

Mineral Assemblages and Genesis of Hornfelses in the Outer Contact Zone of the Khibina Massif, Kola Peninsula, Russia

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Abstract—The paper presents materials on the fabric of the western, southwestern, and southern exocontact zones of the Khibina alkaline pluton and metavolcanic rocks of the Il'mozerskaya Formation of the Paleoproterozoic Imandra–Varzuga riftogenic structure. The volcanics of the Imandra–Varzuga structure were originally metamorphosed to the greenschist facies (at temperatures of $\geq 300^{\circ}\text{C}$ and pressures of ≥ 2.0 – 2.5 kbar) and were afterward metamorphosed to the pyroxene-hornfels facies under the thermal effect the Khibina pluton with the development of a hornfels zone 150–400 m thick. According to their composition, the hornfelses are subdivided into three zones: inner, intermediate, and outer. The inner zone is up to 30 m thick and consists of hornfelses of clinopyroxene–plagioclase composition with olivine as a typomorphic mineral and with variable amounts of amphibole. The intermediate zone occurs at a distance of 30–200 m from the pluton, is separated from the inner zone by the olivine isograd, and consists of amphibole–clinopyroxene–plagioclase hornfelses. The outer zone, 200–400 m away from the contact of the pluton, is made of fine-grained melanocratic hornblende hornfelses. The thermal transformations of the metavolcanics involved the gradual replacement of their low-temperature mineral assemblage (actinolite + albite) by a higher temperature one (clinopyroxene + amphibole + andesine–bytownite \pm olivine). Our data on the chemical composition of the rock-forming minerals of the hornfelses indicate that the olivine is ferrowortonolite–fayalite, the clinopyroxene belongs to the augite–ferroaugite series, and occasional orthopyroxene grains (which were found only in the intermediate zone) are ferrowortonolite. The amphibole in the hornfelses of the intermediate zone and the outermost (farthest from the contact) part of the inner zone is edenite, a Ca amphibole. The amphibole in hornfelses near the contact is kataphorite of the Na–Ca amphibole group. The plagioclase composition generally corresponds to andesine and bytownite and is albite–oligoclase near the contact with the pluton. The hornfelses adjacent to the contact bear rare sanidine grains. The mineral thermo- and barometry of the hornfelses yielded temperatures of 700 – 640°C and pressures of 1–1.5 kbar. The temperature determined for the zone exactly at the contact was approximately 700°C , which corresponds to the initial temperature of the rocks in contact with the magma and is close to the crystallization temperature of the nepheline syenites of the Khibina pluton.

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INTRODUCTION

The structure of the contact zone of the Khibina alkaline pluton has long attracted the attention of several researchers, who examined it in relation with the long-term (during several decades) studying of the pluton itself. The history of this research can be subdivided into two periods. During the first of them, the rocks in the contact aureole of the Khibina pluton were examined by B.M. Kupletskii (in 1924–1932), K.K. Sudislavlev (1934), E.N. Egorova (1935, 1936), I.S. Ozhinskii (1934, 1936), N.I. Soustov (1934, 1935, 1940), and L.I. Pazyuk (1940, 1947). It was then established that fenitization zones developed in granitoids and gneisses of the Kola block and pyroxene hornfelses replaced metavolcanics of the Imandra–Varzuga structure. Simultaneously, E.N. Egorova–Fursenko (1939) presented a fairly thorough and detailed description of the mineralogy of the contact hornfelses. The next period in the history of studying the contact rocks of the Khibina pluton encompasses the 1960s–1970s, when the structure of

the contact zone was examined by S.I. Zak (1963) and V.N. Gorstka (1971), who characterized the geologic setting of these rocks and described the petrography of all rocks composing the contact zone of the Khibina pluton, their mineral assemblages, chemical composition, the composition of gas components contained in them, and their density and porosity. However, these papers left unsettled some issues. For example, the chemistries of hornfels minerals were not examined because of the absence of necessary analytical techniques, and the conditions under which these contact rocks were formed were then not constrained.

This paper is an attempt to bridge this gap. Our research is based on the factual material obtained in the course of the 1 : 200000 follow-up geological survey of the central Kola Peninsula that was recently conducted by the Central Kola Expedition. These operations refined the geology of the structures hosting the Khibina pluton, the character of rock alterations, including the chemical and mineral composition of the contact

hornfels. These data are of certain interest, because quantitative chemical analyses of hornfels minerals made it possible to quantify the conditions under which the hornfels themselves and the adjacent rocks of the Khibina pluton were formed and provide insight into the processes of the thermal effect of the pluton onto the host rocks.

GEOLOGICAL OVERVIEW

The Khibina pluton is the world's largest intrusion of alkaline rocks. In map view, it has a round shape and is up to 45 km in diameter. The Rb–Sr age of the pluton lies within the range of 362–377 Ma [1], which corresponds to the Late Devonian. The pluton consists of two rock series, which are recognized as individual phases: older agpaitic alkaline nepheline syenites (khibinites), which comprise a number of compositional and textural varieties, and younger mafic and ultramafic foidolites (urtites and ijolites). The Khibina pluton is hosted in Archean gneisses and granitoids of the Kola block in the north and in metavolcanics of the Paleoproterozoic Imandra–Varzuga riftogenic structure in the south (Fig. 1). The inner contact zone within a few dozen meters from the contact (more rarely, at a distance of 200–300 m from it) contains abundant xenoliths of the host rocks of broadly varying shape and size. In the outer contact zone, the host rocks of the pluton are fenitized and hornfelsized (depending on their composition). The northern outer contact zone contains fenitized rocks of Archean age, predominantly plagiogranitoids and high-alumina gneisses of the Kola metamorphic complex, which are metamorphosed to the high-temperature amphibolite facies. Fenitization is pronounced as the partial or complete metasomatic replacement of primary minerals (quartz, plagioclase, and biotite) by newly formed alkaline minerals (microcline and aegirine-augite). The width of this zone normally does not exceed 70–100 m [2]. In the western, southwestern, and southern outer contact zones, hornfels developed after the rocks composing the western flank of the Paleoproterozoic Imandra–Varzuga riftogenic structure (Fig. 1), which is made up there of three groups: Strel'ninskaya (combined Kukshinskaya and Seidorechenskaya formations), Varzuga (Polisarskaya and Uмба formations), and Tomingskaya (Il'mozerskaya, Solenozerskaya, and Mitrijarvinskaya formations) [3, 4]. A significant part of the Imandra–Varzuga

structure is obviously truncated by the Khibina pluton (Fig. 1). In its western flank, the stratigraphic succession of the Tomingskaya Group is reduced and includes the Il'mozerskaya Formation alone, which is made up of metamorphosed basalts and basaltic andesites with thin layers of carbonaceous silty pelitic schists of variegated composition. The Khibina pluton has contacts only with the metabasalts of the Il'mozerskaya Formation, which are actinolite–plagioclase schists metamorphosed to the greenschist facies at temperatures of ≥ 300 – 350°C and pressures of ≥ 2 – 2.5 kbar [3]. This paper is devoted to the contact transformations of these rocks. The only reliable U–Pb zircon dates of rocks from the Imandra–Varzuga structure were obtained on acid metavolcanics of the Seidorechenskaya Formation: 2448 ± 8 Ma [5].

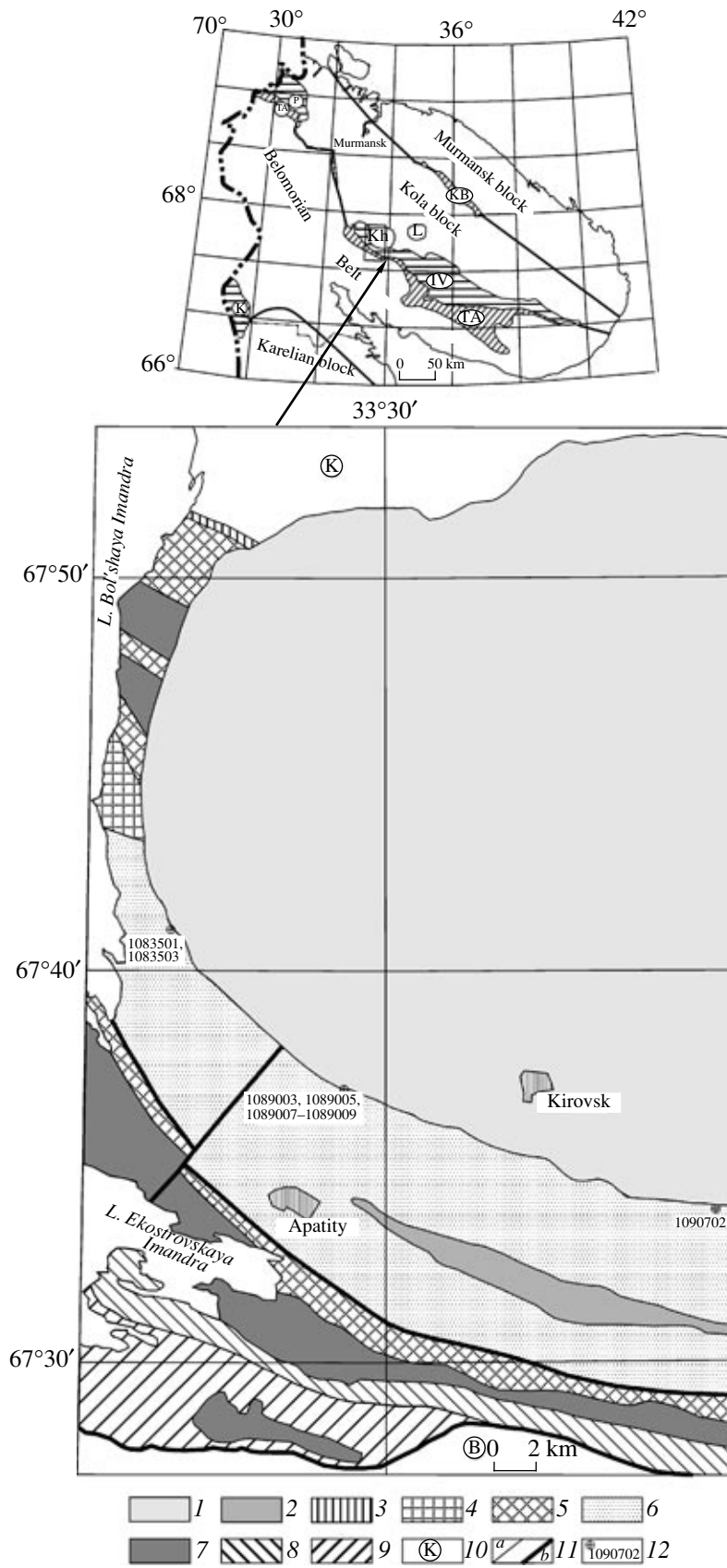
The stratigraphic section of the Imandra–Varzuga structure is significantly complicated by gabbroids of the Imandra lopolith (Fig. 1) [6], whose U–Pb zircon and baddeleyite ages lie within the range of 2446–2437 Ma [7–11]. The rocks of the Il'mozerskaya Formation are cut by the alkaline syenites of the Soustov Massif (Fig. 1), which was dated at 1900 ± 70 Ma by the Rb–Sr method [12] and at 1860 ± 8 Ma by the U–Pb method on zircon [13].

STRUCTURE OF THE CONTACT ZONE

Earlier, the contact belt of hornfels was subdivided into three zones 50–100 m wide each: an inner zone of diopside–plagioclase hornfels with fayalite and barkevikite, an intermediate zone of diopside–plagioclase–amphibole hornfels, and an outer zone of hornblende hornfels with the initial replacement of amphibole by diopside [14]. This structural scheme of the outer contact aureole (150–400 m thick) of the Khibina pluton with metavolcanic rocks was generally confirmed in the course of further research [2, 15, 16].

We refined the structure of the inner hornfels zone immediately adjacent to the Khibina pluton. Our data indicate that this zone is approximately 30 m thick and consists of clinopyroxene–plagioclase hornfels with variable contents of amphibole and olivine as a polymorphic mineral. We documented the contact of the Khibina pluton in a roadway excavation (exposures 1089003, 1089005, 1089007, and 1089008), in which hornfels in contact with medium-grained nepheline syenites of the pluton are exposed within a 2-m interval.

Fig. 1. Schematic geological map of the southern and western contact zones of the Khibina pluton. (1) Nepheline syenites and foidolites of the Paleozoic Khibina pluton; (2) Early Karelian alkaline syenites of the Soustova Massif; (3–6) volcanic–sedimentary rocks of the Paleoproterozoic (Karelian) Imandra–Varzuga riftogenic structure: (3) Kukshinskaya Formation, (4) Seidorechenskaya Formation, (5) Polisarskaya Formation, (6) Il'mozerskaya Formation; (7) gabbroids of the Imandra lopolith; (8, 9) volcanic rocks of the Late Archean (Lopian) Tersko–Allorchka greenstone belt: (8) Kislogubskaya Formation, (9) Vitegubskaya Formation; (10) metamorphic and ultrametamorphic Early (?) and Late Karelian rocks of (K) the Kola block and (B) Belomorian Mobile Belt; (11) (a) geological boundaries and (b) faults; (12) sampling sites and sample numbers. The schematic map of the Kola Peninsula shows the following horizontal hatching—Paleoproterozoic riftogenic structures: (P) Pechenga, (IV) Imandra–Varzuga, and (K) Kuolajarvi; oblique hatching—Lopian greenstone belts: (KV) Kolmozero–Voroninskii and (TA) Tersko–Allarechenskii; plutons: (L) Lovozero and (Kh) Khibina.



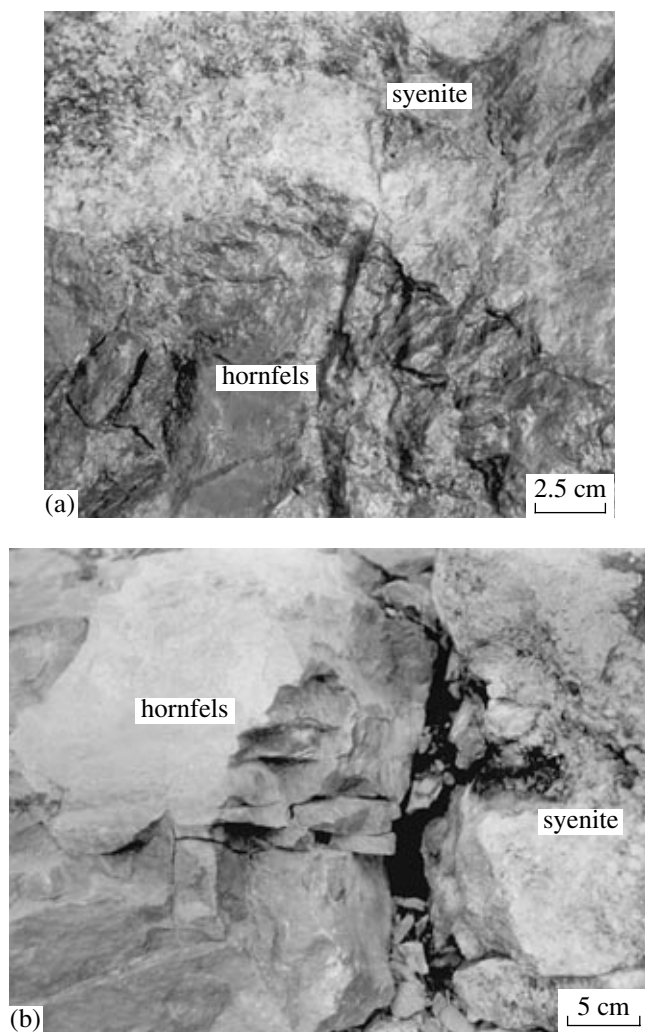


Fig. 2. Contact between hornfels (dark gray) and alkaline syenite (pale gray) of the Khibina pluton in (a) a horizontal plane and (b) vertical wall.

The contact is sharp, weakly twisty in map view (Fig. 2a), generally trends 290° WNW, steep in vertical section (Fig. 2b) with a northeastern dip at angles of $80\text{--}85^\circ$. The hornfels are fine-grained homogeneous rocks of brownish to dark gray color. As can be seen under a microscope, the line of the contact is often marked by a stripe (up to 0.5 mm thick) of separated amphibole and dark brown biotite grains and their aggregates, with individual mineral grains ranging from 0.2 to 0.5 mm. Biotite was found only in this stripe, in which it occurs as thin rims around amphibole and individual grains (Fig. 3a). Amphibole and biotite sometimes contain inclusions of clinopyroxene, olivine, and ore minerals and thus acquire a poikilitic texture. The development of the biotite–amphibole stripe was likely controlled by an increase in the activities of alkalis due to their diffusion from the melt at a temperature peak. When the biotite–amphibole stripe disappears, the granoblastic olivine–amphibole–clinopyroxene–plagioclase horn-

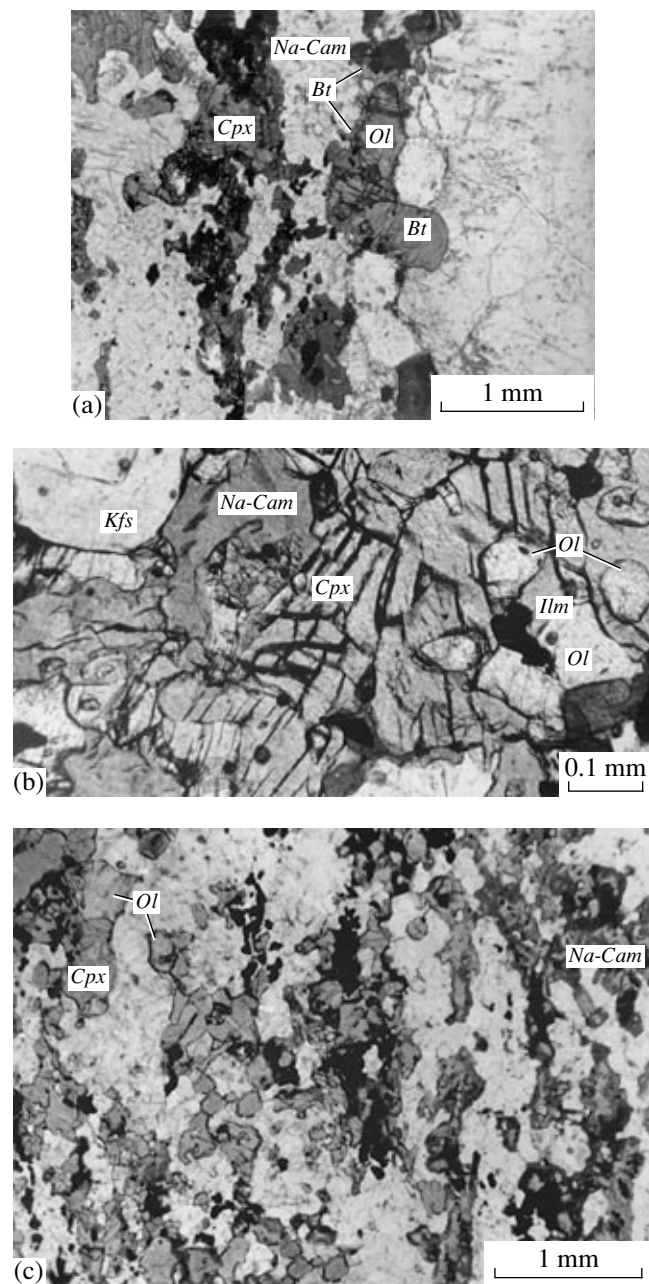


Fig. 3. Microphotographs of hornfels from the zone of immediate contact with the Khibina pluton (sample 1089005). (a) Contact of fine-grained hornfels with medium-grained alkaline syenite. The contact is marked by a discontinuous chain of amphibole–biotite aggregates with rare olivine grains; (b–c) olivine–amphibole–clinopyroxene–plagioclase hornfels: (b) adjacent to the contact, (c) at a distance of 2–6 mm from the contact.

fels (with elements of a poikilitic texture in the form of olivine inclusions in clinopyroxene, Fig. 3b) occur in immediate contact with the rocks of the pluton. Nepheline syenites within 5 mm from the contact often contain assimilated individual grains of olivine, amphibole, and biotite. The hornfels adjacent to the contact

Table 1. Mineral composition of outer-contact hornfelses around the Khibina pluton

Minerals	Zone							
	Inner				Intermediate		Outer	
	Sample no.							
	1089005	1089003	1089007	1089008	1083503	1090702	1083501	1089009
Distance from contact, m								
0.0	0.2	1	2	30	50	190	250	
Content (%)								
Olivine	2–3	r.g.	r.g.	r.g.	r.g.	–	–	–
Orthopyroxene	–	–	–	–	–	r.g.	–	–
Clinopyroxene	33	35	30	30	45	40	60	5
Na–Ca amphibole	2–3	–	r.g.	–	–	–	r.g.	–
Ca amphibole	–	–	–	–	10	7	10	80
Plagioclase	60	60	68	70	40	50	30	10
Ore mineral	5	5	2	r.g.	5	3	r.g.	5
Biotite	r.g.	–	–	–	–	–	–	–

(within a zone 2 to 5 mm thick) have an amphibole–clinopyroxene–plagioclase composition with olivine and contain up to 10–15% brown amphibole. The latter usually developed in the form of rims around clinopyroxene, and the plagioclase occurs as aggregates of small (0.1–0.2 mm) grains (Fig. 3c). At a 20-cm distance from the contact, the clinopyroxene–plagioclase hornfelses with variable contents of brown amphibole (Table 1) contain olivine only as rare grains (sample 1089003, Fig. 4a), and its amount does not change within a distance of up to 30 m from the contact (Fig. 4b), where the content of amphibole (of pale green color) increases to 10% (Table 1).

The intermediate and outer zones of the hornfelses were examined only fragmentarily because of their poor exposure. The boundaries of the intermediate zone pass at 30 and 200 m, respectively, from the contact with the pluton, and this zone is separated from the inner zone by the olivine isograd. The zone consists of fine-grained (grains 0.1–0.5 mm across) amphibole–clinopyroxene–plagioclase hornfels (samples 1090702 and 1083501; Table 1) with a microgranoblastic (hornfels) texture (Figs. 4c, 4e), which sometime grades into a thin porphyroblastic due to the presence of larger amphibole grains (Fig. 4d). The rock is often banded because of alternating lamina (~2 mm thick) with various amphibole contents. The outer hornfels zone occurs at distances of 200–400 m from the contact and consists of fine-grained melanocratic hornblende hornfels (sample 1089009, Table 1) of nematoblastic texture, sometimes with up to 5% newly formed clinopyroxene.

MINERALOGY OF HORNFELSES

The chemical compositions of minerals from the hornfelses were determined on a Cameca MS-46 microprobe at the Kola Research Center of the Russian Academy of Sciences in Apatity (analyst Ya.A. Pakhomovskii). The results are listed in Table 2.

Olivine was found only in the hornfelses of the inner zone, in which it occurs as irregularly shaped (without clear crystallographic outlines) grains up to 0.21–0.5 mm across (Fig. 3). Chemically (Table 2), it corresponds to ferrohortonolite–fayalite (Fa_{85-91}). Fayalite is a quite common mineral of regional and contact-metamorphic iron-rich rocks [20–24], but we are not aware of its other finds in contact-metamorphic metabasites. Many contact aureoles of metabasic rock metamorphosed to the pyroxene-hornfels facies were described to contain much more magnesian olivine [25–27]. For example, the outer contact zone of ultramafic volcanic rocks around the Laramie anorthosite complex in Wyoming (whose inner-contact rocks are ferrosyenites and ferromonzonites) contains hortonolite (Fa_{64}) [26]. At the same time, mineral assemblages with fayalite are quite typical of Fe-rich rocks in the Kola block that are regionally metamorphosed to the granulite facies [20]. Our fayalite is richer in Mg and Mn than this mineral in the Fe-rich rocks of the Kola block.

The pyroxenes of the hornfelses are strongly dominated by clinopyroxene (according to observations in thin sections). The orthopyroxene is ferrohypersthene ($En_{45.7}Fs_{52.1}Wo_{2.2}$, Fig. 5a) and was found in the form of rare grains only in one hornfels sample from the intermediate zone (sample 1090702, 50 m from the contact). Clinopyroxenes from the hornfelses of the intermediate zone and from the part of the inner zone most distant from the contact correspond to augite, and this

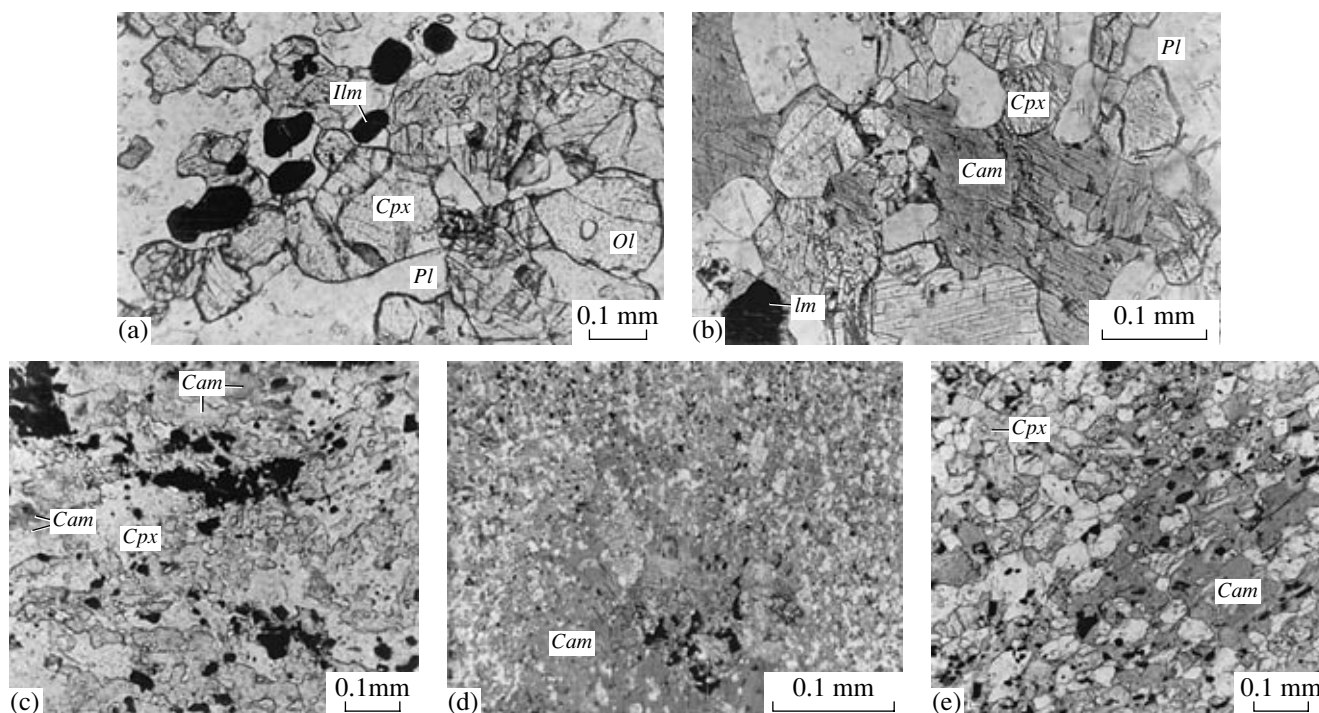


Fig. 4. Microphotographs of hornfels from various zones of the contact aureole. (a, b) Hornfels from the inner zone: (a) olivine-clinopyroxene-plagioclase hornfels at a 20-cm distance from the contact (sample 1089003), (b) amphibole-clinopyroxene-plagioclase hornfels at a 30-m distance from the contact (sample 1083503); (c–e) amphibole-clinopyroxene-plagioclase hornfels from the intermediate zone: (c) 50 m from the contact (sample 1090702), (d, e) 190 m from the contact (sample 1083501).

mineral from the hornfels of the inner zone near the contact is ferroaugite (Fig. 5a). The latter is highly ferrous and bears elevated Ti concentrations (Fig. 5b). It is worth noting that the highest Na concentrations (1.11 wt % Na₂O, Table 2, Figs. 5c, 5d), coupled with elevated concentrations of Fe₂O₃ (1.54 wt %, Table 2), were detected only in clinopyroxene from the vicinity of the contact with the pluton. This is explained by the presence of the aegirine end member in this clinopyroxene. However, already at a distance of 20 cm from the contact, the Na concentration in the clinopyroxene decreases to 0.26% (Table 2), which suggests that alkaline metasomatism occurred only locally. This clinopyroxene shows no correlation between its Na and Al concentrations (Fig. 5c) and between these components and the Fe mole fraction (Figs. 5c, 5d). This implies that the clinopyroxene contains no such end members as jadeite, a fact that is also corroborated by the usual absence of Fe³⁺. The possible exception is clinopyroxene from the outer zone (sample 1083501, Table 2), but its elevated Fe₂O₃ concentrations are not coupled with elevated contents of Al₂O₃ and are likely related to a relatively high oxygen fugacity.

Clinopyroxene is a typomorphic mineral of metabasites metamorphosed to the pyroxene-hornfels facies [25]. Clinopyroxene in hornfels from the outer contact zone of the Khibina pluton notably differs from this mineral in typical contact hornfels [28] in having higher Si and Al^{IV} concentrations. Our clinopyroxenes are compositionally fairly close to this mineral in analogous hornfels of the Laramie anorthosite complex [26] but differ from them in having higher Fe mole fractions (Fig. 5) and somewhat lower Al concentrations. The comparison of the compositions of clinopyroxene from the outer-contact hornfels of the Khibina pluton and from the rocks of the Kola block (which were regionally metamorphosed to the granulite facies [20]) indicates that the former are characterized by broader variations in the Ti and Al concentrations (and, often, by their higher values) and in the Fe mole fraction (Fig. 5b) and show more clearly pronounced negative correlations between the Fe mole fraction and the concentrations of Ti, Al, Na, and, to a lesser extent, between Al and Na.

Fig. 5. Chemical composition of pyroxenes from the hornfels. (a) *Wo-Fs-En* classification diagram; (b) $\text{TiO}_2\text{-(Fe}^{2+} + \text{Fe}^{3+})/(\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mg})$ diagram; (c) Na₂O–Al₂O₃ diagram; (d) correlation of Na₂O and Al₂O₃ with $(\text{Fe}^{2+} + \text{Fe}^{3+})/(\text{Fe}^{2+} + \text{Fe}^{3+} + \text{Mg})$. (1–3) Compositional fields (with data points) of clinopyroxene: (1) from outer-contact hornfels around the Khibina pluton, (2) from outer-contact hornfels around the Laramie anorthosite complex [26], (3) from crystalline schists of the Kola block [20]. Here and in Figs. 6, 7, the numbers of data points correspond to sample numbers in Table 2.

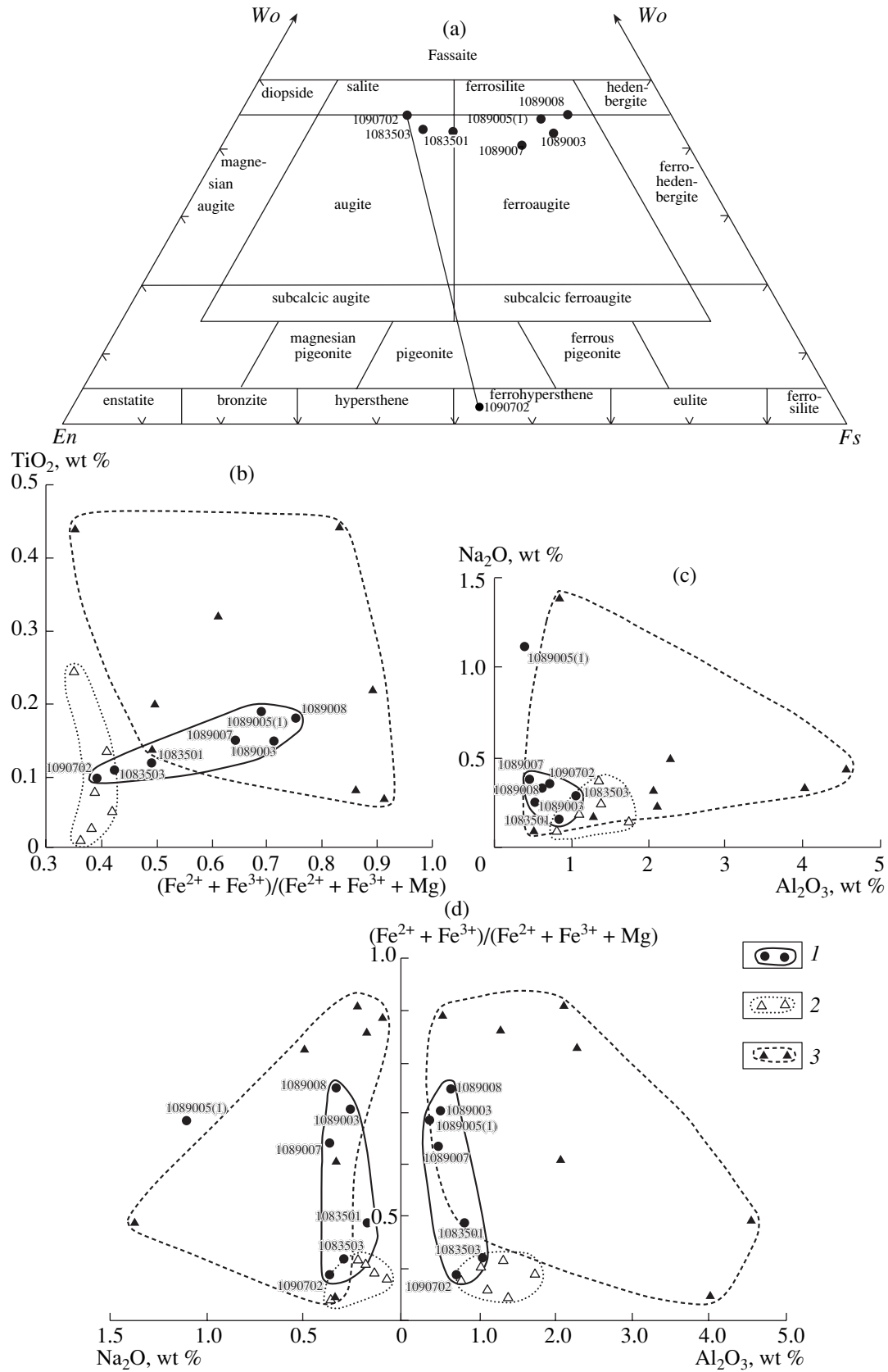


Table 2. Chemical composition of rock-forming minerals in outer-contact hornfelses around the Khibina pluton

Sample	1089005										1089003			1089007			
	<i>Ol</i> (1)	<i>Ol</i> (2)	<i>Cpx</i> (1)	<i>Na-Cam</i> (1)	<i>Na-Cam</i> (2)	<i>Bt</i> (1)	<i>Kfs</i> (1)	<i>Pl</i> (2)	<i>Ilm</i> (1)	<i>Ilm</i> (2)	<i>Ol</i>	<i>Cpx</i>	<i>Pl</i>	<i>Ol</i>	<i>Cpx</i>	<i>Pl</i>	<i>Ilm</i>
SiO ₂	30.45	30.28	50.25	45.99	42.66	33.96	66.63	66.62	0.22	0.28	30.40	50.36	65.28	30.94	49.93	64.37	0.28
TiO ₂	0.04	0.07	0.19	1.55	2.10	3.43	0.00	0.00	50.96	50.78	0.04	0.15	0.00	0.00	0.15	0.00	52.32
Al ₂ O ₃	0.00	0.21	0.37	4.50	6.60	14.15	18.60	20.34	0.13	0.12	0.12	0.49	20.58	0.00	0.46	21.51	0.15
Fe ₂ O ₃	0.00	0.00	1.54	0.00	0.00	—	—	—	—	—	0.00	0.00	—	0.00	0.00	—	—
FeO	62.14	63.15	19.11	21.67	24.43	—	—	—	—	—	62.68	23.45	—	60.97	21.10	—	—
FeO*	—	—	—	—	—	29.72	0.19	0.00	46.45	45.94	—	—	0.27	—	—	0.09	43.14
MnO	2.01	2.72	0.59	0.30	0.44	0.39	0.00	0.00	1.10	1.66	1.28	0.60	0.00	0.97	0.44	0.00	0.52
MgO	3.95	2.08	5.43	7.92	5.59	5.20	0.00	0.00	0.13	0.07	4.37	5.27	0.00	6.27	6.56	0.00	0.63
CaO	0.42	0.33	20.16	9.08	9.88	0.00	0.00	1.07	0.09	0.06	0.12	19.14	1.53	0.19	17.86	3.57	0.04
Na ₂ O	0.00	0.00	1.11	4.10	3.63	0.24	5.36	10.51	0.00	0.00	0.00	0.26	9.29	0.00	0.37	10.42	0.00
K ₂ O	0.00	0.00	0.00	1.15	1.34	8.85	8.93	0.31	0.00	0.00	0.00	0.00	1.24	0.00	0.00	0.86	0.00
Total	99.01	98.84	98.75	96.26	96.67	95.94	99.71	98.85	99.08	98.91	99.01	99.72	98.19	99.34	96.87	100.82	97.08
<i>Numbers of cations</i>																	
Si	1.007	1.013	2.004	7.201	6.802	2.728	3.008	2.949	0.006	0.007	1.003	2.007	2.924	1.004	2.022	2.840	0.007
Ti	0.001	0.002	0.006	0.183	0.252	0.207	0.000	0.000	0.979	0.978	0.001	0.004	0.000	0.000	0.005	0.000	1.008
Al ^{IV}	0.000	0.000	0.000	0.799	1.198	1.272	0.990	1.051	—	—	0.000	0.000	1.076	0.000	0.000	1.118	—
Al ^{VI}	0.000	0.008	0.017	0.032	0.043	0.068	0.000	0.010	—	—	0.005	0.023	0.010	0.000	0.022	0.000	—
Al _{tot}	0.000	0.008	0.017	0.831	1.241	1.340	0.990	1.061	0.004	0.004	0.005	0.023	1.086	0.000	0.022	1.118	0.005
Fe ³⁺	0.000	0.000	0.050	0.000	0.000	—	—	—	—	—	0.000	0.000	—	0.000	0.000	—	—
Fe ²⁺	1.718	1.766	0.683	2.838	3.258	—	—	—	—	—	1.730	0.782	—	1.655	0.714	—	—
Fe*	—	—	—	—	—	1.997	0.007	0.000	0.993	0.984	—	—	0.010	—	—	0.003	0.924
Mn	0.056	0.077	0.020	0.040	0.059	0.026	0.000	0.000	0.024	0.036	0.036	0.020	0.000	0.027	0.015	0.000	0.011
Mg	0.195	0.104	0.323	1.848	1.328	0.622	0.000	0.000	0.005	0.003	0.215	0.313	0.000	0.303	0.396	0.000	0.024
Ca	0.015	0.012	0.861	1.523	1.688	0.000	0.000	0.051	0.002	0.002	0.004	0.818	0.073	0.007	0.775	0.169	0.001
Na	0.000	0.000	0.086	1.245	1.122	0.038	0.469	0.902	0.000	0.000	0.000	0.020	0.807	0.000	0.029	0.892	0.000
K	0.000	0.000	0.000	0.230	0.273	0.907	0.514	0.017	0.000	0.000	0.000	0.000	0.071	0.000	0.000	0.049	0.000
<i>En</i> (<i>Ab</i>)	—	—	16.7	—	—	—	47.7	93.0	—	—	—	16.2	84.8	—	20.8	80.4	—
<i>Fs</i> (<i>An</i>)	—	—	38.9	—	—	—	0.0	5.3	—	—	—	41.5	7.7	—	38.4	15.2	—
<i>Wo</i> (<i>Or</i>)	—	—	44.4	—	—	—	52.3	1.7	—	—	—	42.3	7.5	—	40.8	4.4	—
<i>f</i>	0.90	0.94	0.69	0.61	0.71	0.76	—	—	—	—	0.89	0.71	—	0.85	0.64	—	—

Table 2. (Contd.)

Sample	1089008					1083503					1090702					1083501																			
	Ol	Cpx	Na-Cam	Pl	Ilm	Cpx	Cam	Pl	Opx	Cpx	Cam	Pl	TiMg	Cpx	Cam	Pl	Si	Ti	Al ^{IV}	Al ^{VI}	Al _{tot}	Fe ³⁺	Fe ²⁺	Fe*	Mn	Mg	Ca	Na	K	En (Ab)	Fs (An)	Wo (Or)	f		
SiO ₂	30.51	50.03	40.26	56.67	0.18	51.38	45.13	56.99	50.69	52.76	47.31	49.28	0.66	50.19	43.13	55.86	1.014	2.002	6.510	2.595	0.005	0.005	0.005	0.955	0.040	0.166	0.013	0.000	0.000	0.000	0.000	0.000	0.000		
TiO ₂	0.00	0.18	2.45	0.00	52.16	0.11	1.62	0.00	0.14	0.10	0.85	0.00	11.81	0.12	1.74	0.00	0.000	0.298	0.298	0.000	0.000	0.000	0.016	0.040	0.166	0.013	0.000	0.000	0.000	0.000	0.000	0.000	0.000		
Al ₂ O ₃	0.00	0.60	8.93	26.06	0.00	1.04	8.70	25.49	0.55	0.69	6.27	30.37	1.47	0.82	9.26	27.44	—	0.000	1.490	0.212	—	—	0.007	0.000	0.166	0.013	0.882	1.894	0.887	0.000	0.000	0.000	0.000		
Fe ₂ O ₃	0.00	0.00	0.00	—	—	0.00	12.96	—	0.00	0.00	7.29	—	—	0.00	14.11	—	—	0.000	0.212	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.026	0.887	0.000	0.000	0.000	0.000	0.000		
FeO	63.10	23.27	24.51	—	—	14.39	8.18	—	30.30	12.99	11.56	—	—	13.94	7.21	—	—	—	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.026	0.887	0.000	0.000	0.000	0.000	0.000		
FeO*	—	—	—	0.20	44.52	—	—	0.24	—	—	—	0.32	78.60	—	—	0.25	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—		
MnO	1.41	0.49	0.34	0.00	0.74	0.38	0.30	0.00	0.76	0.35	0.20	0.00	0.20	0.37	0.19	0.00	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—		
MgO	3.35	4.25	4.46	0.00	0.18	10.94	9.49	0.00	14.94	11.31	11.12	0.00	0.19	10.18	8.32	0.00	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
CaO	0.37	20.57	10.93	8.33	0.06	20.06	10.92	9.30	1.02	21.17	10.85	14.17	0.18	21.15	10.66	9.94	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Na ₂ O	0.00	0.33	2.83	6.33	0.00	0.29	1.81	6.09	0.00	0.36	0.98	2.75	0.00	0.16	2.10	6.06	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
K ₂ O	0.00	0.00	0.55	0.07	0.00	0.00	0.35	0.05	0.00	0.00	0.46	0.08	0.00	0.00	0.30	0.01	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Total	98.74	99.72	95.26	97.66	97.84	98.59	99.46	98.16	98.40	99.73	96.89	96.97	93.11	98.68	97.02	99.56	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Numbers of cations																																			
Si	1.014	2.002	6.510	2.595	0.005	1.986	6.796	2.603	2.002	2.004	7.171	2.313	0.030	1.955	6.698	2.525	1.014	2.002	6.510	2.595	0.005	0.005	0.005	0.955	0.040	0.166	0.013	0.000	0.000	0.000	0.000	0.000	0.000	0.000	
Ti	0.000	0.005	0.298	0.000	1.006	0.003	0.183	0.000	0.004	0.003	0.097	0.000	0.399	0.004	0.203	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.016	0.040	0.166	0.013	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Al ^{IV}	—	0.000	1.490	—	—	0.014	1.204	1.372	0.000	0.000	0.829	1.680	—	0.035	1.302	1.462	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Al ^{VI}	—	0.028	0.212	—	—	0.033	0.341	0.000	0.026	0.031	0.291	0.000	—	0.003	0.393	0.000	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Al _{tot}	0.000	0.028	1.702	1.407	0.000	0.047	1.545	1.372	0.026	0.031	1.120	1.680	0.078	0.038	1.695	1.462	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Fe ³⁺	0.000	0.000	0.000	—	—	0.000	1.421	—	0.000	0.000	0.816	—	—	0.057	1.590	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Fe ²⁺	1.754	0.779	3.315	—	—	0.465	0.998	—	1.001	0.413	1.439	—	—	0.506	0.904	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Fe*	—	—	—	0.007	0.955	—	—	0.009	—	—	—	0.012	2.951	—	—	0.009	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Mn	0.040	0.017	0.047	0.000	0.016	0.012	0.038	0.000	0.025	0.011	0.026	0.000	0.007	0.012	0.025	0.000	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Mg	0.166	0.253	1.075	0.000	0.007	0.631	2.130	0.000	0.879	0.640	2.512	0.000	0.013	0.591	1.926	0.000	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Ca	0.013	0.882	1.894	0.409	0.002	0.831	1.762	0.455	0.043	0.862	1.762	0.713	0.008	0.883	1.774	0.481	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Na	0.000	0.026	0.887	0.562	0.000	0.022	0.528	0.540	0.000	0.026	0.288	0.250	0.000	0.012	0.632	0.531	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
K	0.000	0.000	0.113	0.004	0.000	0.000	0.067	0.003	0.000	0.000	0.089	0.005	0.000	0.000	0.059	0.002	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
En (Ab)	—	13.1	—	57.6	—	32.5	—	54.1	45.1	33.2	—	25.8	—	28.8	—	52.4	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Fs (An)	—	41.2	—	42.0	—	24.6	—	45.6	52.7	22.0	—	73.7	—	28.1	—	47.4	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Wo (Or)	—	45.7	—	0.4	—	42.9	—	0.3	2.2	44.7	—	0.5	—	43.1	—	0.2	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
f	0.91	0.75	0.76	—	—	0.42	0.53	—	0.53	0.39	0.47	—	—	0.49	0.56	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	

Note: Numerals in parentheses for sample 1089005 mean: 1—minerals at contact with alkaline syenite, 2—minerals assimilated by alkaline syenite, Fe²⁺ and Fe³⁺ were estimated by the method [17] for pyroxenes and olivine and by the method [18, Appendix 1] for amphiboles; Fe* is all Fe in the form of Fe²⁺; f = (Fe²⁺ + Fe³⁺)/(Fe²⁺ + Fe³⁺ + Mg). Here and in Table 4, in the text, and in figures, mineral symbols are according to [19]: Ab—albite, And—andesine; Act—actinolite, Aug—augite, Cam—calcic amphibole, Na—Cam—sodic-calcic amphibole, Chl—chlorite, Cpx—clinopyroxene, Ed—edenite, Ep—epidote, Fa—fayalite, Ilm—ilmenite, Kfs—potassic feldspar, Opx—orthopyroxene, Ol—olivine, Pl—plagioclase, Qtz—quartz, TiMag—titanomagnetite.

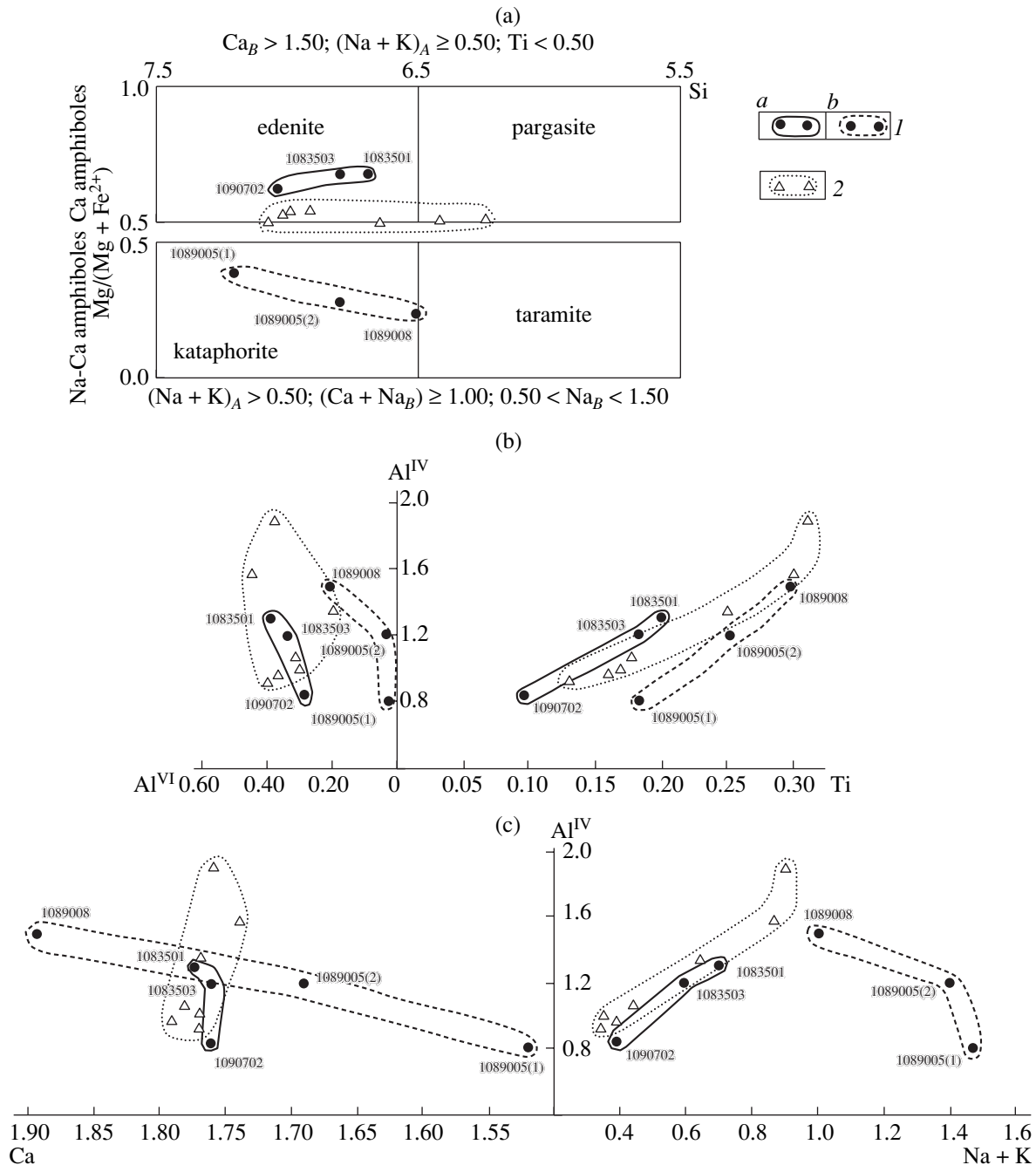


Fig. 6. Chemical composition of amphiboles from the hornfelses. (a) $Mg/(Mg + Fe^{2+})$ –Si diagram [18]; (b) correlation of Al^{VI} and Ti with Al^{IV} ; (c) correlation of Ca and (Na + K) with Al^{IV} . (1) Compositional fields of amphiboles from the outer-contact hornfelses of the Khibina pluton: (a) calcic, (b) sodic–calcic; (2) compositional fields of calcic amphiboles from the outer-contact hornfelses of the Laramie Complex [26].

According to the latest systematics proposed by Leake et al. [18], the contact-metamorphic amphiboles belong to the calcic and sodic–calcic groups. It should be mentioned that our amphiboles were ascribed to the sodic–calcic group provisionally, because their classification parameters, such as $(Ca + Na)_B > 1.50$ and $Na_B < 0.50$, correspond to the group of calcic amphiboles, and only amphiboles from the inner zone in the vicinity of the

contact with the Khibina pluton additionally satisfy the condition $(Na + K)_A > 0.50$ and were thus ascribed to the sodic–calcic group. The composition with calcic amphiboles in Leake’s diagram [18] corresponds to edenite, and the sodic–calcic amphiboles are kataphorites (Fig. 6a). These minerals commonly occur in the form of anhedral grains ranging from 0.1 to 1 mm across and, occasionally, as poikiloporphroblasts,

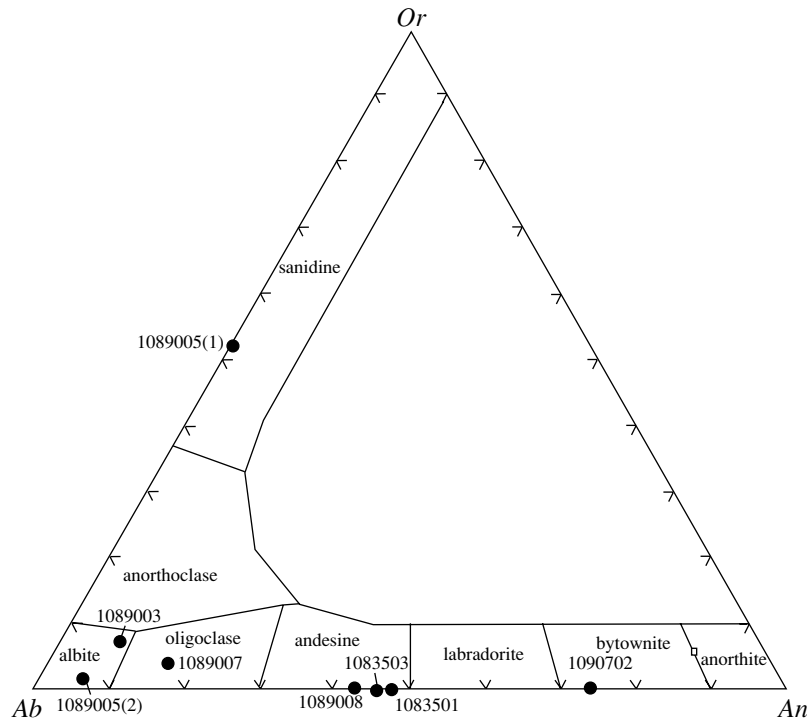


Fig. 7. *Ab-Or-An* classification diagram for feldspars from hornfelses.

which occur predominantly in the hornfelses of the outer zone. This highlights the later crystallization of amphibole compared to other minerals of the hornfelses. The composition of the amphiboles shows strong correlation with their position within the hornfels zone. For example, the most ferrous amphibole was found in the inner hornfels zone, within 2 m from the contact with the pluton ($f = 0.61-0.76$, Table 2), and they also bear the highest concentrations of alkalis (2.83–4.10 wt % Na_2O , 0.55–1.35 wt % K_2O , Table 2), whereas amphibole in the inner zone (at a 30-m distance from the contact) and the intermediate zone has f of 0.47–0.56 (Table 2), and its alkali contents decrease (0.98–2.10 wt % Na_2O , 0.30–0.46 wt % K_2O , Table 2). As in the situation with the clinopyroxene, this implies that alkaline metasomatism affected the rocks only locally. We revealed correlations between the concentrations of some elements and Al^{IV} , which is thought to be a temperature-sensitive parameter. For example, the calcic and sodic-calcic amphiboles display positive correlations of the contents of Al^{IV} with Ti and, to a lesser extent, Al^{VI} , which is a pressure-sensitive parameter (Fig. 6b). Moreover, the calcic amphiboles show a positive correlation between Al^{IV} and $(\text{Na} + \text{K})$, whereas this correlation in the sodic-calcic amphiboles is negative (Fig. 6c). A closely similar correlation was established for the amphiboles of this group between the concentrations of Al^{IV} and Ca, whereas the calcic amphiboles exhibit no such correlation (Fig. 6c). Our calcic amphiboles appear to be compositionally very similar to this mineral in the outer-contact hornfelses of the Laramie anorthosite com-

plex [26] and differ from them only in having a slightly higher Mg mole fraction (Fig. 6).

The composition of plagioclase in the hornfelses significantly varies. The intermediate-zone hornfelses contain andesine and bytownite ($Pl_{47.4-73.7}$) poor in K (Table 2, Fig. 7), whereas the inner-zone hornfelses bear andesine-albite ($Pl_{45.6-5.3}$), which clearly becomes less calcic closer to the contact with the Khibina pluton (Fig. 7) and shows a simultaneous notable increase in the K_2O concentration (from 0.05 to 1.24 wt %, Table 2). The highest K_2O concentrations were detected in feldspar sampled at the immediate contact with the pluton (8.93 wt % K_2O , Table 2). This feldspar corresponds to sanidine in the *Ab-An-Or* diagram (Fig. 7). A notable fact is the relatively low Al concentration in plagioclase from hornfelses at the immediate contact (20.34–20.58 wt % Al_2O_3) compared to plagioclase from the intermediate and outer zones (25.49–30.37 wt % Al_2O_3).

The minor minerals are biotite and ore minerals (ilmenite and titanomagnetite). The biotite occurs as platy grains that compose discontinuous rims along the contact of the hornfelses and Khibina pluton. The biotite chemically corresponds to lepidomelane, which is characterized by an elevated TiO_2 concentration (3.43 wt %) and a high Fe mole fraction (0.76) owing to its fairly high FeO content (29.72 wt %) and a low concentration of MgO (5.20 wt %, Table 2). The ore minerals show certain correlations between their compositional parameters and position in various hornfels zones. For example, the ore mineral of the inner-zone hornfelses is ilmenite, which is contained in these rocks in fairly signif-

Table 3. Equilibrium temperatures and pressures calculated for the outer-contact hornfels of the Khibina pluton

Sample no.	<i>Ol-Cpx</i>	<i>Cpx-Am</i>	<i>Am-Pl</i>	<i>Am</i>	<i>Al-in Am</i>
	1	2	3	4	5
	Temperature, °C				Pressure, kbar
1089005	691	–	–	–	–
1089003	708	–	–	–	–
1089007	754	–	–	–	–
1089008	679	–	–	–	–
1083503	–	680	645	800	1.0
1090702	–	680	660	660	1.0
1083501	–	640	685	815	1.5

Note: (1–4) Geothermometers: (1) olivine–clinopyroxene [29]; (2) clinopyroxene–amphibole [30]; (3) amphibole–plagioclase [30]; (4) amphibole [31]; (5) geobarometer based on Al concentration in amphibole.

icant amounts (up to 5%). The mineral bears elevated Mn concentrations and Al and Ca as minor components (Table 2). The ore mineral of the intermediate-zone hornfels is titanomagnetite (11.81 wt % TiO₂, Table 2) with elevated concentrations of Al and admixtures of Mn, Mg, and Ca.

MINERAL ASSEMBLAGES OF HORNFELSES

The mineral assemblages of mafic metavolcanics metamorphosed under greenschist-facies conditions are gradually replaced by the mineral assemblages of hornfels of the pyroxene-hornfels facies closer to the Khibina pluton. The characteristic mineral assemblages of the mafic metavolcanics are *Act + Ab + Ep + Chl ± Qtz*, *Act + Ab + Chl ± Qtz*, and *Act + Ab ± Qtz* [3] with fairly high contents of magnetite, owing to the relatively high Fe contents of the rocks and their origin under an elevated oxygen fugacity. In the outer-zone hornfels, actinolite is replaced by common hornblende, chlorite and plagioclase gradually disappear, and plagioclase becomes more calcic (up to andesine). In the hornfels of the intermediate zone, common hornblende is replaced by augite and calcic amphibole of edenite composition with the origin of the *Aug₄₉ + Pl₄₇ + Cam₅₆* assemblage (subscript indices near the symbols of Fe–Mg silicates denote their Fe mole fractions, and those near plagioclase denote its anorthite content). Near its boundary with the inner zone, the hornfels contain rare orthopyroxene grains, the Fe mole fractions of the edenite and augite decrease, the plagioclase becomes more calcic, and the mineral assemblage is *Pl₇₄ + Aug₃₉ + Cam₄₇ + Opx₅₃*. In the inner-zone hornfels away from the contact with the pluton, orthopyroxene disappears, olivine appears, the augite and edenite become more ferrous, the plagioclase becomes less calcic,

and the mineral assemblage *Aug₄₂ + Pl₄₆ + Cam₅₃ + Ol* is formed. Closer to the contact with the pluton, the inner-zone hornfels contain no edenite but bear newly formed Na–Ca amphibole of kataphorite composition, the plagioclase becomes less calcic, and the augite becomes more ferrous (up to ferroaugite), so that the *Pl₅₋₄₆ + Aug₄₂₋₇₅ + Ol₈₅₋₉₁ ± Na-Cam₆₁₋₇₆* assemblage is formed.

A disputable issue concerning contact metamorphism is the completeness of metamorphic reactions in relation to the limited time of the metamorphic processes and, consequently, the possibility for minerals existing in a rock to reach equilibrium relations. According to Reverdatto [25], equilibrium between mineral phases newly formed under pyroxene-hornfels facies conditions is fully reached everywhere within this facies, a fact significantly facilitated by the long duration of metamorphic processes and the presence of water and other volatile components in the system. This seems to be fully applicable to the inner-zone hornfels in the outer contacts of the Khibina pluton. Equilibrium could be reached in these rocks because their high temperature was maintained for a long time by the slowly cooling giant intrusion. This also follows from the facts that the newly formed minerals (amphibole, clinopyroxene, and olivine) compose clearly distinguishable crystals and that the hornfels of this zone contain no relict minerals of the metavolcanics. Correspondingly, the mineral assemblages of the rocks can be used to quantify the conditions under which the inner-zone hornfels were formed. At the same time, the hornfels of the intermediate and outer zones may sometimes contain metastable and not completely equilibrated mineral associations.

PT PARAMETERS IN HORNFELSES

The parameters of prograde metamorphic transformations in the hornfels were calculated on the basis of chemical analyses of coexisting minerals, with these parameters calculated using only the outermost zones of mineral grains. The composition of the hornfels imposes certain limitations on the application of various mineralogical thermometers and barometers. We determined that the conditions under which hornfels in contact with the pluton were produced can be evaluated only by the olivine–clinopyroxene geothermometer of Fonarev [29], because all minerals of the rocks are very ferrous. The rock-forming minerals of the inner and intermediate zones are not as ferrous, and their *PT* parameters can be estimated by clinopyroxene–amphibole and amphibole–plagioclase geothermometers [30] and the semiquantitative geothermometer [31]. The most realistic pressure values were yielded by the latter geothermometer [31], which makes use of Al concentration in amphibole. It should be mentioned that *PT* parameters calculated based on equilibria with amphiboles were evaluated using only the composition of calcic amphiboles. The results are listed in Table 3.

Table 4. Chemical composition of rocks from the outer contact zone of the Khibina pluton with the metavolcanics of the II' mozerskaya Formation of the Imandra–Varzuga structure

Sample no.	1	2	1089003	1089007	1089008	6	1089009	8
Structure			3	4	5		7	
SiO ₂	54.52	55.60	56.34	47.31	53.31	49.08	48.11	47.30
TiO ₂	0.92	1.07	1.61	2.93	1.39	1.63	0.76	1.64
Al ₂ O ₃	21.48	20.23	11.10	12.55	11.88	11.04	10.49	12.87
Fe ₂ O ₃	2.42	2.80	0.20	1.47	0.99	4.32	0.95	3.45
FeO	1.43	1.60	16.28	20.62	15.51	12.68	10.12	10.60
MnO	0.25	0.27	0.29	0.30	0.30	0.23	0.18	0.23
MgO	0.59	0.82	2.13	3.21	1.69	6.16	13.90	7.70
CaO	1.32	1.09	8.14	7.10	8.73	10.76	12.43	9.99
Na ₂ O	9.80	9.09	3.15	3.52	4.71	2.62	0.63	2.75
K ₂ O	5.86	5.45	0.42	0.42	1.03	0.15	0.10	0.17
SO ₃	0.05	0.03	0.14	0.84	0.26	–	2.89	–
P ₂ O ₅	0.15	0.12	0.30	0.27	0.28	–	0.07	–
LOI	0.78	1.02	0.19	0.11	0.14	1.42	1.55	2.41
H ₂ O	0.25	0.27	0.19	0.11	0.14	0.26	0.49	0.55
Total	99.82	99.46	100.33	100.69	100.24	100.35	102.40	99.66
<i>f</i>	0.87	0.86	0.89	0.93	0.91	0.73	0.44	0.65

Note: (1) Khibinite from the outer part of the Khibina pluton (average of 21 analyses) [32]; (2) contact fine-grained khibinite (average of two analyses) [2, 32]; (3–5, 7) outer-contact hornfelses whose composition and position relative to the contact are listed in Table 1; (6) hornfels, 60 m from the contact [2]; (8) actinolite–plagioclase schists (metadiabases) of the II' mozerskaya Formation, 500 and 2200 m from the contact (average of two analyses) [2]. *f* is the total Fe mole fraction: $(\text{Fe}_2\text{O}_3 + \text{FeO})/(\text{Fe}_2\text{O}_3 + \text{FeO} + \text{MgO})$.

Our results indicate that the temperature values were sometimes obviously overestimated, perhaps because some mineral assemblages in the intermediate and outer hornfels zones were not in equilibrium and because of the significant inaccuracy of the results yielded by the geothermometer [31]. According to the olivine–clinopyroxene geothermometer [29], the crystallization temperature of the hornfelses in immediate contact with the Khibina pluton was approximately 700°C (Table 3). Farther away from the contact, the temperature of the hornfelses gradually decreases to 640°C, which corresponds to typical prograde zoning (Table 3). The pressure in the inner- and intermediate-zone hornfelses ranged from 1.0 to 1.5 kbar (Table 3). The mineralogy of the hornfelses and their *PT* metamorphic parameters correspond to those of the typical pyroxene-hornfels facies [25].

With regard for the simultaneous character of the origin of hornfelses in immediate contact with the Khibina pluton and the crystallization of the pluton itself, it is reasonable to assume that the temperature of the magma during its emplacement could hardly be much higher than 700°C, and the pressure was 1–1.5 kbar, i.e., the pluton crystallized at a depth of 3–5 km.

CHEMICAL VARIABILITY OF ROCKS IN THE CONTACT ZONE

The hornfelses were analyzed by the conventional analytical techniques of silicate analysis at the Kola Geological Information–Laboratory Center in Apatity.

The chemical composition of hornfelses, khibinites, and mafic metavolcanics of the II' mozerskaya Formation that were not hornfelsized are reported in Table 4, and the compositional variability of the rocks is illustrated in Fig. 8. The khibinites and metavolcanics in contact with them have principally different compositions. The khibinites are rich in Al and alkalis and poor in Mg, Fe, and Ca, whereas the metavolcanic rocks are, conversely, rich in Mg, Fe, and Ca and poor in Al and alkalis (Table 4, Fig. 8).

The inner-zone hornfelses are significantly different from the intermediate- and outer-zone hornfelses in having strongly elevated concentrations of FeO and, to a lesser degree, alkalis at lower contents of MgO, CaO, H₂O, and volatile components. At the same time, the hornfelses at a distance of 60 m from the contact have practically the same concentrations of all major components as in the metavolcanic rocks that were not hornfelsized. The only exception is sample 1089009, which

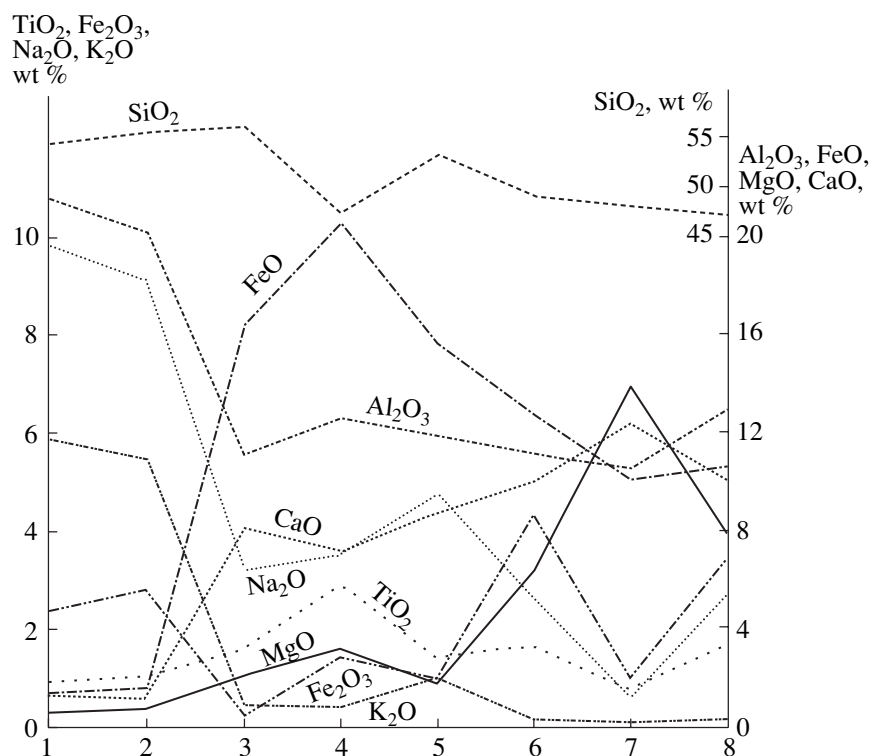


Fig. 8. Chemical variability of rocks in various zones of the contact aureole around the Khibina pluton in metavolcanic rocks of the Il'mozerskaya Formation. Sample numbers from Table 4 are plotted along the horizontal axis.

is hornfels from the outer zone and differs from the metavolcanics and other hornfels in concentrations of most components. The reasons for these differences are still obscure. Our data indicate that the thermal metamorphism of the rocks was associated with only very insignificant metasomatic alterations in the outer-zone hornfels. These alterations involved the slight introduction of K and, to an even lesser degree, Na and resulted in the crystallization of biotite and sanidine at immediate contact with the pluton and in the elevated contents of Na in the clinopyroxene and of K and Na in the amphibole and plagioclase. This conclusion about very weak metasomatism in the contact hornfels of the Khibina pluton is completely consistent with the earlier data [2].

DISCUSSION

The emplacement of such a large intrusive body as the Khibina pluton, which possesses huge reserves of heat and fluids, should have been associated with the significant heating of the wall rocks by the dissipating heat of the intrusion. Traces of this process are quite obvious in the greenschist-facies metavolcanic rocks of the Imandra-Varzuga structure in contact with the pluton, because the temperature difference between the wall rocks and magma of the pluton was about 400°C. The metavolcanics were heated with the development of a hornfels zone 150–400 m thick. The original low-

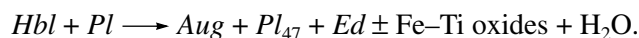
temperature mineral assemblage (actinolite + albite) of these rocks was gradually replaced by the higher temperature assemblage clinopyroxene + amphibole + andesine-bytownite ± olivine.

The thickness of the thermal aureole and the extent of the metamorphic transformations were controlled by several factors, first of all, the readiness of the wall rocks (their grain size, porosity, temperature, and fluid contents) and the characteristics of the magmatic body (its shape, depth, temperature, and water concentration in the magma). The metabasalts of the Il'mozerskaya Formation that host the Khibina pluton are fine-grained, low porous, and hence, weakly permeable [2]. They were formed at relatively low temperatures [3] and contain relatively much water (Table 4).

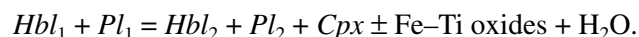
Geophysical data [33] and the results of density modeling [34] indicate that the Khibina pluton has a cylindrical shape with steep (70–80°) dips of its western and southern contacts toward the center. Our data provide indirect evidence that the Khibina pluton was emplaced at shallow depths and at relatively low temperatures. Relatively small thicknesses of contact aureoles are typical of hypabyssal intrusions, in contrast to the contact aureoles of abyssal bodies [25]. According to Kogarko [35], apaitic magmas contain little water due to its loss in the course of the long-lasting differentiation processes. Indeed, the rocks of the Khibina pluton have relatively low water contents [32, Table 4].

Operating together, all of these factors were unfavorable for the development of a thick contact hornfels zone in the host metavolcanic rocks around the Khibina pluton (in spite of its huge size), with these rocks affected only by thermal metamorphism within a fairly thin zone in immediate contact with the pluton.

The transition from the unaltered metavolcanics to the outer-zone hornfelses was associated with actinolite replacement by common hornblende. The formation of clinopyroxene and edenite in the intermediate-zone hornfelses was caused by the decomposition of the *Hbl-Pl* association by the reaction



This reaction is similar to that experimentally examined by Spear [36] at 1 kbar pressure and a variable oxygen fugacity



Note that *Hbl*₂ is richer in pargasite than *Hbl*₁, and *Pl*₂ is more calcic than *Pl*₁, which is consistent with our data on the naturally occurring rocks.

One of the important results of Spear's experiment [36] is the established temperature dependence of the appearance of the very first newly formed minerals on the oxygen fugacity. In particular, the temperature at which clinopyroxene first appears decreases from >780°C at the wuestite–magnetite buffer to 768°C at the fayalite–magnetite–quartz buffer and 720°C at the hematite–magnetite buffer. Correspondingly, as the oxygen fugacity increases, the Fe mole fractions of the newly formed mineral phases (clinopyroxene and amphibole) also increases [36].

The formation of orthopyroxene in the intermediate hornfels zone and fayalite in the inner zone can be realistically enough explained by the decomposition of the earlier assemblage at increasing temperature [36]. For example, the decomposition of amphibole in the experiments at the fayalite–magnetite–quartz buffer and a fluid pressure of 2 kbar results in the formation of olivine at 720°C, and the temperature of this transition notably decreases as the amphibole becomes more ferrous [37]. In application to our situation, olivine can be formed via the decomposition of the original amphibole at 700°C if it contains 67% of the Fe₇Si₈O₂₂(OH)₂ end member [37], which is close to the composition of this mineral in the inner-zone hornfelses. The results of the aforementioned experimental research suggest that the temperature of the prograde metamorphic reactions was largely controlled by the Fe mole fraction of the original compositions and the redox conditions.

The general rarity of orthopyroxene in the hornfelses (this mineral was found only in the intermediate zone) was previously described in [2, 14–16] and could hardly be caused only by the inadequately poorly known mineralogy of these rocks. Orthopyroxene replacement by olivine in the inner-zone hornfelses (which have relatively high silica contents compared to

those in the intermediate-zone hornfelses, Fig. 8) was controlled, first of all, by the narrowness of the orthopyroxene stability field in the fayalite-bearing system at the elevated Fe mole fractions of the hornfelses, as was also proved by studying the phase relations of orthopyroxene and olivine in iron formations [38].

The influence of metasomatic processes on the genesis of hornfelses was discussed by Gorstka [2], who arrived at the conclusion that the effect of contact metamorphism on hornfelses is very insignificant. This conclusion is in complete agreement with our data. The limited mass transfer during the contact metamorphism of metavolcanic rocks around the Khibina pluton can be well enough explained within the framework of the conductive model alone [25, 39]. Traces of metasomatism were identified only in the hornfelses adjacent to the pluton. These traces include the enrichment of the outer-contact rocks in K₂O and Na₂O, which were removed from the pluton and resulted in newly formed biotite and single sanidine grains, as well as in the albitization of the plagioclase and the enrichment of the clinopyroxene in Na and the amphibole in Na and K. The contents of other components in minerals from rocks in immediate contact with the pluton were, thereby, changed insignificantly and cannot appreciably affect, for example, the calculated *PT* parameters of the hornfelses. For example, the elevated Fe mole fractions of the rocks, as well as those of their coexisting minerals, were not caused by metasomatic processes. It is hardly probable that Fe could be taken from the Khibina pluton, whose Fe concentrations are very low (Table 4). The enrichment of the newly formed Fe–Mg silicates in the inner zone during a temperature increase in the course of the isochemical processes likely occurred at the expense of the clinopyroxene and calcic amphibole in the intermediate zone, as also follows from the notable decrease in the contents of these minerals in the inner zone compared to the intermediate zone (Table 1). The other Fe source could be titanomagnetite (78.60 wt % FeO, Table 2) that was transformed into ilmenite (45.94–46.45 wt % FeO, Table 2).

The calculated temperature distribution within the hornfels zone is generally consistent with typical prograde zoning developing when contact hornfelses are formed, which are characterized by a gradual temperature decrease away from the intrusion [25]. The temperature at which the hornfelses were formed at immediate contact with the Khibina pluton is directly dependent on the crystallization temperature of the marginal nepheline syenites in the pluton. The crystallization temperature of nepheline syenites in the Khibina pluton was estimated at 700–720°C on the basis of the direct measurement of the homogenization temperature of primary gas–liquid inclusions in the nepheline [40]. The melting of the Khibina nepheline syenites has never been modeled experimentally, and the temperatures at which the lujavrites from the Lovozero Massifs (which is close to the Khibina pluton) completely melt were estimated at 1070°C (anhydrous conditions)

and 910°C (under a water pressure of 1 kbar). The respective solidus temperatures are 750 and 600°C [41]. Similar results were also obtained on the melting of nepheline syenites from the Blue Mountain Complex in Ontario, Canada, whose solidus temperature was 770°C at a pressure of 1 kbar [42]. With regard for all data presented above, our estimates for the temperatures at which the hornfelses were formed in the immediate contact of the Khibina pluton seem to be realistic enough.

CONCLUSIONS

1. New data were obtained that made it possible to refine the inner structures of the western, southwestern, and southern outer-contact aureoles of the Khibina alkaline pluton in the metavolcanic rocks of the Il'mozerskaya Formation of the Paleoproterozoic Imandra-Varzuga riftogenic structure. According to the composition of the rocks, the aureole can be subdivided into three zones. The inner zone (up to 30 m thick) consists of clinopyroxene-plagioclase hornfelses with olivine as a typomorphic mineral and variable contents of amphibole. The intermediate zone (from 30 to 200 m from the contact with the pluton) is separated from the inner zone by the olivine isograd and consists of amphibole-clinopyroxene-plagioclase hornfelses. The outer zone (at a distance of 200–400 m from the contact with the pluton) is made up of fine-grained melanocratic hornblende hornfelses.

2. New data were obtained on the chemical composition of the rock-forming minerals of the hornfelses. The olivine is ferrohorthonolite-fayalite, and the clinopyroxene is augite-ferroaugite. Single orthopyroxene grains were found exclusively in the intermediate-zone hornfelses and corresponded to ferrohypersthene in composition. The brown amphibole in hornfelses in the intermediate hornfels zone and the part of the inner zone most distant from the contact belongs to the group of calcic amphiboles and corresponds to edenite. The hornfelses of the inner zone near the contact with the pluton contain brown amphibole of kataphorite composition, belonging to the group of sodic-calcic amphiboles. The composition of the plagioclase generally corresponds to andesine and bytownite or to albite-oligoclase near the contact. Single sanidine grains were found at immediate contact with the pluton.

3. We evaluated the *PT* metamorphic parameters of the hornfelses. These rocks were formed by thermal metamorphism at temperatures of 700–640°C and pressures of 1–1.5 kbar. The temperature calculated for the development of hornfelses at immediate contact with the Khibina pluton (approximately 700°C) was likely close to the solidus temperature of the melt.

4. The chemical composition of the contact rocks indicates that the hornfelses of the inner zone principally differ from those in the intermediate and outer zones and from the unaltered (unhornfelsized) metavolcanic rocks in having elevated contents of iron and

alkalis and lower concentrations of magnesium, calcium, water, and volatile components.

5. The relatively insignificant thickness of the contact aureole, similar to those typical of aureoles around hypabyssal intrusions that had relatively low temperatures, was predetermined, on the one hand, by the initial composition of the metabasalts (which were fine-grained rocks of low porosity) and, on the other hand, by the cylindrical shape of the Khibina pluton with nearly vertical contacts with the wall rocks and by the low water concentrations in the magma.

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