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Model *P–T* Conditions of High-Magnesia Magma Generation in the Precambrian of the Fennoscandian Shield

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Abstract—Highly magnesian volcanics (komatiites, picrites, and Mg-rich basalts) are most abundant in Precambrian complexes. Their parental magmas were generated through high to maximum degrees of melting of the mantle. Hence, it is interesting to analyze the conditions of magma generation in the upper mantle using data on high-magnesia volcanics. Our calculations were based on the data obtained during experimental data on the *P–T* conditions of magma generation as a function of its composition. The research subjects were high-magnesia volcanics from the Archean greenstone belts and Proterozoic structures of Karelia, Finland, and Kola Peninsula dated by Sm–Nd and U–Pb methods. Data on basalts from the Early Proterozoic ophiolitic complexes of the Kola Peninsula and central Finland were used for comparison. The calculations were made with 165 analyses of chill zones of lava flows and subvolcanic intrusions in 21 Early and Late Archean and Early Proterozoic structures. Their results show that the Archean stage of magma generation (3.4–2.5 Ga) occurred under the highest temperatures (1700–1850°C), and temperature decreased considerably (down to 1300°C) during the Early Proterozoic stage (2.5–1.9 Ga). The difference between the generation and eruption temperatures was higher for komatiitic than for picritic magmas, which explains their different subsequent differentiation. According to our calculations, the pressure of mantle melting decreased from 7–6 GPa at the Archean stage to 1–2 GPa at the Early Proterozoic stage. The generation of Archean komatiitic magmas occurred at depths of 220–180 km, and Early Proterozoic komatiitic magmas were derived from depths of 180–160 km. Basaltic magmas were generated at depths of 30–50 km, which is consistent with experimental data. Compared with the model presented by Richter [1], our temperature values for the Archean and Early Proterozoic upper-mantle magma generation are higher by 200–300 and 100°C, respectively.

INTRODUCTION

High-magnesia magmas (MgO > 12 wt %) are comparatively rare in nature. As products of high-degree mantle melting, they are its closest analogues in chemical and isotopic compositions, which explains why they are of interest for geochemists and petrologists. The crystallization products of high-magnesia magmas are komatiites, picrites, high-magnesia basalts, alkali picrites, and kimberlites. Komatiites were most abundant in the Archean. Picrites, alkali picrites, and high-magnesia basalts have been generated since the Early Proterozoic, and the oldest kimberlites are Paleozoic [2–7]. Experimental data [8–10] show that komatiitic magmas are formed under the maximum degree of partial melting (up to 50%), and picritic and basaltic magmas, under moderate and comparatively low degrees of partial melting, respectively. This empirical regularity in the evolution of mantle magmatism can be related to a decrease in the degree and depth of mantle melting or a change in its character. These changes could be caused by the global cooling of the Earth's interiors and, as a consequence, by a decrease in its heat flow. Therefore, geochemical and petrologic data on high-magnesia volcanics provide insights into the changes of heat flow in the Earth's interiors.

Precambrian komatiitic, picritic, and basaltic volcanics and subvolcanic intrusions, as well as Paleozoic alkali picrites and kimberlites, are widespread in the eastern part of the Fennoscandian Shield (Kola–Lapland–Karelian province) [11]. This allows one to characterize the evolution of this high-magnesia magmatism [5, 6, 12–14] and model *P–T* conditions of magma generation and eruption. We calculated previously similar models for single structures [15, 16]. This paper is based on a much larger data set, including the analyses of calc-alkaline volcanics from more than 20 structures. Analyses of alkaline volcanics (alkaline picrites and kimberlites) were not used for the calculations of the *P–T* conditions of magmatism, because these rocks cannot be directly derived from the unaltered mantle. Hence, only Precambrian structures were studied.

The choice of structures for modeling was based on the following criteria: (1) well studied facies of high-magnesia volcanism, which is necessary to detect rocks, whose composition corresponds to that of the parental magma; (2) availability of representative chemical and geochemical data; and (3) reliable

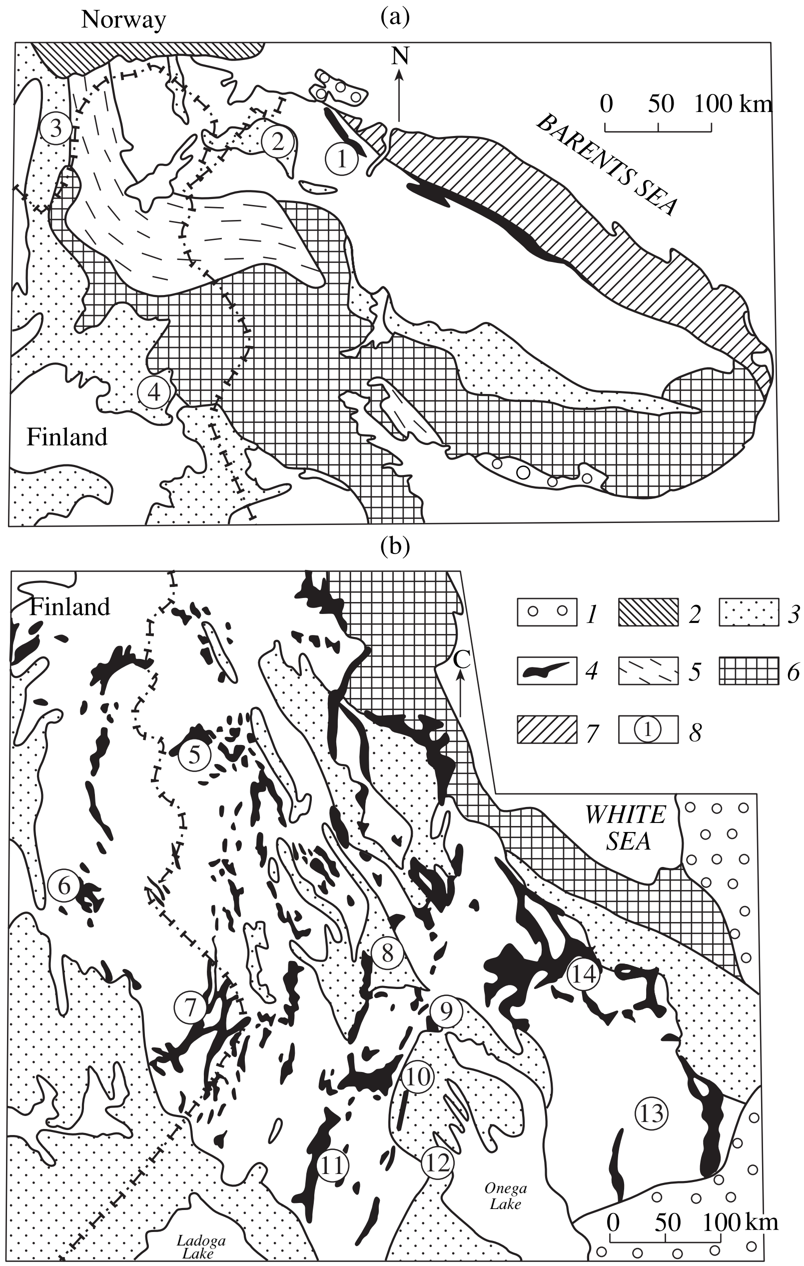

Fig. 1. Distribution of Archean greenstone belts and Early Proterozoic structures in the eastern Fennoscandian Shield including (a) the Kola Peninsula and (b) Karelian Craton. *1*—Terrigenous rocks of the Riphean platform cover; *2*—Caledonides; *3*—sedimentary–volcanogenic rocks of Karelian and Svecofennian complexes; *4*—sedimentary–volcanogenic rocks of Saamian (Voldozero massif) and Lopian (Early and Late Archean greenstone belts) complexes; *5*—Early Proterozoic Lapland–Kolvitsa granulite belt; *6*—Belomorides; *7*—Archean Murmansk terrain; *8*—occurrences of high-magnesia volcanics.

Sm–Nd and U–Pb ages of the rocks. We studied most structures in Karelia and the Kola Peninsula ourselves, and we had an opportunity to visit some structures in Finland owing to Russian–Finnish scientific collaboration.

GEOLOGY OF THE OBJECTS STUDIED

The modeling was carried out for the komatiites and picrites of the Early and Late Archean and Early Proterozoic structures of the Kola Peninsula, northern Norway, central and northern Finland, and Karelia, which were dated by Sm–Nd or U–Pb methods (Fig. 1). Data on the tholeiitic basalts of the Pechenga zone and Jormua ophiolitic complex (Finland) and the Pechenga gabbros and wehrlites hosting Cu–Ni sulfide ores were used for comparison. The objects are described in order of regions and from older to younger complexes.

Kola Peninsula

Ura-guba structure. This structure is situated in the northwestern Kola Region (Fig. 1a, 1) at the boundary of the Murmansk and Central Kola terrains. It is a relict of the Late Archean greenstone belt with a considerably reduced geologic section beginning from basal conglomerates [16]. The lower part of the structure and the belt is formed of basalts and komatiites overlain by intermediate and silicic volcanics. Massive and pillow lavas, layered lava flows, lava breccias, and tuff agglomerates occur as 3- to 20-m-thick and 100- to 500-m-long horizons and lenses in the middle reaches of the Ura, Zapadnaya Litsa, and Titovka rivers. Larger subvolcanic bodies of komatiites (peridotites) with a thickness of 20–50 m and a length of 200–1000 m also occur there. All these rocks were metamorphosed under amphibolite facies conditions. Cumulate zones with occasional relicts of primary olivine (13–15% fayalite), zones with disseminated and platy spinifex textures, and breccia zones are distinguished in the layered lava flows. Rocks from quench and brecciated zones are the closest compositional analogues of the parental magma, and their characteristics were used in the calculations. The komatiites of the Ura-guba and Kolmozero–Voron'ya structures occur on the same stratigraphic level [5]. According to geological data, the Kolmozero–Voron'ya komatiites are older than quartz porphyries dated by the U–Pb method at 2828 ± 8 Ma [17].

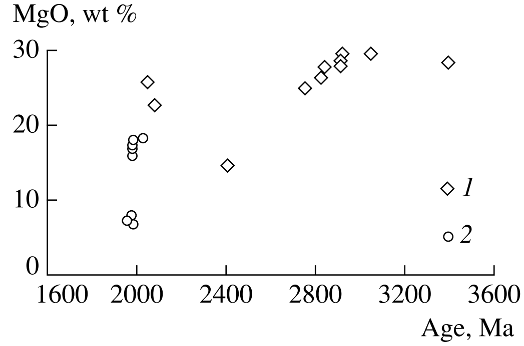

Fig. 2. Average MgO contents of komatiitic and picritic volcanics of the Fennoscandian Shield versus their age (Ma). *1*—Komatiite; *2*—picrite.

Table 1. Chemical compositions of ferropicritic volcanics of the Matert Formation (northern Pechenga zone) and model *P–T* conditions of generation of their parental magmas

Component	Lower Matert Formation						Upper Matert Formation					
	c-2728/749.2	c-2986/96.70	c-3056/513.0	1746/1	1748/1	average	c-3077/310.0	c-3077/313.0	c-3077/314.3	c-3077/322.5	c-VI/376.8	average
CR	1.12	1.08	1.08	1.06	1.07	n.a.	1.11	1.12	1.15	1.08	n.a.	n.a.
SiO_2	47.54	48.11	45.43	46.37	46.79	n.a.	49.33	49.27	47.05	45.50	46.83	n.a.
TiO_2	2.00	1.57	2.40	2.57	2.39	n.a.	2.31	2.26	2.20	2.27	2.48	n.a.
Al_2O_3	6.68	6.07	7.24	8.34	7.21	7.10	8.59	7.16	8.07	8.74	9.27	8.36
Fe_2O_3	4.11	4.05	1.44	1.69	1.94	n.a.	2.67	3.79	2.53	3.95	15.09	n.a.
FeO	11.30	9.98	13.57	14.29	12.47	n.a.	12.83	10.88	12.96	12.45	n.a.	n.a.
MnO	0.24	0.15	0.15	0.25	0.20	n.a.	0.19	0.19	0.18	0.21	0.23	n.a.
MgO	13.30	18.92	19.11	16.53	18.93	17.35	12.51	13.53	15.75	18.06	13.97	14.76
CaO	14.40	10.80	10.30	9.53	9.69	n.a.	10.95	12.59	10.68	8.31	11.62	n.a.
Na_2O	0.13	0.12	0.16	0.13	0.09	n.a.	0.13	0.09	0.16	0.16	0.16	n.a.
K_2O	0.09	0.03	0.00	0.06	0.03	n.a.	0.28	0.04	0.19	0.15	0.13	n.a.
P_2O_5	0.20	0.20	0.20	0.23	0.26	n.a.	0.22	0.20	0.22	0.22	0.22	n.a.
Sum	100.00	100.00	100.00	100.00	100.00	n.a.	100.00	100.00	100.00	100.00	100.00	n.a.
T_{liq} (°C)	1298	1399	1402	1356	1399	1370	1284	1302	1342	1383	1310	1324
T_{pot} (°C)	1422	1569	1574	1508	1569	1529	1400	1428	1487	1547	1440	1461
P_{Al}, GPa	6.4	6.9	6.0	5.3	6.0	6.0	5.1	6.0	5.4	5.0	4.7	5.3
H_{crust}, km	15	28	29	22	28	24	13	15	20	26	16	18

Note: The analyses were taken from [5]; CR is the coefficient for recalculation to a water-free basis; T_{liq} (°C) is the temperature of eruption; T_{pot} (°C) is the potential temperature; P_{Al} (GPa) is the pressure in the magma generation center; and H_{crust} is the thickness of crust above the mantle center of magma generation.

Table 2. Chemical compositions of tholeiitic basalts of the Matert Formation (northern Pechenga zone) and model P–T conditions of generation of their parental magmas

Component	Tholeiitic basalts						Pilgujarvi ore-bearing intrusion					Northern Kotselvaara ore-free intrusion			Average of 15 quench zones
	C-2986	87-17	87-18	87-21	87-26	average	CA-48	CA-49	927/1	927/2	average	CA-243	CA-245	average	
CR	1.05	1.07	1.08	1.07	1.08	–	1.08	1.06	1.09	1.07	–	1.08	1.08	–	1.01
SiO_2	49.63	48.77	47.27	49.30	48.59	–	41.86	45.24	39.27	43.36	–	44.10	44.76	–	44.57
TiO_2	1.14	1.72	1.68	1.64	1.56	–	2.71	2.11	1.58	2.65	–	2.84	1.96	–	2.06
Al_2O_3	14.86	14.47	15.33	14.72	14.33	14.74	10.30	7.52	11.93	8.15	9.47	9.53	6.49	8.01	9.12
Fe_2O_3	–	–	–	–	–	–	2.32	2.23	2.10	1.79	–	3.65	7.05	–	3.34
FeO	12.82	14.75	15.68	14.54	14.67	–	15.24	14.72	16.72	16.29	–	11.60	11.12	–	12.57
MnO	0.23	0.24	0.23	0.25	0.25	–	0.29	0.21	0.33	0.38	–	0.25	0.18	–	0.21
MgO	7.28	6.16	6.69	6.05	6.16	6.46	17.05	17.82	18.73	17.87	17.86	18.72	19.43	19.08	18.59
CaO	10.50	9.44	8.99	9.88	11.53	–	9.32	9.16	7.40	8.64	–	8.29	8.63	–	8.92
Na_2O	3.19	3.79	2.96	2.53	2.68	–	0.55	0.71	0.07	0.15	–	0.19	0.17	–	0.32
K_2O	0.24	0.49	1.01	0.93	0.10	–	0.09	0.06	0.03	0.03	–	0.54	0.02	–	0.09
P_2O_5	0.11	0.16	0.15	0.16	0.13	–	0.26	0.20	1.85	0.69	–	0.29	0.19	–	0.21
Sum	100.00	100.00	100.00	100.00	100.00	–	100.00	100.00	100.00	100.00	–	100.00	100.00	–	100.00
T_{liq} (°C)	1191	1171	1180	1169	1171	1176	1365	1379	1396	1380	1379	1395	1408	1401	1393
T_{pot} (°C)	1253	1221	1236	1217	1221	1229	1521	1541	1564	1542	1542	1564	1582	1573	1560
P_{Al} (GPa)	2.4	2.5	2.2	2.4	2.5	2.4	4.2	5.8	3.5	5.4	4.6	4.6	6.5	5.5	4.8
H_{crust}, km	4	3	3	2	3	3	24	26	28	26	26	28	30	29	28

Note: The analyses were taken from [5, 20]; samples 87-17 to 87-26 are from the rock collection of Skuf'in [20].

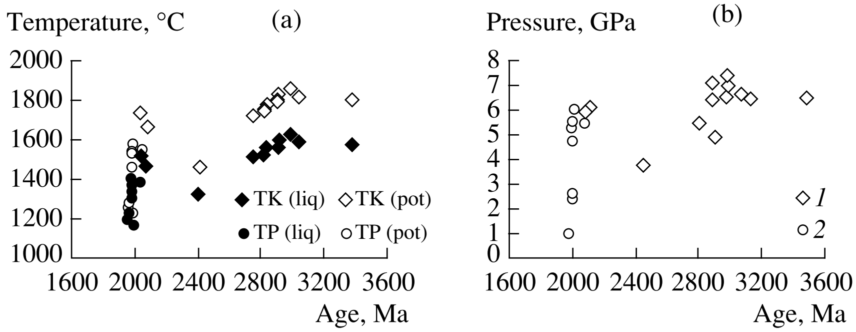

Fig. 3. *P–T* characteristics of eruption and generation of the komatiitic and picritic magmas of the Fennoscandian Shield: (a) temperature (°C) and (b) pressure (GPa) as functions of age (Ma). (a) $TK_{(liq)}$ is the temperature of komatiitic magma eruption; $TP_{(liq)}$ is the temperature of picritic magma eruption; $TK_{(pot)}$ is the potential temperature of komatiites; and $TP_{(pot)}$ is the potential temperature of picrites. (b) *1*—pressure in the mantle source of komatiitic magma; *2*—pressure of in the mantle source of picritic magma. The data are presented in Table 3.

Pechenga structure. This structure is situated in the northwestern Pechenga–Varzuga Belt (Fig. 1a, 2), which is the largest Early Proterozoic rift system of the Fennoscandian Shield [6]. Its central part contains abundant ferropicritic volcanics alternating with tholeiitic basalts of the Matert Formation [5, 18]. The ferropicrites occur as massive and pillow lavas, layered lava flows with a thickness from 3–5 to 25 m, lava breccia and tuff horizons, and dike swarms. Thick layered flows consist of a lower zone of olivine ferropicrite (olivine cumulate), a central zone of fine-grained ferro-picritobasalt (clinopyroxene cumulate), and an upper zone with olivine or, occasionally, pyroxene spinifex textures and globular structures. Rocks of the upper and lower quench zones are preserved in some of the flows. Together with massive and pillow lavas, they can be considered as compositional analogues of the parental magma. Taking into account the long period of ferropicrite generation, the two most representative horizons were characterized (Table 1). The ferropicrites and gabbro–wehrlites show high contents of FeO + Fe_2O_3 (14–16 wt %), TiO_2 (1.3–4 wt %), and P_2O_5 (0.15–0.35, occasionally, up to 0.88 wt %), which is explained by the elevated fayalite fraction of olivine (16%), the occurrence of titaniferous rock-forming (titanaugite, kaersutite, and biotite) and accessory (titaniferous chromite, chromiferous ulvöspinel, and ilmenite) minerals, and the enrichment of the rocks in light rare earth elements relative to chondrite. The ferropicrites of the Matert Formation were dated by the Sm–Nd method (mineral isochron) at 1980 ± 40 Ma with ε_{Nd} = 1.6 [5, 19].

Petrological and isotopic data indicate that the ferropicrites are comagmatic to hypabyssal gabbro–wehrlite nickel-bearing intrusions, which are also confined to the central part of the Pechenga structure and were metamorphosed under greenschist facies conditions [5, 20]. Most of the intrusions occur in the productive sequence of tuffaceous–sedimentary rocks, and they are often differentiated into the lower wehrlite–serpentinite layer, the central clinopyroxenite layer, and the upper gabbro, less usually, orthoclase gabbro layer. Lower quench zones are preserved in some intrusions as monomineral or olivine clinopyroxenite. It is interesting to compare the *P–T* formation conditions of the ferropicritic volcanics and intrusions. For such a comparative analysis, we chose the thickest (to 600 m) Pilgujarvi intrusion hosting the Zhdanov Cu–Ni sulfide deposit and the thin (<200 m) ore-free Severnaya Kotsel'vaara intrusion occurring north of the Kaula and Kotselvaara–Kammikivi deposits (Table 2). The Pilgujarvi intrusion is dated by the U–Pb method at 1982 ± 8 Ma with ε_{Nd} = 1.5 [21]. The gabbro–wehrlitic intrusions and ferropicritic volcanics spatially associate with tholeiitic basalts occurring as pillow, massive, and, occasionally, variolitic lavas and horizons of lava breccia and tuffs. The basalts have a relatively constant

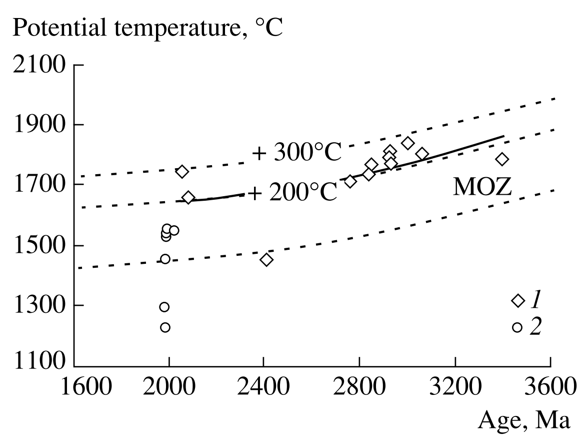

Fig. 4. Model temperature evolution of magma generation for the Early Proterozoic of the Fennoscandian Shield. *1*—Temperature of komatiitic magma generation; *2*—temperature of picritic magma generation. MOZ is the model temperature trend of Earth cooling [1], and +200°C and +300°C are the temperature trends exceeding the MOZ by 200 and 300°C, respectively. The solid line is the model temperature trend of komatiitic magma generation calculated for the Fennoscandian Shield.

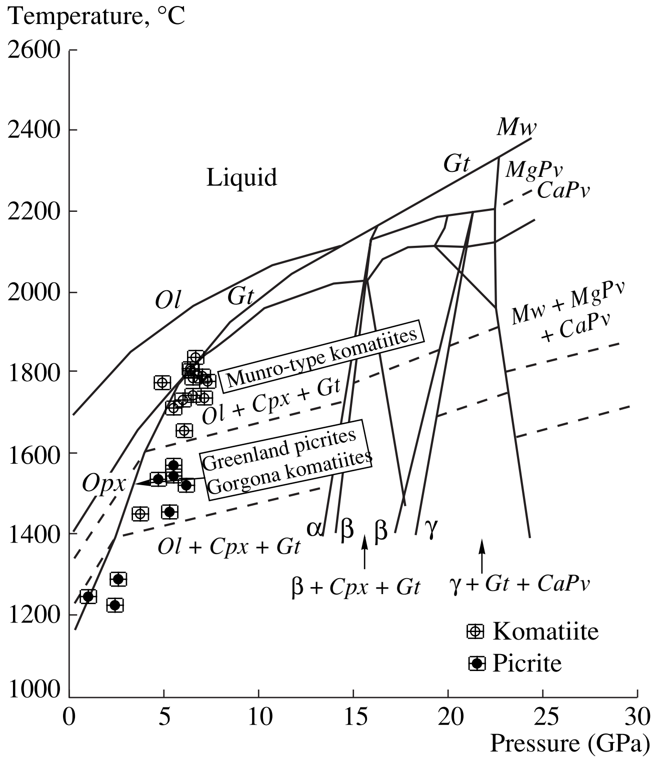

Fig. 5. Model *P–T* diagram for komatiitic and picritic magmas with phase fields of the KLB-1 mantle peridotite [45, 46]. Temperature error is 50°C and pressure error is 1 GPa. *Ol*—olivine; *Opx*—orthopyroxene; *Cpx*—clinopyroxene; *Gt*—garnet; *Mw*—magnesiowüstite; *MgPv*—Mg-perovskite; *CaPv*—Ca-perovskite; α—olivine; β—modified spinel $(Mg,Fe)_2SiO_4$; γ—spinel $(Mg,Fe)_2SiO_4$.

chemical composition (Table 2). Unlike ferropicrites, they are olivine free and show chondritic REE distribution pattern. Their Pb, Nd, and Sr isotope compositions differ from those of the ferropicrites.

Northern Norway

Karasjok belt. The position of the belt is shown in Fig. 1a (3). In its northwestern zone, the central part of the section contains a sequence of chloritized and amphibolized, massive, pillow, and schistose ultrabasic lavas alternating with quartzites and metapelites [22, 23]. Neither spinifex texture nor relicts of primary minerals including chrome spinel are preserved in the volcanics. Nevertheless, their petrochemical characteristics correspond to those of peridotitic and pyroxenitic komatiites enriched in Cr, Ni, and Co and depleted in Al, La, and Ce. Low-Ti and moderately Ti enriched varieties are distinguished among the volcanics. The rocks are dated by the Sm–Nd method at 2085 ± 85 and 2103 ± 87 Ma with $\varepsilon_{Nd} = 4.1$. They present rare examples of Proterozoic komatiites.

Finland

Tipasjarvi structure. This structure occurs in the southern Kuhmo–Suomussalmi (Kuhmo) belt (Fig. 1b, 6), which is the best studied Archean greenstone belt of Finland. Its high-magnesia volcanics of the Kallio Formation are underlain by metabasalts or Fe-rich tholeiites and overlain by metasediments [24, 25]. They are peridotitic and pyroxenitic komatiites occurring as massive lavas, thin (<10 m) layered flows, and brecciated rocks. The lavas and flows show occasional relict zones with olivine cumulates and spinifex textures. The MgO concentration in the breccia zone is 25–27 wt %.

Hattu structure, northern Ilomantsi belt. This structure occurs in southeastern Finland, at the Finnish–Karelian boundary (Fig. 1a, 7). Its high-magnesia volcanics are confined to the lower (containing tholeiitic basalts) and central (containing basic and acid volcanics and gabbro) parts of the general geologic column [26]. They form lenslike bodies with tectonic boundaries, and some of them have relict polygonal jointing. The volcanics contain from 37 to 13 wt % MgO. Their compositions correspond to peridotitic, pyroxenitic, and basaltic komatiites, the latter being prevalent. Zircon from graywacke interbeds in the high-magnesia volcanics is dated by the U–Pb method at 2761 ± 11 Ma [26].

Savukoski structure. This Early Proterozoic structure occurs at the boundary of northern Karelia and central Finland, near Kuolajarvi, Salla, and Savukoski settlements (Fig. 1a, 4). High-magnesia volcanics alternate with silicic and intermediate volcanics and associate with subvolcanic gabbro–peridotitic intrusions at several stratigraphic horizons [13, 27, 28].

The volcanics occur as thin lava flows, horizons of coarse- to fine-fragmental pyroclastics, and rare differentiated lava flows with occasional spinifex textures. They are compositional analogues of picrites and picritobasalts (less usually, komatiites) moderately enriched in the light rare earths and containing from 14 to 24 (up to 29 in cumulate zones) wt % MgO [5].

The northwestern continuation of this structure is the Early Proterozoic Kittila Belt [27]. The upper part of its southeastern segment near the Sattasvaara settlement has a thickness of up to 2 km, contains high-magnesia volcanics alternating with carbon-bearing schists, and is overlain by carbonates. The volcanics occur as prevalent massive and subordinate pillow (occasionally, with a brecciated or scoriaceous roof) lavas and pyroclastic horizons. Volcanic necks and cones are less usual. The MgO content of the volcanics varies from 9 to 30 wt %, which allows one to classify them into basalts, picrobasalts, and picrites (according to some researchers, komatiitic basalts and komatiites).

The age of the high-magnesia rock association is estimated as more than 2050 Ma from a comparative facial analysis of the volcanics and isotopically dated marker horizons [29].

Jormua structure. This complex occurs in central Finland (Fig. 1b, 15) within the Kainu Schist Belt. It separates the Iisalmi–Pidasjarvi and Kuhmo–Ilomantsi Archean terrains. Volcanic rocks are weakly metamorphosed basaltic pillow and massive lavas with 0.5- to 2-m-thick interbeds of hyaloclastites and pillow breccias. The volcanics are crossed by several gabbroic dike systems, including a dike-in-dike series. The rocks of this complex have from 4 to 12 wt % MgO. When compared with intrusive rocks, the lavas are distinguished by elevated concentrations of Cr and Ni. The trace element distribution patterns of the basalts are unfractionated. The intrusive complex is dated by the U–Pb method at 1953 ± 2 Ma [30].

Karelia

Vodlozero structure. This structure, which is often referred to as the Vodlozero terrain, is one of a few Early Archean structures of the Fennoscandian Shield (Fig. 1b, 13). It is formed of tonalite gneisses, amphibole and garnet–amphibole schists, amphibolites, migmatites, and diorites. Relict volcanics occur in the gneiss fields dated by the U–Pb method at 3540–3500 Ma. They are represented by serpentine–cummingtonite–chlorite rocks petrochemically corresponding to komatiites and komatiitic basalts [31]. According to paleogeologic reconstructions, the volcanics form about 50 lava flows with a thickness from 1.5 to 15 m and relict massive, pillow, brecciated, and amygdaloidal structures. They compose four members, in which they alternate with amphibolites or tholeiitic metabasalts. The MgO concentration of the metakomatiites varies from 24.4 to 31.5 wt %, and their Al_2O_3 concentration, from 5.0 to 6.9 wt %. The REE distribution patterns of these rocks are slightly fractionated. Their Sm–Nd age is 3391 ± 76 Ma with $\varepsilon_{Nd} = 1.2$ [32].

Kamennoe Ozero structure. This structure is a composite relict of the Sumozero–Kenozero greenstone belt located southwest of the Vetrennyi Range (Fig. 1b, 14) [13]. Its komatiites, basalts, tuffs, tuffites, schists, and quartzites form the Kumbuka sequence up to 1200 m thick. The volcanic sequence is underlain by metasedimentary rocks, tuffs, and basalts and is subsequently overlain by silicic, basic, and intermediate volcanics alternating with tuffs, tuffites, schists, and quartzites. The komatiites are spatially associated with serpentinite intrusions. About 15 lava flows of komatiites and komatiitic basalts were distinguished in the Zolotye Porogi area. They have a thickness of 10–20 m, a distinct spheroidal jointing, and an occasionally preserved spinifex texture. The komatiites alternate with chlorite schists, tuffs, quartzites, and massive and pillow basalts. Some thick differentiated flows have cumulate, spinifex-textured, and brecciated roof zones. The MgO concentration of the rocks with spinifex textures varies from 23.2 to 28.7 wt %, and their Al_2O_3 concentration, from 4.8 to 6.3 wt %. The komatiites have a Sm–Nd age of 3054 ± 84 Ma, with $\varepsilon_{Nd} = 2.4$ [33].

Sovdozero, Palaselga, Koikary, and Hautavaara structures. These structures occur in the northern and central parts of the Vedlozero–Segozero Archean greenstone belt, which extends approximately N–S from Vedlozero Lake to Segozero Lake over a distance of 300 km and has a thickness of 50–60 km [13, 15, 34]. They contain both komatiites and basic, intermediate, and acid volcanites forming central volcanic edifices.

Sovdozero structure (Fig. 1b, 8) contains a 900-m-thick komatiitic member separating two horizons of tholeiitic basalts and overlain by a sequence of silicic volcanics and sedimentary rocks with interbeds of banded iron formation. The komatiites form 15- to 60-m-thick massive and brecciated lava flows with occasional 0.3- to 1.5-m-thick interbeds of psammitic and agglomerate tuffs. Olivine cumulate, spinifex textured, and brecciated roof zones were observed in two differentiated lava flows. The MgO concentration of these rocks varies from 20 to 37 wt %. Comagmatic intrusive rocks associating with the komatiites are metaolivinites.

Palaselga structure (Fig. 1b, 9). High-magnesia volcanics are confined to the central part of the stratigraphic column, where they form a 700-m-thick sequence among tholeiites. The total thickness of the volcanics is 1.8 km. The komatiites compose mainly massive, pillow, and differentiated lavas, and occasional thin interbeds of aleurolite-sized to agglomerate-tuffs. Their MgO concentration varies from 15 to 30 wt %. Their comagmatic rocks are serpentinites and high-magnesia gabbros.

Koikary (Fig. 1b, 10). Komatiites form the lower 700-m-thick sequence of the unit stratotype (Pitkilampina Formation). The komatiitic sequence is overlain in sequence by tholeiitic basalts, silicic volcanics, and sedimentary rocks. The komatiites are mainly 0.5- to 25-m-thick flows of massive, pillow, and differentiated lavas. A characteristic of the Koikary structure is the wide occurrence of variolites, which usually form individual 5- to 20-m thick lava flows. The MgO concentration of the lavas varies from 9 to 30 wt %, which allows one to distinguish peridotitic, pyroxenitic, and basaltic komatiites. The rocks with a spinifex texture and breccia zones contain 23–28 and 27–29 wt % MgO, respectively, and their Al_2O_3 concentration is low (4–7 wt %). The komatiites are cut by dacite dikes dated at 2935 ± 15 Ma [35]. The Koikary and Palaselga komatiites are dated by the Sm–Nd method at 2921 ± 55 Ma with $\varepsilon_{Nd} = 1.5$ [36].

Hautavaara (Fig. 1b, 11). The komatiite-bearing rock association composes the 600- to 2200-m-thick

Loukhivaara Formation (analogue of the Pitkilampina Formation in the Koikary structure). It lies on andesitic and dacitic lavas and tuffs and metasediments and is overlain by terrigenous and chemogenic rocks and silicic volcanics with iron sulfide mineralization. The komatiites are present as massive lava flows with polygonal jointing and rare spinifex textures. Their MgO concentration varies from 18 to 34 wt %, and their TiO_2 concentration is 0.2–0.7 wt %. The komatiites are crossed by subvolcanic dacite bodies dated by the U–Pb method at 2995 ± 20 Ma [37].

The *P–T* conditions of komatiitic magma generation in the Vedlozero–Segozero greenstone belt were modeled using the analyses of stratigraphically similar volcanics of the Pitkilampina and Loukhivaara formations.

Kostomuksha structure. The Archean Gimoly–Kostomuksha greenstone belt occurs in northwest Karelia (Fig. 1b, 5) and is the Lopian stratotype. Its lower part is formed by ultrabasic, basic, and acid volcanics of the Kontok Group, and its upper part consists of quartzites, banded iron formation, and micaceous schists of the Late Archean Gimoly Group. Komatiites were found at two horizons of the Kontok Group. They associate with basalts and occur as chlorite–amphibole and amphibole schists free of any relicts of primary minerals [13]. Autobreccias are prevalent among the volcanics, while massive, pillow, and brecciated lavas, lava-breccias, thin-layered and brecciated tuffs, and subvolcanic bodies are subordinate. The MgO content of the Kostomuksha komatiites varies widely (from 13 to 36 wt %), and Al_2O_3 often exceeds 9 wt %. Similar MgO concentrations (23–26 wt %) were analyzed in komatiites with microspinifex texture, tuffs, and rocks of subvolcanic bodies. The komatiites show a Sm–Nd age of 2843 ± 39 Ma with ε_{Nd} = 2.9 and 3.4. They are overlain by rhyolites dated by the U–Pb method at 2795 ± 29 Ma [38].

Vetrenyi belt (Fig. 1b, 14). This structure is confined to the Vetrenyi Range and traced along the boundary of the Belomorian block for more than 250 km. Like the Pechenga structure, it was formed during the whole Early Proterozoic. Its high-magnesia volcanics occur at several stratigraphic levels, and their maximum accumulation took place in the Sumian [13]. The Vetrenyi Belt Formation contains abundant high-magnesia volcanics of lava (prevalent), diatreme, and hypabyssal facies associated with massive, pillow, and amygdaloidal basaltic lavas, tuffs, and tuff breccias. Gently sloping lava flows were reported in the area of Myandukha, Bol'shaya Levgora, Shapochka, and Golets mountains. The flows are 14- to 30-m thick (occasionally up to 50-m thick) and consist of picrites and picrobasalts (komatiites and komatiitic basalts by V.S. Kulikov). Scoria- and bomb-rich horizons also occur in this area. The lower parts of lava flows are often enriched in olivine phenocrysts or their skeleton relicts, and their upper parts bear olivine or pyroxene microspinifex textures and variolites. MgO content varies from 27.5 to 23 wt % in the olivine-rich rock varieties, and is no higher than 10 and 8 wt % in the olivine-containing and olivine-free rocks, respectively. The average MgO concentration of lava flows of Myandukha Mountain is 11–16 wt %. The high-magnesia volcanics are dated by the Sm–Nd method at 2448–2410 Ma with $\varepsilon_{Nd} = -1.7$ and –0.9, which indicates their genetic relations to the enriched mantle [39].

Konchozero sill (Fig. 1b, 12). The Konchozero sill is situated in the central part of the Onega Trough filled with the rocks of the Suisar Formation. The latter comprises picrites, picrobasalts, basalts, trachybasalts, their tuffs, and terrigenous sediments and is a stratotype of the Lyudikovii horizon of the Karelian Complex [13, 40]. Picrites and picrobasalts usually occur as thin (0.5 to 9 and occasionally up to 24 m thick) lava flows with a massive, pillow, or amygdaloidal structure, and less common lava breccia and tuff horizons. Picrite–doleritic sills are associated with the volcanics, and the Kochozero sill is the thickest (120 m thick) among them. It is composed of (bottom to top) picrites, peridotites, and gabbro-dolerites. The rocks bear relicts of monoclinic pyroxene and contain from 8.4 to 29.3 wt % MgO. They show elevated TiO_2 concentrations (0.8–1.95 wt %) and are relatively enriched in the light rare earth elements. Petrochemical and isotopic data show that the hypabyssal and volcanic facies are comagmatic. The Konchozero sill is dated by the Sm–Nd isochrone method at 1975 ± 24 Ma with ε_{Nd} = 2.8 [41].

CALCULATION METHODS

The calculation of *P–T* formation conditions was based mainly on the chemical analyses of high-magnesian volcanics, because these rocks had been metamorphosed under widely varying temperature and pressure (from the pumpellyite–prehnite subfacies of the greenschist facies to the amphibolite facies of moderate pressures), and no relicts of primary mineral assemblages had been preserved in the volcanics, except for the Pechenga structure. Primary lava compositions were calculated using the published and our own analyses of rocks sampled in the quench zones of pillow and massive lavas, layered flows and intrusions, as well as in breccia zones (for komatiites) recalculated to a water-free basis. Data on cumulate varieties of volcanic and intrusive rocks were ignored. The composition of quench zones with relicts of primary minerals was compared with the $Mg/(Fe^{2+} + Mg)$ ratio of equilibrium melt calculated from the analyses of early olivine. The olivine composition was studied by electron probe microanalysis. The Mg/Fe^{2+} distribution coefficient between olivine and melt (K_D^{Ol-Liq}) was taken to be 0.31. The calculations were based on the equation $K_D = (Mg/Fe^{2+})_{Liq}/(Mg/Fe^{2+})_{Ol}$ assuming that all the iron of primary melts was Fe^{2+} [42, 43]. The magnesium content of quench zones was always identical to the calculated value.

The temperature of primary melt eruption (T_{liq}, °C) was calculated by the equation approximating the MgO concentration of experimental melts as a function of temperature: T_{liq} = 17.86*(MgO wt %) + 1061°C [43]. Then, T_{liq}°C was converted into the potential temperature of an adiabatically ascending mantle plume (T_{pot}, °C). This temperature can be lower than the real mantle temperature. The potential temperature was calculated by the equation $T_{pot} = 1382.5 + 2.8046T_{liq} - 0.00049671(T_{liq})^2$ [44]. The experimental equations presented above were tested using the analyses of natural basalts, komatiites, and peridotites (from 8 to 32 wt % MgO) in a temperature interval from 1100 to 1900°C, and the data obtained were consistent with model values [4, 16, 43, 44].

Using the value of potential temperature we can calculate the total crust thickness (H_{crust}) above the deep area of magma generation by the equation $T_{pot} = 1177.3 + 38.08(\ln H_{crust}) + 3.4131(\ln H_{crust})^2 + 6.0121(\ln H_{crust})^3$ [44]. The calculated value can differ from the real thickness, because the calculations ignore the tectonic mobility of the Earth's crust. Hence, these data are approximate and should be confirmed by other methods.

Table 3. P–T characteristics of the Precambrian highly magnesian parental magmas of the Fennoscandian Shield

Structure	Age (Ga), reference	Number of samples	Rock name	Average MgO content	Average Al_2O_3 content	T_{liq} (°C)	T_{pot} (°C)	P_{Al}, GPa	H_{crust}, km
Vodlozero	3391 ± 76 [32]	8*	PK	28.29	6.66	1566	1792	6.4	58
Kamennoe Ozero	3054 ± 84 [33]	2	PK	29.50	6.72	1587	1818	6.4	62
Hautavaara	2995 ± 20 [37]	6	PK	30.74	6.42	1615	1845	6.6	67
Sovdozero	2921 ± 55 [36]	5	PK	29.44	6.63	1589	1817	6.4	62
Palaselga	2921 ± 55 [36]	10	PK	28.39	6.01	1568	1794	6.9	59
Koikary	2921 ± 55 [36]	10	PK	27.82	5.62	1557	1781	7.3	56
Kostomuksha	2843 ± 39 [38]	2	PK	27.60	9.04	1553	1776	4.9	50
Tipasjarvi	>2830 [24]	3	PK	26.30	6.70	1530	1747	6.4	51
Ura-guba	>2830 [17]	7	PK	26.05	5.92	1526	1741	7.0	50
Hattu	2761 ± 11 [26]	6	PK	25.03	8.16	1508	1717	5.4	47
Vetrenyi belt	2410 ± 34 [39]	3	PyrK	14.55	11.37	1321	1455	3.7	17
Karasjok	2103 ± 87 [23]	20*	PK	22.62	7.21	1464	1660	6.0	39
Savukoski (Sattasvaara)	>2050 [29]	49*	PK	25.74	7.42	1520	1734	5.9	49
Savukoski (Sotkaselka)	>2050 [29]	8*	P	18.24	8.09	1386	1552	5.4	27
Pechenga (level III)	1980 ± 40 [18, 5, 20]	5	FP	17.35	7.10	1370	1529	6.1	24
Pechenga (level V)	<1980 [20]	5	FP	14.76	8.36	1324	1461	5.3	18
Pechenga (Matert Formation)	<1980 [20]	5	TB	6.46	14.74	1176	1229	2.4	3
Pechenga (Pilgujarvi)	1982 ± 8 [21]	4	OBI	17.89	9.47	1379	1542	4.6	26
Pechenga (Northern Kotselvaara)	>1900 [20]	2	OFI	19.08	8.01	1401	1573	5.5	29
Konchozero	1975 ± 24 [41]	2	P	8.69	14.05	1219	1294	2.7	6
Jormua	1953 ± 2 [30]	3	OFI	7.24	17.42	1190	1252	1.1	4

Note: PK—peridotitic komatiite; PyrK—pyroxenite komatiite; BK—basaltic komatiite; FP—ferropicrite; P—picrite; TB—tholeiitic basalt; OBI—ore-bearing intrusion; and OFI—ore-free intrusion.

*Calculated from the average composition.

Experimental data show that CaO/Al_2O_3 of komatiites is a function of pressure in the magma generation area, which is related to its phase composition [45, 46]. In the case of the *Ol* + *Opx* + *Cpx* ± *Gr* assemblage (with subordinate garnet) on the liquidus, the pressure dependence of the CaO and Al_2O_3 contents of melt is approximated by the following equations: CaO (wt %) = $16.0811 - 2.0724P + 0.1322P^2 - 0.0018P^3$, Al_2O_3 (wt %) = $22.8581 - 4.0110P + 0.2703P^2 - 0.0061P^3$, where P is pressure, GPa [46]. The equation was calibrated using experiments on the melting of mantle peridotite sample KLB-1 under pressures from 2–5 to 22.5 GPa and temperatures from 1400 to 2300°C. The best similarity of calculated and experimental data was established in the pressure interval from 2.5 to 5.0 GPa. Little experimental data were obtained for $P > 5$ GPa, which decreases the accuracy of approximation. According to [45–48], the equations used are correct for picrites, basalts, and komatiites.

Because of the high mobility of CaO during metamorphic processes including widespread late carbonatization, the pressure values calculated from CaO content will be overestimated. More correct values can be calculated from Al_2O_3 concentrations.

When modeling *PT* conditions, we considered first of all analytical errors for major components. Rock samples were analyzed by X-ray fluorescent and atomic absorption spectroscopy in the laboratories of geological institutes of the Karelian and Kola scientific centers, Russian Academy of Sciences (VRA-33 and Perkin Elmer 403), and in the laboratories of the Geological Survey of Finland, Espoo (Philips PW1480). The instrumental analytical error was no higher than 2% for the major components with concentrations above 0.5 wt %. Hence, the analytical error of MgO determination did not exceed 0.7 wt % for its maximum concentration of 31 wt %, and the analytical error of Al_2O_3 determination did not exceed 0.4 wt % for its maximum concentration of 17 wt %. Therefore, the calculation error was ±13°C for liquidus temperatures, ±18°C for potential temperatures, ±0.5 GPa for pressure, and ±3 km for the total crust thickness. The calculated parameters are presented in Table 3. However, the variations of component concentrations in the rocks are considerably higher than the analytical error, and, in order to give more reliable characteristics of *P–T* conditions, we will discuss the temperature and pressure values using 50°C and 1 GPa intervals.

RESULTS AND DISCUSSION

The data presented show that the high-magnesia volcanics were formed during a considerable period of the Earth's early evolution, which lasted from 3.39 to 1.95 Ga, i.e., for 1.43 billion years (Table 3). Based on the Sm–Nd systematics and geochemical classification of the komatiites and picrites of the eastern Fennoscandian Shield, the following time intervals of high-magnesia volcanism were earlier distinguished: 3.4–3.1, 3.1–2.9, 2.9–2.8, 2.8–2.7, and 2.5–1.9 Ga [14, 49]. The komatiitic volcanics analyzed are Al-undepleted with average $CaO/Al_2O_3 < 1$, $Al_2O_3/TiO_2 = 14–23$, and a weak fractionation of the heavy REE, which indicates their generation under relatively similar conditions.

The analysis of the available data (Fig. 2) shows that the MgO content of primary magmas noticeably decreased (from 30.74 to 25.03 wt %) during the Archean (3.4–2.7 Ga), and this occurred in parallel with a decrease in the potential temperature of parental magma from 1850 to 1700°C. The variations of potential temperature (Δ) of comatiitic magmas in different structures do not exceed 150°C and are no higher than 70°C for a single period: 65°C (1845–1780°C) for the period from 3.4 to 2.8 Ga and 59°C (1776–1717°C) for the period from 2.8 to 2.7 Ga (Table 3; Fig. 3).

For the Archean, the highest MgO concentrations (28–31 wt %) and potential temperatures (1800–1850°C) were determined in the older komatiites (3.1–2.9 Ga) of central and eastern Karelia (Kamennoe Lake, Hautavaara, and Sovdozero), while lower MgO concentrations (25–26 wt %) and potential temperatures (1700–1750°C) are characteristic of the younger komatiites (2.8–2.7 Ga) of the Kola Peninsula (Ura Guba) and central Finland (Hatu).

For the Proterozoic (2.5–1.9 Ga), high MgO concentrations (25–26 wt %) and potential temperature (1750°C) were determined in the komatiites of the Savukoski Structure (Kittila belt), and lower MgO concentrations (8–9 wt %) and potential temperature (1300°C) were obtained for the picrites of the Konchozero sill in the Onega Trough with the temperature variations Δ = 440°C (Table 3; Fig. 3). The latter can be due to more contrasting melting temperatures in the Early Proterozoic. The potential temperature of the ferropicritic magmas of the Pechenga structure varied within a relatively wide interval (from 1450 to 1600°C). Lower potential temperatures (1250°C) were determined for the associated tholeiitic basalts, which is consistent with many experimental results. Similar potential temperatures were estimated for the basalts of the Jormua ophiolitic complex. However, they were formed under a much lower pressure (1 GPa) than the Pechenga basalts (2–3 GPa).

All these data are presented in the *T* (°C)–age diagram (Fig. 4), which shows that the model temperature of the upper mantle decreased gradually during the Precambrian. Compared with other model data [1], our values are 300–200°C higher for the Archean and 100°C higher for the Proterozoic. Nevertheless, the model temperature decrease is consistent with the model of cooling Earth and decreasing heat flow.

Our calculations also show that the pressure of mantle source melting decreased from older to younger complexes; i.e., this parameter is a function of time. In the case of komatiites and ferropicrites, pressure varied from 6 to 7 GPa during the period of 3.4–2.8 Ga, from 5 to 7 GPa during the period of 2.8–2.7 Ga, and from 5 to 6 GPa during the period of 2.5–1.9 Ga, whereas in the case of tholeiitic basalts, it varied from 1 to 2 GPa (Table 3; Fig. 2). The minimum pressure of high-magnesia magma generation (4 GPa) was determined for volcanics of the Vetrenyi belt and Onega Trough (Konchozero).

The pressure estimates show that komatiitic magmas were generated in the eastern Fennoscandian Shield mainly at depths of 230–190 km during the Early Archean (Saamian 3.4–2.9 Ga) and 210–150 km during the Late Archean (Lopian 2.8–2.7 Ga). The Early Proterozoic komatiitic and ferropicritic magmas were generated at shallower depths (180–160 km). However, the *P–T* conditions of generation of the Proterozoic komatiitic magmas of the Karasjok belt are comparable to those of Archean komatiites. The parental basaltic magmas of the Jormua ophiolitic complex were formed at a depth of 30 km, which is consistent with the data on younger ophiolites of Phanerozoic structures [3].

The generation characteristics of Archean (Munro, Canada) and Cenozoic (Gorgona, Columbia) komatiites and picrites (Greenland) are plotted in the *P–T* diagram (Fig. 5) with phase fields for the KLB-1 peridotite [45–48]. The data points of Fennoscandian komatiites fall at the boundary of garnet and olivine–clinopyroxene fields, and the data points of picrites and ferropicrites, at the boundary of orthopyroxene and olivine–clinopyroxene–garnet fields at lower temperatures and pressures. Therefore, the mantle mineral assemblage changed during the Archean-to-Proterozoic evolution, which is primarily related to a decrease in magma generation depth.

The data presented allow us to develop an evolution model for the Archean komatiitic magmatism in Karelia and central Finland. They show that (1) older komatiites are dominated in Karelia, while younger komatiites are more common in Finland; (2) the komatiitic magmas of eastern and central Karelia were generated under similar *P– T* conditions with the maximum pressure of 7 GPa in the Koikar structure, while the komatiitic magmas of northern Karelia (Kostomuksha) and Finland (Hattu–Ilomantsi) were formed under lower pressure (5 GPa). The oldest komatiites are confined to the central part of the Archean Karelian craton. Their parental magmas were generated at greater depths than those of the younger komatiites of the craton periphery, which occur at the boundary of the craton with Svecofennian or Belomorian complexes.

According to our calculations, the thickness of Archean crust varied from 50–70 km (3.4–2.8 Ga) to 45–50 km (2.8–2.7 Ga), and the thickness of Early Proterozoic crust (2.5–1.9 Ga) studied mainly in rift systems decreased to 50–20 km.

The *P–T* characteristics of the volcanoplutonic ferropicrite–gabbro–wehrlite association show that the potential temperatures of nickel-bearing intrusions (1550–1600°C) were higher than those of ferropicritic volcanics (1450–1500°C), and the pressure of magma generation decreased in the following order: early ferropicrites (6 GPa), Northern Kotselvaara intrusion (5–6 GPa), late ferropicrites (5 GPa), and Pilgujarvi intrusion (4–5 GPa). These data show that the depth of parental magma generation probably decreased during an undetermined magmatic stage when the Lyudikovian volcanics (Matert Formation) with a total thickness of up to 4 km were deposited.

CONCLUSIONS

The *P–T* conditions of magma generation in the upper mantle were calculated from the analyses of Precambrian high-magnesia volcanics of the eastern Fennoscandian Shield. They indicate a distinct decrease in temperature and pressure in the Earth's interiors during the Precambrian, which is consistent with the model of cooling Earth and decreasing heat flow during this time period. However, our estimates of mantle temperature are higher by 200–300°C for the Archean and by 100°C for the Proterozoic than the values presented by Richter [1]. Hence, our data suggest a higher heat flow in the Precambrian.

Our data confirms that komatiitic magmas prevailed during the Archean and were subordinate during the Proterozoic, when picritic and ferropicritic magmas were dominant. This change of magma composition was related to a decrease in the depth of magma generation and a corresponding change of the phase composition of the mantle source. Evidence for the heterogeneity of magma generation was revealed in Proterozoic complexes and considerably increased in the Phanerozoic, when first kimberlites, alkaline picrites, and other high-magnesia mantle rocks were formed. Extremely high temperatures occurred only in local areas (e.g., Gorgona Island, Columbia), where young komatiites are known.

In the eastern Fennoscandian Shield, the deepest komatiitic magmas were generated in the oldest central part of the Karelian craton during the Archean, whereas magma generation at the craton periphery occurred at shallower depths. The deepest Proterozoic magmas were generated in large rift belts, where extensive rifting took place without the formation of oceanic crust (Pechenga and Karasjork structures).

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