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GEOLOGY

The Ludicovian–Kalevian Boundary in the Northern Ladoga Region: Geological Relations and Isotopic Age

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Supracrustal rocks in the northern Ladoga region are mainly represented by rock associations of the Karelian Complex (2.1–1.65 Ga). The section is divided into the Ludicovian Suprahorizon (metavolcanites of the Sortavala Group) and the Kalevian Suprahorizon (metaturbidites of the Ladoga Group) with the boundary at 1920 ± 50 Ma [1]. The regional stratigraphy and the volume of the Karelian Complex in the northern Ladoga region are debatable issues. Relationships between the Kalevian and Ludicovian supracrustal complexes are ambiguously interpreted, but in general it is postulated that the higher stratigraphic position of the Ladoga Group "stems from the general structure of the region" [2]. The upper age limit for Kalevian rocks is determined by the maximal age of intrusions estimated at 1892 ± 5 Ma (based on zircon from monzodiorites of the Velimyak Pluton [3]). The lower age boundary is based on the minimal age of clastogenic zircons estimated at 1907 ± 15 Ma (metaturbidites of the Tampere Belt [4]).

This work presents new data on the geological structure of the upper part of the Sortavala volcanic section. The data were obtained in the course of detailed geological mapping of the northeastern margin of the Kirjavolakhta dome, the petrological–geochemical analysis of reference objects, and the geochronological study of magmatic complexes. The U–Pb zircon dating was carried out with SHRIMP-II ion microprobe at the Center of Isotopic Investigations of the Karpinskii All-Russia Research Institute of Geology (St. Petersburg) according to technique described in [5]. The content of trace elements in zircons was studied according to methods described in [6] with Cameca IMS-4F microprobe at the Institute of Microelectronics and Informatics, Russian Academy of Sciences (Yaroslavl).

Volcanites of the Sortavala Group in the northeastern part of the Kirjavolakhta dome are subdivided into three sequences, which correspond to successive episodes of volcanic activation [7]. The upper sequence is made up of basaltic lava with komatiite and komatiitic basalt sheets. Lavas are successively replaced upsection by basic tuffs, komatiitic tuffs, and carbonaceous siltstone [7].

The base of the Ladoga Group section is represented by a horizon of pink quartzite (Kontiosari Formation), which is overlain by arkosic sandstones with gritstone lenses and calcareous concretions. The sandstones gradually pass upsection into rhythmically bedded rocks and alternate with aleuropelite interlayers. These rocks are overlain by andalusite-bearing biotite schists.

The contact of Ludicovian and Kalevian supracrustal rocks is mapped 7 km northwest of Kharlu Settlement in the Tervaoya River valley (Fig. 1). All rocks are characterized by northwestern strike and nearly vertical dip. Schistosity in the rocks generally coincides with bedding. Investigation of individual komatiite horizons and boundaries between members of basalts with different textures along the strike of rocks of the Sortavala Group made it possible to decipher early (sublatitudinal) dextral fault deformations, which are lacking in Kalevian rocks. The analysis of geological relations testifies that terrigenous rocks of the Ladoga Group accumulated on the eroded surface of deformed volcanites of the Sortavala Group (Fig. 1).

The contact between rocks of the Sortavala and Ladoga groups is cut by a diorite intrusion (Fig. 1). The Tervaoya diorite is a medium-grained massive mesocratic porphyric rock, which is schistosed in some sublatitudinal and northwest-trending zones. Porphyry phenocrysts (4–6 mm across) represented by zonal plagioclase (andesine no. 25–35) occupy up to 40 vol %. The groundmass is composed of fine-grained quartz–plagioclase aggregate, in which individual crystals of

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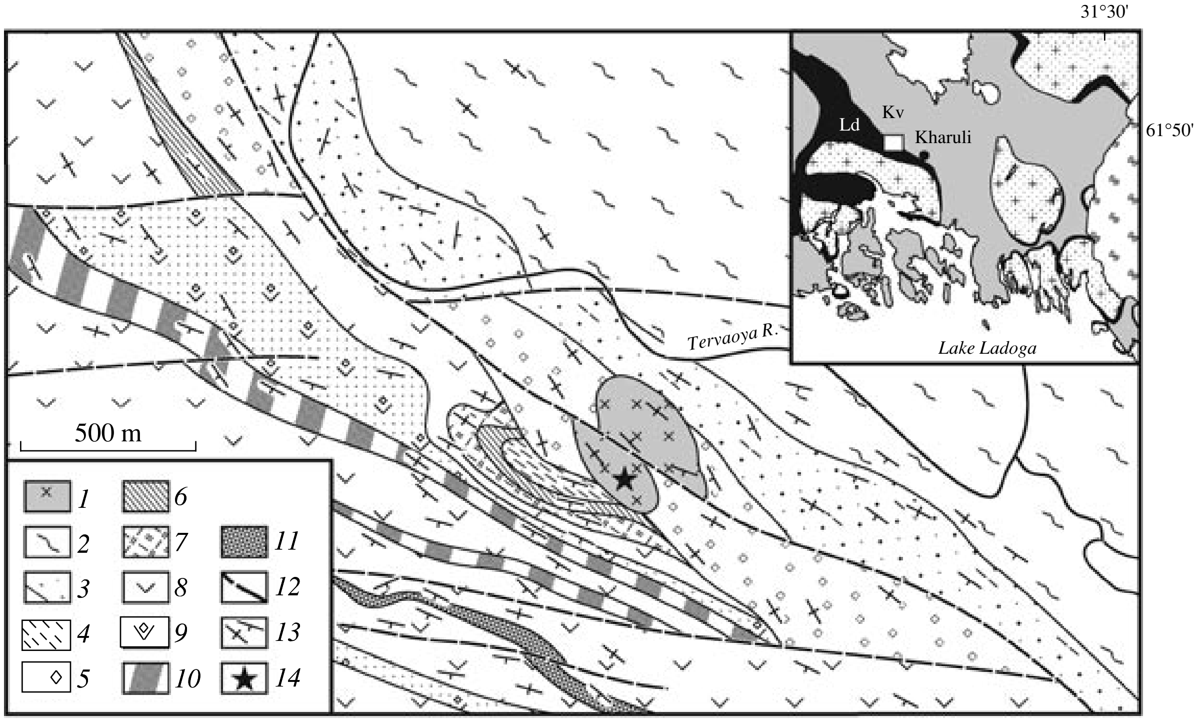

Fig. 1. Schematic geological structure of the northeastern edge of the Kirjavolakhta dome. (*1*) Tervaoya diorite; (*2–4*) Ladoga Group: (*2*) biotite schists with andalusite, (*3*) arkosic sandstones with gritstone lenses and calcareous concretions, (*4*) quartzite; (*5–11*) Sortavala Group: (*5*) carbonaceous siltstone, (*6*) komatiitic tuffs, (*7*) basaltic tuffs, (*8*) massive basalt lava, (*9*) amygdaloidal basalt lava, (*10*) komatiitic basalt lava, (*11*) komatiite lava; (*12*) faults; (*13*) strike and dip; (*14*) sampling sites. Inset: (Ld) rocks of the Ludicovian Superhorizon (Sortavala Group); (Kv) rocks of the Kalevian Superhorizon (Ladoga Group).

zonal plagioclase (andesine no. 25–30) are as large as 0.2–0.4 mm. Dark-colored minerals (~30 vol %) are represented by nearly equal proportions of biotite and amphibole. The chemical composition of diorites is presented in Table 1.

An optical study of accessory zircon from the Tervaoya diorite revealed three morphological varieties of crystals: (1) long-prismatic small transparent crystals; (2) short-prismatic medium-sized transparent light brown crystals; and (3) rounded small transparent pale pink and light brown crystals. Zircons are characterized by the heterogeneous structure of mineral individuals. All crystals consist of an inner core and one or more outer shells that are often unconformable to the core. The parameters of the U–Pb isotopic system were determined in both cores and shells for five zircon crystals from the first and second morphological varieties (Table 2). Seven determinations yielded concordant age values (Fig. 2).

Cores of long-prismatic zircons are, as a rule, well developed, slightly zonal, and conformal to later overgrowths (Fig. 3). Their age is estimated at 1922 ± 16 Ma (Table 2, no. 1). Cores of short-prismatic zircons have a round shape. They contain relicts of primary magmatic zonation, which is truncated by core boundaries (Fig. 3). The zircons yielded two dates: 2610 ± 8 and 2706 ± 12 Ma (Table 2, no. 2). Outer shells of both short-prismatic and long-prismatic zircon crystals yielded 1924 ± 16 and 1841 ± 9 Ma (Table 2, nos. 3–5).

The age of the Tervaoya diorite intrusion, determined by coincidence of ages for the cores of long-pris-

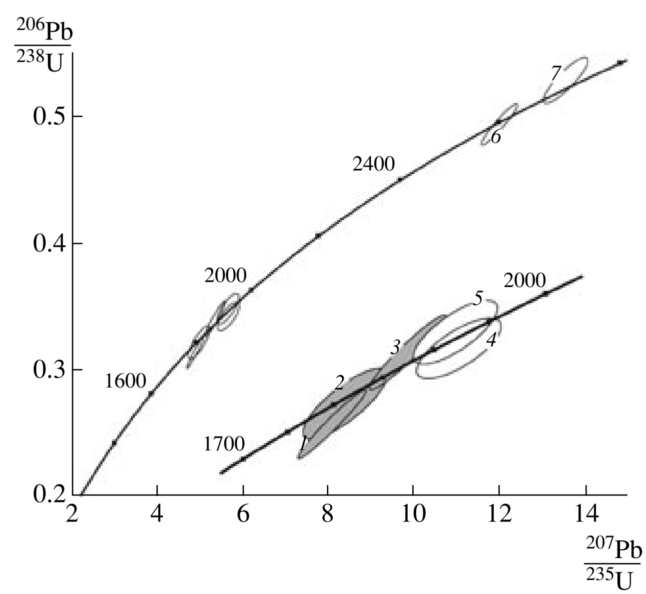

Fig. 2. Diagram with concordia for zircons from the Tervaoya diorite. (*1–7*) See Table 2.

Table 1. Chemical composition of the Tervaoya diorite (83903)

Oxide	Content, wt %	Element	Content, ppm	Element	Content, ppm	Element	Content, ppm
SiO_2	57.60	Li	13	Co	14.5	Eu	1.5
TiO_2	0.74	Rb	52	V	81.6	Gd	3.9
Al_2O_3	17.40	Sr	1164	Cu	304	Tb	0.58
Fe_2O_3	8.36	Ba	772	Zn	94	Dy	3.45
MnO	0.15	Y	20	Mo	1.2	Ho	0.70
MgO	1.80	Nb	10	Hf	2.6	Er	1.89
CaO	6.20	Zr	110	Sn	2.2	Tm	0.30
Na_2O	4.24	Ga	21	Pb	10.5	Yb	1.76
K_2O	1.68	Ta	0.56	La	21.9	Lu	0.27
P_2O_5	0.27	Sc	8.2	Ce	44.6	$^{147}Sm/^{144}Nd$	0.1162
L.O.I.	1.49	Be	2.1	Pr	5.5	$^{143}Sm/^{144}Nd$	0.511477 ± 7
Total	99.93	Cr	66	Nd	21.8	$\varepsilon_{Nd}(1922)$	–2.8
		Ni	31	Sm	4.5		

Note: Contents of rock-forming components were determined by the XFA at the Karpinskii All-Russia Research Institute of Geology, St. Petersburg; contents of rare elements, by MS-ICP at the University of Granada. Isotopic measurements of Nd and Sm were performed with a Finnigan MAT-261 mass spectrometer.

Table 2. Pb and U isotopes in zircons from the Tervaoya diorite (83903)

Ord. no.	Sample	$^{206}Pb_c$, %	U, ppm	Th, ppm	$\frac{^{232}Th}{^{238}U}$	$^{206}Pb_r$, ppm	$\frac{^{238}U}{^{206}Pb}$	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{238}U}{^{206}Pb^*}$	$\frac{^{207}Pb^*}{^{206}Pb^*}$	$\frac{^{207}Pb^*}{^{235}U}$	$\frac{^{206}Pb^*}{^{238}U}$	Corr. coeff.
1	3.2	0.06	453	117	0.27	122	3.178 ± 2.7	0.11315 ± 0.6	3.18 ± 2.7	0.11265 ± 0.64	4.88 ± 2.8	0.3144 ± 2.7	0.973
2	1.2	0.01	418	190	0.47	116	3.104 ± 2.7	0.1117 ± 2	3.104 ± 2.7	0.1116 ± 2	4.96 ± 3.4	0.3221 ± 2.7	0.808
3	4.1	0.14	277	29	0.11	81.2	2.926 ± 2.7	0.11463 ± 0.74	2.93 ± 2.7	0.11344 ± 0.85	5.34 ± 2.9	0.3413 ± 2.7	0.955
4	1.1	1.13	87	28	0.33	26.3	2.834 ± 2.8	0.1269 ± 1.2	2.866 ± 2.8	0.1169 ± 2.4	5.63 ± 3.7	0.3489 ± 2.8	0.755
5	6.2	0.06	393	118	0.31	116	2.915 ± 2.7	0.1201 ± 2.5	2.917 ± 2.7	0.1195 ± 2.5	5.65 ± 3.7	0.3428 ± 2.7	0.732
6	6.1	0.15	240	137	0.59	102	2.019 ± 2.7	0.1771 ± 0.85	2.022 ± 2.7	0.1758 ± 0.88	11.98 ± 2.9	0.495 ± 2.7	0.952
7	3.1	0.36	46	100	2.27	20.8	1.88 ± 2.8	0.1884 ± 1.3	1.887 ± 2.8	0.1852 ± 1.6	13.53 ± 3.2	0.53 ± 2.8	0.873

Note: Intensity of the primary beam of oxygen ions 4 nA, spot (crater) diameter 18 μm. U/Pb ratios were normalized to 0.0668 (standard zircon TEMORA [9]). (c) Common Pb; (r) radiogenic Pb; asterisks denote isotopic ratios corrected for measured ^{204}Pb. Errors of single analyses (ratios and ages) are given at the 1σ level; errors of calculated concordant ages, at the 2σ level.

matic zircons and the oldest overgrowths on short-prismatic zircons, is 1922 ± 11 Ma.

The morphology and age of cores in short-prismatic zircons suggest their xenogenic nature related to removal of the restite material during the melting of the Archean crustal protolith. The primary Nd isotopic composition of diorites (ε_{Nd} = –2.8) also indicates the crustal origin of their protolith. The age of outer shells on zircon crystals (1841 ± 9 Ma) defines the timing of superimposed high-temperature metamorphism.

Magmatic zircon of the Tervaoya diorite differs from the Archean protolithic zircon by a higher content of Y and REE and lower degree of REE fractionation (Table 3). At the same time, metamorphic overgrowths of zircons are characterized by intermediate concentrations of HREE and Y.

The analysis of the geological structure of the contact zone between rocks of the Sortavala and Ladoga groups suggests the existence of structural–metamorphic unconformity between them. Quartzites from the

Table 3. Contents of trace elements in zircons from the Tervaoya diorite (83903)

Ord. no.	Sample	La	Ce	Nd	Sm	Eu	Gd	Dy	Er	Yb	Hf	Y
1	3.2	4.61	33.55	19.52	14.14	6.16	33.69	133.64	337.59	777.03	7263.25	1965.59
2	1.2	49.23	321.48	425.78	305.30	95.18	377.83	331.51	290.70	588.05	6562.84	2233.11
3	4.1	6.34	55.64	96.91	59.41	12.80	77.74	80.54	117.72	236.84	7796.45	837.58
4	1.1	186.82	809.87	583.94	254.82	58.40	356.75	570.15	987.35	1848.83	5650.57	6515.45
5	6.2	59.47	639.56	961.77	741.03	228.70	934.02	800.99	527.53	870.12	7134.26	4434.56
6	6.1	182.60	365.94	163.32	41.92	8.00	68.43	71.17	103.28	241.70	10623.99	704.44
7	3.1	0.17	41.82	3.14	5.30	1.05	23.97	78.83	158.96	326.47	8483.25	1004.18

base of the Ladoga Group section accumulated on deformed and eroded volcanites of the Sortavala Group. The age of the Tervaoya diorite, which intrude quartzites and arkosic sandstones, indicates that the lower part of the Ladoga section had already formed 1922 ± 11 Ma ago.

Volcanites found in sedimentary rocks of the Ladoga Group and zircon datings testify that Kalevian sedimentation continued in the southern structural domain 1888–1880 Ma ago (with regard to dating errors) [8]. Simultaneously, the northern domain accumulated rocks of the gabbro–plagiogranite complex represented by the Kaalamo, Alattu, and Velimyaki plutons (1897–1880 Ma [3]). Termination of the Ladoga Group sedimentation is indicated by the formation of minor intrusions of the Impiniemi Complex with an age of 1874 ± 13 Ma [3].

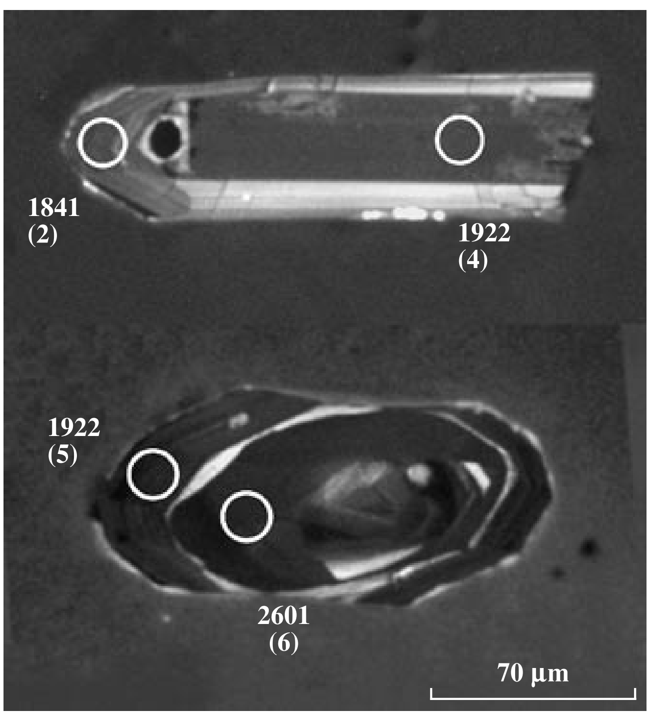

Fig. 3. Microphotographs of accessory zircon from the Tervaoya diorite.

The chronostratigraphic correlation of successions of the Karelian Complex in the northern Ladoga region and the Svecofennian area of Finland suggests the following conclusion. Deposition of the Sortavala volcanosedimentary complex, its metamorphism, erosion, and unconformable overlapping by Ladoga quartzites and arkosic sandstones predated the Svecofennian subduction in the pericraton zone. Accumulation of volcanosedimentary rocks of the Savo Belt (southeastern Finland) and the formation of island-arc granitoids of the central Finland complex was synchronous with late Kalevian volcanism in the southern structural domain of the Ladoga region and the development of the gabbro–plagiogranite complex of the northern Ladoga region.

The geological and isotopic data obtained allow a subdivision of the Ladoga Group rocks into two units of different ages. The lower unit has an upper age limit of 1922 ± 11 Ma, whereas the upper unit formed 1897–1880 Ma ago. The lower unit includes, at least, quartzites of the Kontiosari Formation and arkosic sandstones from the base of the Ladoga section in the northern Ladoga region.

Thus, the Ludicovian–Kalevian boundary can be drawn at 1922 ± 11 Ma for the northern Ladoga region. However, further investigations are required for determining its lower age limit.

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