Mineralogy, Geochemistry, and Sr–Nd–Pb Isotope Systematics of Late Cenozoic Basanites of the Borozdin Bald Mountain (Khentei Ridge, Southern Transbaikalia)

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Abstract—We have estimated the *P–T* conditions of formation of basaltoid melts: *P* = 1.15–1.06 GPa and *T* = 1379–1293 ºC. Olivine pyroxenites (Ol + Cpx + Grt) are assumed to be a mantle source for nepheline-normative basanitic melts. During ascent, the melt trapped mantle xenoliths, which were disintegrated into olivine and augite xenocrysts. A decrease in pressure and temperature led to the crystallization of highly magnesian (Mg# = 86) olivine and diopside and Timgt + Ilm ± Pl phenocrysts. Then, the Ol + Cpx + Timgt + Pl microlite paragenesis formed. Alkaline aluminosilicates (acid plagioclase + nepheline + leucite) were the last to crystallize in the rock interstices. The presence of residual glass indicates that the last stage of crystallization took place under subsurface conditions. The volcanic area of southern Transbaikalia (Khentei Ridge) resulted, most likely, from the mantle plume impact on the lithosphere. The age of this area is estimated at 3.51 Ma. PREMA was the main mantle source for these volcanics, and the contribution of HIMU was strongly subordinate. In geochemical features the studied volcanics correspond to mafic rocks of oceanic islands. They are similar in composition to alkali basalts of the South Baikal volcanic area.

Keywords: basalts, pressure, temperature, parental melt, crystallization sequence, plume, Pb–Nd–Sr isotopy

INTRODUCTION

In the late Cenozoic, volcanism was widely manifested in the southern framing of the Siberian craton. Its products formed a number of fields of different sizes and durations of formation (Laverov, 2008). The volcanic fields extend from the Udokan lava plateau in the north via the South Baikal and South Hangayn areas to the Dariganga Plateau (Mongolia) in the south and form the Central Asian volcanic province (CAVP) (Yarmolyuk et al., 1995). Several stages of magmatic activity (from 34 Ma to recent time) were established in almost all volcanic areas (Yarmolyuk et al., 1998, 2003, 2011). In addition, all areas are characterized by a predominance of mafic lavas, mostly of alkaline and subalkaline types. Almost all Cenozoic volcanics are localized in rift structures.

The Baikal Rift Zone (BRZ) is the largest rift structure in Central Asia (Kiselev et al., 1979; Kovalenko et al., 1999, 2009; Logachev, 2003; Voronstov and Yarmolyuk, 2004; Buslov, 2012). Near the southwestern flank of the BRZ there is the South Baikal volcanic area (SBVA) (350×

450 km), where several stages of magmatism, from late Oligocene (34–24 Ma) to late Pliocene–Pleistocene–Holocene (<3 Ma) (Yarmolyuk et al., 2001), were recognized (Yarmolyuk et al., 2001). The SBVA volcanics are mafic alkaline rocks, from potassic trachybasalts to basanites and hawaiites (Yarmolyuk et al., 2003). In geochemical features they are similar to oceanic-island basalts.

Southern Transbaikalia is one of the areas of Cenozoic volcanics not related to rifting, but such rocks are less spread here as compared with the SBVA. Cenozoic alkali basaltoids with a great number of xenoliths have been long known in the Khentei Ridge in southern Transbaikalia. They occur as lava flows and nappes of small area and thickness in the headwaters of the Chikoi, Chikokon, and Burkal rivers. The first data on these volcanics were reported by Kostyakov et al. (1969). Later, the hosted xenoliths, including garnet and spinel lherzolites, were examined (Rasskazov, 1987; Ashchepkov et al., 1996; Litasov and Taniguchi, 2002).

The chemical composition of the basaltoids has been poorly studied. We found results of only three wet-chemistry analyses of these rocks (Kostyakov et al., 1969; Rasskazov, 1987) and no data on the composition of their minerals. The goal of this research was to determine the composition of parental magmas, the isotope sources, and the *P–T* condi-

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tions of formation of basanites of the Borozdin Bald Mountain. For this purpose we studied the composition of minerals, obtained the first data on the isotope composition of rocks, and determined the absolute age of volcanism.

ANALYTICAL METHODS

Analyses were carried out for sample fragments without visible xenoliths. The compositions of minerals were determined by electron probe microanalysis (EPMA) on a JXA-8200 (JEOL Ltd., Japan) probe. The contents of rock-forming elements were measured by X-ray fluorescence analysis on a SRM-25 (ZAO Nauchpribor, Orel) multichannel X-ray spectrometer according to the technique of Afonin et al. (1984). The spectrometer was calibrated against the ST-1A (trapp, Russia) and JB-1 (basalt, Japan) standard samples (Finkel'shtein et al., 1984). The contents of trace elements were determined by ICP MS on an ELEMENT-2 (Finnigan MAT, Germany) double-focusing mass spectrometer with a magnetic sector. The analytical techniques are given elsewhere (Mysovskaya et al., 2009; Saibatalova et al., 2010).

The samples for isotope analysis were chemically prepared in clean chemical rooms of the Common Use Center for Isotope-Geochemical Studies of the Institute of Geochemistry, Irkutsk. The isotope composition of Sr and the contents of Rb and Sr were determined by the isotope dilution method, using a mixed $^{85}Rb + {}^{84}Sr$ tracer. The isotope composition of Nd and the contents of Nd and Sm were determined by the isotope dilution method, using a mixed $^{149}Sm + {}^{150}Nd$ tracer.

The isotope compositions of Sr and Nd were measured on a Finnigan MAT-262 seven-channel mass spectrometer at the Center for Geodynamics and Geochronology of the Institute of the Earth's Crust, Irkutsk. The measured isotope ratios were normalized to $^{88}Sr/^{86}Sr = 8.37521$. The validity of the obtained Sr isotope data was estimated by analysis of the NBS-987 and BCR-2 standard samples, which yielded $^{87}Sr/^{86}Sr = 0.710254 \pm 7$ (2SD, $n = 45$) and $^{87}Sr/^{86}Sr = 0.705011 \pm 14$ (2SD, $n = 7$), respectively. The validity of the obtained Nd isotope data was estimated by analysis the JNdi-1 and BCR-2 standard samples, which yielded $^{143}Nd/^{144}Nd = 0.512107 \pm 4$ (2SD, $n = 35$) and $^{143}Nd/^{144}Nd = 0.512629 \pm 8$ (2SD, $n = 18$), respectively.

The isotope composition of Pb in the rocks was measured on a Finnigan MAT-262 mass spectrometer. Correction for mass discrimination was made using the double isotope dilution method. The isotope composition was determined in the sample and in its mixture with a $^{207}Pb + {}^{204}Pb$ tracer. The $^{207}Pb/^{204}Pb$ ratio in the tracer was close to 1 in accordance with recommendations (Rudge et al., 2009). The validity of the determined isotope composition was estimated by analysis of the NBS-981 standard sample with the following normalized isotope ratios: $^{208}Pb/^{204}Pb = 36.7029 \pm 35$; $^{207}Pb/^{204}Pb = 15.4911 \pm 16$; and $^{206}Pb/^{204}Pb = 16.9372 \pm 6$ (2SD, $n = 22$).

The $^{40}Ar/^{39}Ar$ isotopic age was determined by the step heating method with measurement of the isotope composition of released argon on an ARGUS VI mass spectrometer (Thermo Scientific, Great Britain) at the Center for Geodynamics and Geochronology of the Institute of the Earth's Crust, Irkutsk. The analytical technique is described in detail by Ivanov et al. (2017).

THE GEOLOGIC POSITION AND PETROGRAPHY OF THE STUDIED ROCKS

The Borozdin Bald Mountain is located in the headstream of the Borozdinaya and Kharchevka rivers, right tributaries of the Burkal River (Fig. 1) (Kostyakov et al., 1969). The peak of this bald mountain is composed of compact black and dark gray volcanics, which overlie the rocks of the Ingoda Formation (C_{1-2}). The volcanics contain mantle xenoliths of spinel and garnet lherzolites composed of high-Mg olivine, augite, spinel, and relict garnet as well as olivine and clinopyroxene xenocrysts. The rocks were sampled from massive volcanics of bedrock exposures at the peak of the Borozdin Bald Mountain. For study we chose samples without visible xenoliths.

The sampled rocks contain large inclusions of olivine and clinopyroxene; their total content varies from 0 to 4–5 vol.%. Massive volcanics are composed of varieties of porphyritic and aphyric textures (Fig. 2). As shown below, the porphyritic texture is due to large phenocrysts and xenocrysts of olivine (0.5–1.0 mm, Fig. 2) and clinopyroxene (0.4–0.5 mm, Fig. 2). The large pyroxene grains are much scarcer than the large olivine grains and amount to ~1.5 vol.%. Seldom, plagioclase phenocrysts ≤0.3 mm in size (sample B11/3) are found. Both the phenocrysts and the xenocrysts of olivine and clinopyroxene are fresh. The rock groundmass is of intersertal texture and is formed by fine grains of olivine (≤0.03 mm), clinopyroxene (≤0.01 mm), ore minerals 40–50 μm in size, thin plagioclase laths (150–200 μm), and a finely crystallized aggregate with glass relics. According to the probe microanalysis results, the aggregate contains nepheline, leucite, apatite, and felsic-plagioclase microlites no larger than 50 μm. The presence of glass relics suggests that the final stage of crystallization proceeded in subsurface conditions. Postmagmatic alterations are minor, as evidenced from the low LOI values.

THE AGE

We performed Ar/Ar dating of the fine-crystalline basaltoids. The Ar release spectrum of the sample B15 shows excess argon, which is present, most likely, in xenogenic pyroxenes (Fig. 3). This is indicated by the significant increase in Ca/K with increasing temperature (Table 1), when a K-containing phase (leucite) is exhausted and a Ca-containing phase (clinopyroxene) is involved in melting (Buikin
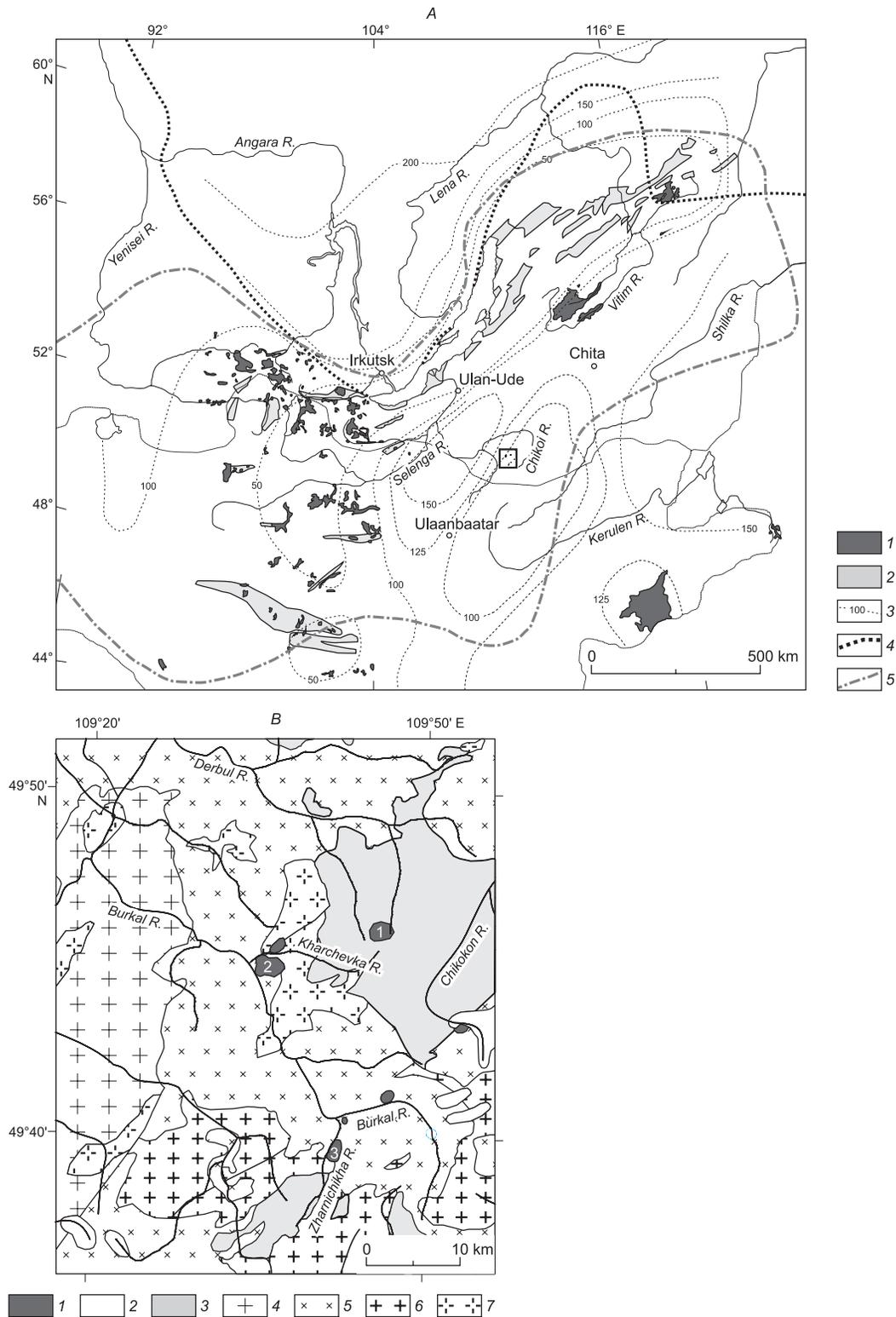
Fig. 1. Position of Cenozoic volcanics in the Baikal–Mongolian region (Litasov and Taniguchi, 2002) (*A*); geological map of the area of Cenozoic alkali basalts of the Dauria–Khentei Ridge, simplified and supplemented after State... (2012) (*B*). *A*: *1*, Cenozoic volcanics; *2*, Cenozoic sedimentary basins and depressions; *3*, lithosphere thickness, km (Zorin et al., 1990); *4*, border of the Siberian craton; *5*, border of the area of anomalous low-velocity mantle (Zorin and Rogozhina, 1978). Rectangle marks the position of alkali basaltoids of the Dauria–Khentei Ridge; *B*: *1*, late Cenozoic alkali basalts; *2*, Quaternary fluvioglacial deposits; *3*, Carboniferous deposits of the Ingoda Group; *4*, granitoids of the Malkhan complex (PZ$_1$); *5*, granitoids of the Daur complex (P$_1$); *6*, granitoids of the Kyra complex (J$_{1–2}$); *7*, granitoids of the Asakan–Shumilovka complex (J$_{2–3}$). Numerals mark outcrops of Cenozoic alkali basaltoids: 1, Borozdin Bald Mountain; 2, Kharchevka; 3, Zharnichikha.
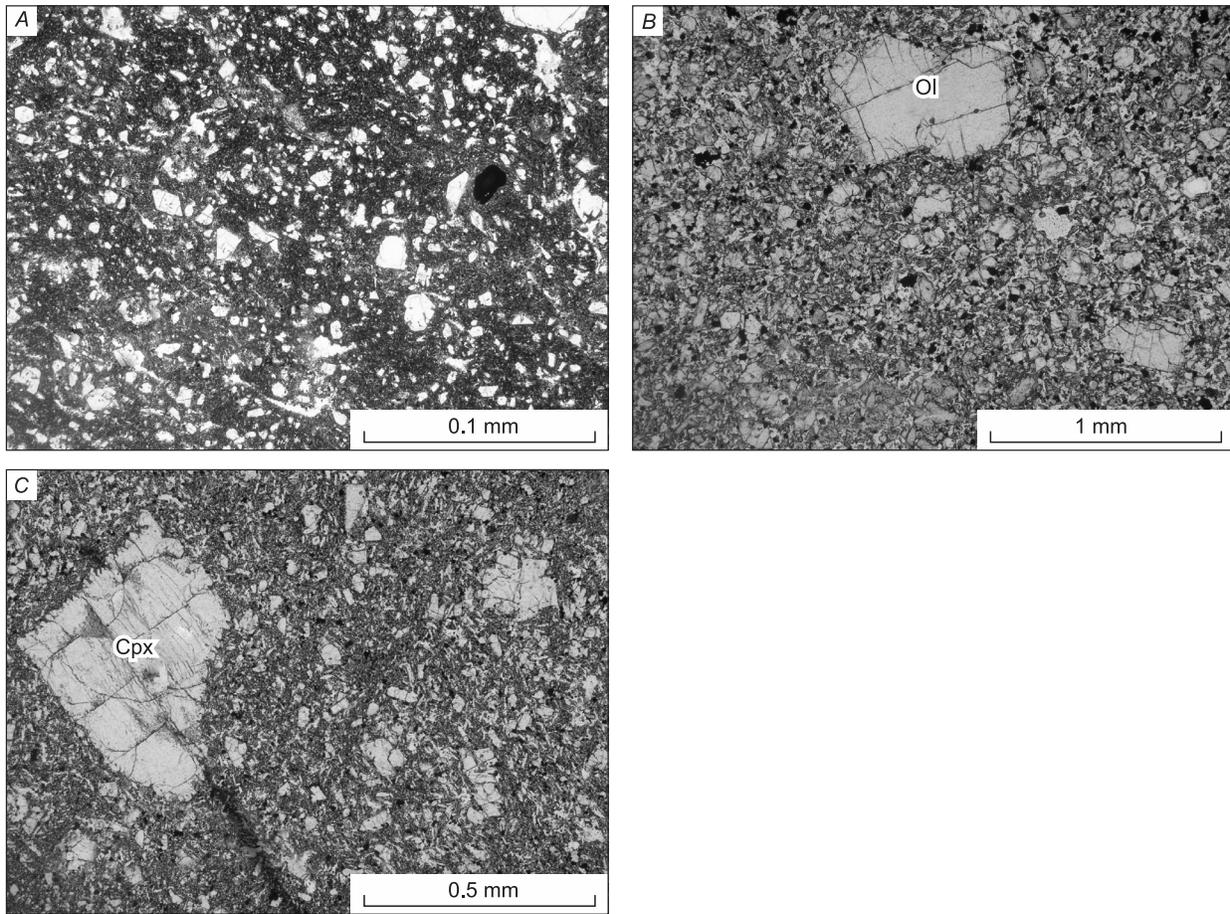
Fig. 2. Thin-section photomicrographs. *A*, Aphyric volcanic rock, sample B11/2, white is fine olivine grains (Ol); *B*, porphyritic volcanic rock with olivine xenocryst, sample B26; *C*, porphyritic volcanic rock with clinopyroxene xenocryst (Cpx), sample B15.

et al., 2010). Thus, the age of the sample is determined from two low-temperature steps. Moreover, olivine might contain microinclusions that crystallized from a basanitic melt (Ivanov et al., 2018). The obtained date of 3.51 ± 0.27 Ma is reliably the upper temporal bound of volcanism or, most likely, corresponds to the time of formation of basanites. The age of the complete melting of the sample, i.e., its K–Ar age, is 7.18 ± 0.37 Ma. Earlier, K–Ar dates of 7.9 and

5.7 Ma were obtained for two melanephelinite samples from the same area, with no analytical error mentioned (Polyakov and Bagdasar'yan, 1986). Obviously, they show an older age because of the presence of excess argon in xenogenic pyroxene in these samples.

MINERAL COMPOSITION

Olivines. Table 2 lists the representative compositions of olivines of three morphologic varieties: (1) ellipsoidal or irregular grains with wavy edges, up to 1 mm in size (Fig. 4), (2) subhedral grains up to 0.4 mm in size, and (3) fine euhedral grains smaller than 50 μm (Fig. 5). Figure 6 shows variations in the compositions of olivine grains of different sizes and shapes. The Mg# value of olivines varies over a wide range of values and gradually decreases from core to rim in large grains. A positive Fo–NiO correlation and a negative Fo–CaO correlation are observed. The content of CaO in the cores of zoned olivine grains is <0.2 wt.% and is higher, within 0.20–0.34 wt.%, in the rims of large olivine grains and in the phenocrysts with Mg# ≈ 0.77.
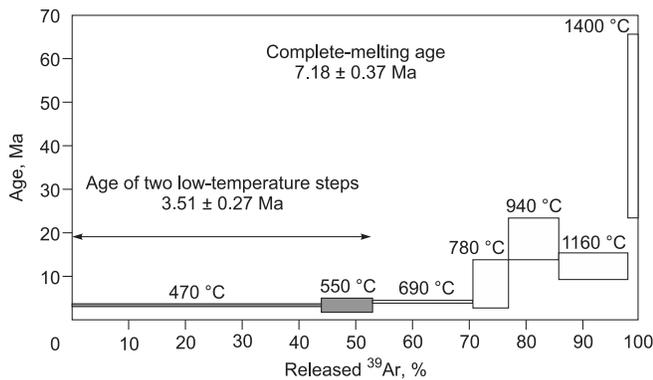
Fig. 3. Results of ^{40}Ar/^{39}Ar dating of the sample B15.

Table 1. Results of $^{40}Ar/^{39}Ar$ dating

Step	T, °C	^{40}Ar	± 1σ	^{39}Ar	± 1σ	^{38}Ar	± 1σ	^{37}Ar	± 1σ	^{36}Ar	± 1σ	Ca/K	$^{40}Ar_{rad}/^{39}Ar$	± 1σ	Age, Ma	± 2σ
1	468	949.10	3.61	888.85	3.03	15.21	0.22	86.22	2.76	1.903	0.196	0.36	0.43	0.02	3.51	0.27
2	552	186.38	0.80	181.70	0.79	5.79	0.23	160.16	2.77	0.369	0.218	3.26	0.42	0.09	3.43	1.51
3	687	1 656.95	5.67	354.79	1.29	6.76	0.21	606.86	3.32	4.918	0.202	6.33	0.53	0.01	4.35	0.17
4	783	167.37	0.71	129.79	0.61	2.00	0.19	275.40	2.80	0.111	0.209	7.85	1.03	0.32	8.45	5.27
5	936	460.15	1.68	174.98	0.79	3.73	0.23	1 724.08	6.57	0.204	0.226	36.46	2.28	0.30	18.60	4.82
6	1156	469.14	1.73	249.23	0.97	6.05	0.23	7 212.48	25.12	0.310	0.238	107.07	1.51	0.19	12.34	3.13
7	1396	229.44	0.84	39.32	0.43	0.80	0.19	1 864.68	7.03	0.045	0.193	175.45	5.49	1.30	44.48	21.09

Table 2. Results of representative analyses of olivine phenocrysts

Component	B15				B11/2					
	Xenogenic olivine Ol 1									
	Large zoned grain									
	core	intermediate zone	rim	core		intermediate zone			rim	
SiO$_2$, wt.%	40.68	41.72	40.63	38.98	41.02	40.38	40.51	40.48	39.24	38.57
FeO*	8.93	8.91	17.15	18.47	9.22	9.22	9.53	13.11	15.75	21.03
MnO	0.13	0.12	0.36	0.40	0.10	0.10	0.10	0.19	0.28	0.41
MgO	49.26	48.64	42.94	42.16	50.08	50.08	50.17	46.99	44.92	39.48
CaO	0.13	0.12	0.17	0.21	0.13	0.13	0.09	0.07	0.15	0.29
NiO	0.39	0.40	0.24	0.18	0.36	0.36	0.34	0.33	0.29	0.15
Total	99.52	99.61	101.5	100.4	100.9	100.3	100.7	101.2	100.6	100.3
Mg#	0.91	0.90	0.82	0.80	0.91	0.91	0.90	0.86	0.84	0.77

Component	B04						B26			
	Olivines of generation I (basaltic), Ol 2						Olivine of generation II, Ol 3			
	Large zoned grain						Fine homogeneous grain			
	core	core	intermediate zone			rim				
SiO$_2$, wt.%	39.01	40.12	39.05	38.53	39.28	38.76	38.27	37.96	38.01	38.41
FeO*	13.54	14.32	15.80	18.37	18.57	19.37	21.28	20.98	21.02	21.10
MnO	0.17	0.19	0.19	0.30	0.30	0.34	0.40	0.41	0.39	0.43
MgO	46.96	46.19	44.26	42.62	42.01	41.85	40.31	40.53	40.37	40.13
CaO	0.19	0.19	0.22	0.22	0.22	0.26	0.27	0.26	0.27	0.27
NiO	0.25	0.20	0.20	0.19	0.16	0.16	0.15	0.14	0.13	0.14
Total	100.1	101.2	99.72	101.2	100.5	100.74	100.7	100.3	100.2	100.5
Mg#	0.86	0.85	0.83	0.79	0.80	0.79	0.77	0.77	0.77	0.77

Note. Electron probe microanalyses were carried out by O.Yu. Belozerova. Mg# = Mg/(Mg + Fe^{2+}), mol.%. Hereafter, FeO* is total iron.

Olivine grains with Mg# > 0.90 in the core have rounded wavy edges (Fig. 4*A*). Similar outlines of olivine grains are observed in the experiments on the interaction of peridotites with SiO$_2$-undersaturated basaltic melts (Shaw and Dingwell, 2008). The grains are zoned: The NiO content and Mg# decrease and the CaO content increases from core to edge (Fig. 4*A*). The core has a zone of homogeneous composition typical of equilibrium olivines of mantle peridotites. The CaO content in this zone is still the same as in mantle peridotite, whereas the NiO and Fo contents are lower. This phenomenon might be due to the different rates of element diffusion at the olivine–basanite contact. The lower rate of CaO diffusion was established during the modeling of assimilation of olivine xenocrysts by basalts, based on the observed olivine composition trends (Costa and Dungan, 2005). Dissolution of crystals can lead to their rounded shape (Donaldson, 1985). The diffusion zoning in olivines indicates that the rate of dissolution was lower than the rate of diffusion (Liang, 2000). These grains are obviously xenogenic olivines (Ol 1) of mantle peridotites.

Euhedral and subhedral olivines show different Fo, NiO, and CaO concentration profiles: The core zone of homoge-
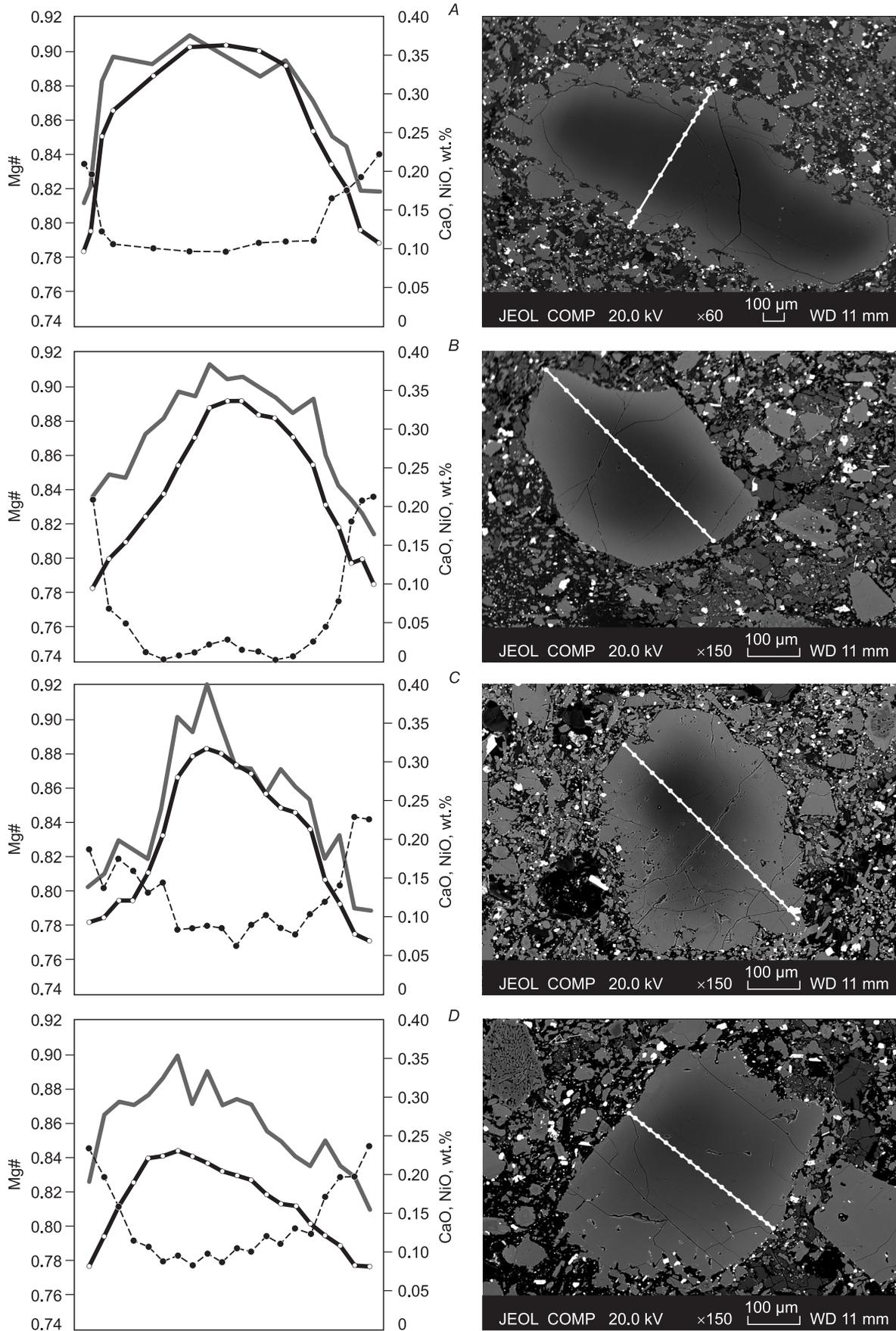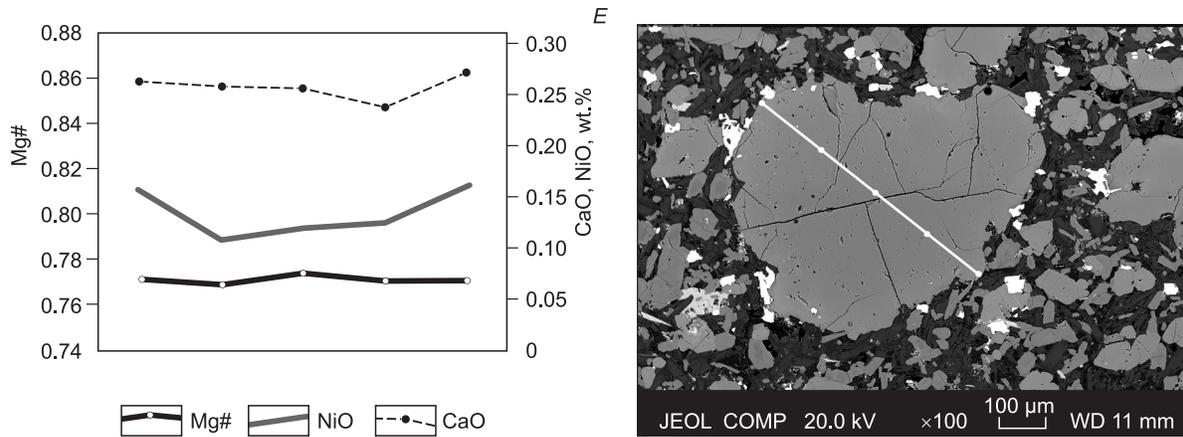
Fig. 4. CaO and NiO contents vs Mg# of olivine and BSE images of xenogenic olivines (Superprobe JXA-8200 (JEOL Ltd, Japan) electron microprobe; accelerating voltage of 20 kV, beam current of 20 nA, and beam diameter of 1 μm). *A*, BB1; *B*, B26; *C*, B15; *D*, B26; *E*, B26. Here and in Fig. 5, white lines and dots mark the scanning profiles and analysis points, respectively.

neous composition has close contents of all elements; a decrease in the Fo and NiO contents to the edge of the grain is accompanied by an increase in the CaO content. These profiles might be due to crystallization of olivine from melt (Fig. 5). In the core of these olivine grains (Ol 2) Mg# varies from 0.86 to 0.81, the CaO content, from 0.10 to 0.18, and the NiO content, from 0.19 to 0.32 wt.%. The fine homogeneous euhedral olivine grains (Ol 3) have Mg# ≈ 0.77, CaO ≈ 0.25 wt.%, and NiO ≈ 0.15 wt.%. In composition they are similar to the edges of zoned xenocrysts and phenocrysts (Fig. 6). Accordingly, these olivines crystallized from the last portion of basaltic melt.

Thus, we discovered several varieties of olivine: xenogenic (Ol 1), basaltic (Ol 2, generation I), and fine homogeneous grains in the groundmass (Ol 3, generation II). Note that the formation of several generations of olivine from basaltic melt is often observed (Al'mukhamedov et al., 1983, 1985; Zolotukhin et al., 1996).

Pyroxenes. All samples contain phenocrysts (<500 μm in size) and megacrysts (0.7 mm) of clinopyroxene. Fine grains of pyroxene are also found in the rock groundmass. Results of representative analyses of these pyroxenes are given in Table 3. The phenocrysts are diopsides (Cpx 2); the large ones are zoned, with Mg# = 0.82–0.74. The contents

Table 3. Results of representative analyses of clinopyroxene phenocrysts

Component	B11-2										B15			B26	
	Cpx 1		Cpx 1			Fine grain of the groundmass	Cpx 1	Cpx 2	Fine grain of the groundmass			Cpx 2	Cpx 1
	core	rim	core	rim	rim		core					core	
SiO₂, wt.%	53.30	52.87	52.23	51.63	50.10	50.38	52.60	50.98	46.88	49.88	47.68	51.97	53.84
TiO₂	0.35	0.41	0.89	1.81	2.50	2.71	1.11	1.01	3.58	3.20	2.99	0.95	0.58
Al₂O₃	6.77	5.47	6.26	3.98	4.86	5.01	7.27	1.73	7.84	4.60	5.12	7.19	0.88
Cr₂O₃	0.10	0.17	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	0.75	0.64	B.d.l.	B.d.l.	B.d.l.
FeO*	5.08	4.18	6.81	5.34	6.43	6.50	6.89	7.93	6.69	6.65	7.58	7.14	6.25
MnO	0.13	0.10	0.12	0.12	0.14	0.11	0.11	0.29	0.10	0.13	0.10	0.12	0.14
MgO	17.12	15.22	15.42	13.43	13.12	12.77	14.69	13.54	11.56	12.99	12.73	14.22	15.77
CaO	16.20	21.64	16.29	22.46	22.30	21.99	15.78	22.09	22.21	21.87	22.57	16.20	21.52
Na₂O	1.94	0.94	2.58	1.10	1.14	1.28	2.09	0.70	1.07	1.10	0.49	2.03	0.81
Total	101.0	101.0	100.6	99.9	100.6	100.8	100.5	98.3	100.7	100.0	99.3	99.81	99.79
Wo, %	36.75	46.90	40.67	47.17	49.89	49.51	37.55	47.96	52.00	49.03	49.89	38.87	44.96
En	54.05	45.91	47.19	46.61	40.84	40.00	48.63	40.91	37.65	40.52	39.16	47.48	45.86
Fs	9.20	7.19	12.71	5.64	9.28	10.50	13.81	11.13	10.35	10.46	10.96	13.65	9.18
Mg#	0.865	0.845	0.816	0.798	0.786	0.778	0.789	0.776	0.784	0.794	0.782	0.777	0.838

Note. Hereafter, B.d.l., below the detection limit of 0.1%.
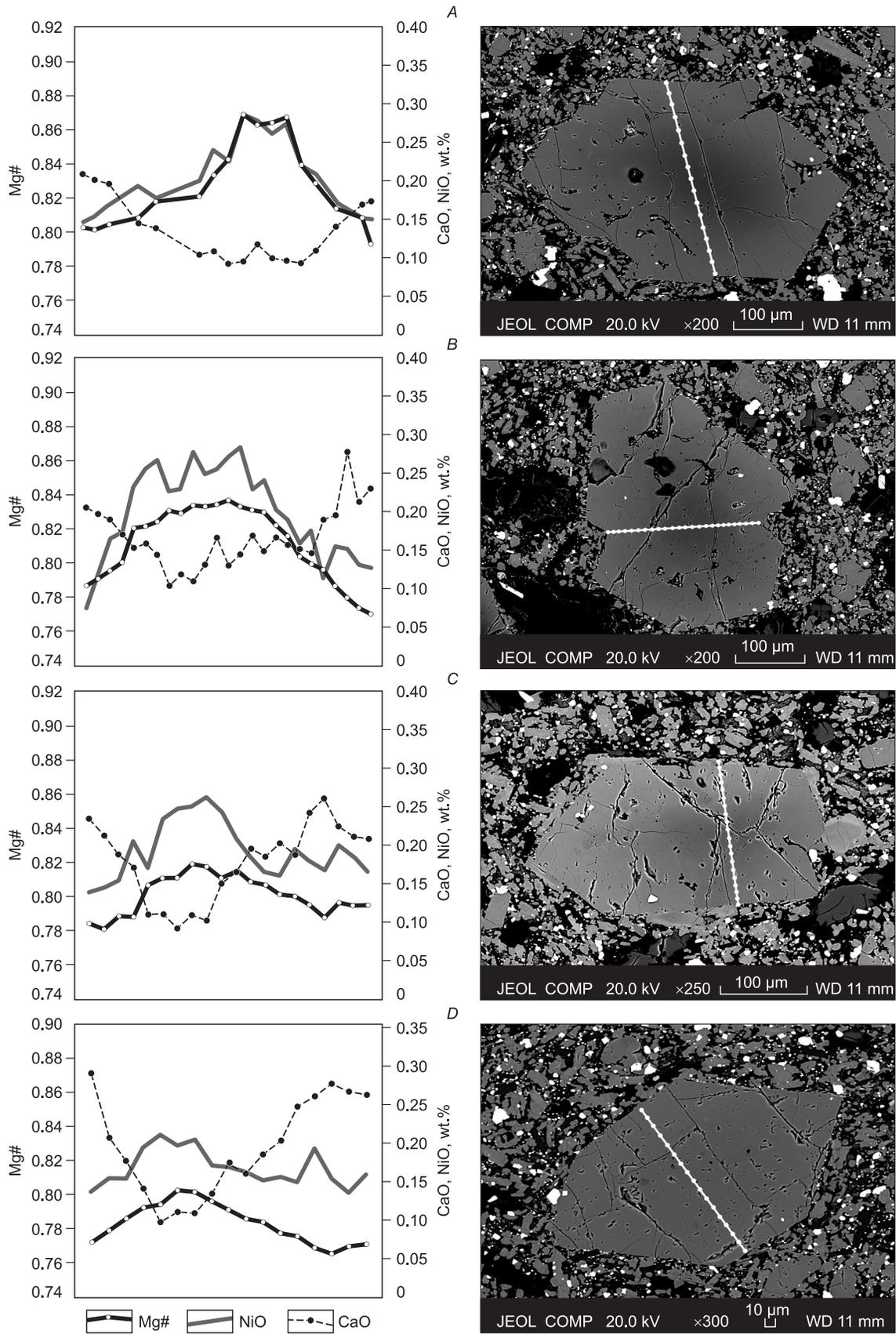
Fig. 5. CaO and NiO contents vs Mg# of olivine and BSE images of euhedral olivines (Superprobe JXA-8200 (JEOL Ltd, Japan) electron microprobe; accelerating voltage of 20 kV, beam current of 20 nA, and beam diameter of 1 μm). *A*, B14-26; *B*, B14-15; *C*, B14-15; *D*, B14-26.

Fig. 6. CaO and NiO contents vs Fo of olivines of different generations. *A*, *B*, sample B26; *C*, *D*, sample B15; *E*, *F*, sample B11/2. *1*, olivine from xenolith; *2*, xenogenic olivine from peridotite (Ol 1); *3*, basaltic olivine (Ol 2); *4*, olivine from the groundmass (Ol 3). Horizontal scale marks the content of forsterite molecule.

Fig. 7. Photomicrograph of a clinopyroxene megacryst.

of Al_2O_3 and TiO_2 increase and the content of SiO_2 and the Mg# value decrease from core to edge of the large pyroxene grains and from large to fine homogeneous grains, whereas the contents of CaO and Na_2O remain constant (Fig. 7). Fine pyroxene grains from the rock groundmass correspond in composition to diopsides.

The clinopyroxene megacrysts (Cpx 1) are augites and seem to be xenocrysts. They have a reaction rim (both over the grains and in their cracks) composed of clinopyroxene of different composition and fine grains of ore clinopyroxene (Fig. 8). The Mg# value of the megacrysts varies from 0.86 to 0.76, i.e., is almost in the same range as the Mg# value of diopsides. The composition points of augites form individual fields in variation diagrams (Fig. 7). The augites differ from diopside phenocrysts in higher contents of SiO_2 and Na_2O and lower contents of CaO and TiO_2. They are rich in Al_2O_3 and Na_2O and are poor in Cr_2O_3. The high Al^{VI}/Al^{IV} ratio of the megacrysts indicates their high-pressure formation. Clinopyroxene megacrysts of similar composition were earlier found in late Cenozoic melanephelinites of the Burkal River basin; their formation was explained by crystallization of alkali basalts in magma-feeding channels (Ashchepkov et al., 1996). Clinopyroxene of the rim is similar in composition to the phenocrysts (Table 3), which indicates its formation by crystallization from basanitic melt. Reaction rims were earlier observed in pyroxene from peridotite and pyroxenite xenoliths from alkali basalts (Wang et al., 2012). Therefore, it is not ruled out that the studied augites are not abyssal phenocrysts of basanites of the Borozdin Bald Mountain but might be minerals of disintegrated mantle pyroxenites. They cannot be clinopyroxenes of mantle peridotites because of the low contents of Cr_2O_3.

Feldspars and feldspathoids. The composition of feldspars and feldspathoids is given in Table 4. Three varieties of these minerals have been recognized: (1) rare unzoned

Table 4. Results of representative analyses of feldspars and feldspathoids

Component	B11/3								B15	B26			BB1		B11/3	
	Plagioclase phenocrysts				Plagioclase laths										Plagioclase microlites	
SiO_2, wt.%	56.19	55.41	56.60	55.77	53.12	52.26	53.40	52.86	54.59	52.69	52.21	51.94	54.29	54.61	63.16	63.14
Al_2O_3	27.00	28.88	26.73	26.83	28.53	29.64	28.84	28.74	27.66	29.78	29.51	29.84	29.27	28.59	22.97	21.07
FeO*	0.52	0.52	0.52	0.75	0.51	0.40	0.35	0.63	0.54	0.59	0.57	0.68	0.48	0.41	0.30	0.27
CaO	10.02	9.82	9.83	10.04	11.89	12.88	12.01	11.66	10.74	12.43	12.33	12.61	11.36	11.00	4.92	3.03
Na_2O	5.47	5.55	5.37	5.20	4.55	3.94	4.50	4.60	4.99	4.01	4.02	3.93	4.78	4.96	6.99	7.25
K_2O	0.39	0.40	0.40	0.25	0.28	0.24	0.20	0.21	0.30	0.23	0.19	0.19	0.21	0.22	2.68	3.71
Total	99.59	100.6	99.45	98.84	98.88	99.36	99.30	98.70	98.82	99.73	98.83	99.19	100.3	99.79	101.0	98.47
Ab, %	49.18	48.29	49.12	50.85	58.12	63.46	58.90	57.60	53.35	62.24	62.31	63.25	56.08	54.36	60.96	63.77
An	48.54	49.35	48.52	47.65	40.27	35.11	39.92	41.16	44.86	36.37	36.67	35.65	42.67	44.36	23.69	14.74
Or	2.28	2.35	2.36	1.50	1.61	1.43	1.18	1.24	1.80	1.39	1.12	1.10	1.25	1.27	15.35	21.49

Component	B15		B26	BB1	B15		B26			B15		BB1	
	Plagioclase microlites				Nepheline					Leucite			
SiO_2, wt.%	62.70	62.42	65.41	48.79	48.63	45.80	45.97	46.03	56.89	57.17	56.52	57.17	56.96	57.06
Al_2O_3	22.35	22.98	21.21	34.35	32.02	33.42	34.45	33.91	22.71	22.61	23.15	22.00	22.98	24.28
FeO*	0.43	0.48	0.37	0.78	0.75	0.72	0.74	0.81	0.32	0.33	0.31	0.40	0.50	0.44
CaO	4.32	5.19	2.32	0.76	0.73	1.81	1.01	0.90	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.
Na_2O	7.34	7.42	7.58	12.27	13.33	13.21	15.02	15.87	B.d.l.	B.d.l.	0.30	0.18	B.d.l.	0.12
K_2O	2.59	1.74	3.84	2.84	3.10	4.02	4.26	4.42	20.43	20.23	18.08	17.89	20.70	19.56
Total	99.73	100.2	100.7	99.79	98.56	98.98	101.4	101.9	100.4	100.4	98.36	97.64	101.2	101.4
Ab, %	64.20	64.88	66.56	–	–	–	–	–	–	–	–	–	–	–
An	20.88	25.11	11.23	–	–	–	–	–	–	–	–	–	–	–
Or	14.92	10.00	22.20	–	–	–	–	–	–	–	–	–	–	–

Fig. 8. Compositions of pyroxene phenocrysts and xenocrysts. *1*, phenocryst cores, *2*, phenocryst rims, *3*, xenocrysts, *4*, pyroxene xenocrysts from melanephelinitic lavas of the Khentei Ridge (Ashchepkov et al., 1996). The pressure/composition fields in the Al^{VI}–Al^{IV} diagram are given after Aoki and Shiba (1973).

Table 5. Results of representative analyses of oxide ore minerals

Component	B11/3					B15	B26	BB1	B11/3	B15	B11/3	B26
	Ilmenite								Titanomagnetite			
									Fine grains			
TiO$_2$, wt.%	52.91	52.20	50.86	52.20	52.46	42.78	50.51	50.53	51.26	9.31	8.96	10.35	9.97	6.21
Al$_2$O$_3$	0.10	0.14	0.21	B.d.l.	B.d.l.	0.48	0.33	0.11	0.28	4.82	3.73	3.26	4.12	3.18
Cr$_2$O$_3$	0.27	0.27	0.17	0.18	0.17	0.72	0.16	0.14	B.d.l.	2.08	1.78	3,93	2.26	1.88
V$_2$O$_3$	B.d.l.	B.d.l.	B.d.l.	0.25	B.d.l.	0.24	B.d.l.	0.18	B.d.l.	0.11	B.d.l.	B.d.l.	0.10	0.10
Fe$_2$O$_3$	–	–	–	–	–	–	–	–	–	50.00	52.48	50.90	49.87	54.49
FeO	40.28	40.38	39.40	37.27	39.40	45.59	38.79	39.83	39.13	29.60	29.21	28.33	30.89	31.73
MnO	0.59	0.56	0.34	0.70	0.34	0.61	0.34	0.66	0.70	0.12	0.13	0,44	0.18	0.16
NiO	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	0.25	0.26	0.22	0.24	0.28
MgO	6.56	6.15	5.24	8.00	5.24	7.30	7.65	7.55	7.10	2.80	2.78	2.93	2.61	1.84
Total	100.8	99.78	96.31	98.68	97.77	97.80	97.85	99.05	98.64	99.09	99.33	100.6	100.2	99.87

Component	B26					B11/3	BB1					B15			
	Titanomagnetite					Titanomagnetite						Cr-spinel			
	Large grains					Large grains						core		rim	
TiO$_2$, wt.%	15.75	13.45	15.82	20.29	19.65	14.00	13.74	18.41	19.23	19.06	18.70	0.10	0.09	5.98	6.28
Al$_2$O$_3$	3.50	3.63	2.35	2.00	1.96	2.45	2.52	2.03	2.43	2.35	2.40	29.48	29.08	15.88	14.87
Cr$_2$O$_3$	0.75	1.37	0.81	1.07	1.11	0.80	0.71	0.59	0.93	0.96	0.99	34.51	34.95	20.20	20.14
V$_2$O$_3$	B.d.l.	0.19	B.d.l.	B.d.l.	0.15	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.	B.d.l.
Fe$_2$O$_3$	48.83	49.17	47.85	44.64	45.05	49.43	49.30	45.25	45.41	46.09	46.23	4.85	5.02	22.38	23.11
FeO	27.18	29.45	28.91	26.91	26.06	28.50	29.58	28.14	27.39	26.66	25.72	19.24	19.31	26.79	27.32
MnO	0.60	0.22	0.55	0.60	0.56	0.51	0.53	0.58	0.57	0.61	0.60	0.57	0.55	0.58	0.58
NiO	0.25	0.20	0.22	0.19	0.17	0.18	0.19	0.16	0.15	0.15	0.16	0.14	0.11	0.12	0.17
MgO	2.50	2.36	4.29	4.87	4.93	3.32	3.41	3.95	4.41	4.59	4.70	11.02	10.98	7.96	7.75
Total	99.36	100.0	100.8	100.8	99.58	99.19	99.98	99.11	100.5	100.4	99.50	99.94	100.1	99.99	100.3
Mg#	–	–	–	–	–	–	–	–	–	–	–	0.51	0.51	0.36	0.35
Cr#	–	–	–	–	–	–	–	–	–	–	–	0.44	0.45	0.46	0.48
Fe#	–	–	–	–	–	–	–	–	–	–	–	0.55	0.57	0.31	0.33

Note. The contents of Fe$_2$O$_3$ and FeO in Cr-spinel and titanomagnetite were calculated from their stoichiometric compositions.

phenocrysts, (2) fine laths in the groundmass, and (3) micro-lites in the finely crystallized matrix. Plagioclase from the phenocrysts contains up to 49% anorthite end-member. The laths are more sodic and have a composition Ab$_{53–63}$An$_{35–45}$Or$_{1–2}$. The interstitial plagioclase microlites have a felsic composition, Ab$_{61–66}$An$_{11–25}$Or$_{10–22}$. The interstices also contain nepheline and leucite (Table 4). The composition of leucite totally corresponds to the stoichiometric formula of the mineral. Nepheline contains up to 4 wt.% K$_2$O. Such contents of K are typical of nepheline (Samsonova, 1973; Ikor-skii, 1980).

Oxide ore minerals are represented by fine grains of il-menite, titanomagnetite, and Cr-spinel. Their composition is given in Table 5.

Ilmenite is scattered in the groundmass. It is picroilmen-ite with MgO ≤ 8 wt.% and impurities of MnO (up to 0.70 wt.%), Cr$_2$O$_3$ (up to 0.70 wt.%), Al (tenths of wt.%), and V (tenths of wt.%) (Table 5). Such contents of impuri-ties are specific to ilmenites of mafic rocks.

Titanomagnetite is present both as individual grains and in intergrowth with ilmenite (Fig. 9). Most of the fine titano-magnetite grains are homogeneous, whereas the larger
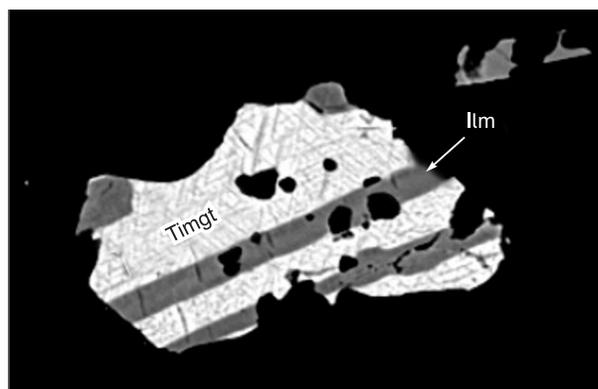
Fig. 9. BSE image of the oxide ore phase. Ilm, ilmenite, Timgt, disin-tegrated titanomagnetite, sample B11/3. Grain size is 40×50 μm.

grains have magnetite–ilmenite disintegration structures (Fig. 9). This might be due to the significantly lower content of TiO_2 in the fine grains (Table 5). The high content of TiO_2 in the large inclusions led to magnetite–ilmenite disintegration with decreasing temperature. The mineral contains many impurities (Table 5): It has high contents of Mg, Al, and Cr and medium contents of Ni and Mn relative to gabbro rocks. A similar mineral composition was observed in hawaiites and basanites of the Heven Plateau (Tsypukova et al., 2014). This leads us to the conclusion that such oxides are typical of alkali basalts.

Cr-spinel. The scarce Cr-spinel grains are zoned. Their cores are similar in composition (Table 5) to Cr-spinel from mantle xenoliths (Ashchepkov et al., 1996; Litasov and Taniguchi, 2002). These are, most likely, relics of garnet and spinel lherzolites. Similar Cr-spinel was found in basanites of the Heven Plateau (Tsypukova et al., 2014).

The present single sulfide grains are pyrrhotite with Fe = 78 wt.% and Ni ≈ 3 wt.%.

MAJOR- AND TRACE-ELEMENT COMPOSITION OF THE ROCKS

The bulk major- and trace-element compositions of the samples are listed in Table 6.

The rocks are highly magnesian, MgO = 11.81–14.77 wt.%, and contain up to 15% normative nepheline. In the $(Na_2O + K_2O)$–SiO_2 diagram, most of their composition points fall in the field of basanites (Fig. 10) (Le Bas et al., 1986). Three samples (B1, B04, and BB2) contain occasional olivine phenocrysts and thus can be considered aphyric. The aphyric and porphyritic basanites differ little in major-element composition. An increase in the number of phenocrysts in the rock is accompanied by an increase in the content of MgO and a decrease in the contents of other major components. An opposite pattern is observed in the aphyric varieties: a slight decrease in the contents of MgO (to

Fig. 11. REE patterns of volcanics from the Borozdin Bald Mountain and other volcanic areas of the South Baikal volcanic area. _1_, volcanics of the Borozdin Bald Mountain; _2_, OIB (Sun and McDonough, 1989); _3–5_, alkali basalts of the SBVA, _3_, Oka sector; _4_, Tunka sector; _5_, Hövsgöl sector (Yarmolyuk et al., 2003); _6_, basanites of the Heven Plateau (Tsypukova et al., 2014).

2 wt.%), Cr, and Ni and a decrease in the content of Al_2O_3. The rocks are rich in TiO_2 (>2 wt.%) and P_2O_5 (>0.5 wt.%). The studied volcanics belong to the Na–K series, because Na dominates over K. All samples are enriched in Cr (up to 1369 ppm), Ni (up to 686 ppm), REE, Nb, Sr, and Ba relative to the average oceanic-island melts (Naumov et al., 2016). We explain the high contents of Cr and Ni by a noticeable impurity of xenogenic minerals: high-Ni olivine and high-Cr Cr-spinel. Other trace elements are minor in the olivines, and their presence does not significantly change the geochemical features of the mineral. The REE pattern of basanites is identical to that of OIB (Fig. 11) (Sun and McDonough, 1989) and is also similar to the REE pattern of alkali basalts of the South Baikal volcanic province (Yarmolyuk et al., 2003; Tsypukova et al., 2014). One more spe-

Fig. 10. $(Na_2O + K_2O)$–SiO_2 composition diagram for volcanics of the Borozdin Bald Mountain (Le Bas et al., 1986).
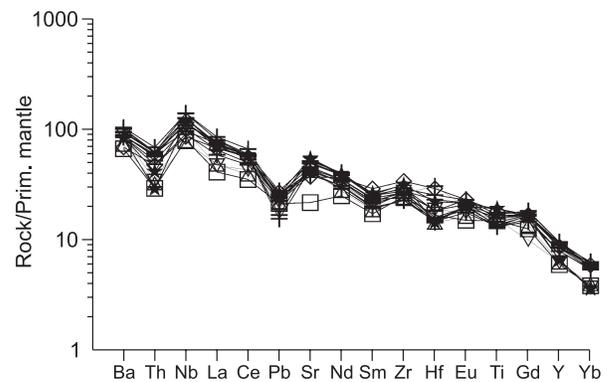
Fig. 12. Element patterns of volcanics from the Borozdin Bald Mountain and other volcanic areas of the South Baikal volcanic area. Designations follow Fig. 11.

Table 6. Chemical and trace-element (ppm) compositions of volcanics

Component	B1	B04	B06	B11/2	B11/3	B12	B15	B23	B24	BB1	BB2
	Basanite				Basalt	Basanite	Basalt	Basanite			Trachybasalt
SiO_2, wt.%	43.93	43.70	43.90	44.84	45.45	44.84	43.69	44.86	44.37	43.96	45.70
TiO_2	3.02	2.56	2.76	2.42	2.00	2.12	2.95	2.35	2.11	2.57	2.76
Al_2O_3	12.51	12.09	11.98	11.86	11.47	11.86	11.88	11.62	11.55	12.04	12.47
Fe_2O_3*	11.73	11.80	11.39	11.28	12.87	12.25	11.58	11.02	12.62	11.47	11.07
MnO	0.19	0.19	0.19	0.19	0.18	0.19	0.18	0.19	0.19	0.18	0.18
MgO	12.27	12.66	13.56	14.02	14.49	14.02	13.29	14.77	14.63	12.84	11.81
CaO	9.94	9.89	9.83	9.32	9.61	9.32	9.96	9.35	9.51	9.81	9.90
Na_2O	3.57	3.56	3.72	3.54	3.15	3.54	3.68	3.31	3.21	3.79	3.72
K_2O	1.68	1.50	1.40	1.44	1.26	1.44	1.51	1.41	1.33	1.67	1.35
P_2O_5	0.68	0.62	0.76	0.61	0.50	0.61	0.62	0.54	0.65	0.65	0.53
LOI	0.34	0.45	0.26	0.24	0.17	0.24	0.22	0.27	0.16	0.25	0.24
Total	99.86	99.02	99.75	99.76	101.15	100.43	99.56	99.69	100.33	99.23	99.73
Sc, ppm	27	25	27	26	28	26	27	28	27	25	30
V	248	218	225	218	223	218	233	210	224	228	214
Cr	832	856	958	1022	1207	1369	1010	1314	979	729	707
Co	67	39	68	67	69	67	65	65	70	64	58
Ni	444	513	538	490	666	674	555	686	615	484	406
Rb	34	32	26	32	43	30	34	34	27	32	34
Sr	753	763	828	685	740	756	685	661	806	799	674
Y	32	32	32	29	21	27	28	27	30	29	29
Zr	243	227	238	210	187	210	220	187	197	217	219
Nb	78	61	78	65	50	65	70	57	64	71	70
Ba	491	452	441	412	432	443	419	404	441	479	515
La	44	43	47	39	41	32	41	36	41	41	39
Ce	85	83	95	77	68	76	81	77	82	77	84
Pr	10.0	9.90	10.50	8.70	7.50	8.70	8.80	7.80	9.30	8.80	9.10
Nd	41	41	43	36	31	36	38	32	37	35	33
Sm	8.60	8.70	8.80	7.30	6.60	7.30	8.20	6.70	8.00	7.80	6.90
Eu	2.99	2.81	2.99	2.57	2.29	2.57	2.71	2.33	2.62	2.63	2.46
Gd	8.30	7.80	8.30	7.70	6.70	7.70	8.00	7.30	7.50	7.90	8.10
Tb	0.99	0.99	1.06	0.96	0.98	1.08	1.00	0.88	1.11	0.98	0.98
Dy	6.70	6.40	6.50	6.20	5.40	6.20	6.00	5.70	6.30	6.30	6.10
Ho	1.24	1.16	1.19	1.13	1.04	1.13	1.12	1.06	1.20	1.11	1.13
Er	3.06	2.97	3.08	2.89	2.70	1.89	2.80	2.25	3.23	2.85	2.97
Tm	0.38	0.36	0.38	0.37	0.35	0.37	0.34	0.35	0.39	0.35	0.38
Yb	2.28	2.04	2.09	2.14	2.32	1.98	2.03	2.01	2.23	2.12	2.28
Lu	0.30	0.27	0.30	0.28	0.27	0.28	0.28	0.28	0.30	0.26	0.34
Ta	4.38	3.61	4.30	3.59	2.91	3.59	4.00	3.21	3.48	3.75	4.01
Hf	4.88	6.17	6.81	3.93	3.93	3.93	4.95	4.17	3.85	4.06	4.27
Pb	3.31	1.99	2.86	2.84	2.70	2.84	1.84	2.86	3.09	2.29	2.65
Th	5.82	5.24	5.21	4.88	4.29	5.10	4.79	4.41	5.30	5.07	5.30
U	1.52	1.38	1.52	1.33	1.13	1.32	1.27	1.15	1.36	1.38	1.29

Note. Analyzed by V.I. Lozhkin, N.N. Pakhomova, and A.L. Finkel'shtein.
Fe_2O_3* is total iron.

cific geochemical feature is a high degree of REE fractionation (La/Yb = 16–20; Dy/Yb = 2.7–3.1). The indicator ratios of trace elements in the volcanics are also close to those in OIB (Ba/Nb = 6.5–9.5; Nb/Zr = 0.29–0.32). Basaltoids of the Borozdin Bald Mountain are enriched in HFSE relative to LILE and LREE and have low Rb/Sr and high U/Pb and Th/Pb ratios.

The spidergram of the studied rocks (Fig. 12) shows their identity to alkali basalts of the South Baikal volcanic province.

Table 7. Isotope characteristics of basaltoids

Sample	$^{87}Rb/^{86}Sr$	$(^{87}Sr/^{86}Sr)_{meas}$	$\pm2SE$	ε_{Sr}	$^{147}Sm/^{144}Nd$	$(^{143}Nd/^{144}Nd)_{meas}$	$\pm2SE$	ε_{Nd}	$^{206}Pb/^{204}Pb$	$\pm2SE$	$^{207}Pb/^{204}Pb$	$\pm2SE$	$^{208}Pb/^{204}Pb$	$\pm2SE$
BB1	0.1391	0.70385	3	−9.2	0.1307	0.512948	22	6.0	18.8412	18	15.5410	14	38.8903	38
B11/2	0.1670	0.70392	2	−8.2	0.1311	0.512888	8	4.9	18.8142	13	15.5430	12	38.8628	31
B15	0.1300	0.70386	1	−9.1	0.1319	0.512919	10	5.5	18.6751	12	15.5325	10	38.6828	29
B26	0.0998	0.70384	4	−9.4	0.1297	0.512925	6	5.6	18.4971	20	15.5281	17	38.4961	45

ISOTOPE CHARACTERISTICS OF BASALTOIDS

According to modern concepts, OIB might form from HIMU, EM I, and EM II reservoirs (Weaver, 1991). To elucidate the nature of the isotope sources of basaltoids of the Borozdin Bald Mountain, we studied the Sr, Nd, and Pb isotope compositions of the rocks. The results of the analyses are listed in Table 7. The isotope characteristics of Cenozoic basaltoids of the South Baikal (SBVA) and South Hangayn (SHVA) volcanic areas are reported elsewhere (Rasskazov et al., 2002; Barry et al., 2003; Yarmolyuk et al., 2003; Savatenkov et al., 2010; Tsypukova et al., 2014). The studies showed that the variations in the Sr, Nd, and Pb isotope compositions in these basaltoids are determined mostly by the interaction between depleted (DM and/or PREMA) and enriched (EM I) mantle sources. The basaltoids of the Boro-

zdin Bald Mountain differ slightly in isotope characteristics. Their Sr–Nd isotope composition (Fig. 13) indicates that the rocks formed from a moderately depleted PREMA-type source, because they are stronger depleted in Nd (ε_{Nd} = 4.9–6.1) and Sr ($^{87}Sr/^{86}Sr$ = 0.7038–0.7039) than the SBVA and SHVA basaltoids ($\varepsilon_{Nd} \leq 4$ and $^{87}Sr/^{86}Sr \geq 0.7041$). At the same time, the studied basaltoids have a more radiogenic Pb isotope composition than the SBVA and SHVA ones and the average MORB and PREMA, as seen from the $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ (Fig. 14*A*) and $^{208}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ (Fig. 14*B*) diagrams. The composition points of basaltoids from the Borozdin Bald Mountain, SBVA, and SHVA form common trends ending with the most radiogenic compositions. Additional radiogenic Pb isotopes might come from EM II and HIMU. In the ε_{Nd}–$^{206}Pb/^{204}Pb$ diagram (Fig. 15*A*), the composition points of the Borozdin basaltoids form a
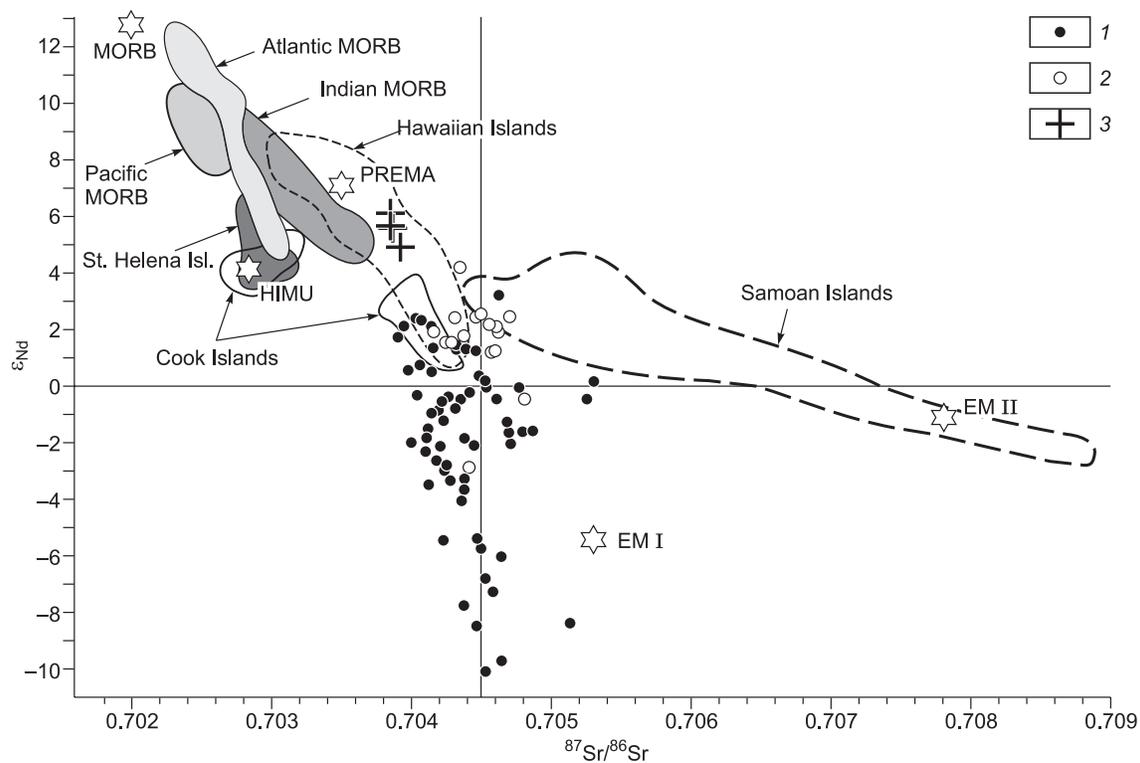
Fig. 13. ε_{Nd}–$^{87}Sr/^{86}Sr$ diagram for Neogene basaltoids: *1*, SBVA (Rasskazov et al., 2002; Yarmolyuk et al., 2003; Tsypukova et al., 2014]; *2*, SHVA (Barry et al., 2003; Savatenkov et al., 2010), *3*, Borozdin Bald Mountain. Fields mark the compositions of EMI, EMII, and HIMU (Armienti and Gasperini, 2007; Jackson and Dasgupta, 2008), PREMA (Zindler and Hart, 1986), Atlantic MORB (Paulick et al., 2010; Skolotnev, 2014), Pacific MORB (Hamilin et al., 2011), Indian MORB (Mahoney et al., 2002), Hawaiian OIB (Stracke, 2012), St. Helena OIB (Chaffey et al., 1989), Cook OIB (Palacz and Saunders, 1986), and Samoa OIB (Palacz and Saunders, 1986; Workman et al., 2004).

compact cluster separated from the points of the SBVA and SHVA basaltoids and shifted relative to PREMA toward more radiogenic $^{206}Pb/^{204}Pb$ values. As in the diagrams of Pb isotope evolution, this shift is minor; therefore, it is impossible to determine the source of radiogenic Pb reliably. To solve this problem, we can estimate varia-

tions in the isotope compositions of the Borozdin basaltoids from the $^{87}Sr/^{86}Sr–^{206}Pb/^{204}Pb$ diagram (Fig. 15B), where their composition points form a clear trend from PREMA to HIMU. This trend shows a constant depleted isotope composition of Sr and an increase in the content of radiogenic Pb isotope.
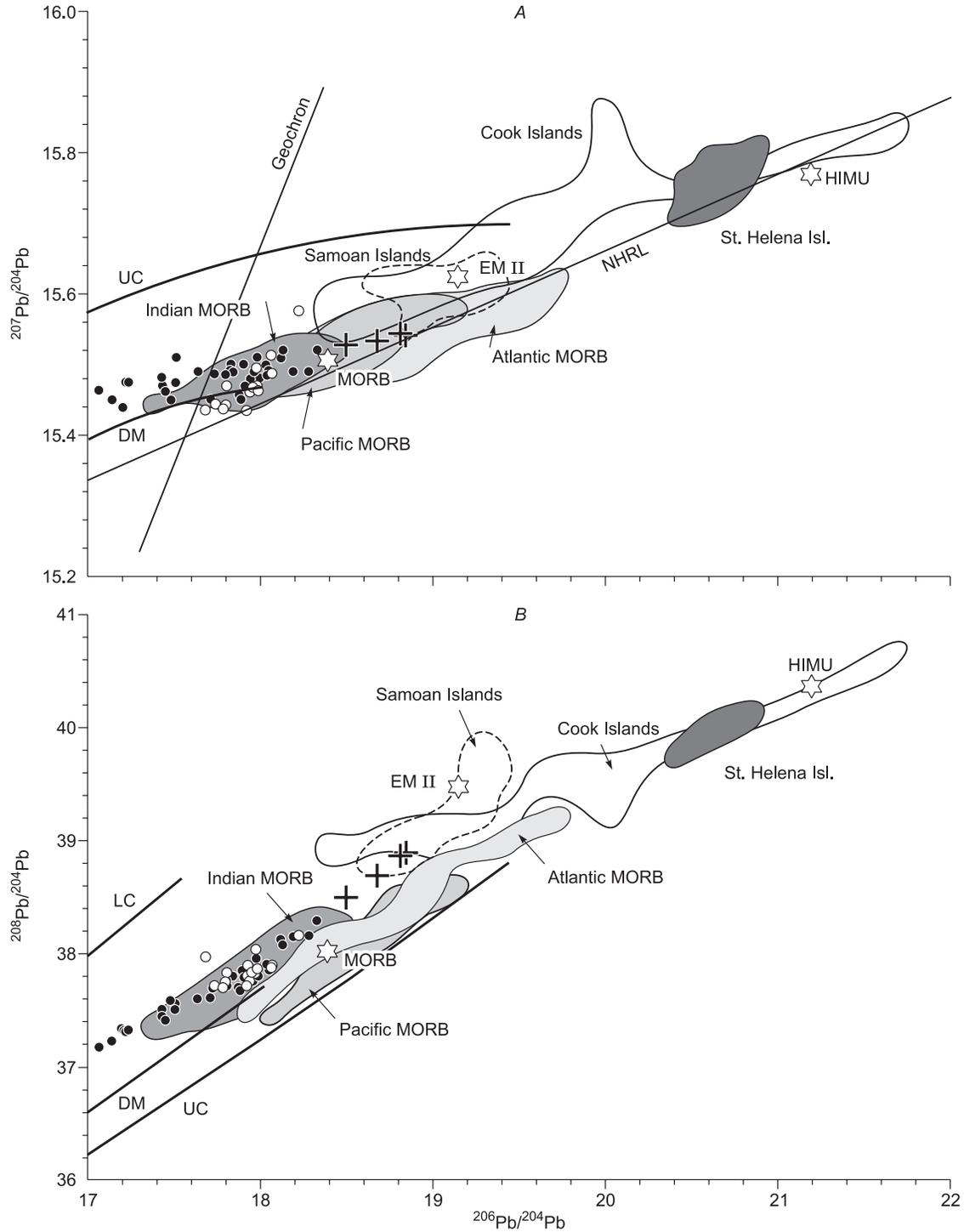
Fig. 14. $^{207}Pb/^{204}Pb–^{206}Pb/^{204}Pb$ (*A*) and $^{208}Pb/^{204}Pb–^{206}Pb/^{204}Pb$ (*B*) diagrams for Neogene basaltoids. Isotope trends of Pb in the upper crust (UC), lower crust (LC), and depleted mantle (DM) (Zartman and Doe, 1981) are shown. Other designations follow Fig. 13.

Thus, the Sr–Nd–Pb isotope data show that PREMA was the main mantle source of substance for the Borozdin basaltoids, whereas HIMU was a subordinate source, in contrast to the SBVA and SHVA basaltoids, whose isotope composition was determined by mixing of PREMA and EM I.

MANTLE SOURCES AND COMPOSITION OF THE PARENTAL MELT

Note that even the minor amount of xenogenic olivine (especially high-Mg one) and augite in the basanites distorted their major-element composition, namely, led to their
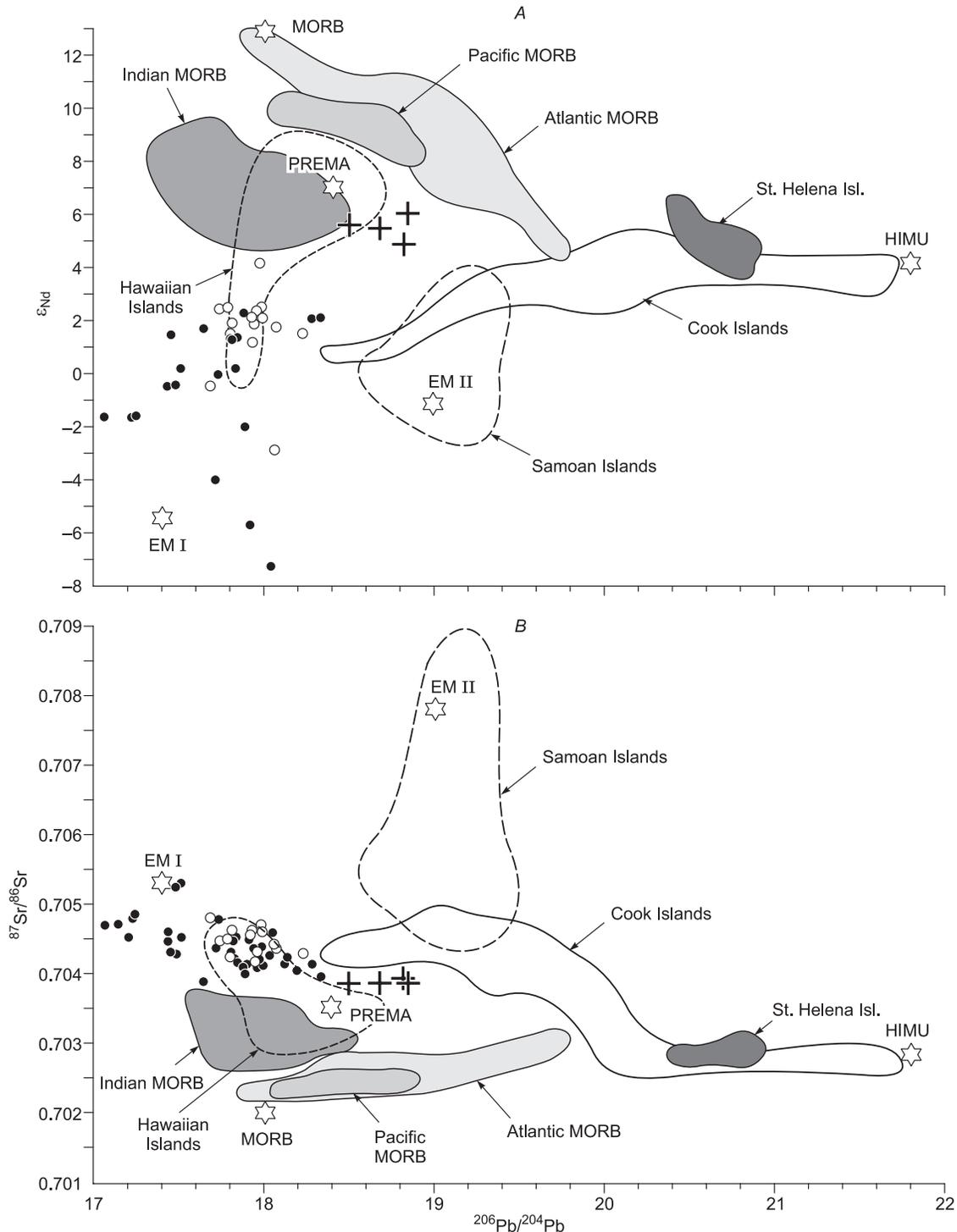
Fig. 15. ε_{Nd}–$^{206}Pb/^{204}Pb$ (*A*) and $^{87}Sr/^{86}Sr$–$^{206}Pb/^{204}Pb$ (*B*) diagrams for Neogene basaltoids. Other designations follow Fig. 13.

Table 8. Calculated composition of the parental melt, wt.%

Sample	SiO$_2$	TiO$_2$	Al$_2$O$_3$	Fe$_2$O$_3$	FeO	MnO	MgO	CaO	Na$_2$O	K$_2$O	P$_2$O$_5$	Total
B1	44.99	3.37	13.98	1.67	9.41	0.20	8.70	11.10	3.96	1.87	0.75	100
B04	45.14	3.27	12.46	1.81	10.29	0.19	9.61	10.79	3.88	1.83	0.73	100
BB1	45.38	2.95	13.77	1.67	9.13	0.20	8.69	11.22	4.32	1.86	0.72	100
B23	46.37	2.80	14.09	1.55	8.70	0.20	8.30	11.39	4.10	1.80	0.70	100
B24	45.81	2.40	13.18	1.65	10.20	0.22	9.85	10.80	3.65	1.50	0.74	100

high Mg# values. This, however, did not significantly influence the distribution of trace elements.

The composition of the parental melt (Herzberg, 2006) can be calculated from the composition of aphyric basalts. Parental melt implies the melt in equilibrium with the most magnesian olivine phenocryst found in the examined basalts. The compositions of the parental melt of the Borozdin basanites are in equilibrium with olivine with Mg# = 0.86. To obtain the parental melt, we removed olivine with Mg# = 0.91 from five rock samples step by step until equilibrium between the melt and olivine with Mg# = 0.86 was reached. The removal of olivine and estimation of the melt–olivine equilibrium were made using special appendices (Putirka, 2008; Danyushevsky and Plechov, 2011). The calculated composition of the parental melt equilibrated with olivine with Mg# = 0.86 is given in Table 8. This melt has a nepheline-normative composition and medium contents of MgO and CaO. The temperatures and pressures of the olivine–melt and diopside–melt equilibria in the cores of olivine grains from basanites were calculated following the technique described by Putirka (2008). The calculation yielded the following crystallization temperatures and pressures of the equilibria between Fo$_{0.86}$ and basanitic melt: 1379–1293 ºC and 1.15–1.06 GPa. Thus, the parental melt began to form at a temperature no lower than 1370 ºC and a pressure no lower than 1.15 GPa. Diopside was the next to crystallize. Following the same technique (Putirka, 2008), we estimated the temperature and pressure of crystallization of the cores of diopside phenocrysts (Wo = 44.96, En = 45.86, Fs = 9.18%): 1174–1122 ºC and 0.94–0.926 GPa. The difference in the P–T parameters of olivine and diopside suggests a rapid ascent of the melt. The same is evidenced by the zoned structure of both olivine and pyroxene phenocrysts. If the melt stayed in the intermediate chambers for a long time, then the crystals would be completely homogenized.

Analysis of experimental data on melting of peridotites and of results of petrological studies showed (Herzberg, 2006, 2011) that primary melts resulting from melting of peridotites have higher contents of CaO than the products of melting of a pyroxenite source. Calculation using the PRIMELT3 software (Herzberg and Asimow, 2015) yielded a low content of CaO in the parental melt of the Borozdin basaltoids, which suggests a pyroxenite source of the melt. This is consistent with the low content of CaO in the most

magnesian olivine phenocrysts in the Borozdin basanites. As shown earlier (Sobolev et al., 2005, 2007), olivine phenocrysts from basalts formed from pyroxenites have lower contents of CaO and MnO and higher contents of NiO than olivines from effusive rocks formed from a peridotite source.

Herzberg (2011) regards olivine pyroxenites (Ol + Cpx + Grt) as a mantle source for nepheline-normative OIB. They might be a cumulative part of recycled oceanic crust or the result of solid-phase reactions between eclogites and peridotites in the lower mantle (Herzberg, 2011). Partial melting of olivine pyroxenites might yield melts with characteristics demonstrated by the Borozdin basanites (low SiO$_2$ and CaO contents and a high MgO content). The presence of olivine in the restite accounts for the "peridotite" contents of NiO in the olivine phenocrysts.

The above hypothesis is confirmed by the localization of the composition points of the studied volcanics and xenogenic olivines in the same curve in the MgO–Ni diagram (Fig. 16), which was earlier assumed by Sobolev et al. (2005, 2007). We calculated the temperature and pressure of crystallization of xenogenic augite, following the technique of Putirka (2008): 1173–1181 ºC and 1.03 GPa. Thus, the basanitic melt trapped these phases, which is confirmed by the P–T conditions of formation of the parental melt (see above).

Some researchers associate the volcanism in this region with the mantle diapir activity (Ashchepkov et al., 1996; Litasov and Taniguchi, 2002). We believe that the studied basaltoid melts resulted from the ascending mantle plume, like the rock melts of the Heven Plateau (Tsypukova et al., 2014).
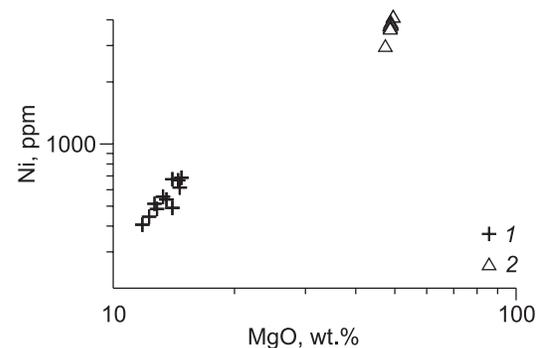
Fig. 16. Ni–MgO correlation diagram. *1*, basanites, *2*, xenogenic olivines.

CONCLUSIONS

We have estimated the *P–T* conditions of formation of basaltoid melts: $P = 1.15–1.06$ GPa and $T = 1379–1293$ ºC. Olivine pyroxenites (Ol + Cpx + Grt) are assumed to be a mantle source for nepheline-normative basanitic melts. During ascent, the melt trapped mantle xenoliths, which were disintegrated into olivine and augite xenocrysts.

The volcanic area of southern Transbaikalia (Khentei Ridge) resulted, most likely, from the mantle plume impact on the lithosphere. The age of this area is estimated at 3.51 Ma. PREMA was the main mantle source for these volcanics, and the contribution of HIMU was strongly subordinate. In geochemical features the studied volcanics correspond to mafic rocks of oceanic islands. They are similar in composition to alkali basalts of the South Baikal volcanic area.

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Editorial responsibility: V.S. Shatsky