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Petrogenesis and Age of the Felsic Volcanic Rocks from the North Baikal Volcanoplutonic Belt, Siberian Craton

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Abstract—Detailed geochemical, isotopic, and geochronological studies were carried out on felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt. U–Pb zircon dating showed that the rocks previously ascribed to a single stratigraphic unit (Khibelen Formation of the Akitkan Group or the Khibelen Complex) have significant age differences. The Khibelen Formation was found out to include both the oldest dated rocks (1877.7 ± 3.8 Ma) of the North Baikal belt and the younger volcanic rocks (1849 ± 11 Ma). Two other dated volcanic rocks have intermediate ages (1875 ± 14 and 1870.7 ± 4.2 Ma). It was established that the volcanic rocks from various areas in the southern part of the North Baikal belt not only have different ages but also differ in geochemical and isotopic signatures. In particular, the felsic volcanic rocks from various sites show the following variations in trace-element composition: from 220–280 to 650–717 ppm Zr, from 8–12 to 54–64 ppm Nb, and from 924–986 to 1576–2398 Ba. The ε_{Nd} obtained for felsic volcanic rocks and comagmatic granitoids from various areas in the southern part of the North Baikal belt vary, respectively, from –1.7 to –2.8 and from –8.0 to –9.2. Based on geochemical and isotopic signatures, the felsic volcanic rocks in various areas of the southern part of the North Baikal volcanoplutonic belt were formed via the melting of a Mesoarchean crustal source of tonalite composition with contribution of variable amounts of juvenile mantle material at different magma generation conditions. Isotopic data indicate that the contribution of juvenile mantle material to their sources varied from ~33–40 to 77–86%. The maximal calculated temperatures of the parent melts for felsic volcanic rocks were 908–951°C, and the lowest temperatures were 800–833°C. The geochemical signatures of dacites with an age of 1877.7 ± 3.8 Ma such as high Th (46–51 ppm) and La (148–178 ppm) contents indicate that these rocks, along with Mesoarchean granitoid and juvenile mantle material, contain an upper crustal component with high Th and LREE contents. Extremely low Y and Yb contents in these dacites implies their formation at pressures of ~ 12–15 kbar in equilibrium with garnet-bearing residue. These rocks were presumably formed in the collisional–thickened crust at the earliest stages of its collapse, possibly during syncollisional collapse, with additional hear input to the lower crust. Other felsic rocks are geochemical analogues of A-type granites and were formed during the subsequent stages of collapse (post-collisional collapse).

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INTRODUCTION

The formation of the Siberian craton by the accretion and collision of the Archean microcontinents and Early Proterozoic island arcs at ~1.9–2.0 Ga (Rosen, 2003; Larin et al., 2003) was finalized by large-scale magmatism duirng the collapse of the collisional system (post-collisional extension). The time span of 1.84–1.88 Ga was marked by the formation of numerous granitoid massifs, which are presently located in the southern marginal salients of the craton, and the emplacement of the North Baikal volcanoplutonic belt (Fig. 1). Larin with co-authors (Larin et al., 2003) combined all magmatic rocks dated at 1.84–1.88 Ga into a single South Siberian post-collisional igneous belt more than 2500 km long.

In recent years, numerous studies (Donskaya et al., 2002, 2003, 2005a; Levitsky et al., 2002; Larin et al., 2000, 2006; Nozhkin et al., 2003; Turkina, 2005;

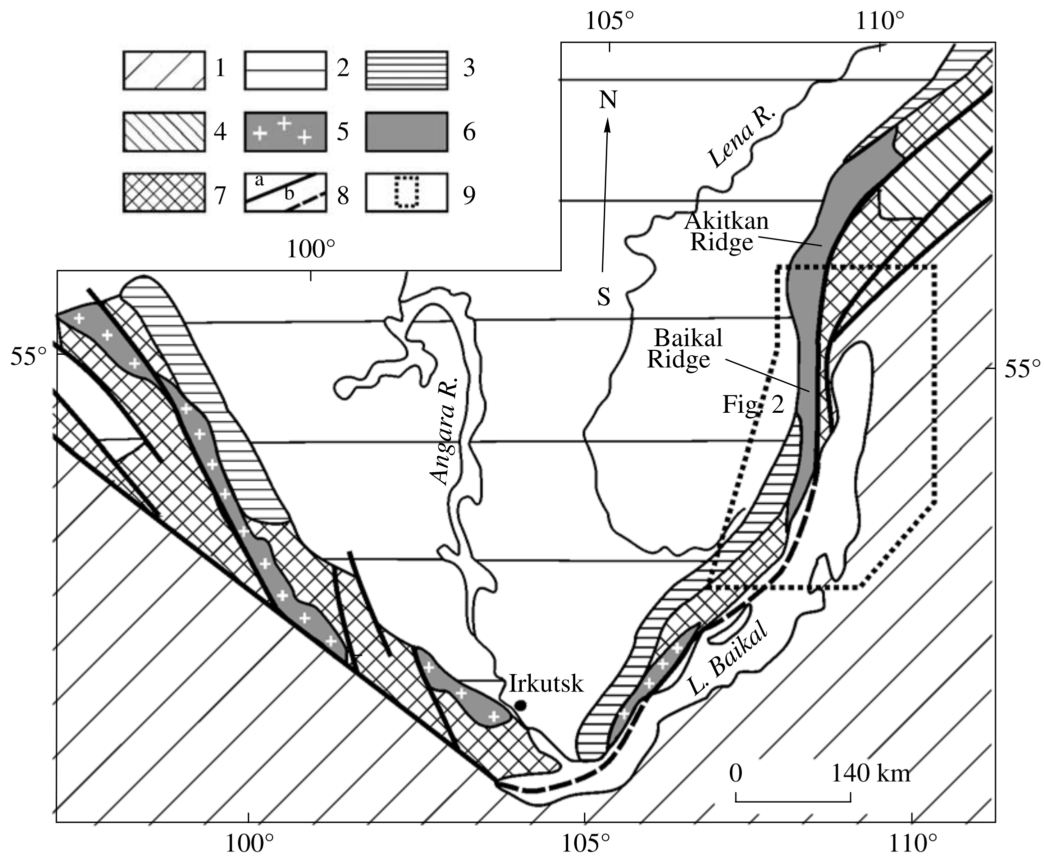

Fig. 1. Schematic geological map of the southern part of the Siberian craton.

(1) Central Asian fold belt; (2) cover of the Siberian platform; (3, 4) deposits of the Riphean pericratonic depressions: (3) unmetamorphosed, (4) highly metamorphosed; (5–7) complexes of marginal basement salients of the Siberian craton: (5) Early Proterozoic post-collisional granitoids, (6) Early Proterozoic North Baikal volcanoplutonic belt, (7) Archean–Early Proterozoic metamorphic and magmatic complexes, (8) main faults: proved (a) and inferred (b); (9) outlines of the area shown in Fig. 2.

Turkina et al., 2003, 2006) showed that the Early Proterozoic granitoids from the South Siberian post-collisional belt widely differ in chemical composition and belong to different geochemical types. A compositional diversity was found even between granitoids ascribed to a single geochemical type, in particular, to the A type according to Whalen's classification (Whalen et al., 1987). They demonstrate significant variations in the contents of major oxides, trace elements, and isotope geochemistry. These reasons for theses differences can be as follows: (1) differences in the composition of the crustal sources; (2) differences in the magma generation conditions (oxygen fugacity, temperature, pressure); (3) a variable contribution of mantle material in the magma generation zone (Donskaya et al., 2005a, Turkina et al., 2006). In addition, granitoid massifs combined in the previously distinguished magmatic complexes appeared to have different age and different sources (Levitsky et al., 2002; Turkina, 2005; Turkina et al., 2006; Larin et al., 2006).

Unlike the Early Proterozoic post-collisional granitoids, no petrological studies of the felsic volcanic rocks from the North Baikal volcanoplutonic belt have been performed over the past decade (Bukharov, 1987; Petrova et al., 1997; Neimark et al., 1998). Recent works in the North Baikal belt were aimed mainly at dating the constituent rocks (Larin et al., 2003; Sobachenko et al., 2005). Moreover, the conclusions obtained from petrological and geochronological data on the volcanic rocks were applied to the entire stratigraphic unit or complex of the belt (Bukharov, 1987; Neymark et al., 1991, 1998; Larin et al., 2003; Buldygerov and Sobachenko, 2005). This resulted in a variety of stratigraphic schemes, in which rocks from the same area were assigned to different formations and complexes. Based on conclusions obtained during studying the Early Proterozoic granitoids from the southern salients of the Siberian craton, the felsic volcanic rocks from various areas of the North Baikal volcanoplutonic belt were subjected to detailed geochemical, isotopic-geochemical, and geochronological studies. Note that some rocks from the studied areas belong, according to all stratigraphic schemes, to a common unit, while the stratigraphic affiliation of other rocks depends on the accepted scheme of the internal subdivision of the North Baikal belt.

In this work, we examined the variations in major, trace element, and isotopic–geochronological compositions for the felsic volcanic rocks from the southern part of the North Baikal belt, considered the possible reasons for their compositional diversity, and estimated the possible sources and conditions of their formation. Our geochronological data constrained the timing of the North Baikal volcanoplutonic belt and defined the stratigraphic position of the studied volcanic rocks in the Precambrian sequence of the Siberian craton.

GEOLOGICAL OVERVIEW OF THE NORTH BAIKAL VOLCANOPLUTONIC BELT

The North Baikal volcanoplutonic belt is located in the southern margin of the Siberian platform and extends NNE from the northwestern extremity of Lake Baikal for 550 km at a width of 5–60 km (Bukharov, 1987). Orographically, the North Baikal belt is confined to the Baikal and Akitkan ridges (Fig. 1). The rocks of the North Baikal volcanoplutonic belt overlay with an angular unconformity the Early Proterozoic metamorphic rocks (Sarma Group, Chuya Group, and their analogues) and Mesoarchean granitoids (Donskaya et al., 2005b) and are overlain by the Upper Riphean sedimentary rocks of the Baikal Group (Fig. 1).

Long-term studies resulted in different viewpoints regarding the tectonic position, the age of constituent rocks, and stratigraphic subdivision of the belt (Brandt et al., 1978; Borukaev, 1985; Bukharov, 1987; Neymark et al., 1998; Petrova et al., 1997; Condie and Rosen, 1994, and others). Recent studies make it possible to resolve some contradictions concerning the tectonic position of the belt in the Siberian craton and its nature. Larin et al. (2003) showed that the belt was formed in a post-collisional extension setting during the final stages of the accretionary–collisional processes, which led to the formation of a common structure of the Siberian craton. U–Pb zircon dating of several samples of volcanic rocks and granites of the North Baikal belt showed their Paleoproterozoic age (1.80–1.87 Ga) (Larin et al., 2003; Neymark et al., 1998; Sobachenko et al., 2005; Poller et al., 2005), instead of a Mesoproterozoic age, as was previously inferred from Rb–Sr isotopic dating (Brandt et al., 1978).

The North Baikal volcanoplutonic belt is made up of the terrigenous–volcanogenic rocks of the Akitkan Group and granitoids of the Irel Complex, which are comagmatic to the felsic rocks of the Akitkan Group (Salop, 1964; Mats et al., 1968). The volcanic rocks of the Akitkan Group are dominated by felsic rocks of elevated alkalinity; volcanic rocks of rhyolite to dacite composition account for ~75% (Neymark et al., 1998).

Salop was the first to distinguished the Akitkan Group time by (Salop, 1964), which combined (from bottom upward) the previously distinguished Malaya Kosa, Khibelen, and Chaya formations. According to this scheme, the Malaya Kosa and Chaya formations consist mainly of terrigenous rocks, while the Khibelen Formation is made up of volcanic rocks; volcanism was thought to have occurred simultaneously in the Akitkan and Baikal ridges. Mats et al. (1968) proposed to divide the Akitkan Group into the Domugda (lower) and Chaya (upper) formations in the Akitkan Ridge, and the Khibelen Formation was correlated with the Chaya Formation in the Baikal Ridge.

Since the early 1970s, most geologists began to distinguish volcanogenic, volcanoplutonic, and volcanogenic–terrigenous complexes within the North Baikal belt (reviews in Bukharov, 1987; Sryvtsev, 1986; Sryvtsev and Buldygerov, 1982), in which volcanogenic rocks were regarded not as part of a stratified sequence but as composing diverse paleovolcanic edifices. Let us consider two of the recent stratigraphic schemes of the North Baikal volcanoplutonic belt, which were proposed by Sryvtsev (1986) and Bukharov (1987). The former scheme combined the volcanogenic rocks of coeval volcanoes into volcanic complexes, while the coeval terrigenous rocks were grouped into formations (Sryvtsev, 1986). Based on these generalizations, the author distinguished the lower Akitkan Group, consisting of the Malaya Kosa, Gol'tsov, Molokolon, Kulenyan, Domugda, and Khibilen formations, and the upper Umbel Group, which included Chaya, Okun formations and Lambor, Revuni, and Parusny volcanic complexes. The stratigraphic scheme proposed by Bukharov (1987) includes a series of volcanogenic complexes, which consist of subvolcanic, pyroclastic, and terrigenous rocks. He recognized the Domugda (lower) and Chaya (upper) volcanogenic complexes in the Akitkan range, and Khibelen Complex in the Baikal range.

Based on newly obtained petrological and U–Pb zircon dates of rocks from the North Baikal belt, Neymark and co-authors (1998) and Larin and coauthors (Larin et al., 2003) developed a generalized stratigraphic scheme (from bottom upward: Malaya Kosa, Domugda, Chaya formation), which differs from the aforementioned stratigraphic schemes of the belt.

GEOLOGICAL–PETROGRAPHIC CHARACTERISTICS OF THE STUDY AREAS

In order to solve the formulated issues, we studied in detail felsic volcanic rocks from five areas in the southern part of the North Baikal volcanoplutonic belt: *Zavorotny, Srednii Kedrovy, Svetly, Khibelen, and Kunerma* (Fig. 2). According to existing stratigraphic schemes, volcanogenic rocks from four studied areas (*Zavorotny, Srednii Kedrovy, Svetly, and Khibelen*) are ascribed to the Khibelen Formation of the Akitkan Group (Salop, 1964; Mats et al., 1968; Neymark et al., 1998) or to the Khibelen Complex (Sryvtsev, 1986 and reference herein, Bukharov, 1987). The volcanic rocks of the *Kunerma* area (the Davan pass) are ascribed either to the Domugda Formation of the Akitkan Group (Mats et al., 1968) or to the Domugda Complex (Sryv-

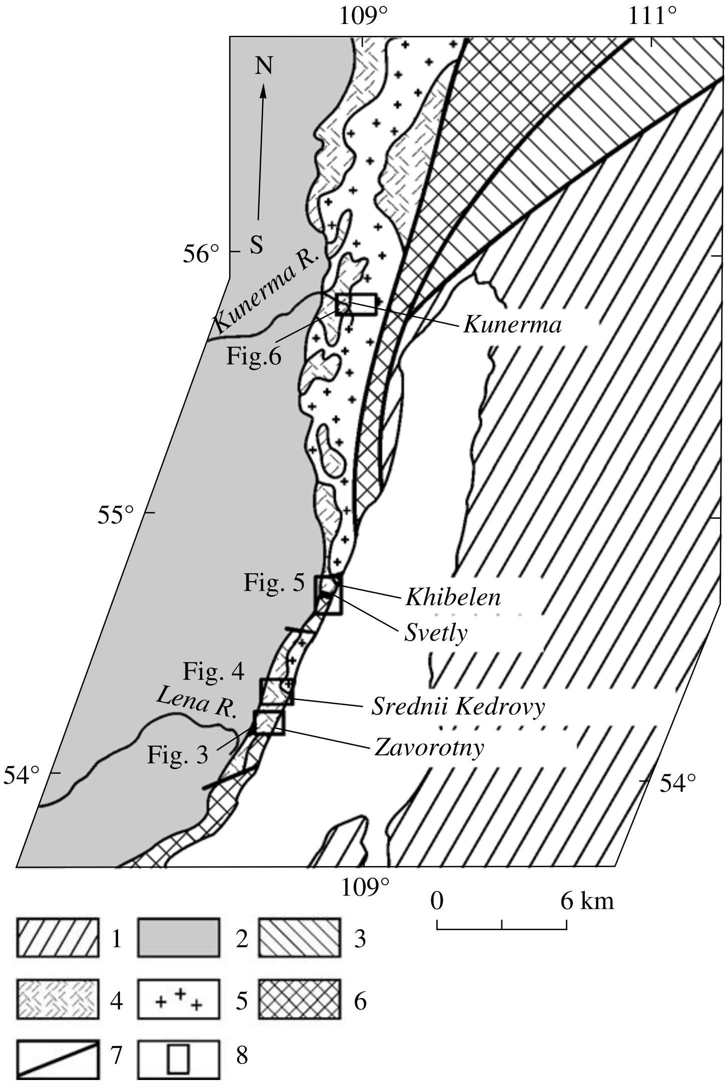

Fig. 2. Schematic geological map of the southern part of the North Baikal volcanoplutonic belt with localities of study areas.
(1) Central Asian fold belt; (2) Late Proterozoic–Early Paleozoic rocks of the cover of the Siberian craton; (3) high-grade deposits of the Riphean pericratonic depressions, (4) terrigenous–volcanogenic rocks of the Akitkan Group, (5) granitoids of the Irel' Complex; (6) Archean–Early Proterozoic metamorphic and magmatic complexes of the Siberian craton; (7) main faults; (8) areas of detailed study.

tsev, 1986, and references herein; Bukharov, 1987; Neymark et al., 1998), while in other publication, these rocks are considered to be part of the Khibelen Formation (Salop, 1964; Sobachenko et al., 2005).

According to concepts of Bukharov (1987), felsic volcanic rocks occur in the southern part of the North Baikal volcanoplutonic belt as flows, sheets, extrusion, and sills of volcanic edifices (Bukharov, 1987). In all the areas, the sequences contain, in addition to volcanic rocks, terrigenous–volcanogenic and terrigenous rocks: tuffites, tuffaceous siltstones, tuffstones, sandstones, and siltstones (Figs. 3–6). In the *Zavorotny* area, they form beds and interlayers among volcanic rocks, while in the *Svetly* and *Khibelen* areas, they were found in the lower parts of the sequence immediately beneath the volcanic rocks. The Zavorotny area contains, in addition to the felsic volcanic rocks, andesites.

Two U–Pb zircon datings were available for the detailed areas. An age of 1866 ± 6 Ma was obtained for volcanic rocks and granites from the *Kunerma* areas (Neymark et al., 1991) and the age of 1864 ± 11 Ma was obtained for granites from the *Srednii Kedrovy* area (SHRIMP, Poller et al., 2005).

The studied felsic volcanic rocks in all the areas are porphyritic rhyolites and dacites. Porphyroclasts comprise variable proportions of plagioclase, K-feldspar, and quartz. All phenocrysts were variably subjected to the secondary alterations. Plagioclase is replaced by sericite, albite, and, occasionally, saussurite; quartz shows wavy extinction; K-feldspar is pelitized and albi-

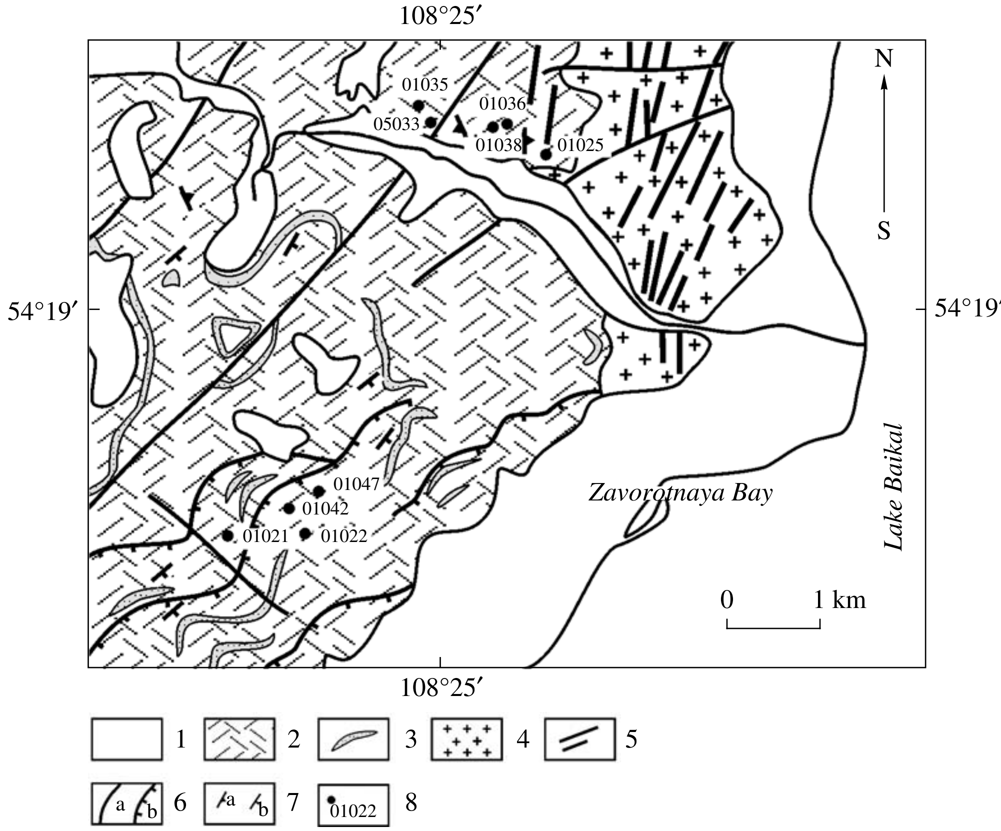

Fig. 3. Schematic geological map of the *Zavorotnaya Bay* area (*Zavorotny* area).
(1) Quaternary deposits; (2, 3) Akitkan Group (Early Proterozoic): (2) felsic volcanic rocks, (3) beds of tuffs, tuffaceous siltstones, and tuffstones; (4) granitoids of the Irel' Complex (Early Proterozoic); (5) dolerite dikes; (6) faults: (a) steep and (b) gentle; (7) dip and strike of the rocks: (a) fluidicity, (b) bedding; (8) sampling sites.

tized. The groundmass is built up of quartz and feldspar, which form felsitic, microfelsitic, poikilitic textures. Some rhyolite samples from the *Zavorotny* area have a spherulitic texture. The spherulites typically consist of fine albite and quartz fibers with an admixture of K-feldspar; in the more complex spherulites, the central part consists of chlorite which grades outward into K-feldspar, quartz, and plagioclase. Chlorite in spherulites and its aggregates in the homogeneous quartz–feldspar groundmass probably developed in some samples after glass. Some of the studied samples contain relics of mafic minerals, which are replaced by chlorite, often with an admixture of ore minerals, titanite, and epidote.

The common accessory minerals in all of the rocks are zircon and apatite. Accessory titanite occurs in felsic rocks from all areas, except for those from the *Khibelen* area. The volcanic rocks of the *Zavorotny* and *Khibelen* areas contain xenomorphic allanite in association with secondary epidote in the former. Single allanite crystals were also found in the felsic rocks of the *Svetly* and *Kunerma* areas. A distinctive feature of the felsic rocks of the *Khibelen* area is the presence of accessory rutile and titanomagnetite. The rocks from the *Kunerma*, *Svetly, and Zavorotny* areas contain xenomorphic titanomagnetite, which complete or partially exsolved into ilmenite and magnetite, in association with xenomorphic titanite aggregates (Fig. 7). At the *Srednii Kedrovy* area, the ilmenite–magnetite intergrowths associate with xenomorphic titanite (Fig. 7). Crystals of pure magnetite were found in felsic volcanic rocks from the *Kunerma* and *Svetly* areas. Monazite was identified in apatite crystal from rhyolite of the *Svetly* area.

ANALYTICAL METHODS

Representative samples of felsic volcanic rocks taken from five chosen areas in the southern part of the North Baikal volcanoplutonic belt were analyzed for major, trace, and rare earth elements. Several samples of felsic volcanic rocks were analyzed for Nd isotopic composition. U–Pb zircon dating was conducted for four samples of felsic volcanic rocks. The sampling sites for petrogeochemical, isotopic–geochemical, and geochronological studies are shown in Figs. 3–6.

Major element composition was determined by silicate analysis at the Institute of the Earth's Crust of the Siberian Branch, Russian Academy of Sciences (analysts G. V. Bondareva, E. G. Koltunova, M. M. Sma-

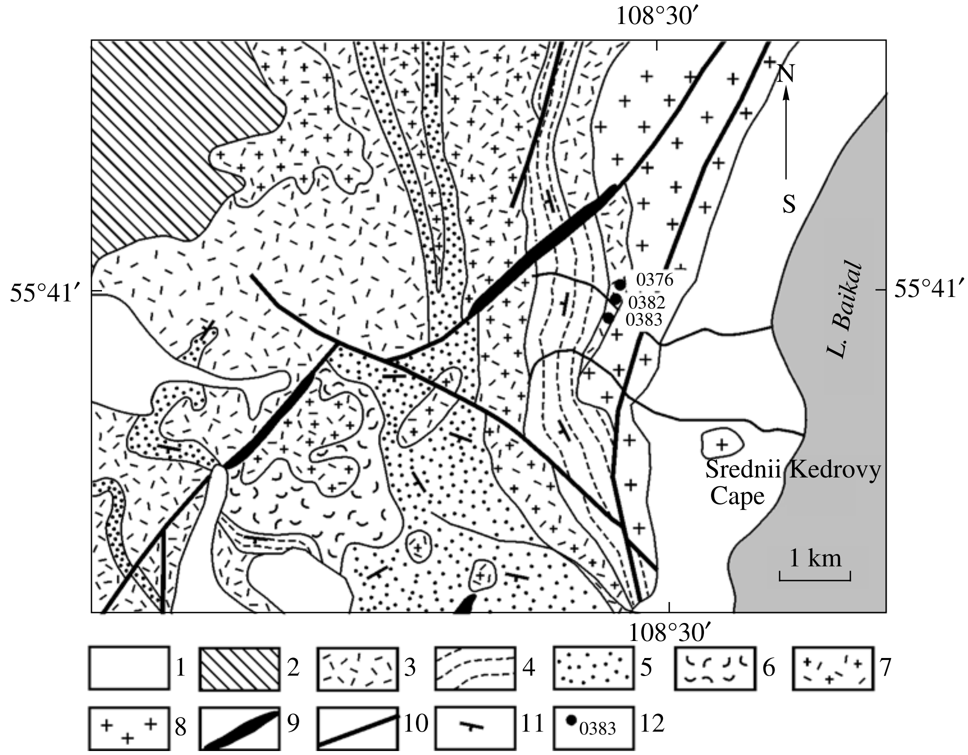

Fig. 4. Schematic geological map of the *Srednii Kedrovy Cape* area (*Srednii Kedrovy* area).

(1) Quaternary deposits; (2) Sedimentary rocks of the Baikal Group (Late Riphean); (3–7) Akitkan Group (Early Proterozoic): (3) porphyritic rhyolite cover, (4) beds of siltstones, tuffaceous siltstones, tuffs with lenses of sandstones and conglomerates; (5) unequigranular polymict to quartz sandstones, (6) ash tuffs, (7) sills and extrusions of porphyritic rhyolites; (8) granitoids of the Irel' Complex (Early Proterozoic); (9) gabbro diabase dikes; (10) faults; (11) dip and strike of bedding; (12) sampling sites.

gunova, and N. Yu. Tsareva) and, in one sample (0234, *Khibelen* area), by X-ray fluorescence at the Analytical Center of United Institute of Geology, Geophysics, and Mineralogy of the Siberian Branch, Russian Academy of Sciences (analyst N.M. Glukhova). Co, Ni, Sc, V, and Cr contents were measured by spectral method at the Institute of the Earth's Crust, Siberian Branch, Russian Academy of Sciences (analysts A.V. Naumova and V.V. Shcherban'). Other trace and rare-earth elements were determine by ICP-MS at the Center of Collective Use of the Irkutsk Scientific Center, Siberian Branch, Russian Academy of Sciences, on a VG Plasmaquad PQ-2 (VG Elemental, England) (analysts S.V. Panteeva and V.V. Markova), in accordance with the technique (Garbe-Schonberg, 1993). The apparatus was calibrated using G-2 and GSP-2 international standards. To reach the complete dissolution of all minerals, the samples for ICP-MS analysis were fused with lithium metaborate using the technique (Panteeva et al., 2003). The errors in the ICP-MS determination of trace and rare-earth elements were no higher than 5%.

The determinations of Sm and Nd and Nd isotopic composition in nine samples were conducted at the Laboratory of Geochronology of the Geological Institute, Kola Scientific Center, Russian Academy of Sciences, in Apatity, following the technique described in (Bayanova, 2004). The Nd isotopic composition and Sm and Nd contents were analyzed on a Finnigan MAT-262 (RPQ) seven-channel mass spectrometer running in a static mode. The laboratory blank was 0.06 ng for Sm and 0.3 ng for Nd. The reproducibility errors were ±0.2% (2σ) for $^{147}Sm/^{144}Nd$ and ±0.003% (2σ) for $^{143}Nd/^{144}Nd$ ratios. The Sm and Nd contents were measured accurate to ±0.2% (2σ). The measured $^{143}Nd/^{144}Nd$ ratios were normalized to $^{148}Nd/^{144}Nd$ = 0.251578, which corresponds to $^{146}Nd/^{144}Nd$ = 0.7219, and were corrected to $^{143}Nd/^{144}Nd$ = 0.511860 in the La Jolla Nd standard. During the measurements, the weighted average $^{143}Nd/^{144}Nd$ ratio was 0.511833 ± 6 (σ) in the La Jolla standard (n =11) and 0.512068 ± 15 (2σ) in the JNdi1 standard (n = 44).

The Nd isotopic composition in two samples was determined at the Geochemical Department of Max-Plank-Institute für Chemie, in Mainz, Germany. 100-m aliquots of powdered samples were decomposed with an HF and HNO_3 mixture in a microwave furnace using the three-stage technique (Todand et al., 1995). REE were extracted by standard cation-exchange chromatography, while Nd was extracted on HDHP columns, following the procedure published in (White and Patchett, 1984). The Nd isotopic composition was determined on a Finnigan MAT-261 multichannel mass-

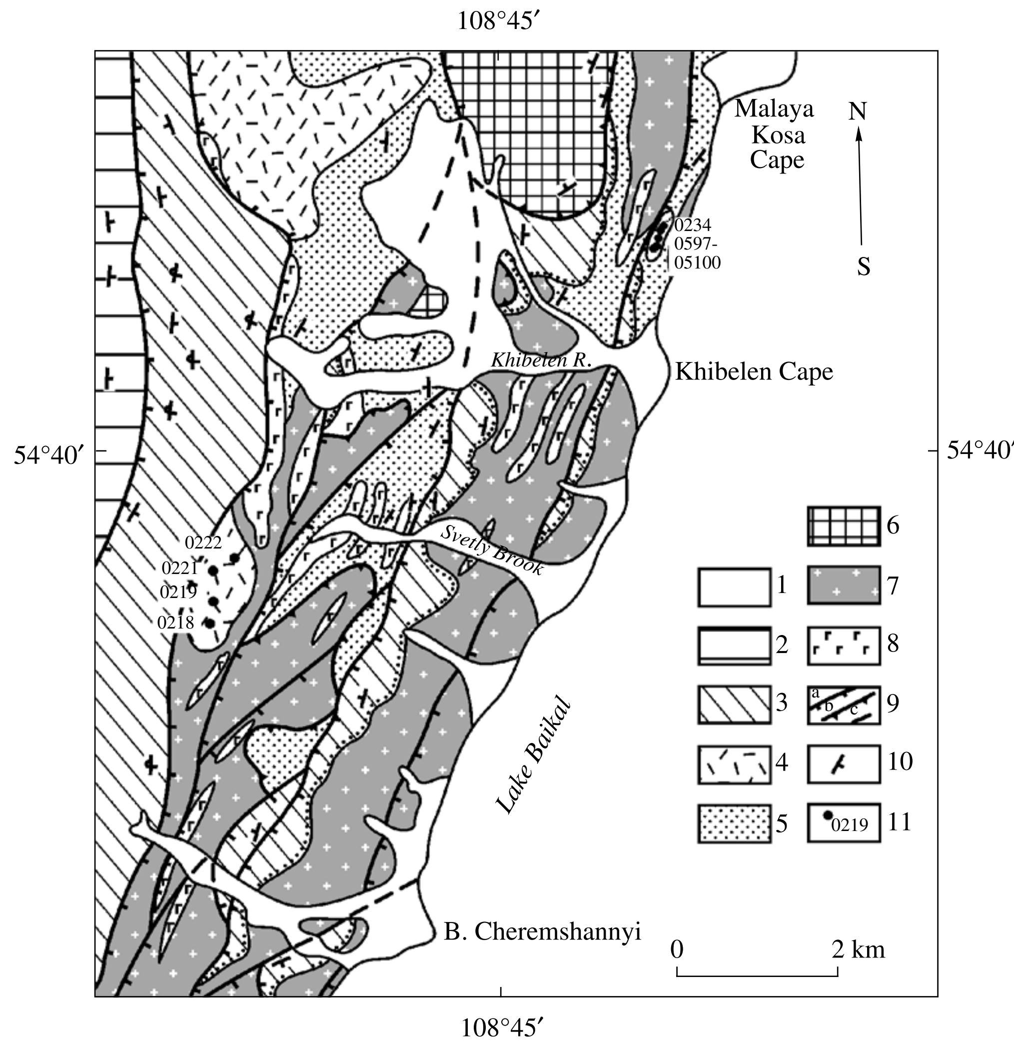

Fig. 5. Schematic geological map of the *Khibelen Cape* area (*Svetly* and *Khibelen* areas).
(1) Quaternary deposits; (2) Vendian–Early Cambrian sedimentary deposits; (3) sedimentary deposits of the Baikal Group (Late Riphean); (4, 5) Akitkan Group (Early Proterozoic): (4) porphyritic rhyolite–dacites, (5) terrigenous rocks; (6) metamorphosed volcanogenic–sedimentary rocks of the Sarma Group (Early Proterozoic); (7) granitoids (Mesoarchean); (8) gabbrodiabase dikes; (9) faults: (a) steep, (b) gentle, (c) inferred; (10) dip and strike of the bedding; (11) sampling sites.

spectrometer in a static mode. The blanks during the measurements were less than 100 pg. $^{147}Sm/^{144}Nd$ were calculated from ICP-MS determined Sm and Nd contents. The measured $^{143}Nd/^{144}Nd$ were normalized to $^{148}Nd/^{144}Nd$ = 0.251578, which corresponds to $^{146}Nd/^{144}Nd$ = 0.7219, and corrected to $^{143}Nd/^{144}Nd$ = 0.511860 in the Nd La Jolla standard. The weighed average $^{143}Nd/^{144}Nd$ ratio in the La Jolla Nd standard during the measurements was 0.511863 ± 13 (n = 30).

The $\varepsilon_{Nd}(T)$ and model ages $T_{Nd}(DM)$ were calculated using the modern values for CHUR after (Jacobsen and Wasserburg, 1984) ($^{143}Nd/^{144}Nd$ = 0.512638, $^{147}Sm/^{144}Nd$ = 0.1967) and DM after (Goldstein and Jacobsen, 1988) ($^{143}Nd/^{144}Nd$ = 0.513151, $^{147}Sm/^{144}Nd$ = 0.2136).

The geochronological U–Pb studies of zircon microsamples from three volcanic samples (samples 0219, 0285, 05100) were conducted at the Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences. The chemical decomposition of zircons and extraction of U and Pb were made using the modified Krogh technique (Krogh, 1973). The U and Pb contents were determined by isotopic dissolution using a mixed $^{208}Pb + ^{235}U$ tracer. The isotopic composition was measured using a Triton multichannel solid-phase mass spectrometer. The blank was no higher than 0.1 ng Pb. The measurement errors in U–Pb isotopic ratios were 0.3%. A correction for common lead was made in accordance to the model values in (Stacey and Kramers, 1975). The experimental data were processed using the PbDAT (Ludwig, 1991) and ISOPLOT (Ludwig, 1999) programs. The ages were calculated using

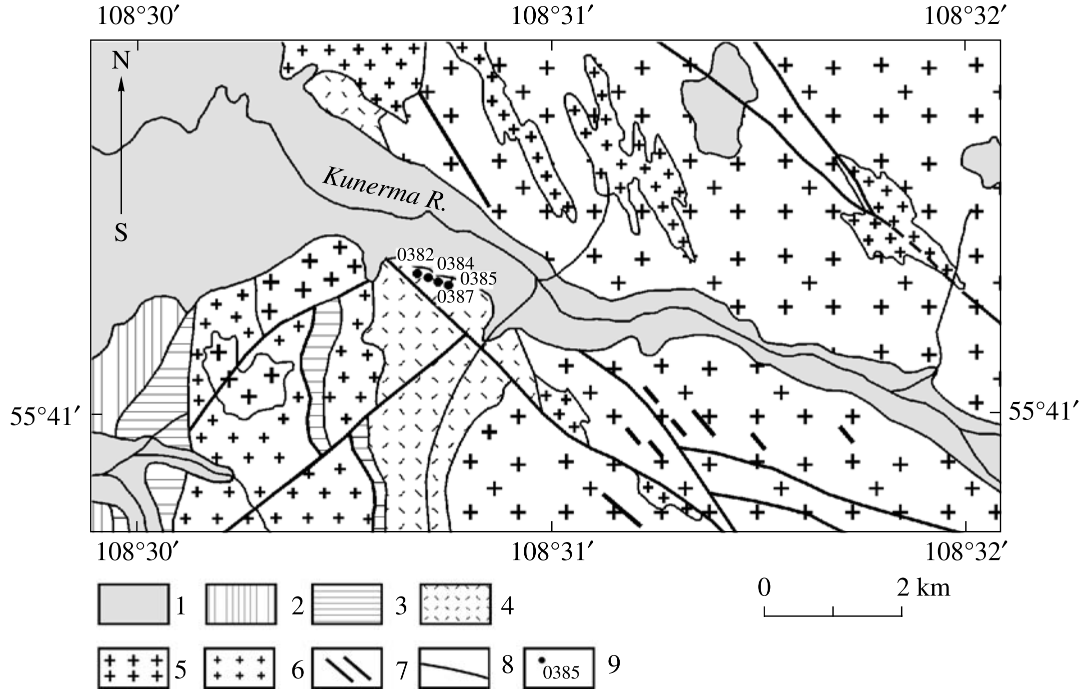

Fig. 6. Schematic geological map of the middle reaches of the *Kunerma River* area (*Kunerma* area).
(1) Quaternary deposits; (2) Vendian–Early Cambrian sedimentary deposits; (3) Sedimentary deposits of the Baikal Group (Late Riphean); (4) felsic volcanic rocks of the Akitkan Group (Early Proterozoic); (5) granitoids of phase I of the Irel' Complex (Early Proterozoic); (6) granitoids of phase III of the Irel' Complex (Early Proterozoic); (7) gabbrodiabase dikes; (8) faults; (9) sampling sites.

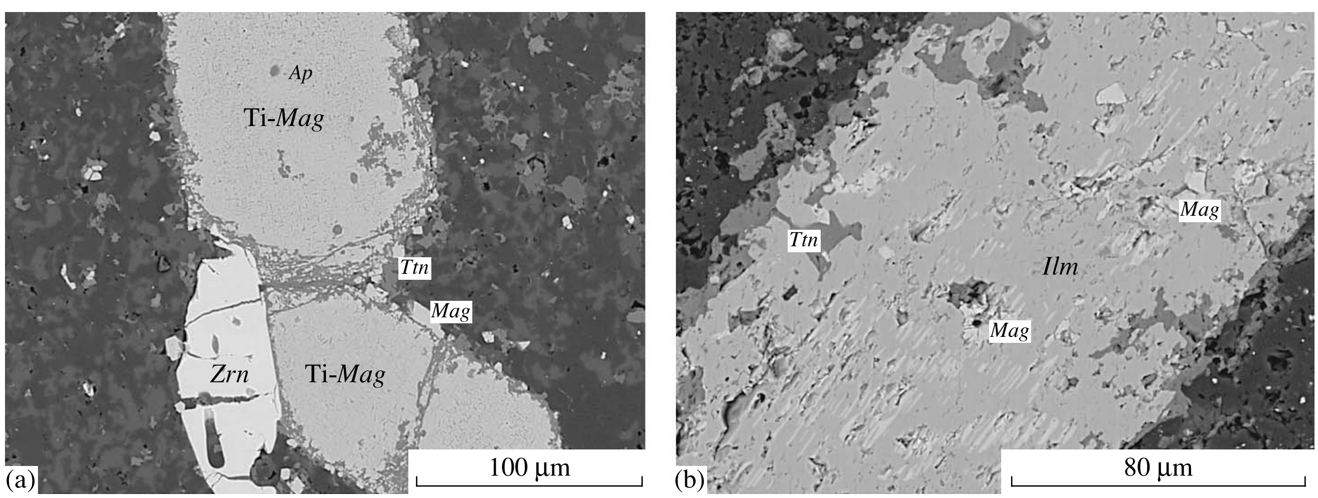

Fig. 7. BSE images of a thin section of (a) trachydacite from the *Kunerma* area, sample 0282 and (b) rhyolite from the *Srednii Kedrovy* area, LEO1430VP SEM.

generally accepted U decay constants (Steiger and Jäger, 1977). All errors are given at a 2σ level.

Single zircon U–Pb dating of volcanic sample (sample 01022) was conducted on an SHRIMP II ion microprobe at the Curtin University of Technology in Perth, Australia. Hand-picked samples were mounted in epoxy with CZ3 standard zirconand were then polished, sputtered with gold, and placed in a vacuum for 24 hours. The spot (crater) diameter was 30 μm. U–Pb ratios were determined relative to CZ3 standard zircon sample (^{238}U = 551, age of 564 Ma; Pidgeon et al., 1994). The measured ratios were corrected for common Pb using measured ^{204}Pb. Extremely low measured ^{204}Pb led to an insignificant correction, and corrections

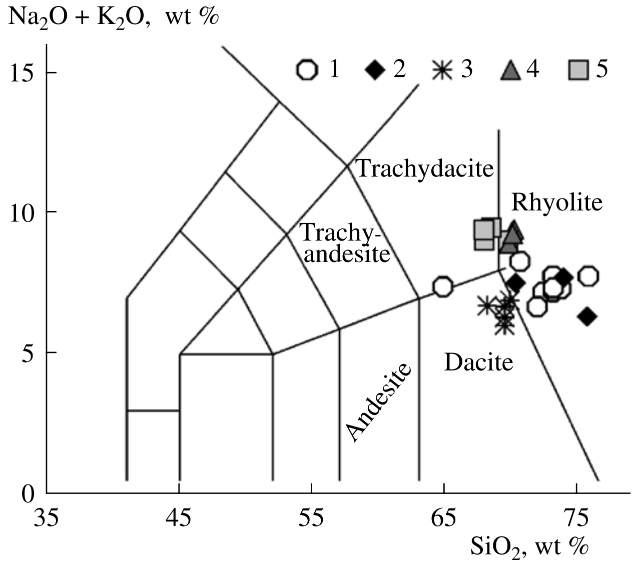

Fig. 8. Diagram (Na_2O + K_2O)–SiO_2 for felsic volcanic rocks from the southern part of the North Baikal volcano-plutonic belt.

(1) Volcanic rocks from the *Zavortny* area, (2) volcanic rocks from the *Srednii Kedrovy* area; (3) volcanic rocks from the *Khibelen* area; (4) volcanic rocks from the *Svetly* area; (5) volcanic rocks from the *Kunerma* area. Fields in the diagram are after (Le Maitre et al., 1989).

for common lead were made in accordance for model values in (Stacey and Kramers, 1975). The calculations were made using generally accepted U decay constants (Steiger and Jäger, 1977). The data were processed using the SQUID (Ludwig, 2001) and Isoplot (Ludwig, 2991b) programs. Concordia diagram for this sample was constructed according to (Tera and Wasserburg, 1972).

GEOCHEMISTRY OF THE VOLCANIC ROCKS

Chemically, the felsic volcanic rocks of the *Zavorotny* area correspond to rhyolites, except for one sample that is close to dacite. Volcanic rocks from the *Srednii Kedrovy* and *Svetly* areas are also classed with rhyolites, those from the *Khibelen area*, with dacites, while *Kunerma* area contains trachydacites and rhyolites (Fig. 8). The highest alkali contents were determined in the rhyolites from the *Svetly* area and in the trachydacites and rhyolites from the *Kunerma* area (Na_2O + K_2O = 8.7–9.1 wt %), while the dacites from the *Khibelen* area have the lowest alkali contents (Na_2O + K_2O = 6.5–7.1 wt %). Practically all of the studied samples of felsic volcanic rocks have $K_2O/Na_2O > 1$ (Table 1).

All volcanic rocks, except for rhyolites from the *Khibelen* area, are meta-aluminous or weakly peraluminous rocks (ASI = 0.75–1.14) with a high *f* (FeO*/(FeO* + MgO) = 0.76–0.96) (Table 1, Fig. 9). The dacites from the *Khibelen* area, they have a strongly peraluminous composition (ASI = 1.2–1.5) and low FeO*/(FeO* + MgO) = 0.65–0.80 (Table 1, Fig. 9).

Volcanic rocks from various areas show some differences in major and trace-element composition (Table 1), in particular, in TiO_2, Ba, Zr, Nb, Y, Th, and REE. As is seen in the Harker variation diagrams (Fig. 10), the volcanic rocks from various areas with similar SiO_2 have different major and trace element contents, which may indicate different sources of these rocks.

The lowest Ba contents were found in volcanic rocks from the *Svetly* (925–986 ppm) and *Kunerma* (995–1152 ppm) areas. The felsic volcanic rocks of the *Zavorotny*, *Khibelen, and Srednii Kedrovy* areas have higher Ba contents: 1577–2398, 1180–2149, and 1456–1813 ppm, respectively. Conversely, the highest Zr, Nb, and Y contents were found in the rhyolites of the *Svetly* area (Zr = 636–655 ppm, Nb = 55–64, Y = 76–108 ppm) and trachydacites and rhyolites of the *Kunerma* area (Zr = 651–766 ppm, Nb = 54–64 ppm, Y = 85–92 ppm) (Fig. 10, Table 1). Rhyolites from the *Srednii Kedrovy* area are noted for the lowest contents of Zr (220–288 ppm), Nb (9–12 ppm), total REE (201–261 ppm), and Th (13.0–18.2) and the highest Sr contents (164.0–188.1 ppm) (Fig. 10). The dacites from the *Khibelen* area show the highest contents of Th (47.0–50.8 ppm), La (148.3–177.8 ppm), Ce (278.7–326.5 ppm) and the lowest contents of Y (9.1–12.9 ppm) and Yb (0.35–0.88 ppm) (Fig. 10).

Except for dacites from the *Khibelen* area, the analyzed rocks show moderately fractionated REE distribution patterns (Fig. 11a). The lowest La_n/Yb_n = 7–11 and Eu/Eu* = 0.3–0.5 are typical of the rocks from the *Svetly* and *Kunerma* areas, while felsic volcanic rocks from the *Srednii Kedrovy* area have elevated ratios of these elements: (La_n/Yb_n = 14–15 and Eu/Eu* = 0.7–0.8).

The multielement spectra of all rocks have negative Nb–Ta, Sr, P, and Ti anomalies (Fig. 11b), with a significant difference in the depth of the Nb–Ta anomaly. Some differences are observed in the incompatible element spectra (Fig. 11b), in particular, Ba demonstrates a positive anomaly in the volcanic rocks of the *Zavorotny* area, its absence in the felsic rocks of the *Srednii Kedrovy* area, and negative anomalies in the rocks of the *Khibelen, Svetly*, and *Kunerma* areas, while Th exhibits a positive anomaly in the volcanic rocks from the *Kunerma, Svetly*, and *Khibelen* areas, and a weak positive anomaly in the rocks from the *Zavorotny* and *Srednii Kedrovy* areas. In addition, a well-pronounced Y–Yb–Lu negative anomaly was identified in dacites from the *Khibelen area*.

Thus, in terms of geochemical characteristics, felsic volcanic rocks from the *Svetly* and *Kunerma* areas seemed to be practically identical to one another and approximate A-type granites (Whalen et al., 1987). The volcanic rocks from the *Zavorotny* area are composi-

Table 1. Chemical composition of representative varieties of felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt and the average composition of the Mesoarchean granitoids from the basement of the belt

Components	*Zavorotny* area								
	01021	01022	01025	01042	01047	01028	01036	05035	05033
SiO_2	73.77	69.86	71.18	70.84	71.74	71.76	72.20	71.76	65.47
TiO_2	0.39	0.41	0.55	0.55	0.39	0.40	0.48	0.45	0.81
Al_2O_3	12.30	14.75	12.50	12.85	12.85	12.36	12.58	11.80	11.73
Fe_2O_3	1.03	0.50	1.18	1.71	0.72	0.73	1.40	1.76	3.01
FeO	2.76	3.34	3.30	3.29	3.10	3.54	2.87	2.69	3.81
$Fe_2O_3^*$									
MnO	0.06	0.06	0.07	0.08	0.06	0.08	0.08	0.08	0.11
MgO	0.17	0.20	0.30	0.61	0.67	0.20	0.40	0.42	2.08
CaO	0.84	0.84	1.54	1.50	0.80	1.40	1.40	1.44	3.15
Na_2O	2.90	4.98	2.62	2.31	2.53	3.01	2.42	2.76	3.15
K_2O	4.88	3.20	4.74	4.65	5.26	4.36	5.03	4.71	4.35
P_2O_5	0.07	0.06	0.13	0.08	0.05	0.07	0.10	0.07	0.14
H_2O	0.03	0.02	0.10	0.25	0.07	0.00	0.02	0.20	0.28
L.O.I.	0.96	0.87	1.22	1.38	1.39	1.12	1.08	1.22	1.88
CO_2	0.18	0.50	0.23	0.26	0.33	0.45	0.12	0.11	0.11
Total	100.34	99.59	99.66	100.36	99.96	99.48	100.18	99.47	100.08
Co	3.2	2.5	6.1	3.1	3.1	2.8	4.0	2.4	19.0
Ni	6.4	3.6	5.1	9.4	3.5	5.7	4.6	8.7	53.0
Sc	9.2	4.4	7.7	9.3	5.5	4.8	7.0	12.0	18.0
V	7.1	7.8	45.0	13.0	9.1	5.7	36.0	8.0	60.0
Cr	96.0	0.0	20.0	8.0	13.0	<6	9.2	9.3	56.0
Rb	171	106	172	144	156	122	173	157	118
Sr	104	86	133	105	67	66	123	92	173
Y	48	88	45	78	74	91	47	39	57
Zr	290	524	283	446	409	564	375	337	374
Nb	17	38	17	29	22	31	17	17	19
Ba	1656	1707	1577	2398	2095	1596	1652	1839	2126
La	78.96	119.95	80.18	140.41	113.86	115.92	82.72	78.47	78.70
Ce	149.92	249.30	152.83	258.10	217.40	240.25	157.86	158.71	155.23
Pr	18.06	30.73	18.22	32.71	26.26	31.56	19.32	20.43	20.11
Nd	67.15	107.46	65.77	120.85	98.71	109.70	67.18	68.65	75.33
Sm	11.95	21.77	10.88	20.18	16.75	22.02	11.11	12.26	12.55
Eu	1.65	3.19	1.69	4.01	3.07	3.01	1.82	1.96	2.53
Gd	9.18	13.00	8.90	17.15	13.73	13.37	9.86	9.27	9.86
Tb	1.36	2.04	1.40	2.46	2.25	2.22	1.58	1.25	1.65
Dy	8.89	11.75	8.56	13.82	13.52	12.17	8.71	8.37	9.39
Ho	1.81	2.39	1.81	2.83	2.80	2.72	1.85	1.78	1.98
Er	5.15	7.20	5.24	9.03	8.06	8.15	5.83	4.92	6.08
Tm	0.72	1.02	0.67	1.16	1.05	1.15	0.79	0.67	1.08
Yb	4.57	6.33	4.63	7.41	6.71	6.82	5.31	4.39	4.23
Lu	0.78	1.11	0.63	1.24	1.12	1.22	0.78	0.85	0.79
Hf	8.33	14.58	8.20	12.19	11.48	14.60	11.66	10.53	9.97
Ta	1.23	3.24	1.25	2.31	1.24	1.51	1.38	2.02	0.94
Th	20.53	17.21	20.24	17.94	16.92	18.75	25.78	23.63	13.08
U	5.46	4.56	5.15	4.28	4.86	5.14	6.61	7.00	3.25
ASI	1.06	1.12	1.02	1.11	1.14	1.01	1.05	0.96	0.75
f	0.96	0.95	0.94	0.89	0.85	0.95	0.91	0.91	0.76
Y/Nb	2.84	2.33	2.61	2.71	3.35	2.89	2.78	2.26	2.99
La_n/Yb_n	11.56	12.68	11.57	12.66	11.34	11.36	10.42	11.96	12.44
Eu/Eu*	0.48	0.58	0.53	0.66	0.62	0.54	0.53	0.56	0.70
T, °C	845	904	836	891	885	906	868	847	808

Table 1. (Contd.)

Components	*Srednii Kedrovy* area			*Khibelen* area				
	0376	0382	0383	0234	05097	05098	05099	05100
SiO_2	72.36	69.62	73.52	69.30	68.99	68.99	68.99	67.99
TiO_2	0.54	0.60	0.40	0.67	0.70	0.72	0.71	0.72
Al_2O_3	12.60	13.05	11.40	14.65	13.38	13.00	13.75	13.85
Fe_2O_3	1.65	2.34	1.17		1.90	2.87	1.89	1.88
FeO	2.58	2.64	2.40		1.89	1.39	1.78	1.56
$Fe_2O_3^*$				3.85				
MnO	0.05	0.05	0.03	0.03	0.04	0.04	0.05	0.04
MgO	0.46	0.73	0.69	0.87	1.52	1.55	1.50	1.77
CaO	1.38	1.92	2.11	0.63	1.16	1.02	1.32	1.04
Na_2O	3.34	3.16	2.88	1.90	3.10	1.67	2.88	1.36
K_2O	4.41	4.44	3.90	5.23	3.86	4.79	3.79	5.63
P_2O_5	0.13	0.12	0.08	0.16	0.17	0.17	0.17	0.17
H_2O	0.27	0.13	0.09		0.27	0.35	0.17	0.42
L.O.I.	0.71	0.87	0.99	2.22	2.49	2.85	2.48	2.95
CO_2			0.50		0.22	0.22	0.22	0.22
Total	100.48	99.67	100.16	99.51	99.69	99.63	99.70	99.60
Co	6.2	4.8	5.3	7.1	5.8	6.9	4.8	5.2
Ni	9.8	9.9	4.2	5.1	7.2	9.2	3.3	2.9
Sc	11.0	8.8	6.2	5.3	4.2	4.0	3.5	2.9
V	26.0	13.0	16.0	31.0	21.0	39.0	37.0	22.0
Cr	7.8	<6	<6	38.0	<6	13.0	8.7	6.7
Rb	152	152	120	234	180	215	182	282
Sr	164	177	188	121	165	111	184	116
Y	32	30	24	11	11	10	9	13
Zr	288	246	220	575	515	456	481	535
Nb	12	11	8	19	17	16	16	17
Ba	1813	1456	1460	1580	1335	1504	1180	2149
La	61.81	53.29	44.82	154.25	177.77	148.30	159.81	176.99
Ce	114.97	104.30	90.99	316.85	326.51	278.74	299.11	326.37
Pr	13.43	12.65	10.48	31.02	39.10	31.14	33.21	35.40
Nd	43.19	42.81	33.79	91.31	108.83	97.50	99.85	108.53
Sm	7.11	7.06	5.60	11.09	13.12	10.24	10.38	12.01
Eu	1.57	1.43	1.27	1.75	1.91	1.53	1.41	1.60
Gd	5.79	5.18	4.18	6.05	6.68	5.30	5.16	5.94
Tb	0.92	0.77	0.59	0.69	0.71	0.65	0.54	0.64
Dy	4.56	4.35	3.58	2.65	3.09	2.24	2.33	3.28
Ho	1.02	0.90	0.74	0.40	0.51	0.42	0.32	0.43
Er	3.23	2.75	2.34	0.97	1.13	0.80	1.08	1.24
Tm	0.31	0.42	0.35	0.16	0.15	0.13	0.15	0.18
Yb	2.93	2.37	2.05	0.69	0.88	0.35	0.48	0.49
Lu	0.47	0.44	0.37	0.10	0.15	0.08	0.08	0.12
Hf	7.40	7.62	7.14	14.09	16.24	11.25	11.85	12.81
Ta	0.77	0.78	0.63	1.05	1.31	0.72	0.80	0.83
Th	18.20	13.80	13.04	47.00	50.76	48.80	46.43	49.65
U	4.90	3.81	4.43	6.03	6.23	5.97	6.77	7.41
ASI	0.99	0.97	0.89	1.48	1.18	1.33	1.22	1.35
f	0.90	0.87	0.83	0.80	0.70	0.72	0.70	0.65
Y/Nb	2.79	2.64	2.87	0.56	0.64	0.65	0.57	0.76
La_n/Yb_n	14.12	15.01	14.62	148.72	135.41	282.87	221.02	239.84
Eu/Eu*	0.75	0.73	0.81	0.66	0.63	0.64	0.59	0.58
T, °C	833	813	800	951	910	912	908	931

Table 1. (Contd.)

Components	*Svetly* area				*Kunerma* area				AR Gr (n = 11)
	0218	0219	0221	0222	0282	0284	0285	0287	
SiO_2	69.24	69.53	69.45	69.41	69.18	67.82	68.22	67.80	68.53
TiO_2	0.46	0.44	0.47	0.47	0.52	0.50	0.47	0.51	0.42
Al_2O_3	13.90	13.90	13.60	14.10	14.60	14.30	14.05	14.20	15.53
Fe_2O_3	1.71	1.55	1.59	2.04	2.54	1.48	1.78	1.78	
FeO	2.83	2.35	2.28	1.80	1.60	3.03	2.52	2.52	
$Fe_2O_3^*$									3.78
MnO	0.05	0.03	0.05	0.05	0.03	0.08	0.07	0.08	0.05
MgO	0.63	0.66	0.39	0.36	0.60	0.53	0.42	0.48	1.11
CaO	1.19	0.85	1.62	1.33	1.56	2.00	1.95	1.80	1.82
Na_2O	3.30	3.63	3.47	3.73	3.83	3.32	3.62	3.74	4.23
K_2O	5.37	5.41	5.45	5.25	4.92	5.42	5.47	5.29	2.04
P_2O_5	0.08	0.08	0.08	0.08	0.11	0.11	0.10	0.10	0.12
H_2O	0.14	0.21	0.19	0.19	0.14	0.07	0.03	0.18	
L.O.I.	0.89	1.22	1.05	1.01	0.81	0.90	0.75	1.05	1.80
CO_2	0.66	0.22	0.66	0.11	0.00	0.11	0.11	0.11	
Total	100.45	100.08	100.35	99.93	100.44	99.67	99.56	99.64	99.43
Co	3.9	3.5	3.2	2.9	3.2	3.8	3.0	3.5	8.4
Ni	6.6	5.8	6.9	2.6	4.2	8.2	7.3	6.0	14.3
Sc	4.0	5.0	6.7	7.1	9.2	7.9	6.3	7.5	4.4
V	5.2	6.0	6.2	5.4	12.0	11.0	12.0	16.0	45.8
Cr	6.3	7.5	8.4	11.0	9.3	6.1	<6	<6	12.3
Rb	237	237	256	199	196	232	222	217	86
Sr	51	70	68	81	78	93	95	82	362
Y	108	99	101	76	88	92	87	85	8
Zr	639	642	655	636	663	717	764	651	211
Nb	63	55	64	59	64	63	54	62	6
Ba	943	925	949	986	995	1153	1085	1150	921
La	144.38	150.24	125.75	95.15	120.13	116.12	140.22	108.96	50.81
Ce	263.72	286.88	258.06	192.83	248.94	233.47	260.66	220.67	85.04
Pr	34.33	34.55	31.49	26.19	28.74	27.39	30.26	25.24	9.02
Nd	121.53	112.72	108.75	95.34	94.68	91.85	96.17	85.20	27.53
Sm	22.46	21.03	20.76	15.99	17.79	17.84	16.80	14.57	3.76
Eu	2.39	2.19	2.34	2.01	2.11	2.33	2.53	2.08	0.92
Gd	15.67	17.20	14.01	14.16	12.05	11.52	14.61	12.40	2.49
Tb	2.47	2.89	2.35	2.06	2.05	1.91	2.52	1.95	0.32
Dy	14.78	18.11	14.28	14.51	12.43	11.96	16.37	12.41	1.61
Ho	3.06	3.62	2.78	3.01	2.51	2.36	3.37	2.80	0.30
Er	8.75	11.08	8.42	9.31	7.30	6.78	9.71	8.84	0.82
Tm	1.35	1.85	1.28	1.63	1.20	0.98	1.64	1.23	0.12
Yb	8.88	9.95	8.21	8.52	8.18	6.85	9.54	7.58	0.81
Lu	1.42	1.59	1.41	1.47	1.19	1.18	1.52	1.22	0.13
Hf	18.34	17.96	17.91	20.42	17.29	16.61	20.88	19.12	5.52
Ta	2.06	3.40	4.64	4.99	4.58	3.90	3.57	6.23	0.75
Th	31.29	38.78	28.10	31.59	33.63	25.84	43.11	34.01	12.11
U	8.83	9.07	10.17	12.42	10.73	8.49	11.03	10.51	1.88
ASI	1.04	1.04	0.93	0.99	1.01	0.96	0.91	0.94	
f	0.87	0.85	0.90	0.91	0.87	0.89	0.91	0.90	
Y/Nb	1.72	1.80	1.59	1.30	1.38	1.45	1.61	1.37	
La_n/Yb_n	10.87	10.10	10.24	7.46	9.82	11.34	9.82	9.61	
Eu/Eu*	0.39	0.35	0.42	0.41	0.44	0.50	0.50	0.48	
T, °C	916	917	904	909	914	914	914	900	

Notes: ASI (mol) = Al_2O_3/(CaO + Na_2O + K_2O), f = FeO*/(FeO* + MgO) = (FeO + 0.9Fe_2O_3)/(FeO + 0.9Fe_2O_3 + MgO), T, °C is the melting temperature of the parent melts (Watson and Harrison, 1983), AR Gr are Mesoarchean granitoids. Oxides are given in wt %, elements are in ppm.

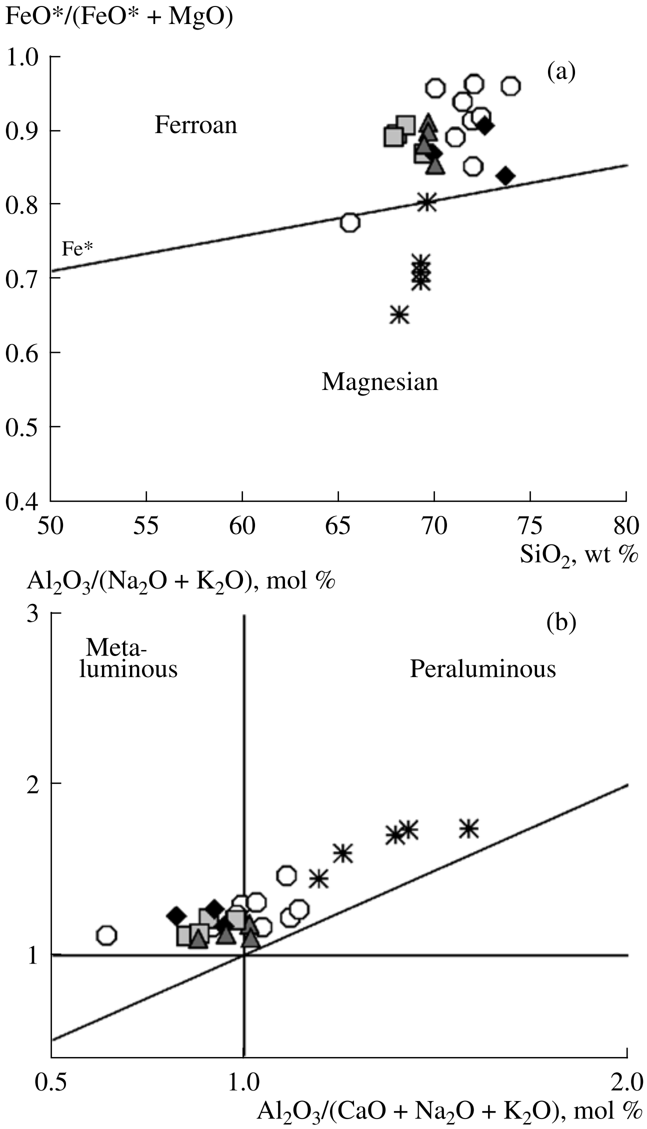

Fig. 9. Classification diagrams (a) $FeO^*/(FeO^* + MgO)$–SiO_2 (Frost et al., 2001) and (b) $Al_2O_3/(Na_2O + K_2O)$–$Al_2O_3/(CaO + Na_2O + K_2O)$ (Maniar and Piccoli, 1989). Symbols are the same as in Fig. 8.

tionally different from the volcanic rocks of the *Svetly* and *Kunerma areas*, but also more closely resemble A-type granites than granites of other geochemical types (Fig. 12). The rhyolites of the *Srednii Kedrovy* area have the lowest Zr, Nb, and REE contents, which makes them similar to I-type rather than A-type granites, though they are plotted in the field of A-type granites in the (FeO^*/MgO)–$(Zr + Nb + Ce + Y)$ diagram (Whalen et al., 1987) (Fig. 12). The dacites from the *Khibelen* area have extremely low Y and Yb contents, strongly elevated La_n/Yb_n= 135–284, and are peraluminous, which makes them similar to modern adakites and Archean TTG associations (Martin, 1999; Turkina, 2000).

Nd ISOTOPIC GEOCHEMISTRY

Sm–Nd isotopic studies were made for felsic volcanic rocks from each of the studied areas, as well as for Mesoarchean granitoids from the belt basement (Table 2). In addition, we also analyzed komagmatic granitoids from the *Srednii Kedrovy* and *Kunerma* areas.

Nd isotopic data (Table 2) indicate that all analyzed felsic rocks and komagmatic granitoids share many common features. All samples have negative initial ε_{Nd}(1850 Ma) and similar T_{Nd}(DM), varying from 2.43 to 2.91 Ga, i.e., they are at least 600 Ma older than the volcanic rocks, which indicates their formation from sources with a long crustal prehistory. At the same time, the rocks from different areas show some difference in isotopic composition. The highest ε_{Nd} were found in the volcanic and granitoid rocks from the *Kunerma* area (ε_{Nd} from –1.7 to –2.8). Rhyolites from the *Svetly* area have an insignificantly less radiogenic Nd composition (ε_{Nd} = –3.7). Similar ε_{Nd} were obtained for rhyolite from the *Zavorotny* area and dacites of the *Khibelen* area: –6.3 and from –5.2 to –5.9, respectively. The rhyolites and granites from *the Srednii Kedrovy* area have the lowest ε_{Nd} from –8.0 to –9.2.

The felsic rocks of the *Khibelen* area have the lowest $^{147}Sm/^{144}Nd$ ratio (0.0666–0.0736) and a more than 300-Ma difference between their one- and two-stage ages (Table 2). The rocks of the other areas have higher $^{147}Sm/^{144}Nd$ varying from 0.1010 to 0.1123 and a much lower age difference between one- and two-stage model ages.

The granitoids having an age of 2.88 Ga (Donskaya et al., 2005b) and located at the base of the southern part of the North Baikal belt have a lower ε_{Nd} (from −13.3 to −16.6) calculated for the formation age of felsic volcanic rocks (1850 Ma).

RESULTS OF U–Pb GEOCHRONOLOGICAL STUDIES

The Zavorotny area. Isotope dating was applied to rhyolite (sample 01022), sampling site for dating is shown in Fig. 3. Accessory zircon extracted from this sample is represented by colorless or yellowish euhedral bipyramidal crystals from 100 to 300 μm in size, with an aspect ratio of 1 : 2 and 1 : 3. Cathodoluminescence shows its well expressed magmatic zoning. Some zircons contain quartz inclusions. The measurement results of six zircon grains on a SHRIMP-II ion microprobe are shown in Table 3 and Fig. 13. The studied zircons contain 50–132 ppm U and 34–89 ppm Th. The $^{232}Th/^{238}U$ ratios vary from 0.44 to 0.72, which is typical of magmatic zircons. In the $^{207}Pb/^{206}Pb$–$^{238}U/^{206}Pb$ isotopic diagram (Tera and Wasserburg, 1972), six points of the zircons define an age of 1849 ± 11 Ma (MSWD = 0.67). Since the morphological and geochemical features of the zircon indicate its magmatic origin, the obtained age can be interpreted as the

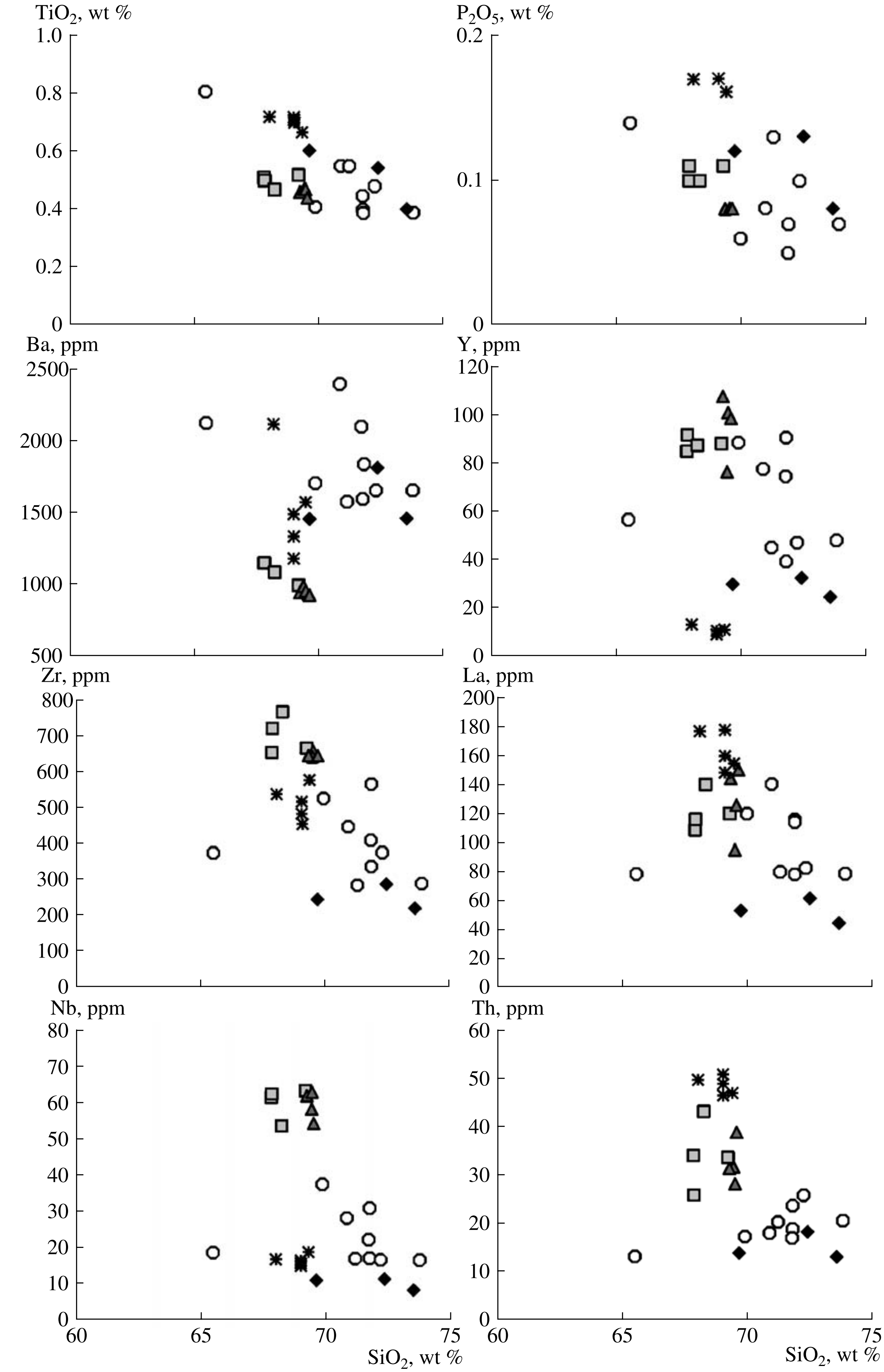

Fig. 10. Harker variation diagrams for the felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt. Symbols are the same as in Fig. 8.

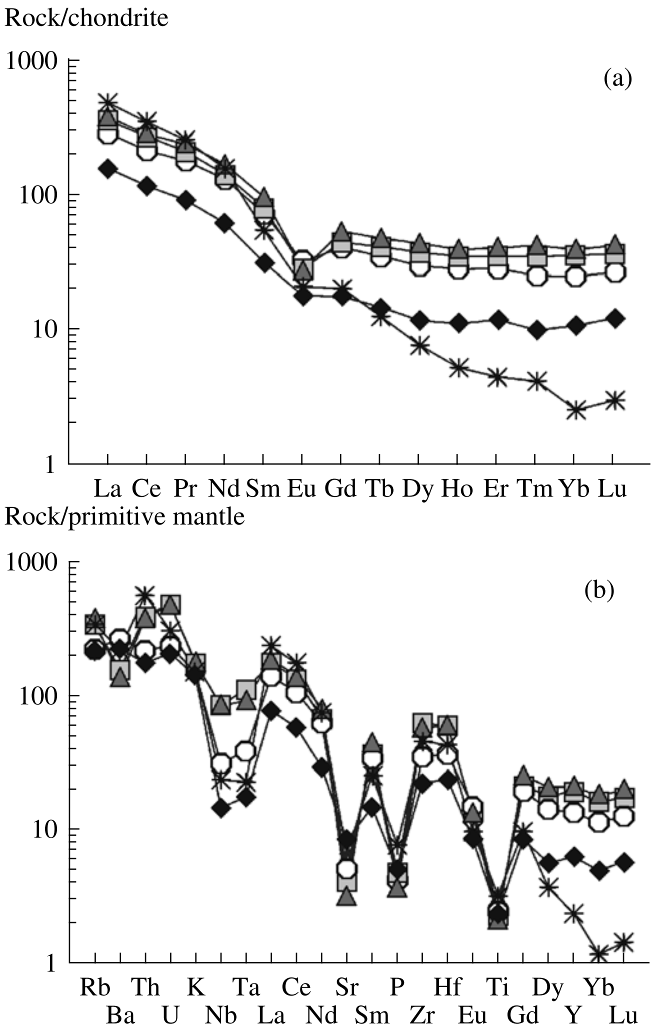

Fig. 11. (a) Chondrite-normalized (Sun and McDonough, 1989) REE patterns and (b) primitive mantle-normalized (Sun and McDonough, 1989) multielement spectra for the average compositions of felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt.

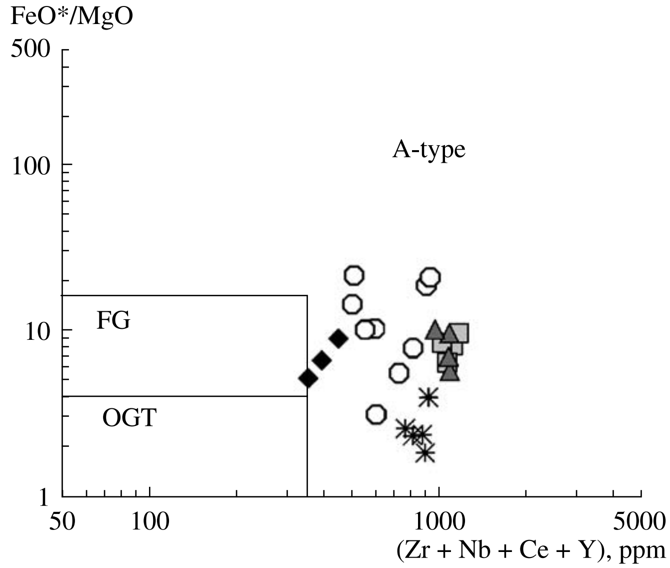

Fig. 12. Diagram FeO*/MgO–(Zr + Nb + Ce + Y) (Whalen et al., 1987) for felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt.

FG—fractionated M-, I-, and S-type granites; OGT—unfractionated M-, I-, and S-types, A-type are granites of A-type. Symbols are the same as in Fig. 8.

most accurate crystallization age of the zircons, and correspondingly, the rhyolite age.

The Khibelen area was sampled by dacite (sample 05100) (Fig. 5). Accessory zircon extracted from the rock comprises short-prismatic, semitransparent, poorly faceted yellowish crystals. The most transparent zircons without inclusions and fissures were hand-picked for analysis. An isotopic study was performed for three size fractions of the zircon, with one fraction subjected to preliminary selective dissolution following technique (Mattison, 1994). The results of the isotopic studies are shown in Table 4 and an U–Pb concordia diagram (Fig. 14a). The three-point discordia yields an upper intercept at 1877.7 ± 3.8 Ma and a lower intercept at 145 ± 29 Ma (MSWD = 0.094). Note that point 1 corresponding to the coarsest zircon fraction was rejected from isochron calculation. Zircon presumably contained an internal core with older lead. Since the zircon morphology suggests a magmatic origin, the 1877.7 ± 3.8 Ma can be interpreted as the most accurate age of zircon crystallization, and correspondingly, the age of the studied dacite.

The Svetly area was represented by a rhyolite sample (sample 0219, Fig. 5 for sampling site). Accessory zircon extracted from the sample has a prismatic euhedral habit, some grains reveal thin internal zoning. Zircons are brownish, transparent. The isotopic study was performed on five size fractions, with one size fraction subjected to selective dissolution using technique (Mattison, 1994). The results of isotopic dating are shown in Table 4 and in a U–Pb concordia diagram (Fig. 14b). The five-point discordia defines an upper intercept at 1875 ± 14 Ma and a lower intercept at 325 ± 210 Ma (MSWD = 2.2). The finest fraction was rejected from the discordia calculations. The high U content in zircon indicates the possible presence of a foreignmineral. Taking into account the zircon morphology, which suggests its magmatic origin, the upper intercept age of 1875 ± 14 Ma can be interpreted as the most accurate crystallization age of zircons and, correspondingly, as the formation age of the rhyolites.

Kunerma area. For dating, we took trachydacite (Sample 0285, sampling site in shown in Fig. 6). Accessory zircons extracted from sample are prismatic semitransparent brown grains. Our isotopic study was carried out on three size fractions of zircon, with one size fraction subjected to preliminary selective dissolution using the technique (Mattison, 1994). The results of the isotopic study are shown in Table 4 and in a U–

Table 2. Sm–Nd isotopic data on the felsic volcanic rocks from the North Baikal volcanoplutonic belt, comagmatic granitoids, and Mesoarchean granitoids from the belt basement

Sample No.	Content, ppm		$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ ±2σ	$\varepsilon_{Nd}(T)$	$T_{Nd}(DM)$, Ma	$T_{Nd}(DM\text{-}2st)$, Ma	Rock
	Sm	Nd						
Kunerma area								
0285	14.989	85.422	0.106073	0.511394 ± 23	–2.8	2478	2572	Trachydacite
0292	18.433	99.240	0.112285	0.511526 ± 29	–1.7	2433	2481	Quartz monzonite
Svetly area								
0219	18.327	103.808	0.106726	0.511352 ± 25	–3.7	2552	2652	Rhyolite
Khibelen area								
0234	11.238	92.317	0.073590	0.510839 ± 28	–5.9	2504	2828	Dacite
05100*	12.01**	108.53**	0.066621	0.510790 ± 39	–5.2	2437	2770	Dacite
Zavorotny area								
01022	16.967	96.298	0.106511	0.511218 ± 16	–6.3	2735	2863	Rhyolite
Srednii Kedrovy area								
0376	6.923	40.652	0.102950	0.511027 ± 19	–9.2	2907	3099	Rhyolite
0371*	10.44**	62.21**	0.101009	0.511065 ± 21	–8.0	2807	3000	Granite
Mesoarchean granitoids from the base of the north Baikal belt								
0265	4.102	34.572	0.071723	0.510270 ± 24	–16.6	3074	3701	Tonalite
0253	2.272	14.201	0.096708	0.510652 ± 23	–15.1	3234	3577	Plagiogranite
05104	2.566	16.912	0.091716	0.510679 ± 20	–13.3	3070	3437	Plagiogranite

Note: the values of $\varepsilon_{Nd}(T)$ and $T_{Nd}(DM\text{-}2st)$ were calculated for an age of 1850 Ma. Samples marked with (*) were analyzed at the Geochemical Department of Max-Planck-Institut für Chemie, Mainz, Germany. Other samples were analyzed at the Geological Institute of the Kola Scientific Center, Russian Academy of Sciences, Apatity. The Sm and Nd contents marked with (**) were determined by ICP-MS.

Table 3. Results of U–Pb studies of zircons from rhyolite from the *Zavorotny* area, southern part of the North Baikal volcanoplutonic belt (sample 01022)

Crystal number	$^{206}Pb_c$, %	U, ppm	Th, ppm	$^{232}Th/^{238}U$	Isotopic ratios				Age, Ma	
					$^{238}U/^{206}Pb^{*1}$	±1σ	$^{207}Pb^*/^{206}Pb^{*1}$	±1σ	$^{206}Pb/^{238}U^1$	$^{207}Pb/^{206}Pb^1$
1	0.092	114	80	0.72	3.01793	0.03903	0.11300	0.00125	1845 ± 21	1848 ± 20
2	0.119	66	34	0.53	2.97744	0.04131	0.11230	0.00102	1867 ± 22	1837 ± 16
3	0.078	85	49	0.59	2.92510	0.04121	0.11352	0.00101	1896 ± 23	1856 ± 16
4	0.192	101	65	0.66	2.98771	0.03937	0.11233	0.00094	1861 ± 21	1838 ± 15
5	0.066	132	89	0.69	3.01031	0.03846	0.11287	0.00070	1849 ± 21	1846 ± 11
6	0.025	50	21	0.44	3.05446	0.04488	0.11347	0.00118	1826 ± 23	1856 ± 19

Note: Errors are given at 1σ level; Pb_c and Pb* denote common and radiogenic Pb, respectively. The error in the calibration of CZ3 standard was 0.85% (2σ). [1] Correction for common lead using measured ^{204}Pb.

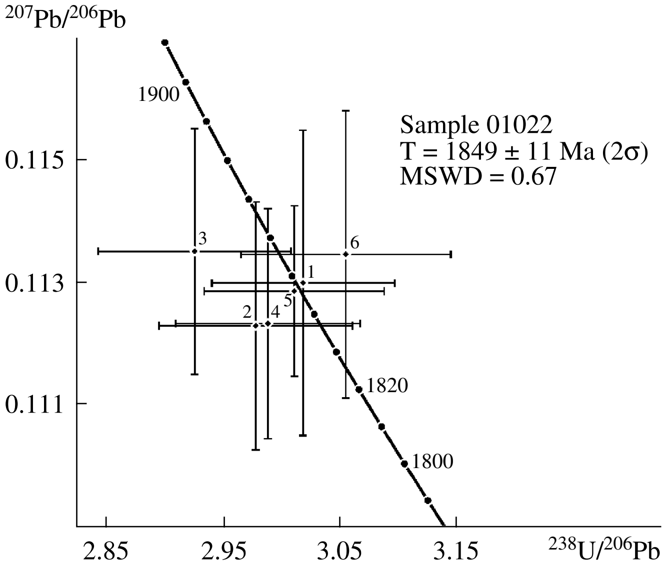

Fig. 13. Diagram $^{207}Pb/^{206}Pb$–$^{238}U/^{206}Pb$ (Tera and Wasserburg, 1972) for zircons from the rhyolite of the *Zavorotny* area, southern part of the North Baikal volcanoplutonic belt. Sample numbers correspond to ordinal numbers in Table 4. (a) Dacite from the *Srednii Kedrovy* area (sample 05100), (b) rhyolite from the *Svetly* area (sample 0219), (c) trachydacite from the *Kunerma* area (sample 0285).

Pb concordia diagram (Fig. 14c). The three-point discordia defines an upper intercept at 1870.7 ± 4.2 Ma and a lower intercept at 339 ± 44 Ma (MSWD = 0.55). The coarsest fraction has a significantly higher U contents and was not taken into account in the discordia calculations. Taking into account the magmatic morphology of the studied zircon, the age of 1870.7 ± 4.2 Ma can be interpreted as the most accurate age estimates of zircon crystallization and, respectively, the age of the trachydacites.

DISCUSSION

Qualitative Constraints on the Possible Sources of the Felsic Volcanic Rocks

Since the felsic volcanic rocks are extrusive analogs of granitoids, the methods applied in reconstructing granitoid sources can be used to determine the sources of volcanic rocks.

The negative ε_{Nd} in the felsic volcanic rocks from all studied areas (Fig. 2) could indicate a long crustal prehistory of their sources. Additional evidence in support of a sialic rather than mantle (OIB) source of the analyzed rocks is provided by elevated (>1.2) Y/Nb ratios (Fig. 15), which points to a source composition close to A-type granites (Eby, 1992). The exception is dacites from the *Khibelen* area, which have low Y/Nb (Table 1) related to the conspicuously low Y contents in these rocks.

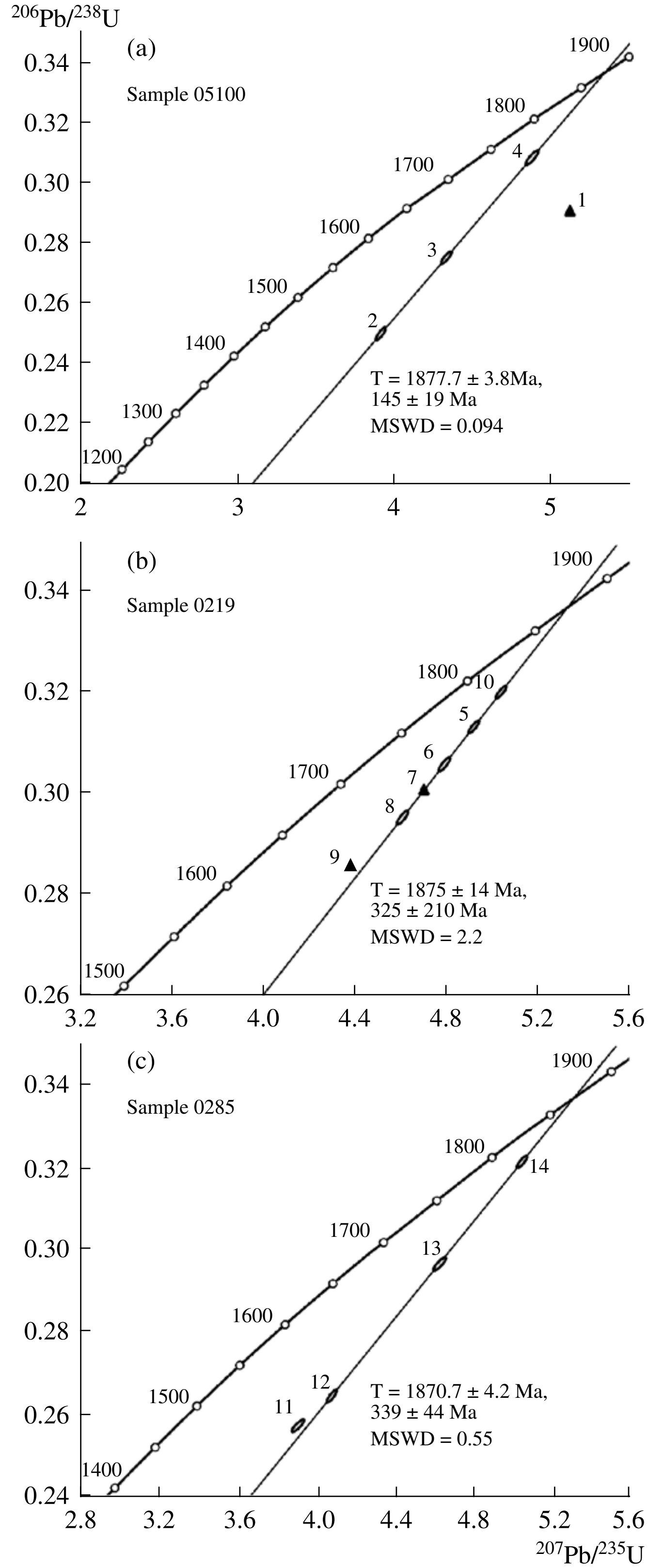

Fig. 14. U–Pb concordia diagram for zircons from felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt. Sample numbers correspond to ordinal numbers in Table 4. (a) Dacite from the *Srednii Kedrovy* area (sample 05100), (b) rhyolite from the *Svetly* area (sample 0219), (c) trachydacite from the *Kunerma* area (sample 0285).

Table 4. Results of U–Pb isotopic studies of zircons from felsic volcanic rocks of the North Baikal volcanoplutonic belt

Ordinal No.	Size fraction (μm) and its characteristics	Weight, g	Content, μg/g		Pb isotopic composition			Isotopic ratios		Rho	Age, Ma
			U	Pb	$^{206}Pb/^{204}Pb$	$^{206}Pb/^{207}Pb$	$^{206}Pb/^{208}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$		$^{207}Pb/^{206}Pb$
Dacite from the *Khibelen* area (sample 05100)											
1	+100	0.00114	218.59	82.43	1416	7.3264	2.8220	0.2911	5.1093	0.94	2060.8 ± 1.0
2	+ 75	0.00110	258.29	91.02	534	7.2304	2.2207	0.2504	3.9077	0.96	1850.9 ± 1.2
3	–75	0.00100	225.50	82.96	1050	7.9031	2.4694	0.2755	4.3238	0.96	1861.0 ± 2.0
4	–75SD	–	–	–	4610	8.4841	3.1101	0.3209	5.0358	0.95	1861.2 ± 0.9
Rhyolite from the *Svetly* area (sample 0219)											
5	+ 160	0.0023	145.43	50.26	2571	8.4010	6.3019	0.3132	4.9179	0.96	1862.4 ± 1.0
6	+ 125	0.0016	170.80	56.79	6950	8.6595	6.7767	0.3057	4.7900	0.93	1858.3 ± 1.3
7	–125…+100	0.0012	184.06	60.30	7180	8.7098	6.7339	0.3012	4.6946	0.93	1848.6 ± 1.3
8	–100…+ 75	0.0011	188.52	60.94	4560	8.6311	6.4831	0.2954	4.6030	0.94	1848.1 ± 1.2
9	–75…+ 60	0.0009	202.56	63.44	3260	8.6813	6.4286	0.2857	4.3782	0.93	1818.1 ± 1.4
10	+ 75 SD	–	–	–	10880	8.6848	7.3466	0.3202	5.0335	0.93	1864.1 ± 1.4
Trachydacite from the *Kunerma* area (sample 0285)											
11	+ 100	0.0013	383.82	118.79	870	7.9872	4.0269	0.2570	3.8890	0.78	1795.0 ± 3.0
12	+ 75	0.0014	198.53	59.05	1459	8.3071	5.8321	0.2650	4.0635	0.87	1819.2 ± 2.1
13	+ 60	0.0011	133.30	44.42	1400	8.1640	5.8050	0.2966	4.6191	0.87	1847.4 ± 2.1
14	+ 75 SD				21200	8.7443	6.6286	0.3209	5.0358	0.96	1861.2 ± 0.9

Note: [SD] is the selective dissolution of zircons using the technique (Mattison et al., 1994). [Rho] is the correlation coefficient of the $^{207}Pb/^{235}U$–$^{206}Pb/^{238}U$ isotopic ratios. All errors are given at a 2σ level.

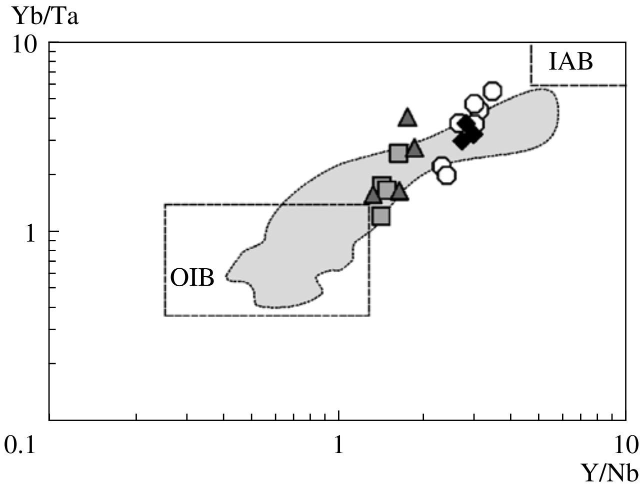

Fig. 15. Diagram Yb/Ta–Y/Nb (Eby, 1990) for felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt. Shaded in gray is the compositional field of A-type felsic rock (Cadoux et al., 2005). Symbols are the same as in Fig. 8.

At the same time, the volcanic rocks notably differ in the depth of their Nb–Ta minimum, ε_{Nd}, and T_{Nd}(DM) (Fig. 11b, Table 2). Two scenarios can be proposed to explain the isotopic variations of these volcanic rocks: (1) they could be formed from different crustal sources; and (2) the volcanic rocks were formed from a common source, while differences in their composition are related to the various contribution of juvenile mantle material to the magma generation area.

Geochemically, all analyzed volcanic rocks resemble A-type granites, though the volcanic rocks from the *Srednii Kedrovy* and *Khibelen* areas also exhibit geochemical features of I-type granites. According to experimental data (Creaser et al., 1991; Skjerlie and Johnston, 1993; Singh and Johannes, 1993), A- and I-type felsic rocks can be derived by the melting of tonalite or granodiorite material of almost identical composition. The formation of one or another type of the melt from a similar source is controlled mainly by the melting temperature and redox conditions (Turkina et al., 2006). Among the host rocks of the North Baikal volcanoplutonic belt in its southern part, only gneissic granitoids of tonalite and plagiogranite composition with an age of 2.88 Ga can be regarded as the crustal source of the felsic volcanic rocks (Donskaya et al., 2005b). No other potential crustal sources for felsic volcanic rocks were found in the North Baikal belt. Volcanic rocks with different isotopic characteristics occur not far from one another (in particular, volcanic rocks from the *Svetly* and *Khibelen* areas, Fig. 5) and, consequently, could not be derived from different crustal sources. A common crustal source for the volcanic rocks is also supported by the similar isotopic and geochemical characteristics of the volcanic rocks from the *Svetly* and *Kunerma* areas that occur at a significant distance from one anohter (Tables 1, 2). It is hardly possible that the rocks having a practically identical composition were derived from different crustal sources. As we believed earlier (Donskaya et al., 2005a), the granitoids of the *Kunerma* area that are comagmatic to the felsic volcanic rocks in this area were formed via the melting of an Early Proterozoic crustal source of monzodiorite composition. However, as was already mentioned above, the rocks of this composition and age were not found among the basement rocks of the North Baikal belt. Thus, the second scenario should be chosen to explain the isotopic differences between felsic volcanic rocks, i.e., that the rocks were formed from a common crustal source with a different contribution of mantle material to the magma generation area. The possibility of the involvement of basic material in the sources of felsic volcanic rocks is also supported by the presence of mafic magmatic rocks in the North Baikal belt. Terrigenous rocks ascribed to the Malaya Kosa Formation of the Akitkan Group contain basaltoid beds (Salop, 1964; Sryvtsev, 1986), which are thought to be among the oldest magmatic rocks of the belt.

Quantitative Constraints for Crust–Mantle Interaction during Formation of Volcanic Rocks

The granitoids that compose the basement of the North Baikal belt and are considered as the possible crustal source for felsic volcanic rocks have a less radiogenic composition (ε_{Nd} from –13.3 to –16.6, Table 2) for an age of belt formation (1.85 Ga), even as compared to the rhyolites and granites from the *Srednii Kedrovy* area with the lowest ε_{Nd} (from –8.0 to –9.2) (Fig. 16). Thus, the contribution of juvenile mantle material is required to form the studied volcanic rocks.

The contribution of mantle and crustal material to the sources of the felsic volcanic rocks can be estimated using the model of two-component mixing (Jahn et al., 2000):

$$X_m = [(\varepsilon_c - \varepsilon_{mc})Nd_c]/[\varepsilon_{mc}(Nd_m - Nd_c) - (\varepsilon_m Nd_m - \varepsilon_c Nd_c)],$$

where X_m is the content of the mantle component, Nd_m and Nd_c are the Nd concentrations in the mantle and crustal components, respectively; ε_{mc}, ε_m, and ε_c are the ε_{Nd} isotopic compositions for the obtained mantle–crustal mixture, mantle, and crustal components, respectively.

The average composition of Mesoarchean granitoids calculated for an age of 1.85 Ga (ε_{Nd} = –15.0, Nd = 30 ppm) was taken as the composition of the crustal component. The isotopic characteristics of CHUR (ε_{Nd} = 0) and Nd content of 38.5 in ocean-island basalts (Sun and McDonough, 1989) were taken as the mantle component.

The calculations showed that, at the parameters specified herein, at least ~ 33–40% of the mantle component is required to form the felsic rocks of the *Srednii Kedrovy* area and as much as ~70% and ~77–86% to produce rhyolites from the *Svetly* and *Kunerma* areas,

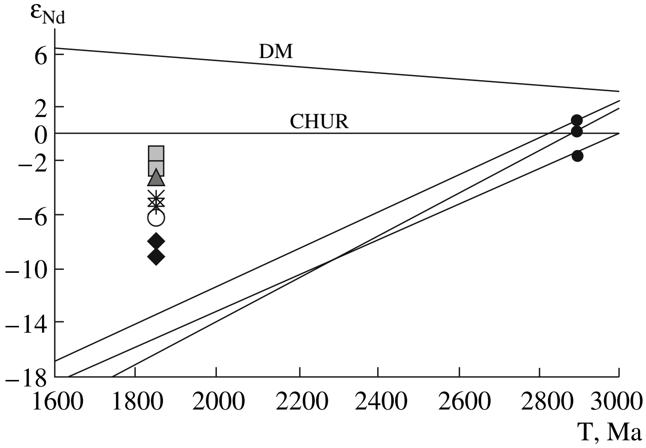

Fig. 16. Diagram ε_{Nd}–T for the felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt and Mesoarchean granitoids from the belt basement.
Symbols are the same as in Fig. 8. Solid symbols show the Mesoarchean granitoids.

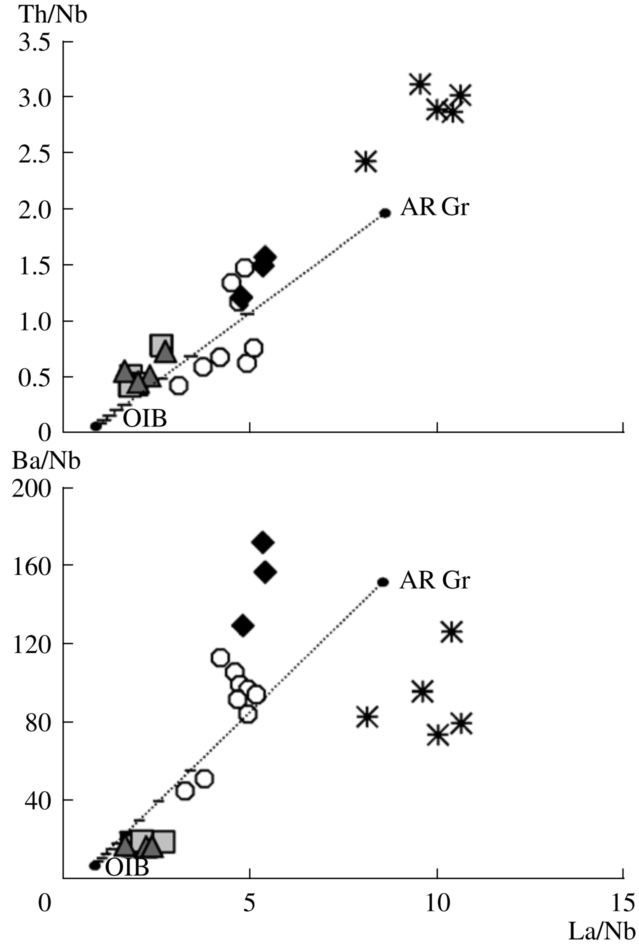

Fig. 17. Diagrams Th/Nb–La/Nb and Ba/Nb–La/Nb for felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt. The dashed line shows the mixing of Mesoarchean granitoids (AR Gr) from the belt base (crustal component) and OIB-type basalts (mantle component). Tick marks intervals correspond to 10% increments in mixing proportions. The average composition of Mesoarchean granitoids is shown in Table 1, the OIB composition is after (Sun and McDonough, 1989). Symbols are the same as in Fig. 8.

respectively. The contribution of mantle material to volcanic rocks of the *Zavorotny* and *Khibelen* areas at given parameters is estimated at ~52 and ~55–59%, respectively.

In order to check the plausibility of the calculated models of crust–mantle interaction based on isotopic data, we calculated the variations in the trace element ratios (Th/Nb, La/Nb, and Ba/Nb) during the mixing of the mantle and crustal components. These elements are not fractionated during crystallization differentiation, and hence, their ratios reflect those of the source. The trace-element composition of Mesoarchean granitoids from the basement of the North Baikal belt that applied in calculations is shown in Table 1, while the composition of OIB representing the mantle component was taken from (Sun and McDonough, 1989). The calculations were conducted using the model of simple two-component mixing (Fourcade and Allégre, 1981):

$$C_{mc} - C_c = X_m(C_m - C_c),$$

where C_{mc}, C_c, and C_m are the concentrations of elements in the obtained mantle–crust mixture, crustal, and mantle components, respectively, X_m is the content of the mantle component.

The results of the calculations are shown in Fig. 17. A satisfactory agreement was obtained between the resulted yielded by isotopic and trace-element data for the felsic rocks of the *Srednii Kedrovy*, *Zavorotny*, *Svetly*, and *Kunerma* areas. The data points of felsic volcanic rocks of the *Svetly* and *Kunerma* areas, which contain, according to the calculations, a predominant mantle component, are plotted closer to the OIB point in the Th/Nb–La/Nb and Ba/Nb–La/Nb diagrams. The data points of volcanic rocks from the *Srednii Kedrovy* area, for which the highest content of the crustal component is inferred from Nd isotopic data, are closer to the average composition of Mesoarchean granitoids. The data points of the volcanic rocks from the *Zavorotny* area fall between points of the *Srednii Kedrovy* and *Svetly–Kunerma* areas in the Th/Nb–La/Nb and Ba/Nb–La/Nb diagrams. Thus, the volcanic rocks from the *Srednii Kedrovy, Zavorotny, Svetly*, and *Kunerma* areas were formed from a common crustal source (Mesoarchean granitoids) with a variable contribution of juvenile mantle material to the magma generation area. The different contribution of crustal component to source region is also responsible for variations in the Ba and Th contents in the rocks of the *Srednii Kedrovy, Zavorotny, Svetly*, and *Kunerma* areas. At the same time, the higher Th, La, and Ce contents in the dacites of the *Khibelen* area as compared to those from other areas cannot be explained by the addition of a certain

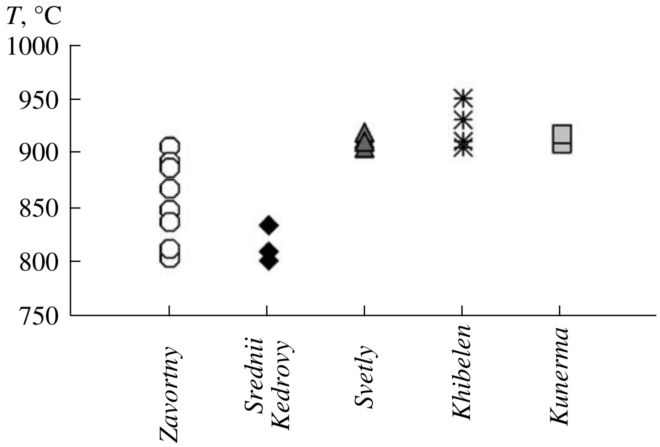

Fig. 18. Calculated melting temperatures of the parental melts of felsic volcanic rocks from various areas of the southern parts of the North Baikal volcanoplutonic belt. Temperatures were calculated after (Watson and Harrison, 1983).

amount of juvenile mantle material to the Mesoarchean granitoids considered as the crustal source but require the addition of one more component to the magma generation area. This component can be upper crustal rocks that have high Th and LREE contents and were subsided to the magma generation depth during collision. The addition of one more crustal component to the source of the Khibelen rocks can also explain the difference in Nd isotopic composition, in particular, the lower Sm/Nd ratios in these dacites as compared to the volcanic rocks from other areas.

More complex mixing models, for instance AFC (combined assimilation and fractional crystallization) were not used in our calculations, because the main purpose of our calculations was to qualitatively test the conclusions concerning the sources of felsic volcanic rocks.

Conditions of the Formation of Parent Melts and their Influence on the Compositions of Felsic Volcanic Rocks

The approximate melting temperature of parent melts of felsic volcanic rocks was estimated by zircon thermometer of Watson and Harrison (1983):

$$\ln D_{\mathrm{Zr}}^{\mathrm{zircon/melt}} = \{-3.80 - [0.85(M-1)]\} + 12\,900/T,$$

where $D_{\mathrm{Zr}}^{\mathrm{zircon/melt}}$ is the ratio of Zr content in stoichiomteric zircon to that in the melt, T, K is the temperature, $M = (\mathrm{Na} + \mathrm{K} + 2\mathrm{Ca})/(\mathrm{Al}\ \mathrm{Si})$.

The calculations showed that the highest temperatures were recorded in the felsic volcanic rocks from the *Khibelen, Svetly*, and *Kunerma* areas: 908–951, 904–917, and 900–914°C, respectively (Table 1, Fig. 18). The lowest temperatures of 800–833°C were obtained for rhyolites from the *Srednii Kedrovy* area. The volcanic rocks from the *Zavorotny* area define temperatures from 808 to 906°C (Table 1, Fig. 18).

The calculated temperatures are well correlated with the geochemical characteristics of the volcanic rocks. The highest temperatures were calculated for the volcanic rocks of the *Svetly* and *Kunerma* areas, which are geochemically the most proximate to classic A-type granites (Whalen et al., 1987) and have extremely high Zr, Nb, Y, and REE contents (Table 1). The elevated contents of these elements could be related to the high solubility of accessory phases at such high temperatures, which facilitates significant enrichment of the melts in HFSE and REE (Creaser et al., 1991). The lowest temperatures were calculated for the rhyolites from the *Srednii Kedrovy* area, which bear some geochemical signatures of I-type granites and have the lowest HFSE and REE contents (Table 1). The parental melts from which the rhyolites from the *Srednii Kedrovy* area were presumably formed under more oxidizing conditions and at an elevated H_2O activity as compared to the parental melts of other volcanic rocks. This is also supported by the presence of ilmenite–magnetite intergrowths as an accessory ore phase in the volcanic rocks from the *Srednii Kedrovy* area instead of titanomagnetite common in other volcanic rocks (Fig. 7), a feature indicating a lower water activity and melting under more reducing conditions.

The pressure in the magma generation area cannot be accurately constrained using mineral and chemical compositions of the volcanic rocks. Note only that the extremely low Y and Yb contents in the dacites of the Khibelen area and their strongly elevated $\mathrm{La_n/Yb_n}$ ratios can indicate the formation of the parent melts at pressures of ≥12–15 kbar in equilibrium with a garnet-bearing residue (Turkina, 2000).

In terms of geochemistry and formation conditions, the dacites from *the Khibelen* area are unusual rocks rarely found in nature. A distinctive feature of the dacites is a combination of geochemical signatures that are typical of high-temperature A-type granites and calculated high melting temperatures, with characteristics of peraluminous I-type tonalites, and, correspondingly, high pressures in the magma generation areas. According to experimental data (Skjerlie and Johnston, 1993), the major element composition of A-type magmas could be formed by the melting of tonalite gneisses at high temperatures (950–975°C) and high pressures (14 kbar). Under these conditions, garnet is a mandatory residual phase, which is responsible for low Y and Yb contents in the melt. The scarcity of rocks similar to the studied dacites is presumably related to the fact that their parent melts could be formed only in the collisionally thickened crust, during the earliest stages of its collapse owing to the additional heating of the lower crust. The geochemical characteristics of typical A-type granites (Whalen et al., 1987) point to their formation at high temperatures (>900°C) and pressures less than

Table 5. Ages of the magmatic rocks from the North Baikal volcanoplutonic belt (U–Pb zircon dating)

Geographical assignment	Rock	Age, Ma	Reference
Southern part of the belt (Baikal Ridge)	Dacite	1877.7 ± 3.8	This work
	Rhyolite	1875 ± 14	This work
	Trachydacite	1870.7 ± 4.2	This work
	Dacite, trachyrhyodacite, quartz monzonite, quartz diorite	1866 ± 6	(Neymark et al., 1991)
	Rhyolite	1869 ± 6	(Larin et al., 2003)
	Granite	1864 ± 11	(Poller et al., 2005)
	Rhyolite	1849 ± 11	This work
	Metasomatite	1821 ± 7	(Sobachenko et al., 2005)
Northern part of the belt (Akitkan Ridge)	Rhyodacite	1863 ± 9	(Donskaya et al., 2007)
	Trachyrhyolite	1854 ± 5	(Larin et al., 2003)
	Quartz latite, quartz monzonite porphyry	1823 ± 7	(Neymark et al., 1991)
	Trachyrhyodacite	1801 ± 22	(Sobachenko et al., 2005)

Note: the names of the dated rocks are given in accordance with the reference papers.

10 kbar (the residue lies in the stability field of orthopyroxene). Such conditions in collisional systems are provided during the final stages of their evolution, under conditions of post-collisional extension and additional heat supply caused by underplating of basic magmas at the lower crust (Whalen et al., 1987; Sylvester et al., 1989). In contrast, high pressure required to form lower temperature I-type melts with low Y and Yb contents is created in the crust thickened owing to collisional–accretionary processes, whereas additional heat is a less important parameter. Thus, a combination of several factors is required to provide favorable setting for the formation of melts with geochemical characteristics similar to those of the studied dacites of the Khibelen area: thickened crust (high pressures) and additional heating (high temperature). In addition, these rocks are formed only during the initial stages of collapse of a collisional system, since its subsequent break-up leads to a drastic decrease in loading and, respectively, pressure, thus favoring the formation of typical I-type melts. This is consistent with data obtained for the studied part of the North Baikal volcanoplutonic belt (*Evolution of the Southern…*, 2006). The dacites of the Khibelen area dated at 1878 Ma are the oldest felsic rocks of the belt, whereas other analyzed rocks with geochemical characteristics of common A-type granites have relatively younger age.

Tectonic Implications

The U–Pb zircon ages obtained for the felsic volcanic rocks from the North Baikal volcanoplutonic belt allow us to revise some aspects concerning the internal structure of the belt and its age position in the Siberian craton. At present, the obtained age of dacites from *the Khibelen* area (1877.7 ± 3.8 Ma) is the oldest age among the dated rocks of the North Baikal belt (Neymark et al., 1991; Larin et al., 2003; Donskaya et al., 2007; Poller et al., 2005; Sobachenko et al., 2005; this work) (Table 5). The ages of the dacites coincide within the measurement error with the U–Pb SHRIMP zircon age of granulites from the Kaltygei Cape (1876 ± 6 Ma, Poller et al., 2005) in the Western Baikal region (50 km south of the North Baikal belt). The age of the granulites of the Kaltygei Cape, together with U–Pb age of syncollisional granites of the Kocheryak Complex (1910 ± 30 Ma; Bibikova et al., 1987), is considered as the time of the main collisional events in the Western Baikal part of the Siberian craton (Gladkochub et al, 2006). Thus, the onset of formation of North Baikal volcanoplutonic belt was almost coeval to the termination of the collisional events, i.e., the development of the belt was initiated during the syncollisional collapse of the collisional orogen (Fedorovskii and Sklyarov, 2007). The dacites of the *Khibelen* area rest without any visible unconformity on the terrigenous rocks of the lower parts of the Akitkan Group, which are ascribed to the Malaya Kosa Formation of the North Baikal belt. Hence, the age of these terrigenous rocks is close to that of the dacite.

The examination of the ages obtained on the felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt, as well as on the granites that are comagmatic to volcanic rocks on the *Srednii Kedrovy* area (Poller et al., 2005), showed that they were formed within a range of about 30 Ma. As we already mentioned, the oldest ages were obtained on the dacites from the *Khibelen* area. The rhyolites from the *Zavorotny* area have the youngest age (1849 ± 11 Ma) among all analogous rocks in the southern part of the North Baikal belt. This age practically completely overlaps with the age of rhyolites of

Gol'tsovskii volcano (1854 ± 5 Ma; Larin et al., 2003). The latter cuts across the sedimentary rocks of the Chaya Formation of the Akitkan Group, whose terrigenous rocks are traditionally considered as the youngest stratified rocks of the belt. The ages younger than ~1850 Ma obtained for individual small volcanic bodies in the North Baikal belt reflect the attenuation of volcanism within the belt after its formation (Donskaya et al., 2007). Taking into account that all of the studied rocks were previously ascribed to the common Khibelen Formation (Khibelen complex), the internal structure of the belt in its southern part is controversial. We have a high confidence only in the time of initial development of the belt and the termination of large-scale magmatic activity within it. However, a correlation of rocks of the belt between these events is complicated, because their ages overlap within the measurement errors (Table 5). Therefore, the internal correlation of the rocks of the North Baikal belt is a separate problem, which is beyond the scope of this paper.

CONCLSUIONS

(1) The U–Pb zircon dating of felsic volcanic rocks from the southern part of the North Baikal volcanoplutonic belt showed that the rocks previously ascribed to a single stratigraphic unit (Khibelen Formation of the Akitkan Group or Khibelen Complex) significantly differ in age. The rocks of the Khibelen Formation appeared to include the oldest dated rocks of the North Baikal belt (dacite, *Khibelen* area, 1877.7 ± 3.8 Ma) and one of the youngest dated volcanic rocks (rhyolite, *Zavorotny* area, 1849 ± 11 Ma). The age values obtained on two other samples of volcanic rocks (rhyolite from the *Svetly* area and trachydacite from the *Kunerma* area) fall between these two dates (1875 ± 14 Ma and 1870.7 ± 4.2 Ma, respectively). Our U–Pb geochronological studies indicate a longer duration of the formation of most of the magmatic and terrigenous rocks than was assumed previously (Larin et al., 2003).

(2) It was established that the felsic volcanic rocks previously ascribed to the single Khibelen Formation (Complex) differ not only in age but also in geochemical and isotopic characteristics. The felsic rocks from the *Svetly* and *Kunerma* areas have the highest analyzed Zr, Nb, Y, REE and ε_{Nd}. The rhyolites from the *Srednii Kedrovy* areas have the lowest HFSE and REE contents and ε_{Nd}. The felsic volcanic rocks from the *Zavorotny* area have intermediate values of indicated geochemical parameters between those of the volcanic rocks of the *Svetly–Kunerma* and *Srednii Kedrovy* areas. Dacites from the *Khibelen* area have specific geochemical features: extremely high LREE and Th contents at lowest Y and Yb contents, which make them different from volcanic rocks from other studied areas.

(3) The felsic volcanic rocks from various areas of the southern part of the North Baikal volcanoplutonic belt could be formed by the melting of a Mesoarchean crustal source of tonalite composition with a variable contribution of juvenile mantle material, under various conditions of magma generation. Based on isotopic data, the contribution of juvenile mantle material accounted for ~33–40% for the rhyolites of the *Srednii Khibelen* area, ~52% for the volcanic rocks of the *Zavorotny* area, ~55–59% for the dacites of the *Khibelen* area, ~70% for the rhyolites of the *Svetly* area, and ~77–86% for the rocks of the *Kunerma* area. The temperatures of the parent melts for felsic volcanic rocks of the *Khibelen, Svetly, and Kunerma* areas were estimated at 908–951, 904–917, and 900–914°C, respectively; the melting temperatures for rhyolites from the *Srednii Kedrovy* area were within 800–833°C, while those of the *Zavorotny* area, 808–906°C. The parent melts of the volcanic rocks of the *Srednii Kedrovy* area were formed under higher oxidizing conditions than the volcanic rocks from other studied areas.

(4) The geochemical characteristics of the *dacites* from the *Khibelen* area indicate an input of an upper crustal component with high Th and LREE contents, in addition to the Mesoarchean granitoids and juvenile mantle material. The extremely low Y and Yb contents in the dacites of the *Khibelen* area imply the generation of their melts under pressures of ~12–15 kbar in equilibrium with a garnet-bearing residue. The parent melts of the dacites of the *Khibelen* area were presumably formed in the collisionally thickened crust during the earliest stages of its collapse at the additional significant heating of the lower crust. The parental melts for felsic volcanic rocks from other studied areas in the southern part of the North Baikal volcanoplutonic belt are closer to the classic A- and I-type granites and were formed during subsequent stages of the collapse (post-collisional collapse).

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