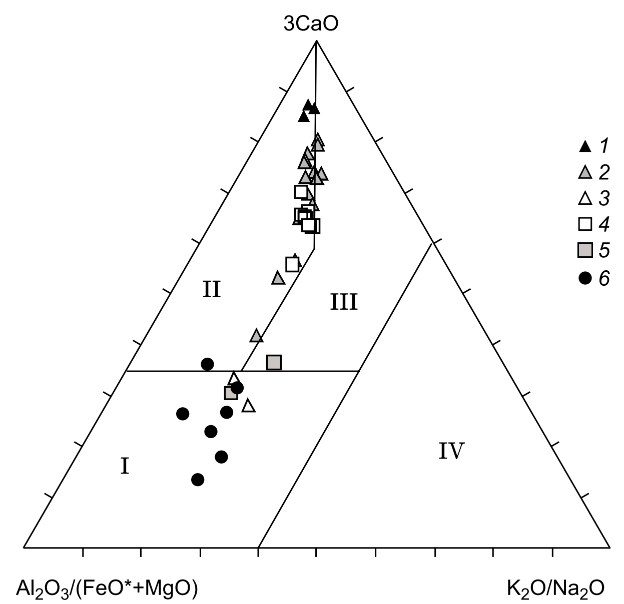

Fig. 8. Al_2O_3/(FeO* + MgO)–3CaO–K_2O/Na_2O diagram for Paleoproterozoic granitoids. *1–3*, Toisuk pluton: *1*, monzodiorites, *2*, granodiorite-granosyenites, *3*, vein granites; *4*, *5*, Nizhnii Kitoi pluton: *4*, granodiorite-granites, *5*, vein granites; *6*, Malaya Belaya pluton. Fields of the sources, after Laurent et al. (2014): I, tonalitic, II, mafic low-K and III, high-K, IV, metasedimentary.

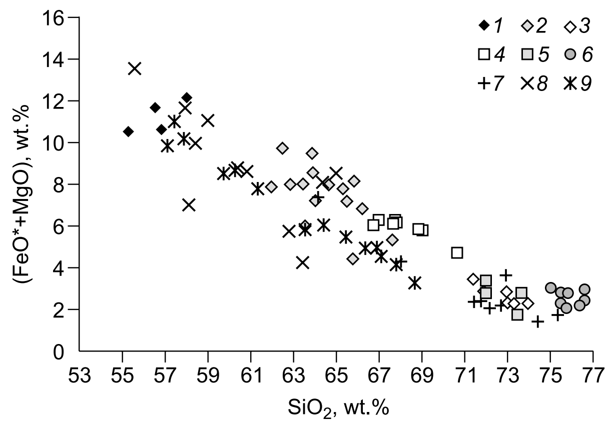

Fig. 9. SiO_2–(FeO* + MgO) diagram for Paleoproterozoic granitoids. *1–3*, Toisuk pluton: *1*, monzodiorites, *2*, granodiorite-granosyenites, *3*, vein granites; *4*, *5*, Nizhnii Kitoi pluton: *4*, granodiorite-granites, *5*, vein granites; *6*, Malaya Belaya pluton; melts of different sources; *7*, quartz–feldspar (granodiorites) (Bogaerts et al., 2006); *8*, *9*, mafic: *8*, low-K (Beard and Lofgren, 1991; Rapp and Watson, 1995), *9*, high-K (Sisson et al., 2005).

mits them to be assigned to crystallization products of residual melts formed during the fractional crystallization of subalkalic mafic magma. Probably the extreme product of this fractionation is melanocratic inclusions with SiO_2 = 58 wt.% and the maximum contents of TiO_2 (2 wt.%) and P_2O_5 (0.86 wt.%).

The high contents of incompatible trace elements in the monzodiorites and granitoids of major phase favor the formation of the parental mafic magmas/sources from the enriched mantle. This agrees with the isotope composition of the monzodiorites (ε_{Nd} = –10.2 to –7.2) and their zircons (ε_{Hf} = –6.0 to –10.7). The high contents of Ba (>1000 ppm) and Sr impose additional constraints on the composition of the sources and the melting conditions. We performed numerical simulation using the experimental data on the melting of K-enriched mafic rocks (Sisson et al., 2005). At 7 kbar and 975–875 ºC and with the melting degree (*F*) of 15–30%, melts with SiO_2 = 57–66 wt.% are in equilibrium with orthopyroxene–amphibole–plagioclase restite (Opx : Hb : Pl = (0–5) : (45–46) : (50–54)). The most melanocratic (SiO_2 = 68 wt.%) melts from a quartz–feldspar source are produced with a high degree of melting (*F* = 60%) at 950 ºC in equilibrium with orthopyroxene–clinopyroxene–plagioclase restite (Opx:Cpx:Pl = 10:23:45) (Bogaerts et al., 2006). The numerical simulation shows[1] that melting of sialic sources (Archean TTG, upper crust, and Archean gneiss-granites of the Irkut terrane) with 630–800 ppm Ba and 320–450 ppm Sr resulted in melts with 850–1100 ppm Ba and 200–510 ppm Sr. These contents of Ba and Sr are much lower than those in most of the Toisuk granitoids (Fig. 10). The maximum contents of Ba (3900–1500 ppm) and Sr (460–1140 ppm), close to its contents in these rocks, can be obtained in melts from a mafic source similar to continental intraplate basalts (Ba = 750 ppm and Sr = 800 ppm), for example, high-Ti basalts of the Karoo province (Jourdan et al., 2007). Close contents of Sr (900–990 ppm) were found in dike gabbro-dolerites of the Kitoi terrane (Gladkochub et al., 2013).

In contrast to the monzodiorites, the vein granites are clearly peraluminous and contain inherited Archean zircon cores, which makes it possible to interpret the negative ε_{Nd} values of rocks and the negative ε_{Hf} values of zircons as a result of the melting of an ancient crustal source. In major elements the granites (Figs. 8 and 9) are similar to melts from quartz–feldspar (tonalitic and granodioritic) substrate. Compared to Archean granulites of intermediate–felsic composition of the Irkut terrane (ε_{Nd} from –7.6 to –15.2 at 1.85 Ga), the granites have higher ε_{Nd} values (–5.5) (Fig. 6), which suggests the contribution of juvenile material to their formation.

As mentioned above, the rocks of major phase are divided into two groups according to trace-element and Nd isotope compositions. Granodiorites of group I are strongly enriched in LREE and Ba and show a low ε_{Nd} value (–8.2), being similar in these parameters to monzodiorites. This suggests that they resulted from the differentiation or partial melting of mafic rocks, which are compositionally similar to monzodiorites and derived from an enriched mantle source. Most of granodiorite-granosyenites and one monzodiorite sample assigned to group II are, on the contrary, depleted in LREE and Ba and are similar in ε_{Nd} values (–5.3 and –4.6) to granites (ε_{Nd} = –5.5). The strong difference in their contents of incompatible trace elements and in Nd isotope composition argues for the formation of these rocks from a distinct magma portion produced from a less enriched mafic source. The

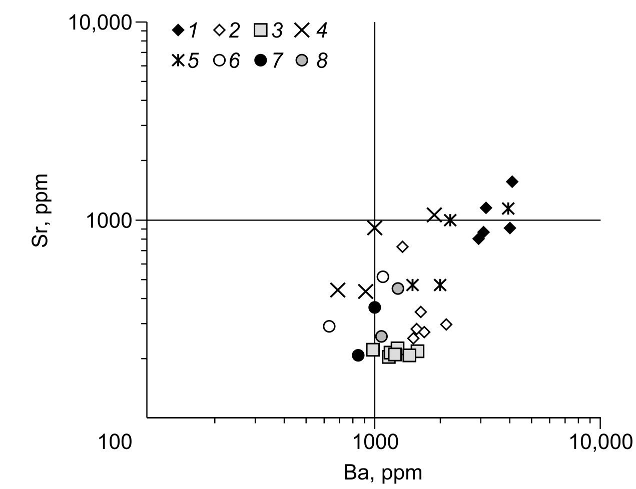

Fig. 10. Ba–Sr diagram for Paleoproterozoic granitoids. *1*, *2*, Toisuk pluton: *1*, monzodiorites, *2*, granodiorite-granosyenites; *3*, Nizhnii Kitoi pluton; calculated contents in melts from: *4*, *5*, mafic sources: *4*, OIB (Sun and McDonough, 1989), *5*, Karoo basalts (Jourdan et al., 2007); *6–8*, quartz–feldspar sources: *6*, Archean TTG (Martin, 1994), *7*, upper crust (Rudnick and Gao, 2003), *8*, Archean granite-gneisses of the Irkut terrane. For explanation, see the text.

[1] On calculation, we used the minimum and maximum values of the mineral/melt distribution coefficients from Laurent et al. (2013).

contents of Ba and Sr in the group II granitoids are close to the calculated contents of these elements in model melts from a less enriched mafic source (Ba = 350 ppm and Sr = 660 ppm) (Fig. 10), e.g., lower crust basites or ferrodiorites. Thus, rocks of the Toisuk pluton might have formed through the differentiation of subalkalic mafic magma derived from the enriched mantle, the partial melting/differentiation of a less enriched mafic source, and the melting of crustal quartz–feldspar material.

Granodiorites and granites of **the Nizhnii Kitoi pluton** are distinguished by a narrower range of SiO_2 contents (67–71 wt.%) and a metaluminous (except for vein granites) calc-alkalic composition. According to Frost and Frost (2011), *A*-type granites with such characteristics are products of differentiation of tholeiite-basaltic magmas or melting of a mafic source in combination with assimilation of crustal material or mixing with crustal-derived melts. In major-element composition (Figs. 8 and 9) the granodiorites and granites correspond to products of partial melting of mafic sources. As follows from the results of numerical simulation, these rocks (with Ba = 990–1500 ppm) can form both from mafic substrates produced from the moderately enriched mantle and from a quartz–feldspar source compositionally corresponding to Archean gneiss-granites (Fig. 10). Zircons from the granodiorite are similar in ε_{Hf} (–5.0 to –8.1) to zircon from the Toisuk monzodiorite but have a more radiogenic Hf isotope composition than Archean crustal rocks of the Irkut terrane (Fig. 7*A*). Thus, the geochemical and isotope data suggest that granitoids of the Nizhnii Kitoi pluton formed through the melting of an enriched mafic source, e.g., lower-crustal gabbroids/ferrodiorites and crustal quartz–feldspar substrate, whose role probably increases for peraluminous vein granites.

The origin of enriched mantle sources. Analysis of the composition and isotope parameters of the Paleoproterozoic granitoids shows the formation of their initial magmas at different levels of the lithosphere of the southwestern Siberian craton, which evolved before granitoid magmatism.

Formation of the monzodiorite–granodiorite series rocks with high contents of incompatible trace elements and negative ε_{Hf} and ε_{Nd} values, which prevail in the Toisuk and Nizhnii Kitoi plutons, implies the participation of sources derived from the enriched mantle. Metasomatism and formation of the enriched subcontinental lithospheric mantle (SCLM) might have taken place in the Neoarchean. This stage in the Irkut terrane corresponds to the formation of protoliths of mafic and intermediate–felsic granulites (2.7–2.66 Ga), which are characterized by high contents of LILE and Nb depletion typical of subduction volcanics (Turkina et al., 2012). The involvement of this lithosphere in the processes of Paleoproterozoic magmatism is recorded by dike gabbro-dolerites (Gladkochub et al., 2013) and gabbronorites of the Malyi Zadoi intrusion (Mekhonoshin et al., 2016), which are variably enriched in K_2O, LREE, Rb, Ba, Sr, and Th. The isotope composition of the Paleoproterozoic SCLM of the Siberian Craton can be judged, to a first approximation, from gabbroids of the Chinei pluton in the west of the Aldan Shield (~1.86 Ga, ε_{Nd} = –5.0 to –4.4 (Gongal'skii et al., 2008)). The isotope data permit estimation of the parameters of the subcontinental lithospheric mantle enriched during the Neoarchean subduction processes. The enriched mantle source is modelled by mixing of 95% depleted mantle (ε_{Nd} = 4.2 and Nd = 0.58 ppm) and 5% crustal material (Nd = 27 ppm and ε_{Nd} = –5 to +1). The average content of Nd in the upper crust (Rudnick and Gao, 2003) and the range of ε_{Nd} values for Archean intermediate–felsic granulites and paragneisses of the Irkut terrane (Turkina et al., 2012; Turkina and Sukhorukov, 2015) were taken as the crustal parameters. The change in the isotope composition of the mixture from 2.70 to 1.84 Ga is calculated by taking $^{147}Sm/^{144}Nd$ = 0.125 (Fig. 11). The ε_{Nd} values for the model mantle source vary from –6.3 to –10.5, which is generally close to the range of the ε_{Nd} values for rocks of the Toisuk and Nizhnii Kitoi plutons (–10.2 to –4.0). The higher ε_{Nd} values (–5.5 to –4.0) for some granitoid samples are close to those for gabbroids of the Chinei pluton and might be due to the mantle source with a lower portion of crustal component.

Paleoproterozoic granitoids and stages of evolution of the Early Precambrian crust in the southwestern Siberian craton. Continental crust in the southwestern Siberian craton formed since Paleoarchean till Neoarchean (3.40–2.55 Ga). Archean crustal rocks served as the sources of the inherited cores of zircons (2.40–2.55 and 2.80–2.86 Ga, T^{c}_{Hf}(DM) = 3.2–3.4 Ga) in vein granites of the Toisuk pluton (Fig. 7*A*). Compared with the Archean crust, which was characterized by ε_{Hf} = –8.5 to –15.7 and ε_{Nd} = –7.6 to –15.2 in Paleoproterozoic time (1.84 Ga), magmatic zircons (ε_{Hf} = –6.7 to –10.1) and granites (ε_{Nd} = –4.6 and –5.5) from the Toisuk pluton have a slightly more radiogenic isotope composition, which suggests the contribution of juvenile material to the generation of granitic melt (Figs. 6*A* and 7*A*). The major contribution of the Archean crust is assumed for Paleoproterozoic granitoids of granite–greenstone blocks in

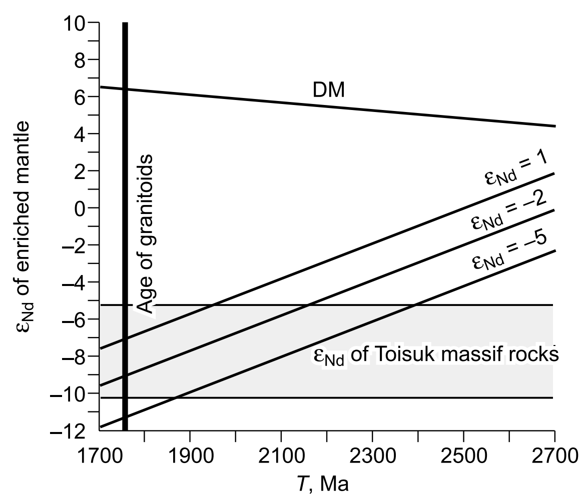

Fig. 11. *T*–ε_{Nd} diagram showing the calculated composition trend of the enriched mantle. For explanation, see the text.

the northwestern Sharyzhalgai uplift (Alar and Shumikha plutons) (Turkina and Kapitonov, 2017). In contrast to these granitoids, the isotope composition of leucogranites of the Malaya Belaya pluton argues for their juvenile Paleoproterozoic crustal source. In major- and trace-element contents and Hf isotope composition the Malaya Belaya granites are similar to highly ferroan leucogranites of the Toporok pluton (1.88 Ga) of the Biryusa terrane in the southwestern Siberian craton (Turkina and Priyatkina, 2017). Zircons from the Toporok leucogranites are characterized by ε_{Hf}= 3.0–0.8 and T^{c}_{Hf}(DM) = 2.3–2.5 Ga (Fig. 7*B*). Juvenile Paleoproterozoic crust has not been revealed within the southwestern Siberian craton. The crust that supplied detrital zircons for Paleoproterozoic paragneisses of the Irkut terrane might have been the source of the melt for the leucogranites of the Malaya Belaya and Toporok plutons. Detrital zircons from paragneisses with an age of 2.00–1.95 Ga are characterized by mostly positive ε_{Hf} values (from +10 to –3) overlapping with those of the zircons from leucogranites (Fig. 7*B*) (Turkina et al., 2016). Thus, the isotope composition of granitoids of the Sharyzhalgai uplift indicates both the Archean and the Paleoproterozoic stages of crustal growth preceding the granite magmatism. The subsequent growth and juvenilization of continental crust correspond to the time of the Late Paleoproterozoic collisional magmatism. This stage included the emplacement of both mafic dikes and intrusions and monzodiorite–granodiorite plutons formed through the differentiation/melting of mafic sources derived from the enriched lithospheric mantle.

CONCLUSIONS

Granitoids of the Toisuk (1838 ± 6 and 1827 ± 9 Ma), Nizhnii Kitoi (1846 ± 7 Ma), and Malaya Belaya (1863 ± 16 Ma) plutons in the Sharyzhalgai uplift (southwestern Siberian craton) formed at the Late Paleoproterozoic collision stage, simultaneously with mafic magmatism.

The studied rocks show a wide range of SiO_2 contents. The Toisuk pluton is composed of a range of rocks from monzodiorites to granodiorites (granosyenites) and granites; the Nizhnii Kitoi pluton, of granodiorites and granites; and the Malaya Belaya pluton, of leucogranites. A characteristic feature of melanocratic granitoids of the Toisuk and Nizhnii Kitoi plutons is extremely high contents of Ba: 4080–1500 ppm and 1560–990 ppm, respectively. Based on analysis of experimental data on the melting of various substrates and the results of numerical simulation, we assume that monzodiorite–granodiorites of the Toisuk pluton and granodiorites of the Nizhnii Kitoi pluton resulted from the differentiation/melting of a mafic source similar in Ba and Sr contents to intraplate continental basalts. The isotope compositions of zircon and melanocratic granitoids of the Toisuk (ε_{Hf} from –6.0 to –10.7 and ε_{Nd} from –5.3 to –10.2) and Nizhnii Kitoi (ε_{Hf} from –5.0 to –8.1 and ε_{Nd} = –4.0 and –5.1) plutons argue for the generation of their mafic sources from the enriched lithospheric mantle resulted from Neoarchean subduction processes. Vein granites of the Toisuk pluton and leucogranites of the Malaya Belaya pluton formed through the melting of quartz–feldspar (granodiorite) source. The contrasting isotope parameters of the Toisuk vein granites (ε_{Hf} from –6.7 to –10.1, zircons, and ε_{Nd} = –5.5, rock) and Malaya Belaya leucogranites (ε_{Hf} from 2.9 to 5.9, zircons, and ε_{Nd} from +0.7 to –1.9, rocks) indicate melting of the Archean and Paleoproterozoic crust, respectively. The more radiogenic Hf isotope composition of zircons from vein granites as compared with rocks of the Archean crust of the Irkut terrane is evident of the contribution of juvenile material to the granite formation.

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