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High-Temperature Metasomatism of the Layered Mafic–Ultramafic Massif in Kiy Island, Belomorian Mobile Belt

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Abstract—The paper presents results of a detailed petrologic study of metasomatites and their host metagabbroids in the northwestern part of Kiy Island, Onega Bay, White Sea. The first evidence is acquired that coronitization and amphibolization of the host rocks took place at the peak of Svecofennian metamorphism at $T = 700–640°C$, $P = 9–10$ kbar, and $a_{H_2O} = 0.2–0.3$. Accompanying metasomatism has formed a number of long (up to several meters long) melanocratic hornblendite and garnet–amphibole veins 0.3–2 m thick. In this area, metasomatites of another type make up single relatively thin amphibole–zoisite lenses that sometimes host ruby-like corundum. The fluid phase that induced metasomatism was poor in salts (Na,K)Cl, and hence, the rocks do not contain sodic plagioclase, and their amphibole is tschermakite but not pargasite. The compositions of the metasomatites of the two types are proved to be complementary, and this indicates that they were most likely produced by high-temperature metasomatism but not via the removal of components by fluid from migmatization zones.

Keywords: metasomatites, hornblendite, metagabbro, fluid, corundum

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INTRODUCTION

In classic interpretations of migmatization, transformations of the protolithic rocks into granitoids are explained by the influx of alkaline fluid into these rocks and the complementary leaching of Ca, Mg, and Fe from the rocks (Korzhinskii, 1952). Most of the excess (with respect to the granitic eutectic) components are removed and scattered outside migmatization regions. However, in certain instances these elements are not scattered but are redeposited around the granitic anatectites as rocks sporadically formed by Mg–Fe–Ca metasomatism, whose intensity is much lower than the complementary migmatization and debasification. These are *Grt*[1]*–Hbl, Grt–Cpx–Opx, Cpx–Grt–Pl–Mag*, and other melanocratic rims, lenses, and veins that range from 2 cm to 1.5 m in thickness and are richer in Mg, Fe, and Ca than the pristine host rocks (Belyaev and Rudnik, 1980; Utenkov, 1989; Ronenson, 1989; Sudovikov, 1964; Khodorevskaya and Korikovsky, 2007; Khodorevskaya, 2010; and others). In these situations, genetic relations between migmatization and the melanocratic rims are evident.

However, in certain instances, melanocratic metasomatites (autonomous basificate veins) are formed far away from migmatization regions. The mineral assemblages and mineral chemistries of such metasomatites and melanocratic accumulations around migmatization/charnockitization zones are similar. This provides grounds to suggest (Khodorevskaya and Korikovsky, 2007; Korikovsky and Aranovich, 2010) that such autonomous basificate veins and melanocratic rims were formed in genetic relations with migmatization processes. The only difference between the two types of basificates is the distance from the regions from which Mg, Fe, and Ca are removed: this distance is of the order of centimeters in the former instance and dozens and hundreds of meters in the latter.

At the same time, it has been proved that zonal melanocratic veins sometimes found in gabbroids are not always basificate veins but were sometimes formed by high-temperature metasomatism of the host rocks

[1] Mineral symbols: *Cpx*—clinopyroxene, *Di*—diopside, *Hd*—hedenbergite, *Opx*—orthopyroxene, *Aug*—augite, *Hbl*—amphibole, *Grt*—garnet, *Prp*—pyrope, *Grs*—grossular, *Alm*—almandine, *Jd*—jadeite, *Pl*—plagioclase, *An*—anorthite, *Ab*—albite, *Kfs*—potassic feldspar, *Zo*—zoisite, *Chl*—chlorite, *Ap*—apatite, *Ttn*—titanite, *Crn*—corundum, *Dsp*—diaspore, *Ilm*—ilmenite, *Mag*—magnetite, *Rt*—rutile, *Qtz*—quartz, *Py*—pyrite, *Act*—actinolite, *Tr*—tremolite, *Ed*—edenite, *Rit*—richterite, *Eck*—eckermannite, *Arf*—arfvedsonite, *Ktp*—cataphorite, *Gln*—glaucophane, *Prg*—pargasite, *fPrg*—ferropargasite, *Ts*—tschermakite, *Rbk*—riebeckite, *Mrg*—margarite, *Ms*—muscovite. f = FeO/(FeO + MgO) is the Fe mole fraction.

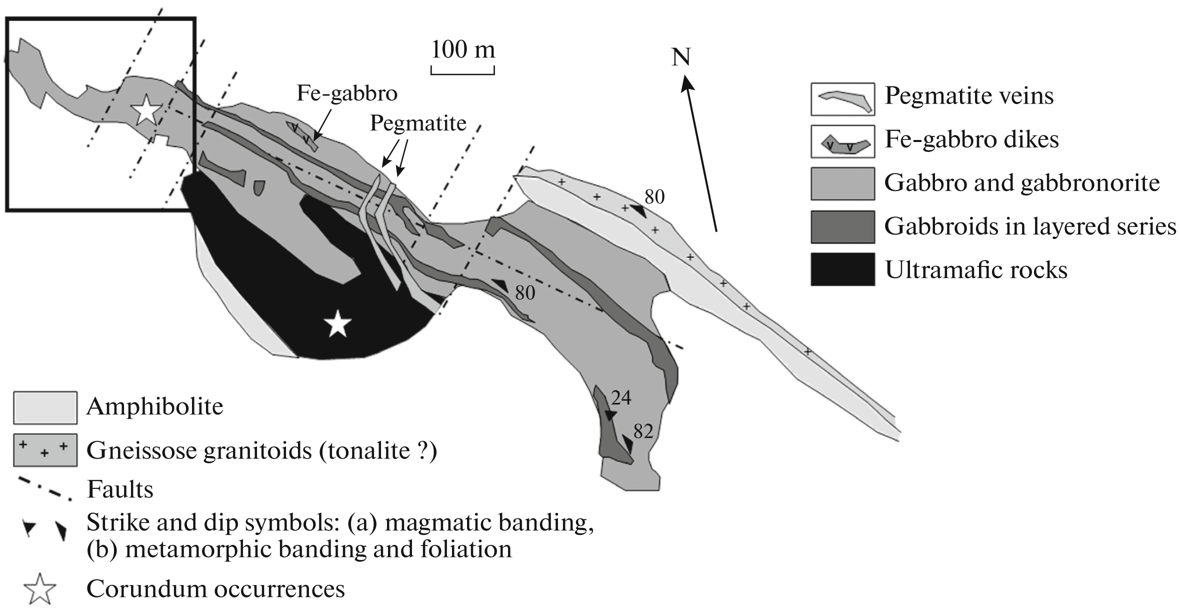


Fig. 1. Schematic geological map of Kiy Island (Kulikov and Kulikova, 1990). The rectangle shows our study area.

and were not anyhow related to the distant regions where charnockite and migmatites were generated (Khodorevskaya, 2012).

Regretfully, there are so far no clearly formulated criteria for reliable distinguishing between so-called autonomous melanocratic veins, which were formed by material removed from migmatization regions and redeposited at significant distances from these regions, and melanocratic veins, which were produced by high-temperature metasomatism of the host metabasites and were not anyhow related to migmatization. The reasons for the absence of such criteria is both the lack of data on the solubility of Ca–Fe–Mg silicates (such as pyroxenes, garnets, and amphiboles) in fluids of various composition and the paucity of data acquired by field studies of such rocks.

Our research was aimed at studying the mineral assemblages and mineral chemistries of melanocratic metasomatic veins and their host metagabbroids. These data were utilized to correlate the composition of the fluid phase and the host rocks with the types of the metasomatites to understand the principal mechanisms and relations governing the origin of melanocratic veins at high-temperature metasomatism of mafic rocks, including the complementarity of these processes and migmatization.

GEOLOGICAL OVERVIEW

The Belomorian Mobile Belt (BMB) hosts widespread Paleoproterozoic mafic–ultramafic intrusive and volcanic rocks. The southernmost body of these rocks in BMB is the layered peridotite–gabbro–anorthosite massif cropping out in Kiy Island in the southern part of the Onega Bay of the White Sea (Fig. 1). This massif is likely a fragment of the larger (40 × 20 km) intrusion that crops out on the Kem’ shore and the islands of Purluda, Shogly, Kiy, and Feresovy Ludy in the southern part of the White Sea (Kulikov and Kulikova, 1990). According to data of these authors, the rocks exposed on Kiy Island range from ultramafites to gabbroids and compose a layered suite. In natural outcrops, the rocks display magmatic layering trending to the northwest, with beds richer in mafic minerals (pyroxenes and amphiboles) alternating with leucocratic units, which contain no more than 20–30% mafic minerals. The U–Pb age of the massif is 2.437 Ga (Slabunov et al., 2006).

Granatite veins and amphibolized and zoisitized rocks found on the island are conformable with northwest-trending shear zones. The metasomatic zones host single corundum crystals (Kulikov and Kulikova, 1990, 2004, 2005). Chemical analyses of the metasomatites were presented in (Terekhov, 2003), where also various models of the origin of corundum-bearing metasomatites in Belomorie are discussed. However, the aforementioned researchers did not focus on detailed petrologic studies of metasomatites on Kiy Island. In this paper, we report our original newly obtained estimates of the genesis of these metasomatites.

Petrographic characteristics of these rocks, including their micrographs and descriptions, were obtained by studying representative samples of the pristine host rocks of the metasomatites and these metasomatites themselves. Relations between minerals in samples and their chemical composition were studied in thin sections manufactured using these samples.

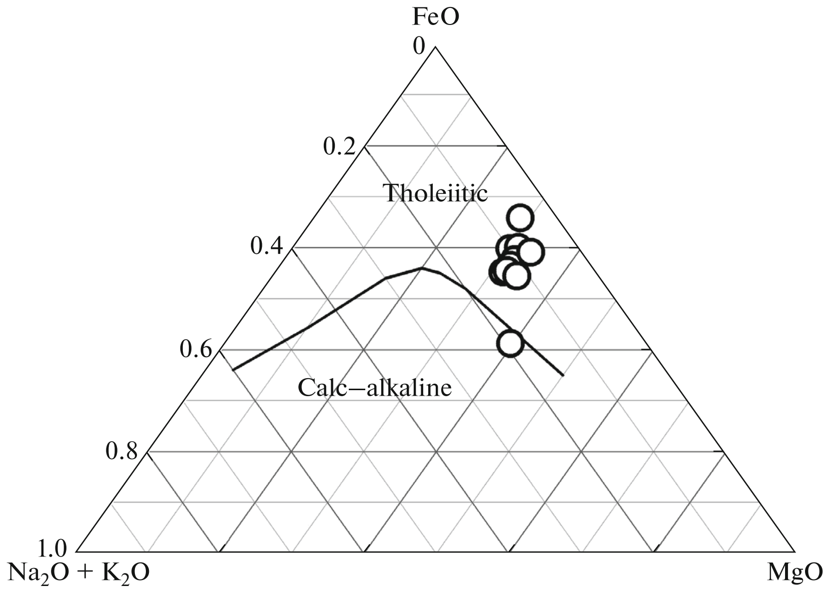


Fig. 2. AFM diagram (Irvine and Baragar, 1971) for coronitic gabbro in Kiy Island.

PRISTINE HOST ROCKS

The least modified rocks found in the northwestern part of the island are coronites (also referred to as drusites), which occur as relics. The dominant rocks are amphibolites, whose pyroxenes are completely replaced by hornblende (Fig. 1). As mentioned in (Terekhov, 2003), it is quite possible that some of the amphibolites that were thought to be ancient (Archean) rocks are, in fact, transformed coronites, as follows from close similarities between their chemical compositions.

Coronites (Drusites)

The least modified rocks found among our samples are represented by samples 9 and 10 (Table 1). These are amphibolized gabbro ($f = 0.23–0.27$) that is very locally found in the northwestern part of the island among amphibolites. The amphibolized gabbro is poorer in Fe and Ti (6–7 wt % Fe_2O_3 and 0.21 wt % TiO_2 in sample 9, Table 1) than the average unaltered rocks (13.57 wt % Fe_2O_3 and 1.42 wt % TiO_2; Levitskii, 2005). The rocks are typically relatively rich in Al_2O_3 (16–18 wt %), as compared to the average concentrations of ≈13.5 wt % (Levitskii, 2005), and usually contain such minerals as kyanite, garnet, corundum, as was mentioned in (Terekhov, 2003). In the AFM diagram (Irvine and Baragar, 1971), the composition points of the rocks plot within the field of tholeiite-series rocks (Fig. 2).

The amphibolized gabbro preserves its coronitic texture. Sample 9 (Fig. 3, Table 1) represents a typical coronite with primary orthopyroxene, clinopyroxene, and plagioclase.

The orthopyroxene occurs as large (locally larger than 1 mm) single crystals. The mineral contains no clinopyroxene lamellas but occasionally hosts amphibole and quartz inclusions. The orthopyroxene is compositionally homogeneous, and its composition is the same in grain cores and margin: $f = 0.23$, Al = 0.07 a.p.f.u, Ti = 0.2–0.3 wt %, and $Cr_2O_3 = 0.3$ wt %.

The grains of the primary plagioclase Pl_1 are as large as 1 mm. The cores of the grains are $An_{77–90}$, and

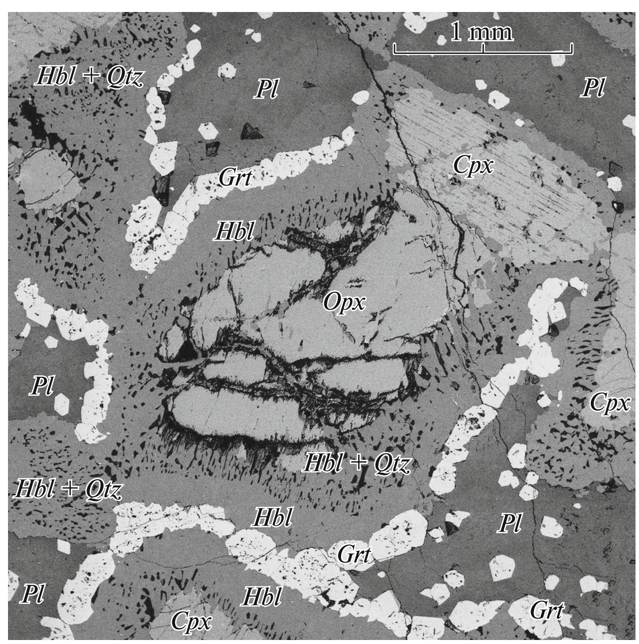


Fig. 3. Coronitic texture of the host rocks, sample 9.

Table 1. Composition of minerals in gabbroids and metasomatites from Kiy Island

Sample	Rock	Mineral composition	Accessory and secondary minerals	Cr_2O_3, wt % in mineral	Cl concentration (a.f.p.u.) in apatite
			Host rocks		
9	Metagabbro (coronite)	*Opx*, *Grt*, *Hbl*, *Pl*	*Ap*, *Qtz*	*Cpx* = 0.42%, *Hbl* = 0.41%	0.10–0.11
10	Metagabbro (coronite)	*Grt*, *Hb*, *Pl*, *Cpx*	*Ap*, *Rt*, *Qtz*	*Cpx* = 0–0.6%, *Grt* = 0–0.6%	0.16
12	Gabbro-amphibolite	*Hbl*, *Pl*, *Cpx*	*Rt*, *Ttn*	*Hbl* = 0–6%	*Ap* not found*
155	Amphibolite	*Hbl*, *Pl*	*Chl*, *Ttn*	*Hbl* = 0–6%	Same
153	Amphibolite with garnet	*Hbl*, *Grt*, *Bt*	*Ttn*, *Zo*, *Ab*, *Kfs*, *Ms*, *Chl*, *Ilm*, *Rt*	*Grt* = 0–4.6%	Same
147	Amphibolite	*Hbl*, *Pl*	*Ttn*, *Bt*, *Chl*	*Hbl* = 0–6%	Same
130	Amphibolite	*Hbl*, *Pl*	*Ap*, *Chl*, *Ttn*	*Hbl* = 0–6%	0.10
			Metasomatites		
132	Hornblendite	*Hbl*, *Pl*	*Ap*, *Ttn*, *Zo*, *Ab*, *Ms*, *Chl*, *Ilm*, *Rt*,	*Hbl* = 0.4% *Chl* = 0–0.6%	*Ap* not found*
131	Hornblendite	*Hbl*, *Grt*, *Pl*	*Ap*, *Ttn*, *Zo*, *Ab*, *Ms*, *Chl*, *Ilm*, *Rt*,	*Hbl* = 0.3% *Chl* = 0–0.65%	0.08
133	Garnetite	*Grt*, *Hbl*	*Ap*, *Ttn*, *Zo*, *Chl*, *Ilm*, *Rt*,	*Hbl* = 0–1.8%	0.05–0.06
134	Garnetite	*Grt*, *Hbl*	*Ap*, *Ttn*, *Zo*, *Chl*, *Ilm*, *Rt*,	*Hbl* = 0–0.5%	0.05–0.06
135	Hornblendite with *Pl*	*Hbl*, *Pl*,	*Ttn*, *Zo*, *Chl*, *Ilm*, *Rt*	*Hbl* = 0–3.0%	*Ap* not found*
149	*Zo–Hbl* lenses in metagabbro	*Hbl*, *Zo*, *Pl*	*Ap*, *Chl*	*Hbl* = 0–0.5%	0.07–0.08
183	Hornblendite with *Grt*	*Hbl* + *Grt–Pl*, *Bt*, *Czo*, *Sph*, *Ap*, *Ilm*, *Chl*	*Ilm*, *Chl*, *Zo*, *Kfs*	*Hbl* = 0–1.3%	0.14–0.20
170	*Crn*-bearing *Czo–Hbl* rock	*Hbl–Czo*, *Crn*	*Ttn*, *Ms*, *Chl*, *Mrg*, *Kfs*, *Py*	*Crn* = 0–0.7% *Hbl* = 0–0.6%	*Ap* not found*
174	*Crn*-bearing *Czo–Hbl* rock	*Hbl–Czo*, *Crn*	*Ttn*, *Ms*, *Mrg*, *Chl*	*Crn* = 0–0.6% *Hbl* = 0–0.4% *Chl* = 0–0.4 *Ms* = 0–0.5%	"

**Ap* not found means that no apatite was identified in the thin sections.

the composition of plagioclase in contact with garnet and amphibole coronas and with clinopyroxene is An_{60-64}. The presence of more sodic plagioclase than of original one (Pl_2) suggests that garnet and clinopyroxene were formed with the involvement of the anorthite end member of plagioclase.

The size of Cpx_1 grains significantly vary from small relics among amphibole and quartz grains to large single crystals, occasionally as large as 2–3 mm (Fig. 3). The mineral hosts Fe–Ti phases (rutile and ilmenite) and thin (3–4 μm) orthopyroxene lamellas up to 40–50 μm long. The composition of orthopyroxene in these lamellas does not differ from that of the aforementioned large grains. In cleavage planes and in fractures, the clinopyroxene is replaced by hornblende. The composition of the mineral corresponds to high-Mg augite. The composition of the mineral insignificantly varies from grain cores to margins: the Fe mole fraction decreases from 0.10 to 0.05, the *Jd* concentrations decreases from 2 to 0.5 mol % (recalculation to molar percentages according to Cawthorn and Collerson 1974), and the concentrations of Cr_2O_3 and TiO_2

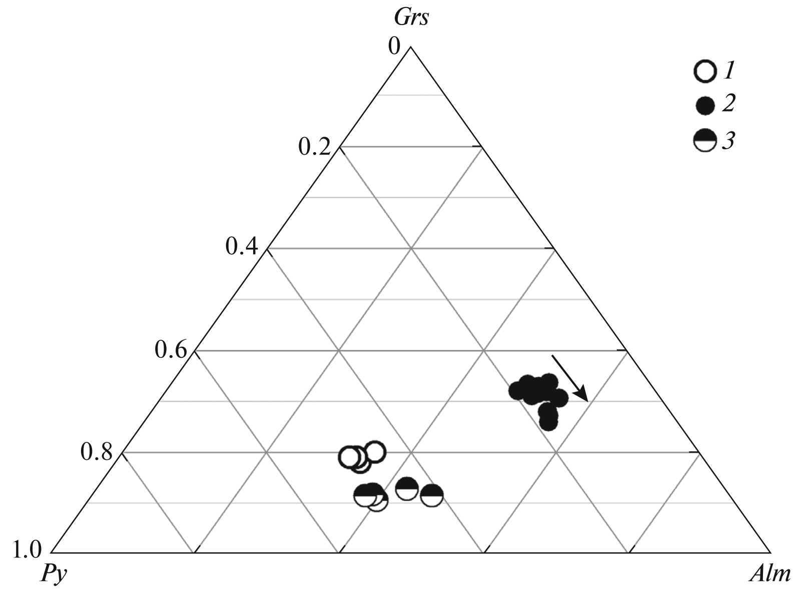


Fig. 4. Ternary *Grs–Alm–Prp* diagram for the garnet: (*1*) garnet from the host coronites (samples Ki-9 and Ki-10), (*2*) garnet from the amphibolite (sample 153); (*3*) garnet from the melanocratic veins (samples 131, 133, 134, 183). The arrow indicates changes in the *Grs–Alm–Prp* proportions from garnet cores to margins.

simultaneously increase from 0.3 to 0.6 and from 0.2 to 0.3 wt %, respectively.

In contact with plagioclase, large pyroxene grains are rimmed by coronas of later minerals (hornblende and garnet). The corona rims are zoned: *Opx* → *Hbl* → *Grt* → *Pl* and/or *Cpx* → *Hbl* → *Grt* → *Pl* (Fig. 3). Chains of garnet crystals (whose size is no larger than 50–80 μm) occur along the boundaries of plagioclase grains. The garnet is weakly zoned: its almandine concentration increases from cores to margins, and the composition changes from $Grs_{15}Alm_{30}Prp_{44}$ to $Grs_{16}Alm_{33}Prp_{39}$ (Fig. 4, *1*). The garnet hosts small (no larger than 10–20 μm) acicular inclusions of Al_2SiO_5, *Qtz*, *Hbl*, and *Chl*.

Amphibole in the amphibole coronas (Fig. 3) is Mg-*Hbl* (here and below, amphibole nomenclature is according to Leake et al., 1997), $f = 0.10–0.12$ (Table 3). The composition of the amphibole depends on which minerals occur in contact with it. For example, amphibole in contact with garnet is the poorest in Si and the richest in Al and Na; and that in contact with pyroxenes is, conversely, the richest in Si and the poorest in Al and Na (Table 3). The Cr_2O_3 concentration of the mineral varies from 0 to 0.4 wt %. Amphibole in inclusions in the garnet has the same composition as amphibole at outer contacts with the garnet (Table 3). The amphibole locally abounds in minute quartz inclusions (Fig. 3, *Hbl* + *Qtz*). The *Hbl* + *Qtz* domains are distributed not randomly but contour orthopyroxene and clinopyroxene grains. No quartz is found in amphibole near garnet. In fact, *Hbl* + *Qtz* occur in the setting of clinopyroxene (Cpx_2) coronas, as described in Belomorian coronites (Larikova, 2001). It seems to be reasonable to suggest that Cpx_2 in the clinopyroxene coronas in sample 9 was completely replaced by amphibole with quartz inclusions in the course of overprinted amphibolization. The primary clinopyroxene Cpx_1 is locally preserved in this sample.

The accessory minerals of the amphibolized gabbro are titanite, rutile, and ilmenite.

In sample 10, retrograde alterations are more intense than in sample 9 (Fig. 5): they involved partial

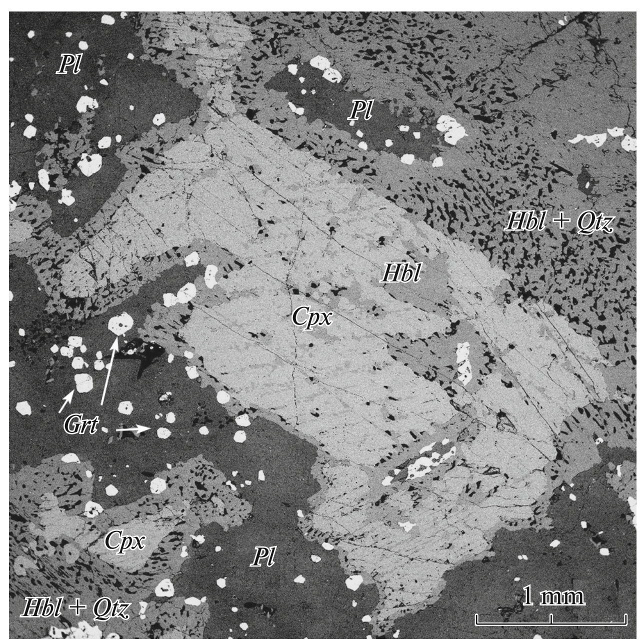


Fig. 5. Decomposition of the coronitic textures, sample 10.

Table 2. Chemical composition of metasomatites and their host rocks in Kiy island

Component	Host rocks					Metasomatites						
	9	10	12	130	147	131	132	133	153	137	170	174
SiO_2	51.28	49.62	48.1	47.91	53.76	46.76	44.06	38.1	42.25	53.09	41.41	42.98
TiO_2	0.21	0.28	0.25	0.31	0.61	0.13	0.87	3.18	2.32	0.08	0.2	0.02
Al_2O_3	16.85	17.14	18.32	17.78	14.82	19.9	14.80	18.1	15.96	25.39	31.67	35.98
FeO*	5.8	6.38	6.22	6.01	10.6	6.98	11.77	19.38	17.00	2.44	3.21	0.78
MnO	0.08	0.12	0.1	0.11	0.16	0.05	0.16	0.14	0.25	0.09	0.02	0.01
MgO	10.11	10.23	11.11	11.00	6.87	11.09	12.59	6.9	7.72	2.87	2.68	0.12
CaO	12.1	12.2	12.13	13.41	10.03	11.69	11.48	12.1	12.39	9.88	16.01	17.21
Na_2O	2.13	1.85	1.7	1.75	2.01	2.12	1.53	0.73	0.98	4.68	0.62	0.51
K_2O	0.28	0.34	0.35	0.15	0.54	0.32	0.45	0.1	0.41	0.23	0.50	0.97
P_2O_5	0	0.06	0.02	0.03	0.03	0.02	0.01	0.8	0.61	0.01	0.15	0.03
LOI	0.51	1.32	1.3	0.82	0.71	1.21	2.06	0.57	0.41	0.62	3.64	1.78
Total	99.35	99.54	99.6	99.28	100.14	100.27	99.78	100.1	100.29	99.38	100.11	100.39
Cr	425	300	320	300	35	80	111	63	42	39	99	231
V	200	131	100	180	200	52	40	375	217	1	200	111
Ni	120	53	42	41	39	150	179	117	30	27	29	51
Zn	100	62	45	42	141	60	59	86	98	14	69	17
Sr	200	250	296	130	151	210	200	22	310	680	202	251
Zr	35	29	20	30	68	22	15	260	310	25	67	78
Ba	190	161	11	19	121	111	95	30	63	100	154	109

Rock samples: 9, 10—Coronite; 12—gabbro-amphibolite; 130 and 147—amphibolite; 131, 132 hornblendite; 133—garnetite; 153—garnet- and biotite-bearing hornblendite; 137—anorthosite; 170, 174—corundum-bearing amphibole–zoisite rock.

* FeO means that all Fe was determined as FeO.

Oxides are in wt %, trace elements are in ppm.

decomposition of the garnet coronas and are evident from the absence of orthopyroxene. The primary Cpx_1 contains minute inclusions of Fe–Ti minerals and titanite; the rock is amphibolized. The clinopyroxene is in places practically completely replaced by amphibole and is found merely as relics. The chemical compositions of the *Grt*, Cpx_1, and *Hbl* are identical to those of the respective minerals in sample 9.

The rock contains primary calcic plagioclase, but it becomes progressively more sodic, up to An_{57}. The rock also contains rutile and apatite with 1.1 wt % Cl.

Amphibolite

The dominant rocks in the area are amphibolite (samples 130, 155, Tables 1, 2). These are typical amphibole–plagioclase rocks without orthopyroxene but sometimes with relics of Cpx_1. The amphibole is magnesian hornblende or tschermakite (Leake et al., 1997), i.e., is relatively poor in alkalis, with $(Na + K)_a < 0.5$, and slightly higher $f = 0.20–0.25$ (Table 3) than that of amphibole in the coronites. The TiO_2 concentration is 0.6–0.83 wt % on average. The plagioclase is $An_{60–85}$, and the rock occasionally contains quartz. The low-temperature minerals are mostly chlorite.

The amphibolites locally contain megalenses of coarse-grained amphibolite (Table 1, sample 153). In these rocks, single garnet grains are found between large tschermakite and plagioclase (An_{60}) laths. These garnet grains are larger than in the coronites (up to 2 cm). The composition of this garnet also differs from the composition of garnet in the coronites in being 5–7 mol % poorer in the grossular component (Fig. 4, *2*) and having a pyrope/almandine proportion varying from grain to grain. No compositional variations were detected within single grains (from their cores to margins). The low-temperature minerals are chlorite with rare thin potassic feldspar veinlets (this feldspar contains approximately 2 wt % BaO). An accessory mineral is apatite with 0.8 wt % Cl. The garnet hosts quartz, hornblende, chlorite, muscovite and plagioclase ($An_{33–45}$) inclusions.

Sample 147 (Table 1) represents another amphibolite variety. The rocks are found as small (<1 m) lenses in which, similar to the previous sample, the rock is recrystallized, and mineral grains (amphibole and plagioclase $An_{60–80}$) are larger. The rocks do not contain

Table 3. Microprobe analyses (wt %) of amphibole in rocks from Kiy island

Rock	Coronite			Amphibolite				Melanocratic veins	
Sample	9	9	10	130	147	153	153	131	131
	2	3*	9	1	1	24*	32	2c**	4m**
SiO_2	50.00	47.36	48.90	46.17	46.22	45.56	46.36	43.04	43.41
TiO_2	0.55	0.75	0.80	0.44	0.22	0.26	0.36	1.21	1.02
Al_2O_3	9.55	13.31	10.04	13.95	16.33	17.14	15.89	15.84	16.33
FeO	5.16	4.81	6.63	7.94	7.32	8.67	9.48	13.00	12.96
MnO	0.07	0.05	0.26	0.28	0.20	0.21	0.21	0.16	0.07
MgO	19.18	17.85	16.58	15.43	14.42	14.34	13.80	11.57	11.16
CaO	12.19	12.74	12.90	12.30	11.73	10.77	10.76	11.70	11.56
Na_2O	1.03	1.05	0.91	1.43	2.17	1.90	1.91	1.57	1.81
K_2O	0.44	1.06	0.92	0.47	0.38	0.11	0.18	0.68	0.60
Total	98.17	98.98	97.94	98.41	98.99	98.95	98.96	98.78	98.92
Cations per 23 O									
Si	6.92	6.56	6.91	6.51	6.41	6.32	6.45	6.20	6.22
Ti	0.06	0.08	0.09	0.05	0.02	0.03	0.04	0.13	0.11
AlIV	1.08	1.44	1.09	1.49	1.59	1.68	1.55	1.80	1.78
AlVI	0.48	0.73	0.58	0.82	1.08	1.12	1.05	0.89	0.98
Fe^{3+}	0.33	0.19	0.08	0.25	0.42	0.49	0.53	0.32	0.37
Fe^{2+}	0.26	0.37	0.7	0.68	0.42	0.52	0.57	1.25	1.18
Mn	0	0	0.03	0.04	0.02	0.02	0.02	0.02	0
Mg	3.96	3.68	3.49	3.24	2.98	2.96	2.86	2.48	2.38
Ca	1.81	1.89	1.95	1.86	1.74	1.6	1.6	1.80	1.77
Na	0.27	0.28	0.25	0.39	0.58	0.51	0.51	0.44	0.5
K	0.08	0.19	0.17	0.08	0.07	0.02	0.03	0.13	0.11
f	0.06	0.09	0.17	0.17	0.13	0.15	0.17	0.33	0.33
$(Na + K)_A$	0.28	0.41	0.37	0.4	0.43	0.27	0.26	0.45	0.45
Type of *Hbl*	Mg-*Hbl*	Mg-*Hbl*	Mg-*Hbl*	Mg-*Hbl*	*Tc*	*Tc*	*Tc*	*Tc*	*Tc*

Rock	Melanocratic veins								Lenses in metagabbroids	
Sample	132	133	133	133	134	134	135	183	170	170
	4	6	1*	3*	1*	17	4	1*	6	15
SiO_2	44.51	43.73	41.32	41.55	45.01	43.82	44.15	42.77	42.81	41.88
TiO_2	0.88	1.24	0.76	0.90	1.11	0.86	0.64	1.36	0.03	0.17
Al_2O_3	14.95	13.98	18.69	17.71	14.38	14.34	15.11	14.79	18.93	21.72
FeO	11.89	15.30	16.05	16.02	14.11	15.56	12.99	14.30	5.28	5.13
MnO	0.16	0.04	0.01	0.07	0.00	0.00	0.00	0.12	0.19	0.26
MgO	12.72	10.22	8.10	8.14	10.46	10.02	11.70	9.87	15.34	14.11
CaO	11.59	12.07	11.77	12.15	11.94	11.88	12.14	12.40	12.97	12.36
Na_2O	1.55	1.20	1.58	1.85	1.55	1.65	1.61	1.66	1.76	1.22
K_2O	0.46	0.48	0.71	0.60	0.45	0.63	0.62	0.49	0.38	0.45
Total	98.72	98.26	98.99	98.98	99.01	98.76	98.94	97.76	97.7	97.30
Cations per 23 O										
Si	6.35	6.40	6.02	6.09	6.47	6.39	6.35	6.31	6.10	5.89
Ti	0.09	0.14	0.08	0.10	0.12	0.09	0.07	0.15	0.00	0.02
AlIV	1.65	1.60	1.98	1.91	1.53	1.61	1.65	1.69	1.90	2.11
AlVI	0.86	0.82	1.23	1.14	0.91	0.85	0.91	0.88	1.18	1.50
Fe^{3+}	0.37	0.19	0.27	0.15	0.26	0.24	0.21	0.06	0.02	0.29
Fe^{2+}	1.04	1.68	1.69	1.8	1.43	1.65	1.34	1.7	0.67	0.31
Mn	0.02	0	0	0	0	0	0	0.02	0.02	0.01
Mg	2.7	2.23	1.76	1.78	2.24	2.18	2.51	2.17	3.13	2.96
Ca	1.77	1.89	1.84	1.91	1.84	1.86	1.87	1.96	2.01	1.86
Na	0.43	0.34	0.46	0.53	0.43	0.46	0.45	0.47	0.61	0.35
K	0.08	0.09	0.13	0.11	0.08	0.12	0.11	0.09	0.08	0.08
f	0.28	0.43	0.49	0.5	0.39	0.43	0.35	0.44	0.18	0.10
$(Na + K)_A$	0.38	0.37	0.48	0.56	0.37	0.48	0.48	0.53	0.68	0.43
Type of *Hbl*	*Tc*	*Tc*	*Tc*	*Prg*	*Tc*	*Tc*	*Tc*	*Prg*	*Prg*	*Tc*

* Inclusion in garnet.
** Core and margin of amphibole.

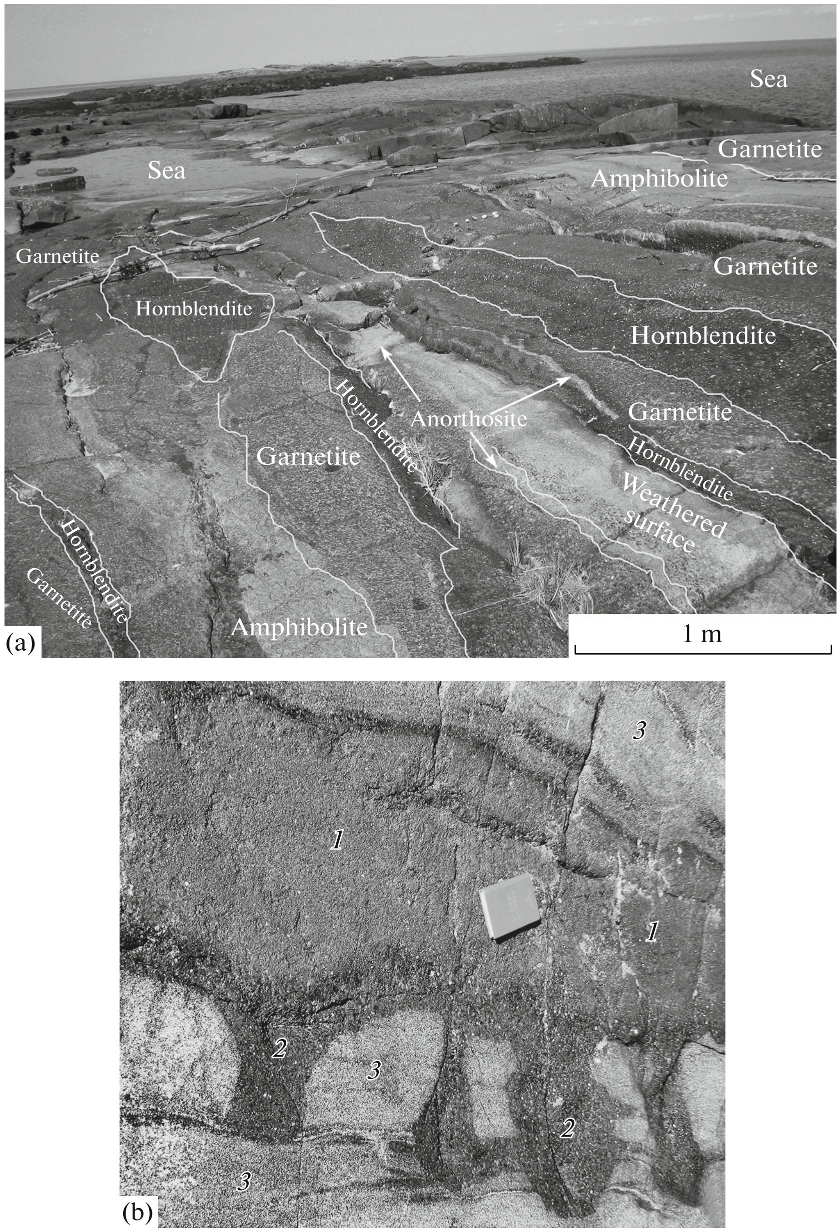


Fig. 6. Melanocratic veins in amphibolites in the northwestern part of Kiy Island: (a) natural outcrop, and (b) zonation of a metasomatic vein. (*1*) Garnetite zone; (*2*) hornblendite zone; (*3*) host rock. Thin lines approximately mark zone boundaries.

garnet but instead bear blade-shaped biotite crystals <1 mm (f = 0.2, TiO_2 = 1 wt %, Cr_2O_3 = 0.5–0.8 wt %), which cut across the hornblende and are obviously younger.

Amphibolites in these lenses (sample 153 is amphibolite with large garnet grains and sample 147 is amphibolite with biotite) were produced in zones of the pristine rocks that were more permeable to fluids.

The melanocratic veins and host amphibolite locally contain single thin (2–7 cm) leucocratic beds dominated by calcic plagioclase (An_{55-80}) (*anorthosite* in Fig. 6a). The composition of amphibole in the rocks is as in the amphibolite (tschermakite, f = 0.2), and its content in the rock does not exceed 10%. The absence of quartz and sodic plagioclase from the plagioclasite indicates that the rocks were not produced by migmatization of rocks (this process is induced by influx of alkalis and silica) but are rather magmatic leucocratic layers in the intrusion.

METASOMATITES

Rocks in the northwestern part of the island were affected by intense metasomatism, which resulted in alternating thick veins of northwestern trend in the pristine host amphibolite (Fig. 6a). The veins are reddish because of garnet and are clearly seen in the gray amphibolite (Fig. 6a, *1*). These veins are sometimes darker than the amphibolite because of black amphibole (Fig. 6a, *2*). The thickness of the veins reaches 2 m. Their contacts are sharp and are not accompanied by any alterations in the outer-contact zones.

Some of the veins are symmetrically zoned (Fig. 6b). The central parts of the veins consist of garnet with strongly subordinate amphibole amounts (garnetite, samples 133, 134, Table 1). Garnet grains in the melanocratic veins are larger than in the coronite: 2–3 mm and 0.05–0.08 mm, respectively (Fig. 7a). From the central parts of the veins toward their host rocks (sample 130 in Table 1), the content of garnet strongly diminishes, and garnetite gives way first to amphibole zones with garnet (sample 131 in Table 1 and Fig. 7b) and then to garnet-free amphibole zones (sample 132 in Table 1). The garnet zone is sometimes absent, and the veins consist of large amphibole crystals (hornblendite). All of the veins abound in titanite (Figs. 7a, 7b), which is found as individual grains and their accumulations in an amphibole matrix and as numerous inclusions in the garnet. The rocks contain rutile and ilmenite, as well as minor amounts of small albite, potassic feldspar, and muscovite grains along the contours of garnet crystals.

The garnet usually contains numerous inclusions, whose size varies from a few micrometers to >500 μm (Fig. 7a). The inclusions commonly consist of a single phase: hornblende, quartz, chlorite, titanite, and zoisite. The rare polymineralic inclusions are 200–300 μm across and consist mostly of hornblende with plagioclase grains along the margins (the right-hand large garnet grain). *P*–*T* estimates on hornblende inclusions in the garnet (see below) show that the inclusions could be formed at the metamorphic peak or during retrogression.

Amphibole in the metasomatites, particularly in the hornblendite zones (Fig. 6a), occurs as large, sometimes larger than 2 mm, grains with abundant titanite inclusions (Fig. 7b) and sometimes with accumulations of ilmenite and apatite (with 0.6–0.7 wt % Cl) grains. As seen in the right-hand part of Fig. 7b, amphibole replaces Pl_1, which is sometimes preserved only as small inclusions.

Plagioclase practically never occurs as a major mineral of the garnetite, except only plagioclase inclusions in the garnet. The hornblendite zones (samples 131, 132, Table 1) contain rare plagioclase grains, which are sometimes as large as 2 mm (Fig. 7b). The content of plagioclase in the metasomatites increases toward the host rocks, and the large size of its grains and their composition indicate that this is a preserved mineral of the host amphibolite.

Along with the melanocratic veins, we have found two lenses in the host amphibolite. The lenses are approximately 1–1.5 m thick and consist of a porphyroblastic hornblende–zoisite rock (Fig. 8a) with large (sometimes larger than 1 mm) amphibole crystals. The zoisite occurs between amphibole grains and forms the granonematoblastic groundmass. The rock contains 20–30 wt % zoisite. The small grains of the accessory minerals are barite, titanite, Fe and Cu sulfides, and rare Ba-bearing potassic feldspar ($K_{0.88}Ba_{0.12}AlSi_3O_8$), but never quartz. The low-temperature transformations of the hornblende–zoisite metasomatites involve the development of chlorite and thin rims of albite (dominant mineral), potassic feldspar, and muscovite. As seen in Fig. 8b, the rocks contain very little secondary minerals. Sometimes

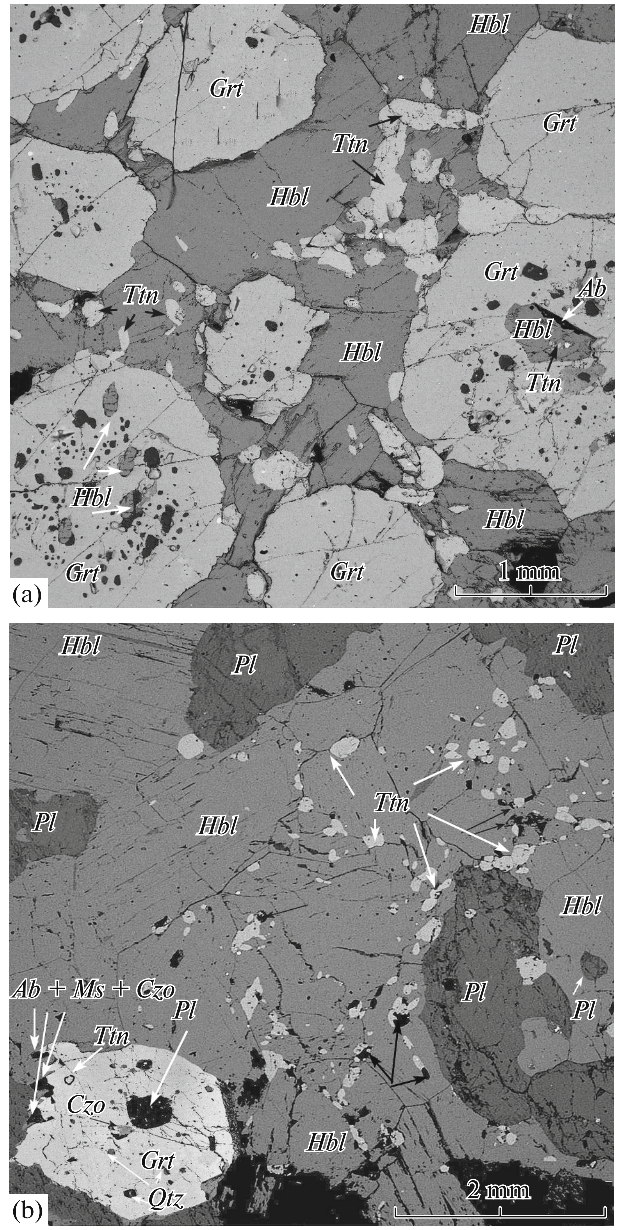


Fig. 7. Samples of veins of the melanocratic metasomatites: (a) garnet–hornblende metasomatite, sample 133; (b) hornblendite, sample 131.

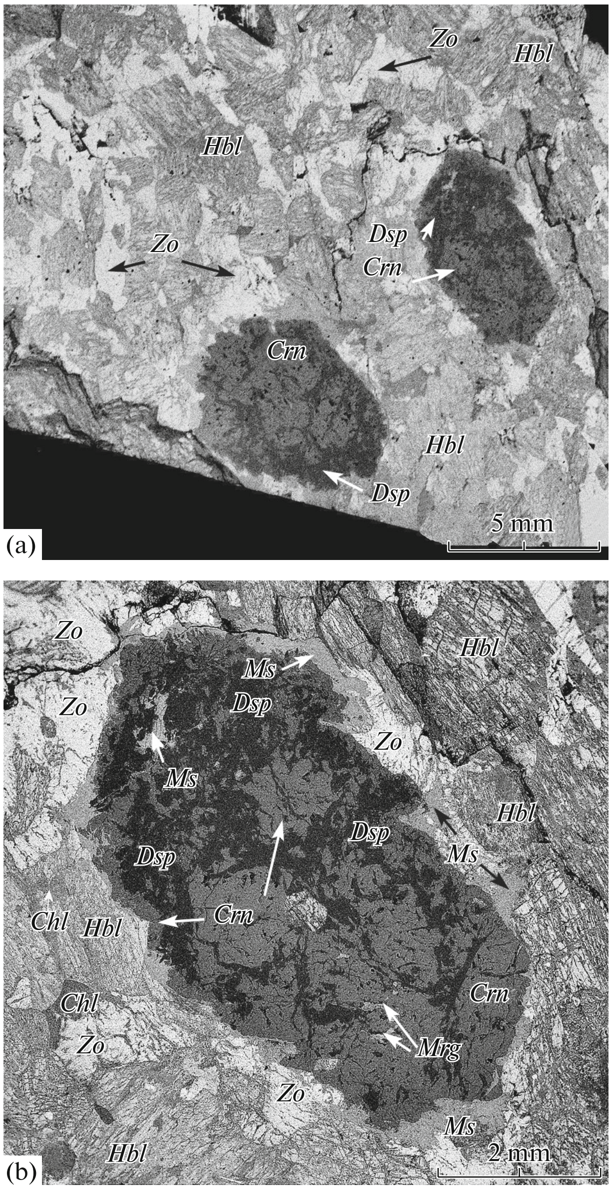


Fig. 8. Corundum crystals in zoisite–hornblende metasomatite: (a) hand-specimen of the rock; (b) relations between corundum and diaspore in a corundum crystal with muscovite and margarite inclusions.

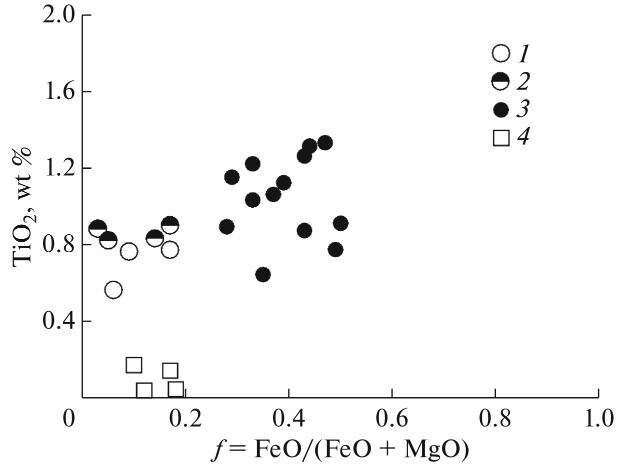


Fig. 9. f–TiO_2 relations in amphibole from various rocks: (*1*) coronite, (*2*) amphibolite, (*3*) veins, (*4*) corundum-bearing hornblende veins.

these rocks bear small titanite and apatite grains, occasionally in aggregates with ilmenite and rutile.

The hornblende–zoisite rocks occasionally host ruby-like corundum (whose crystals are sometimes of gem quality), whose size reaches 2 × 2 × 0.4 cm (Kulikov et al., 2004). The corundum is sometimes cut by diaspore veinlets and contains diaspore domains (Fig. 8b), which spoil the gem quality of the crystals. Inclusion minerals in the corundum are mostly white mica (margarite and muscovite), which also develop along the margins of the corundum crystals. No margarite was found in the groundmass of the hornblende–zoisite rocks, and we have never found the hornblende–zoisite rocks in contact with the anorthosite.

Composition of the Metasomatic Minerals

The composition of garnet in the melanocratic veins (Fig. 6b, *1*) varies from $Grs_{29}Alm_{48}Prp_{14}$ in grain cores to $Grs_{26}Alm_{51}Prp_{16}$ in their margins (Fig. 4, *3*), i.e., the garnet depletes in Ca. This is explained by the origin of numerous small titanite grains and their chains, which are found in practically all metasomatites (Figs. 7a, 7b). As seen in Fig. 4, garnet in the metasomatites is enriched in the grossular and almandine components compared to this mineral in the pristine host rocks.

Amphibole in the metasomatic veins (hornblendite, Fig. 6b, *2*) is tschermakite, $(Na + K)_A \approx 0.4$–0.5. Amphibole in the garnet–hornblende and hornblende veins is richer in Fe (f = 0.3–0.5) and TiO_2 than this mineral in the host metabasites (Table 3, Fig. 9). Amphibole in hornblende + albite inclusions in the garnet (Fig. 7a) has a composition close to that of the matrix amphibole but sometimes contains more Na, $(Na + K)_A \approx 0.5$–0.7. Amphibole in the inclusions is pargasite or tschermakite (Table 3, analyses 3 and 31). Iron-poorest amphibole, which contains practically no titanium, is found in the corundum-bearing rocks (Table 3, analyses 170, 174; Fig. 9).

It is seen in Fig. 10 that the transition from the weakly modified metagabbro (coronite) to amphibolite and metasomatic veins (Fig. 10, *1*–*4*) is associated with an increase in the Al/(Al + Si + Fe + Mg + Mn + Ti) ratio of the amphibole and a slight decrease in its Ca/(Ca + Na + K) ratio. The magnesian hornblende gives way to tschermakite, with pargasite found much more rarely (Table 3). Amphibole with low aluminum and alkali concentrations (Table 3) was found in the

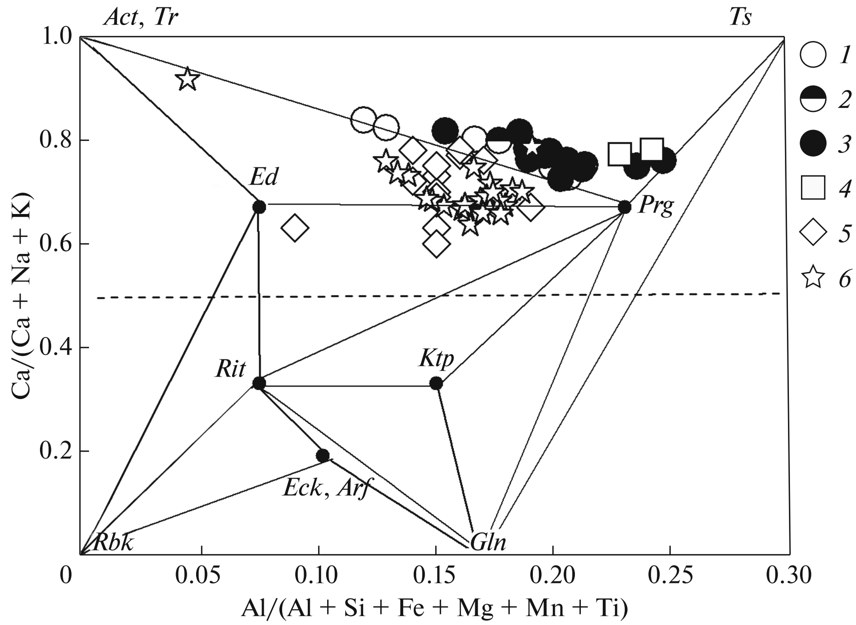


Fig. 10. Al/(Al + Si + Fe + Mg + Mn + Ti)–Ca/(Ca + Na + K) relations in amphibole from various rocks: (*1*) coronite, (*2*) amphibolite, (*3*) veins, (*4*) corundum-bearing hornblende veins, (*5*) data on field amphibole (Khodorevskaya, 2010, 2012; Korikovsky and Aranovich, 2010, 2015), (*6*) experimental data (Safonov et al., 2014). Solid lines contour miscibility regions of three amphibole end members.

corundum-bearing zoisite–hornblende rocks and nearby garnet-bearing metasomatic veins (Table 1, samples 170, 174, and 183).

In Fig. 10, rhombi show the Al/(Al + Si + Fe + Mg + Mn + Ti)–Ca/(Ca + Na + K) parameters of natural amphiboles in basificate veins (Khodorevskaya, 2010, 2012; Korikovsky and Aranovich, 2010, 2015) found elsewhere in the Belomorian region in relation to migmatization regions. These amphiboles typically possess lower Ca/(Ca + Na + K), i.e., are richer in alkalis and are pargasite. As seen in Fig. 10, amphibole from Kiy Island generally has elevated Ca/(Ca + Na + K) ratios, and considered together with the fact that the rocks contain practically no sodic plagioclase, this indicates that the metasomatizing fluid was poor in alkalis.

Plagioclase in the margins of the melanocratic metasomatites is An_{70-80} (Fig. 6b, *2*) and is more sodic when occurring as inclusions in the garnet: from pure albite to An_{40-50}.

The composition of zoisite found as inclusions in garnet in the garnet–hornblende rocks corresponds to $Czo_{0.80-0.74}Ep_{0.20-0.26}$. The composition of this mineral in the hornblende–zoisite rocks is $Czo_{0.96}Ep_{0.04}$, i.e., the concentration of the epidote end member (defined by the Fe^{3+} concentration) in zoisite included in the garnet is higher than in zoisite in the hornblende–zoisite rocks.

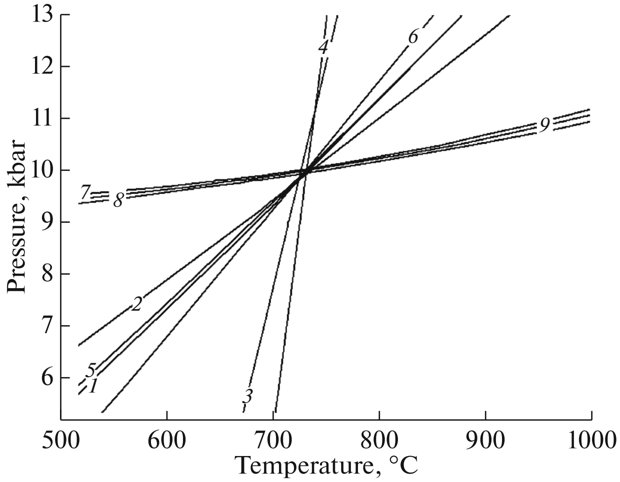
The corundum contains up to 0.7 wt % Cr_2O_3 and 0.16–0.25 wt % Fe_2O_3, and this causes the reddish purple color of the mineral. The diaspore contains practically no Cr.

The composition of muscovite found at the margins of the corundum crystals (Fig. 8b) slightly varies: where occurring in inclusions in the corundum and in contact with this mineral, the muscovite contains up to 0.7 wt % Cr_2O_3 and 1.22 wt % FeO. The father away from contact with corundum, the lower Cr_2O_3 and FeO concentrations in the muscovite, and these parameters decrease to 0.4 and 0.5 wt %, respectively.

The apatite, which was found in most of rocks samples from this area, contains Cl, but its concentrations are relatively low, <0.09 a.p.f.u.

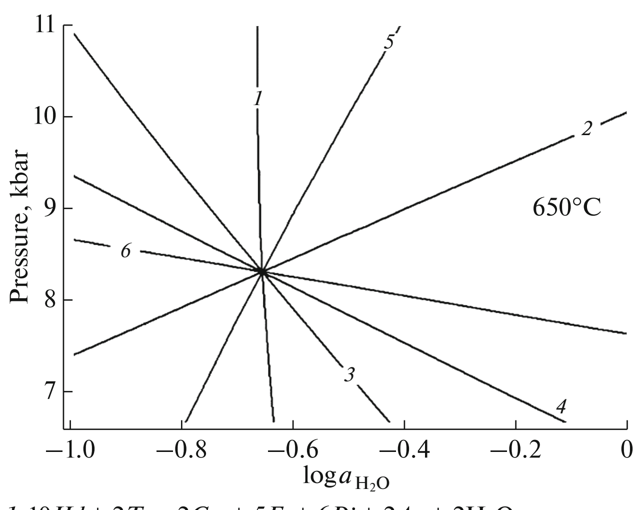
P–T–a_{H_2O} *Parameters*

The *P–T* parameters of petrologic processes were estimated using the TWQ program package and the proprietary bases of consistent thermodynamic data (Aranovich and Berman, 1996; Berman and Aranovich, 1996). In the calculations, we made use of the following end members of mineral solid solutions: almandine, grossular, and pyrope for garnet; ferrosilite and enstatite for orthopyroxene; diopside and hedenbergite for clinopyroxene; anorthite and albite for plagioclase; pargasite, tschermakite, tremolite, and their Fe-bearing analogues for amphibole; as well as minerals of constant composition. The latter were quartz (only for the coronites, because the metasomatites contain practically no quartz), corundum, titanite, rutile, ilmenite, and sillimanite/kyanite (only



1. *Qtz* + *Di* + *Alm* = *An* + *En* + 3*Fs*
2. 2*Alm* + *Grs* + 3*Qtz* = 6*Fs* + 3*An*
3. 3*Di* + *Alm* = 3*En* + 3*Fs* + *Grs*
4. *Alm* + 3*En* = *Prp* + 3*Fs*
5. *Qtz* + *Di* + *Alm* = *An* + *En* + 3*Fs*
6. 3*Qtz* + 3*Di* + 4*Alm* = 3*An* + 12*Fs* + *Prp*
7. *Qtz* + *Prp* + *Di* = *An* + 4*En*
8. 3*Qtz* + 2*Prp* + 2*Grs* = 3*An* + 3*Di*
9. 3*Qtz* + 2*Prp* + *Grs* = 3*An* + 6*En*

Fig. 11. *P–T* parameters of initial coronitization of gabbroids in Kiy Island. Numbers on lines in Figs. 11–13 correspond to the numbers of mineral equilibria listed beneath the figures and calculated with the TWQ program package (see text).



1. 10*Hd* + 2*Ts* = 2*Grs* + 5*Fs* + 6*Di* + 2*An* + $2H_2O$
2. 8*Grs* + 11*Fs* + 2*Ab* + $2H_2O$ = 6*An* + 14*Hd* + 2*fPrg*
3. 8*Ts* + 26*Hd* + 2*Ab* = 14*An* + 24*Di* + 9*Fs* + 2*fPrg* + $6H_2O$
4. 5*Ab* + 9*Grs* + 20*Hd* + 11*Ts* = 5*fPrg* + 33*Di* + 26*An* + $6H_2O$
5. *Ab* + 9*Di* + 13*Fs* + 7*Grs* + $4H_2O$ = *fPrg* + 3*Ts* + 22*Hd*
6. 5*Ab* + 10*Fs* + 13*Grs* + 7*Ts* = 5*fPrg* + 21*Di* + 22*An* + $2H_2O$

Fig. 12. P–a_{H_2O} parameters (at 650°C) of final coronitization.

if these minerals were identified in the rocks). No chlorite was introduced in the calculations. One of the dominant rock-forming minerals in rocks from the island is amphibole, which makes it possible to estimate a_{H_2O} during its crystallization. In view of this, we have not rejected this mineral from the calculations, in spite of that not all thermodynamic data on it are reasonable reliable.

Along with using the TWQ software, we have independently estimated the *P–T* parameters using the TPF program package, which involves a number of garnet–orthopyroxene–plagioclase–quartz thermobarometers: *Grt–Cpx–Pl–Qtz*; *Grt–Hbl–Pl–Qtz; Pl–Hbl–Qtz* (Fonarev et al., 1991, 1994; Powell, 1985; Graham and Powell, 1984; Perchuk and Lavrent'eva, 1989; Holland and Blundy, 1994).

The differences between the TPF and TWQ temperature estimates never did exceed 50°C.

The *P–T* parameters of early caronitization were estimated from the compositions of orthopyroxene, plagioclase, clinopyroxene, and garnet (Table 1, sample 9) in the cores of their grains at 700°C and approximately 9–10 kbar (Fig. 11).

The *P–T* parameters of final coronitization were estimated from the compositions of the grain margins. Because the margins of garnet, orthopyroxene, and clinopyroxene grains are in reaction relations with the amphibole (Fig. 3), equilibrium in this system should be described by three parameters: P–T–a_{H_2O}. The TWQ program package was employed to calculate *T–P*, *a–P*, and *a–T* sections at a constant third parameter. In these situations, the temperature was independently estimated using the margins of hornblende and garnet grains in contact with one another by the *Grt–Hbl* ± *Pl* ± *Qtz* and *Pl* + *Hbl* + *Qtz* thermobarometers. The calculated temperature values were then used to calculate the P–a_{H_2O} relations. Our estimates indicate coronite textures in the gabbroids of Kiy Island ended to develop at $T \approx$ 600–650°C, $P \approx$ 8–8.5 kbar, and $a_{H_2O} \approx 0.2$ (Fig. 12).

Likewise, the maximum temperatures ($T \approx$ 700°C) for the melanocratic metasomatic veins (Table 1, samples 133, 134, 183) were calculated by the TPF program package using the compositions of garnet and hornblende inclusions in this garnet. These temperature values were subsequently utilized to calculate the P–a_{H_2O} relations by the TWQ program package. Our results indicate that the melanocratic veins were produced at $T \approx$ 700°C, $P \approx$ 9–10 kbar, and $a_{H_2O} \approx$ 0.4–0.5 (Fig. 13). The compositions of the margins of the matrix amphibole grains in contact with garnet indicate that the melanocratic metasomatites generally ended to develop at T = 630–660°C, P = 8–10 kbar, and $a_{H_2O} \approx$ 0.3–0.4.

The garnet coronas in metabasites and the melanocratic veins in Kiy Island were thus produced at the metamorphic peak and during early retrogression within a fairly narrow *P–T* range of 700–630°C and 10–8 kbar.

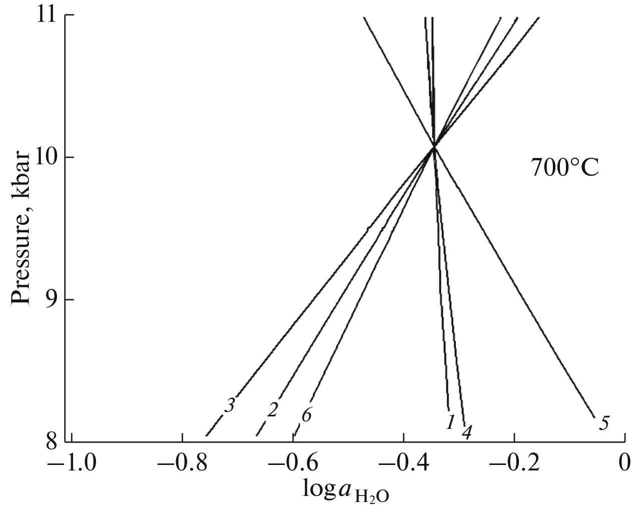
DISCUSSION

As was mentioned above, the mafic–ultramafic massif in Kiy Island was dated at 2.437 Ga. In reviews of all isotopic dates of Belomorian rocks (Bibikova et al., 2001, 2004), it was shown that the emplacement of gabbroids in the metamorphic rocks and the Svecofennian cycle (1.85–1.9 Ga) were separated by a long-lasting interlude of endogenic events. This interlude lasted for approximately 500 Ma, during which neither any significant metamorphic nor granite-forming events occurred. The interlude ended when the Svecofennian tectono-metamorphic cycle began at ~1.9 Ga, when granitoids and pegmatites were emplaced, the Archean rocks were intensely recycled, coronite textures were formed in the gabbroids (Alekseev et al., 1999), synmetamorphic (with granulite-facies metamorphism) charnockitization and enderbitization took place (Bogdanova et al., 1996; and others), and melanocratic metasomatites were produced (Sal'nikova et al., 2009).

The foregoing facts and led us to suggest that coronites were formed in the metabasites and melanocratic veins in Kiy Island at during Svecofennian time. This is consistent with data in (Slabunov et al., 2006) that rocks of the massif in Kiy Island contain zircons yielding strongly discordant ages, reportedly due to the effects of Svecofennian processes.

At the peak of Svecofennian metamorphism at 700°C and 9–10 kbar, coronite textures were formed in the metagabbroids of the Kiy Archipelago and the metagabbroids were amphibolized because the fluid contained H_2O. Concentrations of K and Na chlorides in the fluid were relatively low, as follows from the absence of sodic plagioclase from the rocks. The amphibole, which is mostly tschermakite, also indicates that the fluid was poor in (Na,K)Cl. This conclusion is based on experimental data (Safonov et al., 2014) that (Na + K) concentration in amphibole increases with increasing $X_{(Na,K)Cl}$ in the equilibrium fluid, with even low salt concentrations in the fluid ($X_{(Na,K)Cl} \approx 0.01–0.02$) inducing an increase in concentrations of alkalis in the amphibole to $(Na + K)_A \approx$ 0.6–1.2, and with the amphibole composition changing from edenite–pargasite to barroisite–winchite (Fig. 10, *6*). With regard for experimental data in (Safonov et al., 2014) and considering that tschermakite with $(Na + K)_A < 0.5$ was formed in the Kiy amphibolites, it is reasonable to conclude that salt concentration in the fluid that induced amphibolization of the metagabbroids was insignificant.

The metasomatic transformations of the metagabbroids, which also occurred at the metamorphic peak within a narrow range of T = 700–640°C, $P \approx$ 8–10 kbar, and a_{H_2O} = 0.45–0.35, produced, first and foremost, the thick melanocratic hornblendite bodies and garnet–hornblende veins (Fig. 6a). These veins are typically zoned, and their garnet and hornblende are richer in Fe than these minerals in the host rocks.



1. 3*Ttn* + 15*Sil* + 11*Grs* + *Alm* + 9H_2O = 3*Ilm* + 18*Zo*
2. 3*Ilm* + 5*fTs* = 2*Zo* + 3*Ttn* + *Grs* + 6*Alm* + 4H_2O
3. 6*Ilm* + 3*Sil* + 11*fTs* = 8*Zo* + 6*Ttm* + 13*Alm* + 7H_2O
4. *fTs* + 3*Ttn* + 18*Sil* + 13*Grs* + 10H_2O = 3*Ilm* + 22*Zo*
5. 2*Grs* + 3*Sil* + *fTs* + H_2O = 4*Zo* + *Alm*
6. 6*Ttn* + 3*Sil* + 4*Grs* + 11*Alm* + 9H_2O = 6*Ilm* + 9*fTs*

Fig. 13. $P–a_{H_2O}$ parameters (at 700°C) of the origin of the melanocratic veins.

Another types of metasomatites in the area is rare single lenses of hornblende–zoisite rocks. The smaller thicknesses of these rocks than those of the garnet–hornblende veins are likely explained by that the former were formed at $T \leq 640$°C, with the involvement of residual fluid, from which certain elements, including Fe and Ti, were removed when the garnet–hornblende veins were formed. Because of this, amphibole in the hornblende–zoisite rocks is the poorest in Fe and contains no Ti (Fig. 9), and the zoisite contains practically no Fe^{3+}.

During retrogression (at approximately T = 450–480°C, see below), minor amounts of muscovite and potassic feldspar (sometimes containing BaO) were formed. These minerals were produced near corundum grains (Fig. 8b) and in the intergranular space of the zoisite and hornblende. The muscovite and potassic feldspar crystallized in the presence of final small portions of the residual fluid, which was rich in alkalis, including K_2O. The increase in the mole fractions of alkalis relative to H_2O in the fluid can, perhaps, be explained by the early crystallization of zoisite in these metasomatites.

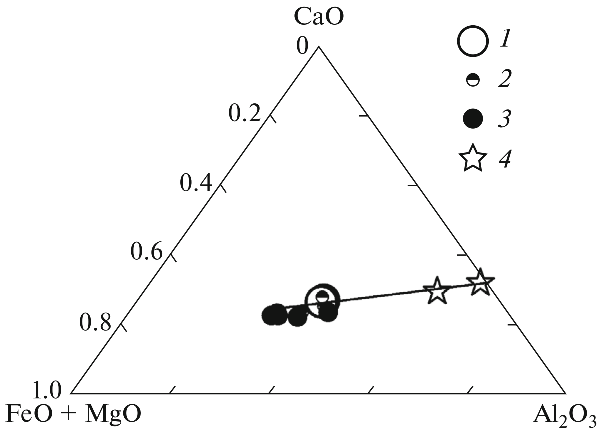


Fig. 14. CaO–Al_2O_3–(FeO + MgO) diagram for rocks in Kiy Island: (*1*) coronite, (*2*) amphibolite, (*3*) melanocratic veins, (*4*) corundum-bearing hornblende–zoisite rocks.

The CaO–Al_2O_3–(FeO + MgO) diagram in Fig. 14 shows the compositions of the host rocks (coronites and amphibolites), the melanocratic veins, and the hornblende–zoisite rocks in which corundum was found (Tables 1, 2). As seen in this figure, the compositions of the melanocratic veins (garnetite, *Grt–Hbl,* or hornblendite) are complementary to those of the aluminous hornblende–zoisite rocks. Thus, the ruby-bearing hornblende–zoisite rocks and aluminous garnet, garnet–hornblende, and hornblende melanocratic veins were produced by high-temperature metasomatism of the host coronite and amphibolite. The presence of Cr in minerals of the garnet–hornblende and hornblende–zoisite rocks also confirms genetic links of the metasomatites with the host rocks, which contain elevated Cr_2O_3 concentrations.

The metasomatic veins were formed at the removal of elements, including Mg, Fe, and Ca, from the host rocks. The elements were redeposited in more permeable regions, for example, along zones of magmatic layering in the intrusion. This follows from the northwestern trends of the metasomatic veins, which coincide with the trend of the magmatic layering of the intrusion.

As was mentioned above, the fluid at the metamorphic peak in rocks in the Kiy Island area had $a_{H_2O} \approx$ 0.2–0.5, which roughly corresponds to X_{H_2O} = 0.45–0.7 (Aranovich and Newton, 1996). The fluid at the metamorphic peak contained a low salt concentration and had X_{H_2O} = 0.45–0.7. Another major component of the fluid was likely CO_2, but it is not possible to estimate X_{CO_2} of the fluid because of the absence of carbonate-bearing mineral associations. The fluid removed elements from the host rocks and produced the metasomatites. Among Ca, Mg, and Fe of the pristine host rocks, Ca was most significantly, and Mg least significantly, extracted to solution (i.e., these elements were the most and the least mobile, respectively). The high Ca solubility maintained merely insignificant variations in its concentrations in both the pristine rocks and the metasomatites, while Mg concentrations in the metasomatites are obviously lower than in the host rocks (Table 2). Our conclusion is confirmed by experimental data (Budanov and Shmulovich, 2000), which indicate that Ca concentration in solution at incongruent dissolution of diopside in H_2O (at 650°C, 2–7.5 kbar) is 200 times higher than the Mg concentration. No experimental data on the solubility of Fe-bearing silicates are available so far, but data in (Khodorevskaya and Aranovich, 2016) and field observations, including those on garnet, hornblende, and clinopyroxene Fe-rich metasomatic rocks (Khodorevskaya and Korikovsky, 2007; Korikovsky and Aranovich, 2010; and others) suggest that Fe^{2+} is more actively than Mg mobilized from host rocks. Thus, Ca, Mg, and Fe can be arranged according to decreasing their mobility (Korzhinskii, 1955) as Ca > Mg > Fe.

Some features of melanocratic veins in Kiy Island (their darker color than that of the host rocks, zoning, and higher Fe mole fractions of the mafic minerals) resemble those of so-called autonomous basificate veins that are formed at distances of dozens to hundreds of meters from migmatization and charnockitization regions (Khodorevskaya and Korikovsky, 2007; Korikovsky and Aranovich, 2010). The volumes of the rocks produced by Ca–Mg–Fe metasomatism, and particularly, the volumes of the distant basificate veins, are commonly incomparably smaller than the complementary granitoid volumes in the migmatization regions (Sudovikov, 1964; Belyaev and Rudnik, 1980; and others). However, no migmatites whose volume is sufficient to produce such large volumes of the melanocratic metasomatic rocks found in the island (Fig. 6a) were mapped anywhere in the Kiy Archipelago, whose total area is 40 × 20 km (Kulikov and Kulikova, 1990). With regard for this and for the aforementioned genetic relations between the metasomatites and their host rocks, we believe that metasomatites in Kiy Island were formed by high-temperature metasomatism of the host rocks but not as a result of removal of components from extensive migmatization regions (which are not found anywhere in the island) by fluid.

In this context, an issue of particular importance is the mobility or inertness of Al and the genesis of the corundum. High mobility of Al was proved for certain metasomatic veins in areas of high-grade metamorphism (Foster, 1977; Kerrick, 1988; McLelland et al., 2002). The high mobility of Al was explained by the composition of the fluids: for example, the presence of NaCl, as well as Si and Ca, in solutions result in Al complexes with Si and Na (Newton and Manning, 2006, 2008), and this, in turn, leads to a tenfold (and more) increase in the solubility of Al-bearing minerals and, hence, Al transfer and redeposition (Newton and Manning, 2006, 2008). However, fluid in the rocks of Kiy Island was poor in NaCl. The corundum solubility in fluids with low salt concentrations at $T = 800°C$ and $P = 10$ kbar is as lows as 0.0013 mol/kg of H_2O (Tropper and Manning, 2007). Hence, Al could hardly be transferred from distant rocks by the filtering fluid. The fact that the compositions of the metasomatites are complementary led us to suggest that intense Ca, Mg, and Fe removal from the host rocks of the Kiy Island metagabbroids and the redeposition of these elements in the form of thick melanocratic veins resulted in Al enrichment in the residual rocks, particularly, in more permeable regions (Table 2). This was favorable for the growth of corundum and the origin of zoisite–hornblende lenses in the typical amphibolite. At the same time, according to (Terekhov, 2003), Al required for corundum crystallization could be mobilized from leucocratic metabasite layers, which consist mostly of calcic plagioclase. Regardless of the mechanism that produced the corundum, the only Al source was the host rocks with elevated (relative to the average, see above) Al_2O_3 concentrations. The fact that corundum in the rocks contains Cr_2O_3 (up to 0.7 wt %), which is responsible for the reddish color of the mineral, also suggests genetic relations of the corundum with the host rocks.

GENESIS OF THE CORUNDUM

Two types of corundum-bearing metasomatic rocks are distinguished in northern Karelia: (1) metasomatites developing after rocks with high Al_2O_3 concentrations, such as plagiogneisses, and (2) metasomatites replacing gabbroids (Serebryakov, 2004). Zoisite–hornblende rocks in Kiy Island obviously affiliate with type 2 of metasomatites. The corundum was likely formed because of the host rocks were relatively rich in Al_2O_3. Genetic links between the corundum and host rocks also follows from the elevated Cr_2O_3 concentrations of the mineral.

No consensus is reached so far concerning the parameters under which corundum-bearing rocks were formed in northern Karelia. According to (Serebryakov and Aristov, 2004), corundum-bearing rocks of type 2 in northern Karelia were formed at $T = 630$–$680°C$, $P = 6$–8 kbar (garnet–hornblende and garnet–cordierite thermometry). The crystallization parameters of these rocks were estimated at no less than 720°C and no lower than 10 kbar by minimizing the thermodynamic potential with the Selector program package (Avchenko et al., 2007). Based on isotopic data, the crystallization temperatures of corundum-bearing rocks at Khit Island, northern Karelia, were estimated within the range of 400–500°C (Krylov et al., 2011; Ustinov et al., 2008), and these authors emphasize that these temperature estimates correspond to the termination of material exchange between the minerals and that the corundum could crystallize at higher temperatures.

The corundum-bearing hornblende–zoisite layers on Kiy island contain no garnet, and hence, the P–T estimates of the beginning of corundum crystallization can be slightly less accurate. However, the fact that the corundum occurs as euhedral crystals (Fig. 8a) indicates that it crystallized in the presence of fluid. It seems to be reasonable to suggest that the corundum crystals grew during the Svecofennian metamorphic peak ($T \approx 700°C$, $P \approx 10$ kbar) and related metasomatism of the gabbroids. The single veinlets and domains of secondary diaspore (Fig. 8b) suggest that this mineral was formed by the reaction $Al_2O_3 + H_2O = 2AlO[OH]$ at a temperature below 500°C (Haas, 1972), after the corundum was formed.

Obviously, to produce corundum in metasomatites replacing metabasite, the metasomatic fluid should have contained low NaCl and KCl concentrations, because otherwise SiO_2 and Al_2O_3 removed from the host rocks would have been incorporated in sodic plagioclase and feldspar, as is often the case with high-temperature metasomatism of mafic rocks, for example, during migmatization and charnockitization (Korikovsky and Khodorevskaya, 2006; Korikovsky and Aranovich, 2010; and others). Indeed, the absence of sodic plagioclase and pargasite–hastingsite amphibole from the host rocks indicates that the fluid that came to the metabasites at the peak of Svecofennian metamorphism did not contain high Na and K concentrations.

It is thus reasonable to suggest that the corundum crystals grew during high-pressure metamorphism at $P \approx 9$–10 kbar and $T \approx 700°C$. The probable minimum parameters of corundum crystallization in these rocks are 460–480°C and $a_{H_2O} \approx 0.20$–0.30.

CONCLUSIONS

(1) Coronite textures in metagabbroids in Kiy Island were formed at the peak of Svecofennian metamorphism and retrogression, starting at $T = 700°C$ and $P = 9.5$ kbar.

(2) The related high-pressure metasomatism of the metagabbroids produced metasomatites in the form of conformable garnet–hornblende and hornblende melanocratic veins and corundum-bearing hornblende–zoisite rocks.

(3) Genetic relations between the metagabbroids, which are noted for elevated Al_2O_3 and Cr_2O_3 concentrations, and the metasomatites follows from that their compositions are complementary and from that the metasomatites contain Cr_2O_3.

(4) The gabbroids were metasomatized and the melanocratic veins were produced with the involvement of fluid that practically always contained low salt

concentrations, and hence, the rocks contain no sodic plagioclase, and the amphibole is tschermakite instead of pargasite.

(5) Calculations indicate that the corundum grew at high-temperature metasomatism of the rocks at $P \approx$ 9–10 kbar and $T \approx 700°C$. The source of Al was the host metagabbroids, which are characterized by elevated Al_2O_3 concentrations.

(6) Low concentrations of alkalis in the fluid were favorable for Al redeposition in the form of corundum (instead of incorporation in sodic plagioclase, a mineral often produced by metasomatism of mafic rocks).

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