RUSSIAN JOURNAL OF EARTH SCIENCES, VOL. 5, NO. 2, PAGES 75–91, APRIL 2003

Genesis of the earliest (3.20–2.83 Ga) terranes of the Fennoscandian shield

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Abstract. Our study of magmatic rocks provides grounds for discussing the successive phases and geodynamic conditions of continental crust generation between 3.2–2.83 Ga within the Fennoscandian shield. Three tectono-magmatic phases, 3.2–3.1 Ga, 3.0–2.92 Ga, and 2.92–2.83 Ga, are established. During the 3.2–3.1 Ga phase, voluminous intrusives were emplaced, creating the large Vodlozero "sialic core." Positive $\varepsilon_{\mathrm{Nd}}(t)$ values for mafites and, partly, granitoids point to a weighty contribution of juvenile material derived from a depleted mantle. At the same time, Nd isotope composition for a number of granitoid massifs and zircon ages suggest the presence of an earlier (as old as 3.5 Ga) crustal component. Endogenic processes that occurred during the second, 3.0–2.92 Ga phase, have been recorded in southeastern and western Karelia and are inferred to have occurred in the Kola Peninsula as well. During this phase, oceanic plateaus and island arcs were formed near, to be accreted onto, the western and eastern margins of the ancient Vodlozero core. Simultaneously, the central part of the Vodlozero "sialic core" was the locus of emplacement of gabbronorite–diorite intrusions and purely dioritic bodies, as well as vigorous tonalite–granodiorite magmatism, to form the Vodlozero domain, the oldest on the shield. In the western Karelian domain, rocks younger than 2.92 Ga are exposed at the current erosional surface. The presence of ancient material in western Karelian crust is pinpointed by Nd model ages for granitoids and volcanites and by a detrital zircon age from granite. The third, 2.92–2.83 Ga phase entailed further reworking of ancient terranes and initiation of new sialic cores. At the northern margin of the Vodlozero domain and within the western Karelian one, a system of rift-related features came into being, eventually to evolve into bimodal greenstone belts largely dominated by mafic and ultramafic volcanites. The Kola province provided the stage for inception of rift-related greenstone belts with their associated komatiite–tholeiite (2.92–2.87 Ga) series followed by the basalt–andesite–dacite (2.88–2.79 Ga) series. Apparently, these belts in their present-day form, just like those of the western and eastern margins of the Vodlozero domain, result from tectonic juxtaposition of rock assemblages that originated from a variety of geodynamic settings. Archean continental crust of the Fennoscandian shield is shown to have formed through both progressive addition of sialic crust over time, mainly at convergent boundaries of ancient plates, and via reworking of ancient fragments, which involved input of juvenile material resulting from rising mantle plumes. In all likelihood, ascending mantle plumes are responsible for the formation of accretionary and collisional orogens, whereas the coeval magmatism at active plate margins was due to subduction (in the context of the plate tectonic mechanism). Generation of rift-related structures and associated magmatism may have been driven by rising mantle plumes. Brittle deformations leading to rifting and associated with the ascent of mantle plumes did not result in break-up of the young continental crust. The main outcome of this mechanism was massive inflow of high-temperature magma into the lithosphere forming within-crust layers of "asthenosphere," in which granite melts originated, eventually to migrate to shallower levels.

Paper number TJE03118.
ISSN: 1681–1208 (online)

The online version of this paper was published 7 March 2003.
URL: http://rjes.wdcb.ru/v05/tje03118/tje03118.htm

Introduction

High precision geochemical and isotope geochemical data amassed over the recent decades enabled the development of criteria for linking geochemical signatures of magmatic rocks to their original geodynamic settings based on case

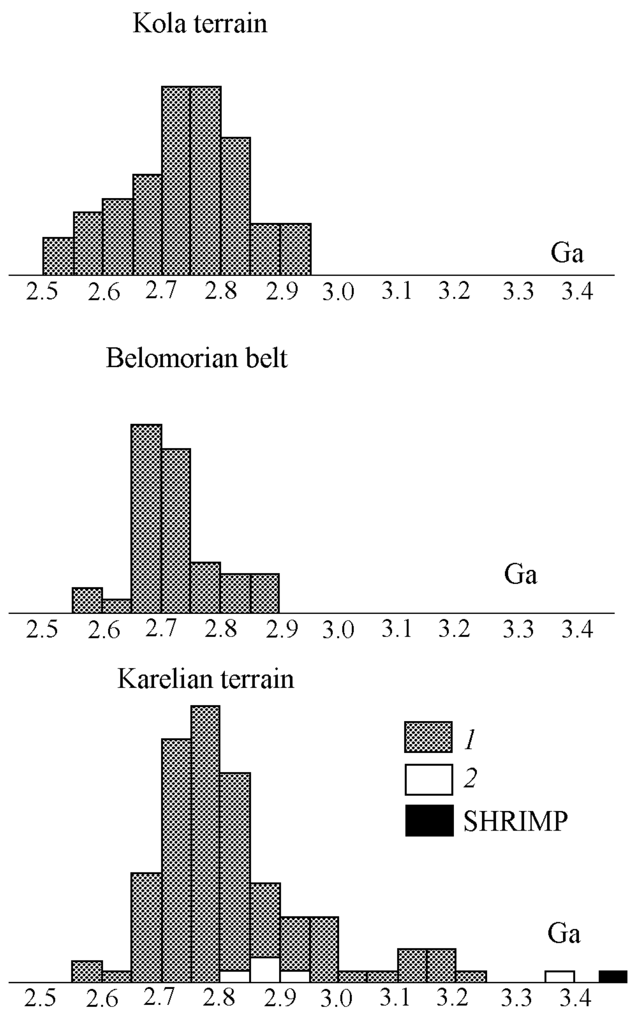

Figure 1. Distribution of U–Pb zircon ages (1), whole rock Sm–Nd ages (2), and ages obtained by ion microprobe (SHRIMP) on individual zircon grains from the Belomorian belt and the Kola and Karelian provinces.

studies in Phanerozoic assemblages [*Kerr et al.,* 2000; *Kerrich and Wyman,* 1997; *Pearce et al.,* 1984, 1996]. Concomitantly, these criteria were intercompared and used to unravel Precambrian geodynamic settings. In the course of the studies of the Early Precambrian evolution, a variety of geodynamic models for continental crust generation were proposed; these models are often mutually opposing. The most commonly accepted tenet, covering the Early Archean [*De Wit et al.,* 1992], draws on the plate tectonic paradigm. According to this model, best refined for Canada [*Card et al.,* 1989; *Hoffman,* 1988; *Kusky and Polat,* 1990; *Percival et al.,* 1994], continental crust was generated at convergent plate boundaries through accretion of island arcs, oceanic crustal fragments, oceanic plateaus, and accretionary wedge sediments. The latest works emphasize the role of oceanic plateaus in accretionary orogens, where plume-derived mafic lavas (komatiitic inclusive) are accumulated. On the other hand, in certain regions mafic and alkaline lavas were piled up in continental margin rift settings (volcanic rifted margins) that cut through both oceanic and continental lithospheres [*Kerr et al.,* 2000; *Marzoli et al.,* 2000]. Protracted evolution of Archean cratons with repeated manifestations of endogenic processes complicates Archean geologic reconstructions considerably. An example of an ambiguously interpreted Archean mafic succession on the Fennoscandian shield is offered by the Kostomuksha greenstone structure, seen by some workers as an oceanic plateau obducted onto the continent [*Puchtel et al.,* 1998] and by others, as a "continental" plateau generated above continental margin rifts [*Lobach-Zhuchenko et al.,* 2000a].

Albarede [1998] proposes three continental crust growth mechanisms: (1) melting of hot (i.e., less than 30 m.y. old) oceanic crust in subduction zone, (2) accretion and subsequent subduction of oceanic plateaus, and (3) formation of plateaus on free-floating plates ("loose-plate loading") followed by their transformation to continental crust, occasionally bypassing subduction. All the models just mentioned imply that the rocks of the tonalite–trondhjemite series, a volumetrically important constituent in the Archean continental crust, originated through partial melting of metamorphosed basalt, the most popular model being the one invoking oceanic crust melting [*Drummond and Defant,* 1990; *Martin,* 1994].

However, there is an essential discrepancy between natural tonalitic compositions and basalt melting experiments; among other things, natural tonalites have higher Mg-numbers [*Evans and Hanson,* 1994; *Lobach-Zhuchenko et al.,* 2000c; *Rudnick,* 1995]. Therefore, the issue of the tonalite source, which is of prime importance to elucidating the nature of continental crust, remains as yet without unambiguous solution, either.

Metamorphic and metasomatic processes, broadly manifested in the Precambrian and rendering a number of elements highly mobile, further complicate comparison of geochemical data for similar Phanerozoic and Precambrian rocks. This study, nonetheless, offers such a comparison based on detailed geologic observations, abundant geochronology data, and geochemical characteristics of the least mobile elements, and discusses continental crust growth on the Fennoscandian shield. Numerous isotope ages from Archean rocks of the shield testify to continuity of crust generation in the Archean (Figure 1). The main pulse of endogenic processes falls within 2.8–2.6 Ga. First, between 2.8–2.7 Ga, enormous granitic masses of tonalite–trondhjemite composition were generated on the shield, and then, between 2.7–2.6 Ga, two-feldspathic granites were emplaced, with concomitant massive migmatization and metasomatism. These endogenic processes brought about a considerable reworking of the preexisting rocks, which obscures recognition of ancient crustal fragments. Nevertheless, analyzing spatial distribution of geochronologic and isotope geochemical data in conjunction with detailed geological observations enables us to (1) establish lithotectonic domains of different age, (2) delineate the oldest continental crust fragments on the Fennoscandian shield, (3) depict the main phases of their formation, and (4) unravel the history for the Archean crust of the shield as a whole.

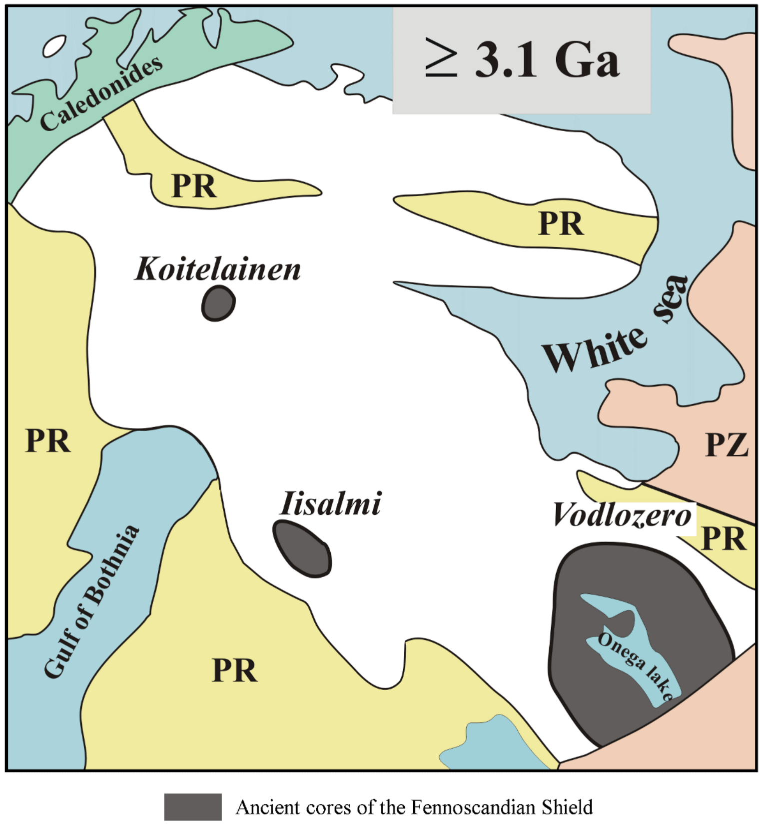

Figure 2. Ancient core (>3.1 Ga) of the continental crust of the Baltic shield. Circles, localities with 3.2–3.1 Ga dates in the Vodlozero core.

Principal Phases of Formation of the Ancient Archean Crust

The early history of generation and development of the Archean crust of the Fennoscandian shield is divided into the following phases: 3.2–3.1 Ga, 3.0–2.92 Ga, and 2.92–2.83 Ga.

3.2–3.1 Ga Phase

This phase is established from the presence of rocks dated with confidence at 3.2–3.1 Ga and forming "sialic cores," the oldest reconstructible entities in the Baltic shield (Figure 2). The largest, Vodlozero core is located in southeastern Karelia and makes the central part of the Vodlozero domain [*Lobach-Zhuchenko et al.,* 2000a]. Minor fragments of rocks generated during this phase are found in northern and central Finland.

In northern Finland, the oldest rocks are exposed near Koitelainen, making a small (1.4×2.6 km) dome overlain by younger sedimentary and volcanic rocks [*Puustinen,* 1977], likely belonging to the Kittelä greenstone belt [*Gaal et al.,* 1976]. The dome is composed of tonalites and trondhjemites dated at 3110±34 Ma, with $\varepsilon_{\rm Nd}(t) = -3.7 \pm 1.8$, which suggests a long-lasting (300–500 m.y.) crustal prehistory for the tonalites [*Jahn et al.,* 1984; *Kröner and Compston,* 1990; *Kröner et al.,* 1981].

In southwestern central Finland, *Höltta* [1997] has established the Iisalmi "microcontinent," distinctive from the rest of central Finland in that it contains ancient tonalitic gneisses and displays young Archean granulite metamorphism. According to [*Höltta,* 1997], this terrane is composed of small enderbite blocks ranging in composition from diorite to tonalite and intercalated by mafic granulites, occurring among rocks of tonalite–trondhjemite composition and migmatites that preserve some amphibolite facies assemblages. The tonalites were dated at 3136±20 and 3095±18 Ma [*Paavola,* 1986]. Later, a similar age (3.2–3.1 Ga) was obtained from melanosome of tonalite composition in migmatites, metamorphosed in the granulite facies [*Huhma et al.,* 2000]. According to these writers, the old zircons ages obtained from migmatized granulite facies rocks correspond to the protolith age, whereas most granulites and enderbites are dated between 2870–2630 Ma [*Huhma et al.,* 1995, 2000; *Paavola,* 1986]. Therefore, the oldest rocks in the Iisalmi region are relics of ancient tonalite crust occurring among the younger and areally predominant dioritic and tonalitic intrusions, overprinted by granulite metamorphism. Compositionally similar rocks affected by granulite metamorphism of the same age are developed in

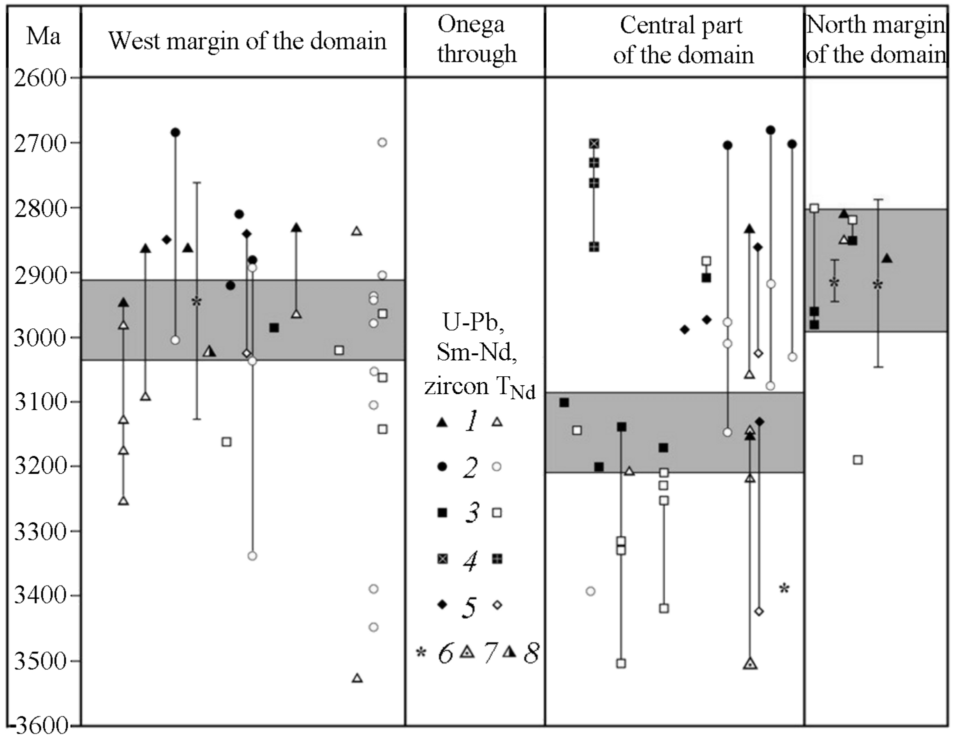

Figure 3. Distribution of the ages of rocks and their protoliths in the Vodlozero domain through the section: western accretionary zone—central part of the domain ("sialic core")—northern zone. 1 – gneisses, acid and intermediate volcanites; 2 – granites; 3 – tonalite–trondhjemite–granite series rocks; 4 – subalkaline series; 5 – gabbro and diorite; 6 –isochron age obtained by the Sm–Nd method from basalts and komatiites, with interval at the point showing analytical error; 7 – oldest zircon ages obtained on SHRIMP; 8 – whole rock Pb–Pb age from andesite. Vertical lines connect ages obtained from the same sample or massif. The age of the shaded area corresponds to the first phase of formation of the structure (marginal zones and sialic core).

western Karelia, in the region of Lake Tulos and the village of Voknavolok, where no ages older than 2.85 Ga have been obtained to date.

The largest fragment of the ancient structural setup is the "sialic core" of the Vodlozero domain with a reconstructible protracted geologic scenario. The oldest firmly dated sialic rocks are 3.2–3.1 Ga in age [*Lobach-Zhuchenko et al.,* 1993]. They are established to form isolated patches across the domain (Figures 3, 4), which suggests their once broad spread. The oldest rocks of the Vodlozero domain are represented by both plutonic (chiefly tonalitic) and calc-alkaline volcanic rocks: gneisses and amphibolites of the Vodlozero assemblage. This metamorphic assemblage, whose U–Pb zircon age is 3151±18 Ma (from a gneiss sample) and 3128±86 Ma (from an amphibolite sample) [*Lobach-Zhuchenko et al.,* 1993], ranges in composition from basaltic andesite to dacite [*Lobach-Zhuchenko et al.,* 1984]. Compositionally similar gneisses have been documented from the central and southern parts of the Vodlozero domain (Figure 5). The generation of mafic rocks and gneisses and the early phase of their deformation, metamorphism, and migmatization predated the emplacement of tonalites and granodiorites. Ancient tonalite and granodiorite ages were obtained from three localities. In the middle reaches of the Vyg River, migmatites and granodiorites have been dated at 3210±12 and 3138±63 Ma, respectively. Tonalites from the Lairuchei River have yielded a 3166±14 Ma age [*Lobach-Zhuchenko et al.,* 1993]. In the vicinity of Lake Palaya Lamba, tonalites make the protolith to a migmatite dated at 3100±70 Ma [*Lobikov and Lobach-Zhuchenko,* 1980]. Needless to say, ancient tonalites prevailing among the rocks of old cratons are pivotal for elucidating the origin of ancient crust. Compositional study of the tonalites shows most of them to have elevated Mg-numbers and some other chemical features that preclude their genesis through dehydration melting of basaltic slab—i.e., in compliance with the most commonly accepted model. The mismatch between Mg-numbers for the tonalites under study and those obtained from metabasalt melting experiments was noted earlier [*Evans and Hanson,* 1997; *Kelemen,* 1995; *Rudnick,* 1995]. The source that would produce melts with characteristics observable for natural tonalites should be chemically similar to high-Mg andesite or boninite [*Lobach-Zhuchenko et al.,* 2000c].

Therefore, during the 3.2–3.1 Ga phase, considerable masses of acid and intermediate intrusive rocks were emplaced in a more ancient country rock, thus giving rise to the

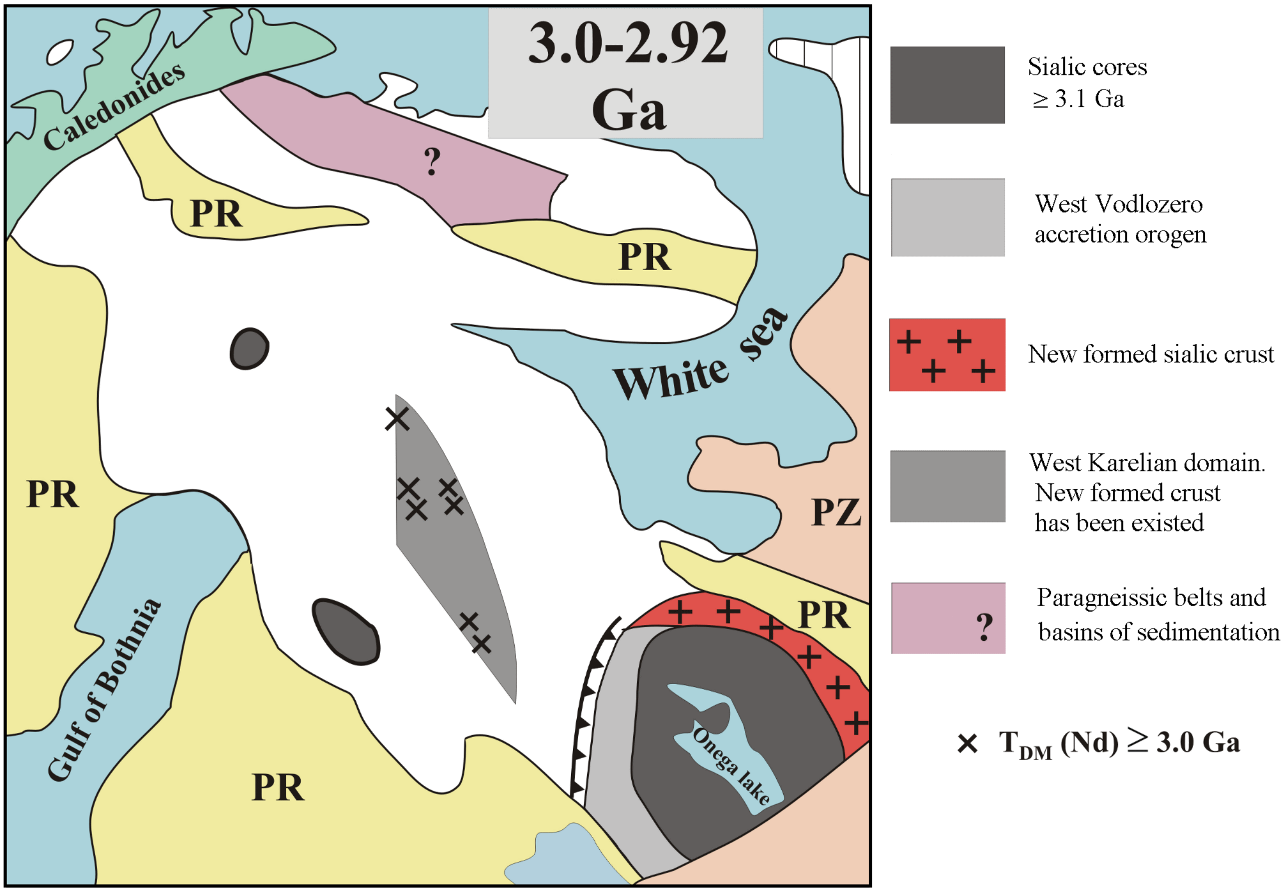

Figure 4. Distribution of sialic crustal fragments generated by the end of the 3.0–2.92 Ga phase. 1 – ancient sialic core (3.2–3.1 Ga); 2 – western Vodlozero accretionary orogen; 3 – newly formed sialic crust; 4 – portion of the western Karelian terrane with likely preexisting new crust; 5 – inferred depositional basin; 6 – available U–Pb zircon ages; 7 – position of rocks with Nd model ages >2.92 Ga.

large Vodlozero "sialic core." The Nd isotope composition of the mafites with positive $\varepsilon_{\mathrm{Nd}}(t)$ values implies a considerable contribution of a juvenile material derived from a depleted mantle (Figure 3). At the same time, the Nd isotope composition of a number of granite massifs suggests that the lower crust contains an older (up to 3.5 Ga) component, which was a constituent in the protolith to these massifs. The presence of an ancient component in the crust of the Vodlozero core is also deduced from a SHRIMP age determination on cores of large zircon grains, equaling 3500±90 Ma [*Sergeev et al.,* 1990]. While dating zircons from tonalites of the Lairuchei massif, a zircon population was detected that plots to the right of the discordia defining a 3.17 Ga age for the tonalites. The position of this data point also suggests that zircon cores host a more ancient component [*Lobach-Zhuchenko et al.,* 1993]. In constraining the age for Vodlozero gneisses from the ^{207}Pb/^{206}Pb ratio, as measured by LA-MC-ICP-MS, two zircon varieties from the same sample yielded older ages (3205±0.95 and 3238±0.95 Ma) than those obtained by the classic method [*Sergeev,* 2000]. On the east of the Vodlozero domain, *Kulikova* [1993] detected basaltic komatiites whose Sm–Nd age is measured at ca. 3.4 Ga [*Puchtel et al.,* 1991]; this age, however, calls for a better grounded geochronologic validation. The totality of the above data unequivocally suggest ages older than 3.1–3.2 Ga for the rocks of the Vodlozero sialic core, which was essentially reworked at a later time.

3.0–2.92 Ga Phase

The endogenic processes manifested during this phase and responsible for further evolution of the ancient continental crust of the Fennoscandian shield have been documented in southeastern and western Karelia (in the Vodlozero and western Karelian domains) and are inferred to have occurred on the Kola Peninsula (in the Kola–Norwegian domain, Figure 4) as well.

A geologic scenario for this phase is restored most fully from the Vodlozero domain (Figure 5). Here, near the western margin of the Vodlozero sialic core, between 3.0 and 2.92 Ga there took shape island arcs, whose accretion onto the core created the shield's oldest accretionary orogen. Currently, relics of rocks from ancient island arcs and oceanic

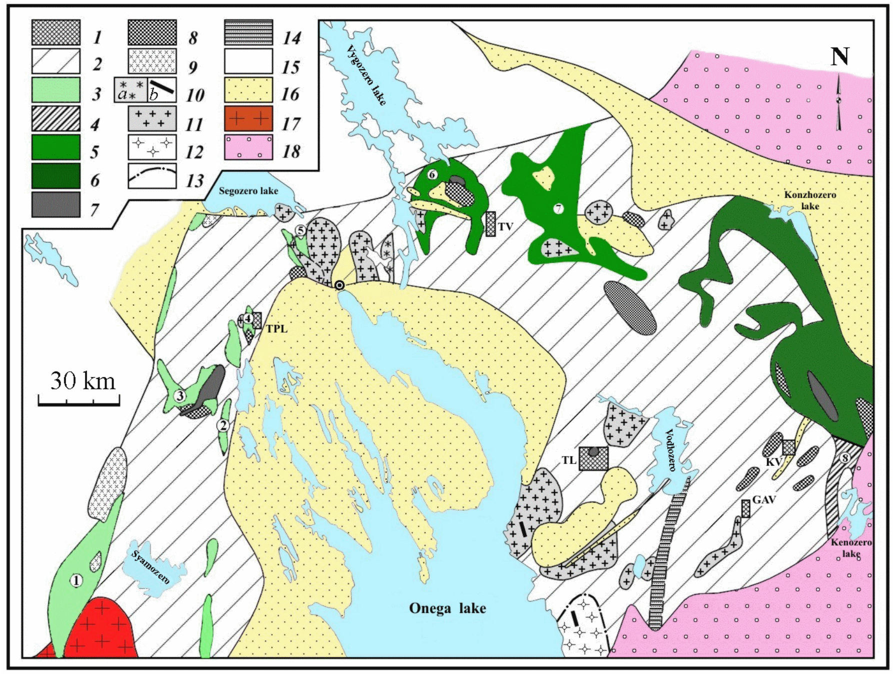

Figure 5. Geologic map showing the Vodlozero domain [*Lobach-Zhuchenko et al.,* 2002]. 1 – localities with the oldest dated rocks in the domain: KV – mafic volcanites at the Vinela and Cherva rivers, TL – tonalites at the Lairuchei River; GAV – Vodla gneisses and amphibolites; TV – tonalites at the Vyg River; TPL – Palaya Lamba tonalites; 2 – tonalitic gneiss, granitic gneiss, and migmatite, undifferentiated. Greenstone belts: the oldest (3.0–2.92 Ga), 3 – with multimodal volcanism, 4 – with bimodal volcanism; 5 – younger (2.9–2.85 Ga), with bimodal volcanism, 6 – with bimodal volcanism, undated. Numerals indicate greenstone belts: 1 = Hautavaara, 2 = Koikary, 3 = Semchensky, 4 = Palaya Lamba, 5 = Oster, 6 = Shilos, 7 = Kamennye Ozera, 8 = Kenozero. Intrusions: 7 – gabbronorite, gabbro, diorites; 8 – tonolites, trondhjemit ; 9 – high-Mg granite; 10 – subalkaline rocks: a, granitods, b, mafic dikes; 11 – granite; 12-granulite facies area with charnockite and enderbite plutons; 13 – boundaries of granulite facies area; 14 – mafites of Matkalachtinskaja zone; 15 – Central-Karelian domen; 16 – Proterozoic rocks; 17 – rapakivi plutons; 18 – Paleozoic rocks.

rock assemblages make large lenses and belts (Hautavaara, Semchensky, Oster, etc., greenstone structures or belts; Figure 2), enclosed in granites and viewed as being part of the continuous Segozero–Vedlozero greenstone belt [*Greenstone Belts...,* 1988; *Sokolov,* 1981; *Svetov,* 1997; *Svetova,* 1988]. The fact that rocks presently found in the western accretionary pile were formed during this phase has been validated by age determinations for andesites of the Oster and Palaya Lamba belts (3020±10 Ma) [*Lobikov,* 1982], trondhjemites cutting through metandesites (2985±10 Ma) [*Belyatsky et al.,* 2000], andesitic dacites of the Hautavaara belt (2945±19 Ma) [*Ovchinnikova et al.,* 1994], as well as komatiites and basalts (2944±170 Ma) [*Svetov and Huhma,* 1999] and rhyolites (2935±15 Ma) [*Bibikova and Krylov,* 1983] of the Koikary belt. The accretionary orogen was formed prior to 2876±21 Ma ago (the age of granites cutting through the newly formed greenstone structure [*Chekulaev et al.,* 2002; *Kovalenko and Rizvanova,* 2001]).

The study of volcanite compositions suggests that the rocks were derived from a variety of sources and are similar to modern rocks generated at contrasting geodynamic settings. Geochemical signatures of komatiites from the Oster and Koikary belts (high MgO and Cr; Table 1) are akin to high-T melts erupted on oceanic plateaus and related to mantle plumes. Komatiites of the Hautavaara belt are light REE enriched and have negative $\varepsilon_{\mathrm{Nd}}(t)$ values, most likely due to crustal contamination (Table 1, Figure 6). They may represent part of a plateau formed on continental crust. Apparently, the Palaya Lamba komatiite–tholeiite assemblage also makes a fragment of a plateau formed on continental crust. This is evidenced by the low angle attitude surviving in volcanic flow units; the overprinting deformations (high angle foliation) leave intact bedding planes, as would be expected in an oceanic plateau obducted onto a continental margin.

Tholeiites of the greenstone belts of the western accretionary zone are classed into three principal types: high-T plateau tholeiites, backarc basins tholeiites, and island arc tholeiites. The Oster greenstone belt contains type 1 and type 3 tholeiites, composing isolated tectonically dispersed blocks. Type 1 tholeiites are found in close association with amphibolite dikes and serpentinite lenses [*Chekulaev et al.,*

Table 1. Geochemical analyses of representative komatiite samples from the greenstrone belts of Vodlosero domain western margin

Belt	Hautavaara			Semchensky	Palaya Lamba					Oster	
Sample no.	427-2 Vr	427-5 Vr	427-7Vr	54 Ar	2103b Ar	2104 Ar	851a Ar	927 Ar	422 Ar	534 Ar	565 Ar
SiO_2	46.49	46.82	46.98	45.85	46.23	47.91	47.84	49.08	52.42	45.11	45.65
TiO_2	0.35	0.35	0.39	0.33	0.23	0.34	0.30	0.37	0.46	0.2	0.23
Al_2O_3	7.02	6.90	7.79	10.74	6.9	6.91	12.32	10.90	9.06	5.79	6.3
FeO	11.91	11.54	11.68	10.34	12.73	11.9	8.84	12.93	9.68	12.63	12.97
MnO	0.20	0.21	0.16	0.16	0.2	0.17	0.17	0.25	0.21	0.19	0.19
MgO	29.64	30.66	28.82	24.79	28.3	25.67	16.01	15.60	15.00	32.83	29.19
CaO	4.32	5.71	5.13	7.45	5.16	5.98	5.95	8.80	10.47	2.58	4.65
Na_2O	0.03	0.05	0.03	0.2	0.17	0.01	2.58	1.32	1.47	0.07	0.11
K_2O	0.01	0.02	0.01	0.1	0.1	0.01	0.11	0.82	0.07	0.02	0.02
P_2O_5						0.01	0.02	0.11	0.09	0.01	0.01
mg	0.83	0.81	0.83	0.81	0.80	0.79	0.76	0.68	0.74	0.82	0.81
Rb	1	3	0	4	1	3	2	3	3	1	1
Sr	16	29	7	8	3	4	29	32	36	23	30
Y	5	6	6	8	4	3	11	10	11	9	8
Zr	15	12	19	26	12	15	20	18	19	13	15
Nb	1	1	1	5	2	1.5	1.5	2	1	1	1
Ti	1323	1235	1653	1997	1380	2040	2564	2177	2567	1235	1283
Ba	7	9	11		22	36	50	12	15	101	90
Cr	1985	2172	2170	2362	3632	3251	1053	2030	1154	5021	4501
Ni	706	644	584	976		929	334	466	356	1148	984
Co				77			77	108	77	126	116
V				149						90	128
Th	0.3	0.33	0.36		0.13	0.04	0.23	0.05	0.09	3	4
La	1.5	2.16	1.58		2.0	1.4	1.4	3.8	0.55	0.63	1.3
Ce	2.96	4.62	3.65		3.1	3.3	3.1	6.8	1.5		
Nd	1.99	3.29	2.37	2.48	2	5.5	2.4	7.7	1.1	0.38	0.57
Sm	0.58	0.82	0.75	0.89	0.82	0.8	0.94	1.4	0.41	0.145	0.21
Eu	0.18	0.3	0.19		0.3	0.12	0.26	3.1	0.065		
Gd	0.92	1.22	0.85		1.0	1.1	1.3		0.6		
Tb					0.18	0.19	0.24	0.2	0.088		
Tm					0.13	0.13	0.17	0.13	0.085		
Yb	0.68	0.87	0.7		0.88	0.88	1.12	0.83	0.51	0.52	0.72
Lu	0.1	0.14	0.1		0.13	0.13	0.14	0.13	0.068		
Ti/Zr	91	100	86	76.8	115	136	128	121	135	95	86
Nb/Y	0.2	0.17	0.17	0.63	0.50	0.50	0.14	0.20	0.10	0.11	0.13
Zr/Y	2.9	2.05	3.2	3.25	3	5	1.8	1.8	1.7	1.4	1.9
Nb/La	0.67	0.46	0.63		1.0	1.1	1.1	0.5	1.09	1.59	0.77

2002]. These are low in the lithophile elements and REE and are enriched in Cr and Ni (Table 2; Figures 7a, 7b). In the Semchensky belt, all three types of tholeiites are brought together. Type 2 tholeiites are low in Ni and have low Nb/Y ratios (Table 2; Figures 7a, 7b). Type 3 tholeiites, compositionally similar to island arc tholeiites, are closely associated with andesites. They are LREE enriched and are marked by negative Nb anomaly and Nb/La <0.9 (Table 2; Figures 7a, 8).

Intermediate and acid volcanites are represented by andesites and andesitic dacites akin to volcanites of plate-margin volcanic arcs in terms of K_2O and K_2O/Na_2O, and to modern volcanites of mature and evolved island arcs in terms of Sr, Ba, TiO_2, and Zr contents and Ti/Zr and La/Sm ratios. They have high REE abundances and rather fractionated REE patterns with $(La/Yb)_N$ = 5–9, $(La/Sm)_N$ = 3.5–4.1, Tb/Yb = 1.6–2, and negative Nb and Ti anomalies. Andesites of the Semchensky and Hautavaara belts are high-Mg (with Mg-numbers of 0.52–0.55, respectively) and have high Cr contents (on average, 146 to 160 ppm). Andesites from the other belts have lower Mg-numbers (0.42–0.44) and low Cr contents (~50 ppm).

The intermediate and acid rocks of this volcanic phase occurring in greenstone belts at the western margin of the Vodlozero domain are inferred to be linked to subduction beneath ancient sialic crust of the Vodlozero block [*Glebovitsky,* 1993; *Lobach-Zhuchenko et al.,* 2000b]. Generation of melts initial to volcanites of the Oster and Palaya

Table 2. Geochemical analyses of representative basalt samples from the greenstrone belts of Vodlosero domain western margin

	Oceanic platean basalts, (Group 1)									Back-arc (Group 2)		Island-arc (Group 3)	
Belt	Hautavaara		Semchensky		Palaya Lamba		Oster			Semchensky		Semchensky	Oster
Sample no.	7 L-Z	8 L-Z	717 V-Ch	61 Ar	517 V-B	76/9 Ar	B8/7b I-K	1736 I-K	348a Ar	1044 Ar	12 Ar	27 Ar	546 Ar
SiO_2	50.78	50.72	50.47	50.66	48.48	49.12	49.41	48.56	50.43	50.66	50.59	50.54	51.47
TiO_2	0.65	0.65	0.85	0.87	1.11	0.90	1.20	0.92	1.10	0.53	0.52	1.71	1.47
Al_2O_3	14.77	14.81	14.48	14.36	14.76	14.20	14.89	14.85	14.37	13.01	14.59	11.82	11.75
FeO	9.61	9.54	9.82	10.64	12.36	11.79	12.66	11.7	12.25	10.11	11.55	15.96	13.30
MnO	0.23	0.16	0.21	0.16	0.20	0.21	0.22	0.15	0.20	0.16	0.20	0.28	0.22
MgO	8.15	8.84	7.61	7.85	6.72	7.91	7.81	7.38	8.51	8.54	8.99	6.04	6.45
CaO	12.08	11.18	14.60	11.38	14.08	9.52	12.3	11.46	9.56	10.53	12.24	10.19	11.91
Na_2O	1.60	1.71	1.73	2.51	1.82	2.80	0.85	2.43	2.8	1.85	2.25	2.81	2.93
K_2O	0.17	0.16	0.19	0.08	0.02	0.26	0.2	0.35	0.64	0.05	0.04	0.52	0.27
P_2O_5	0.06	0.04	0.05	0.04	0.13	0.04	0.25	0.05	0.01	0.01	0.02	0.12	0.23
mg	0.60	0.64	0.58	0.57	0.50	0.54	0.52	0.53	0.55	0.60	0.59	0.40	0.47
Rb	4	5	5	7	3	8	2	14	44	1	2	9	9
Sr	102	85	10	120	112	108	125	163	125	112	119	99	214
Y	17	15	21	20	18	20	22	21	24	13	16	36	41
Zr	39	44	44	52	67	54	56	54	62	37	30	122	147
Nb	2.3	3	3	3	4	4	3.2	4	4.2	1	2	7	8
Ti	4101	4158	5094	5094	6600	5354	7200	5269	7042	3974	3742	9760	8918
Ba	53	50		50	20		128		158				99
Cr	414	403	247	220	144	319	309	307	366	425	544	69	125
Ni	105	107	156	134	97	93	123	121	175	69	109	41	64
Co	39	46	43	50	58	55		60	49	47	48	40	69
V	236	249	254	287		277		280	331	236	248	363	287
Th					0.1		0.21		4		1		11
La	1.4				3.8		3.4		4		2.4		11
Ce	4.8				8.3		10.5		11		6.1		27
Nd				6.64	5.8		8.5		7.6		4.2	11.4	20
Sm	1.54			2,1	2.0		2.9		2.67		1.36	3.89	5.97
Eu	0.544				0.67		1.1		0.7		0.51		1.85
Gd													
Tb	0.36				0.49		0.58		0.64		0.34		1.3
Tm					0.34		0.28						
Yb	1.4				2.2		1.8		2.4		1.5		4.5
Lu	0.21				0.29		0.24		0.37		0.22		0.57
Ti/Zr	105	95	116	98	99	99	129	98	114	107	125	80	61
Nb/Y	0.14	0.2	2.1	0.15	0.22	0.20	0.15	0.19	0.18	0.08	0.13	0.19	0.20
Zr/Y	2.29	2.9	116	2.60	3.7	2.7	2.5	2.57	2.6	2.85	1.88	3.39	3.6
Nb/La	1.64				1.05		0.95		1.05		0.83		0.73

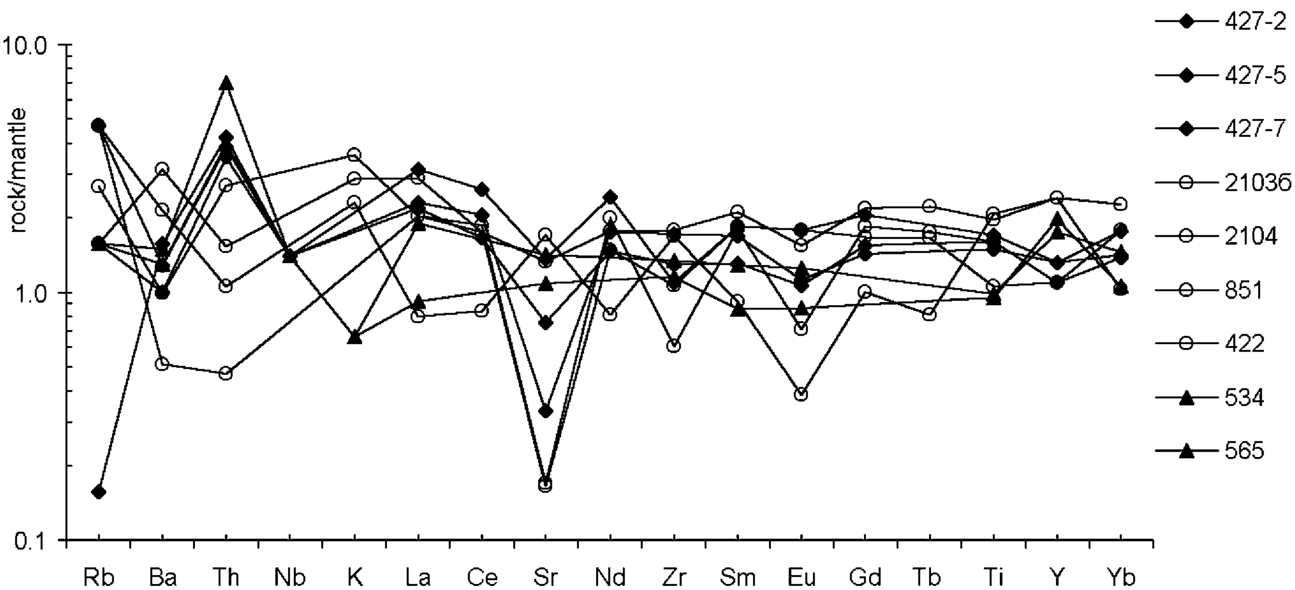

Figure 6. Spidergram for komatiites from the western margin of the Vodlozero domain. Sample numbers on diagrams, as in Table 1.

Lamba belts is modeled assuming 30–50% equilibrium partial melting of amphibolites compositionally similar to island arc basalt. Nd isotope data do not corroborate the presence of ancient (>3.1 Ga) material in the source region [*Chekulaev et al.,* 2002]. Andesitic dacites of the Hautavaara and, partly, Koikary–Semchensky belt are likely derivatives of boninitic liquids generated via melting of the mantle wedge and ancient crustal material [*Matrenichev et al.,* 1990].

The mafic rocks and andesites making up the greenstone belts have been deformed to acquire a high angle pervasive foliation and metamorphosed under the low-T subfacies (T = 500–570°C) of the amphibolite facies, classed with the moderate-P facies series [*Kratz,* 1978]. These early-stage deformations and metamorphism reflect a tectonic process that ultimately resulted, mainly through tangential compression, in spatial juxtaposition (collage) of compositionally diverse rocks originating from a variety of geodynamic settings.

A similar setup, with volcanites of different types being brought together within the same structure, is recorded at the eastern margin of the Vodlozero domain. There, in the Kenozero belt, high-T oceanic plateau rocks (2960±150 Ma) occur in association with coeval island arc rocks.

On the other hand, in the interior of the Vodlozero domain, in its "sialic core," emplaced were gabbronorite–diorite intrusions dated at 2987±11 Ma [*Arestova,* 1997; *Lobach-Zhuchenko et al.,* 1993] and diorite intrusions yielding a 2971±6 Ma age [*Chekulaev et al.,* 1994; *Lobach-Zhuchenko et al.,* 1999], and at the northern margin of the Vodlozero domain vigorous tonalite–granodiorite magmatism took place. This magmatism is dated as 2980±12 Ma by the U–Pb zircon method on tonalite at Lake Chernoe and as 2959±14 Ma on granodiorite from the same locality [*Chekulaev et al.,* 1994; *Lobach-Zhuchenko et al.,* 1999].

At the current erosional surface, the western Karelian domain displays rocks younger than 2.9 Ga. The presence of more ancient sialic rocks as crustal constituents can be deduced from indirect evidence. Thus, Sm–Nd model ages for crustal granites are 3.1–2.9 Ga [*Lobach-Zhuchenko et al.,* 2000b; *O'Brien et al.,* 1993], and for acid volcanites of the Kostomuksha greenstone belt, 3.1–3.3 Ga [*Puchtel et al.,* 1998]. In southeastern Finland, granitoids contain detrital zircons dated at 3027±43 Ma [*Vaasjoki et al.,* 1993], and volcanites of the Suomussalmi belt contain lead with an ancient (3.6 Ga) component [*Vaasjoki and Sakko,* 1991]. An age older than that of the Kostomuksha volcanites has been determined by thermoionic emission on zircon in the vicinity of Voknavolok (2890 Ma) [*Lobach-Zhuchenko et al.,* 2002] and by the U–Pb method on a trondhjemitic gneiss north of Lake Kuito (2887±24 Ma) [*Samsonov et al.,* 2001]. An ancient (>2.9 Ga) age should also be assigned to gneissic tonalites in the vicinity of Lake Tulos, resembling ancient rocks of the Iisalmi region, central Finland, in terms of composition and the character and succession of endogenic processes. Therefore, this region can be viewed as an ancient domain reworked essentially during the formation of greenstone belts and subsequent granitization, metamorphism, and deformations.

Currently, there is no evidence to indicate whether at 3.0–2.92 Ga the ancient western Karelian and Vodlozero domains constituted a single domain, to be separated later, or whether they evolved as independent terranes. The former option, however, is at odds with our own data that suggest that the western part of the Vodlozero domain originated through subduction of oceanic crust to the east, beneath the Vodlozero sialic core, which implies that an ocean existed westward of the core. This is supported by the proto-

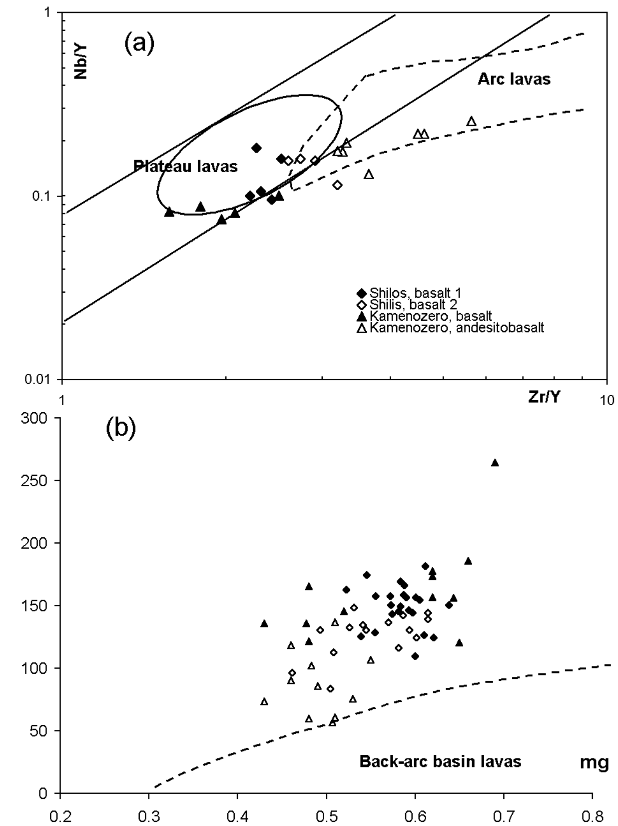

Figure 7. Zr/Y vs. Nb/Y (a) and Mg# vs. Ni (b) diagrams for basalts from the western margin of the Vodlozero domain showing fields for basalts from various geodynamic settings [after *Kerr et al.,* 2000].

ophiolitic assemblage of this age being present in the Oster greenstone structure. Hence, the model invoking independent domains appears more adequate.

In the Kola province, sialic crust of that age was likely in existence as well, but it is poorly preserved because of vigorous endogenic processes between 2.85–2.70 Ga, as recorded by U–Pb dating (Figure 1). The only evidence for the presence of sialic fragments is scanty U–Pb zircon ages from the rocks of the Kola Group and the Kola Superdeep Borehole [2933±54 Ma; *Mitrofanov et al.,* 1997] and from Hompen tonalitic gneisses in northwestern Norway [2902±9 Ma; *Levchenkov et al.,* 1995]. An older age, 2925±6 Ma, has been obtained from gabbro-anorthosite of the Patchemvaraka massif [*Kudryashov and Gavrilenko,* 2000], situated at the boundary shared by the Murmansk block and the Kolmozero–Voronya greenstone belt. The fact that sialic crust was then developed in that area more broadly is evidenced by model Nd ages (T_{Nd}(DM)) for rhyodacite from the Tersky–Allarechka greenstone belt and for rocks of the central Kola Peninsula and Murmansk massif, equaling 2900–2960 Ma [*Timmerman and Daly,* 1995].

2.92–2.83 Ga Phase

During this phase, ancient terranes continued growing and being reworked, and new sialic cores were incepted (Figure 9).

Along the northern margin of the Vodlozero domain, a system of rift-related structures took shape, to eventually evolve into bimodal greenstone belts: Shilos (northern and western parts of the southern Vygozero belt) and Kamennye Ozera [*Lobach-Zhuchenko et al.,* 1999, 2002]. Mafic–ultramafic magmatism at the northern margin of the domain falls in two phases, 2913–2916 Ma [*Lobach-Zhuchenko et al.,* 1999; *Puchtel et al.,* 1999] and 2877–2875 Ma [*Puchtel et al.,* 1999]. Phase 1 mafic volcanites are represented by komatiites and high-temperature basalts and are not associated with any intermediate or acid volcanites. Komatiites are found only in the Kamennye Ozera greenstone belt, being there represented by peridotitic varieties high in MgO, Cr, and Ni (in spinifex textured rocks) (Table 3). Komatiites are LREE depleted ($(La/Yb)_N = 0.6$–0.7), medium and heavy REE concentrations matching primitive mantle abundances (Figure 10). In the komatiites, $(Nb/La)_N = 0.9$–1.0, which is only possible in the absence of melt contamination by crustal material. Geochemical fingerprints of the komatiites point to their genesis from high-T plume-derived melts in oceanic or continental margin rifts.

Tholeiites of the northern margin show a broad compositional spectrum. The Shilos belt displays two groups of basalts (Table 3). Geochemically, both groups are high-T basalts, significant distinctions between them being in terms of Ti and Zr concentrations and degrees of REE enrichment. Group 1 basalts are LREE depleted ($(La/Yb)_N = 0.5$–0.7, $(La/Sm)_N = 0.6$)) and slightly HREE depleted ($(Tb/Yb)_N = 1.2$), their REE abundances being 2.5–3.5 times those of the primitive mantle (Figure 11). Group 2 basalts have $(La/Yb)_N = 1.9$ and $(La/Sm)_N = 1$ and REE abundances 6–8 times those of the primitive mantle. Basalts of both groups lack evidence of crustal contamination, their $(Nb/La)_N$ ratio ranging 0.8–1.5. The early phase tholeiites of the Kamennye Ozera greenstone belt fall in groups with distinctive Mg-numbers (from 0.62 to 0.53) and high Cr and Ni abundances. These tholeiites are LREE

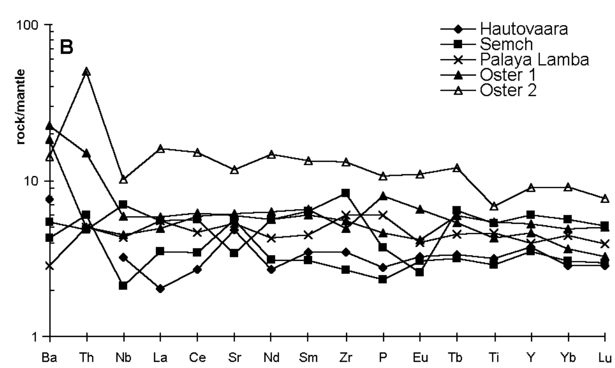

Figure 8. Spidergram for basalts of the western margin of the Vodlozero domain.

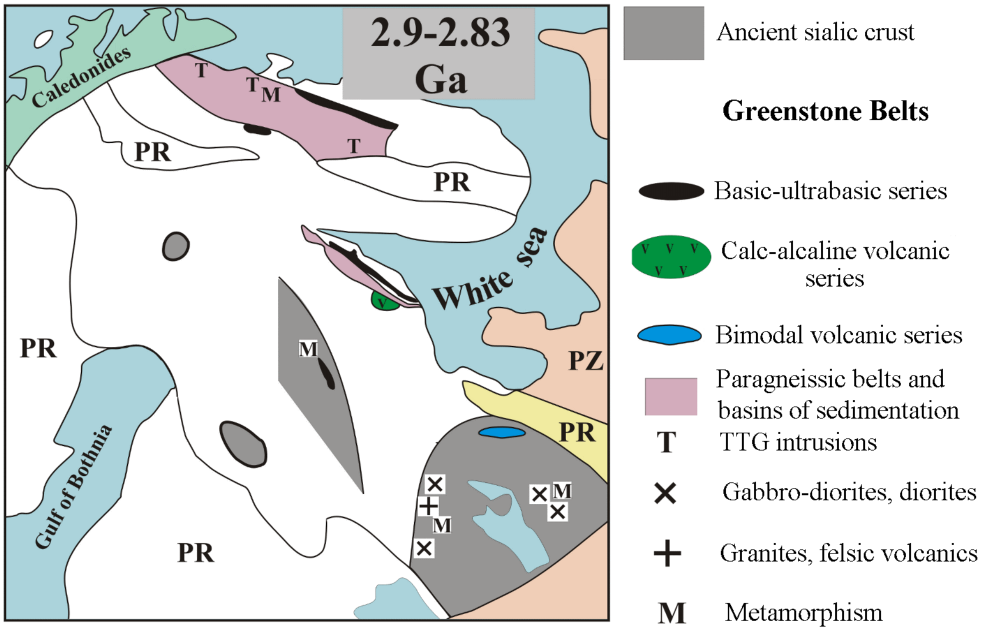

Figure 9. Evolution of ancient segments of sialic crust in the time interval 2.92–2.83 Ga. 1 – crust older than 2.92 Ga; 2–4 – greenstone belts: 2 – with mafic–ultramafic assemblages, 3 – with calc-alkaline volcanism, 4 – with bimodal volcanism; 5 – depositional basins; 6 – tonalite–trondhjemite–granodiorite series intrusions; 7 – gabbro-diorite and diorite intrusions; 8 – granite intrusions; 9 – metamorphism manifested in this time interval.

depleted and show no crustal contamination, their $(Nb/La)_N$ ranging 0.8–1.7

Geochemical characteristics of basalts of both sub-groups of the Shilos belt and early phase basalts of the Kamennye Ozera belt resemble those for present day basalts of oceanic rises or continent margin plateaus (Figure 12). Dissimilarities between the $\varepsilon_{Nd}(t)$ values for isochrons of the Shilos (+1.6) [*Sochevanov et al.,* 1991] and Kamennye Ozera (+2.7) mafites [*Puchtel et al.,* 1999] are due to distinctions between isotope compositions of their parental melts, suggesting separation of these melts from different parts of a heterogeneous source region, which is possible provided the melting took place in a mantle plume head.

The second magmatic phase at the northern margin of the domain, with a ca. 2875 Ma U–Pb zircon age [*Puchtel et al.,* 1999], is represented by contaminated tholeiites (basaltic andesites; Table 3) and rhyolites, interpreted by [*Puchtel et al.,* 1999] to constitute a single basalt–andesite–dacite–rhyolite island arc assemblage, and by adakite series rhyolites. Acid volcanites of the Kamennye Ozera belt and the rhyolitic dike assemblage of the Shilos belt have positive $\varepsilon_{Nd}(t)$ values (+1 to +3) and, hence, the source of the volcanites must have separated from the mantle shortly prior to their generation. Acid rocks of the Kamennye Ozera belt are inferred to have been generated (1) through shallow depth fractionation of andesitic melts in an oceanic

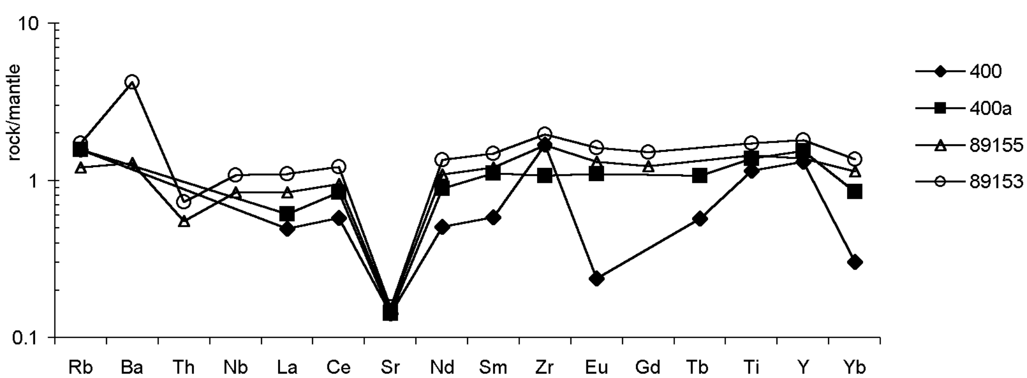

Figure 10. Spidergram for komatiites from the northern margin of the Vodlozero domain. Sample numbers on diagrams, as in Table 3.

Table 3. Geochemical analyses of representative komatiite and basalt samples from the greenstrone belts of Vodlosero domain north margin

	Shilos belt								Kamennye Ozera belt										
	Subgroup 1 basalts				Subgroup 2				komatiites		basalts			basaltic andesites					
Sample no.	41 Ar	60 Ar	5 L-Z	6/12 L-Z	6/2 L-Z	25 Ar	15 Ar	340 Ar	400 Ar	89153	89 Ar	94141	605 Ar	80 Ar	269 Ar	373 Ar	372 Ar	94140
SiO_2	46.38	50.20	47.47	51.27	46.53	50.36	50.07	51.13	45.15	45.90	51.19	51.94	49.79	56.22	54.83	54.06	49.95	51.9
TiO_2	0.77	0.89	0.95	0.67	1.04	1.09	1.03	1.42	0.29	0.37	0.47	0.85	0.79	0.81	0.83	1.38	1.61	1.91
Al_2O_3	16.26	15.78	15.39	15.15	15.68	16.39	15.80	15.15	7.62	7.3	15.67	14.10	15.10	15.30	15.49	13.49	14.53	14.40
FeO	12.25	10.41	11.48	10.18	11.18	11.89	11.19	12.72	9.42	11.43	9.56	10.89	12.06	10.29	10.28	14.40	13.84	12.78
MnO	0.19	0.21	0.19	0.17	0.19	0.20	0.21	0.21	0.28	0.21	0.21	0.19	0.20	0.16	0.16	0.16	0.17	0.18
MgO	9.76	8.11	9.38	8.13	9.16	6.89	6.96	7.26	30.32	27.5	8.75	9.97	7.50	6.41	6.14	6.95	6.48	6.1
CaO	11.26	10.75	12.63	10.26	12.84	9.30	12.09	6.91	5.34	5.8	13.10	10.56	11.04	6.48	9.27	6.29	9.95	10.45
Na_2O	2.41	2.53	1.91	2.42	2.42	2.53	2.09	3.91	0.04	0.01	1.57	1.09	1.66	3.77	2.74	3.07	2.45	2.04
K_2O	0.08	0.25	0.19	0.27	0.31	0.01	0.31	0.51	0.07	0.02	0.04	0.35	0.07	0.03	0.03	0.13	0.17	0.13
P_2O_5	0.06	0.07	0.04	0.04	0.06	0.08	0.08	0.09	0.22	0.07	0.04	0.06	0.06	0.11	0.10	0.04	0.06	0.11
mg	0.59	0.58	0.59	0.59	0.59	0.51	0.53	0.50	0.85	0.81	0.62	0.62	0.53	0.53	0.52	0.46	0.46	0.46
Rb	4	4	9	12	10	7	4	4	1	1.09	< 5	11.6	< 5	6	< 5	< 5	< 5	5
Sr	75	150	122	169	421	159	218	77	3	3.3	103	42	109	87	196	59	124	197
Y	19	21	22	19	20	22	22	27	7	8	13	18	18	18	22	31	34	23
Zr	42	51	50	48	52	64	62	88	12	22	27	44	45	83	124	88	113	84
Nb	2	2	4	3	3.1	3.4	4	4	1	0.76	1.5	1.9	1.8	3.9	5.6		7	10.7
Ti	4030	5150	5434	4652	5637	6624	6110	8245	1740		3165		5284	5074	5586	7319	9660	11400
Ba	< 100	< 100	< 100	< 100					< 30	29.5	< 30	22	< 30	< 30	< 30			14.7
Cr	371	379	374	386	375	281	342	203	2527	3370	471	553	400	165	87	111	200	156
Ni	158	145	146	166	130	112	132	83	1437	1396	173	151	145	75	52	85	90	73
Co										111	60	51	56	37	40	39	40	49
V	298	291	265	236	268	318	280	408		159	262	249	312	313	317	462	311	170
Th										0.062		0.167						2.7
La	0.98	0.73			2.00	4.01			0.42	0.764		2.25	1.080					16.3
Ce	3.2	2.35			6.0	11.0			1.50	2.17		6.33	4.00					37.5
Nd			2.75	3.47	5.73	8.54	9.43	9.63		1.83		5.39						21.5
Sm	1.32	0.87	1.18	1.65	2.05	2.85	3.02	3.37	0.496	0.659		1.800	1.740					5.41
Eu	0.51	0.37			0.69	1.04			0.186	0.271		0.694	0.700					1.59
Gd										0.902		2.44	2.58					5.58
Tb	0.36	0.24			0.53	0.71			0.116				0.40					
Tm																		
Yb	1.3	1.1			2	1.9			0.420	0.675		1.82	1.62					2.51
Ti/Zr	96	101	109	97	108	104	99	94	145		117		117	61	45	83	85	136
Nb/Y	0.11	0.10	0.18	0.16	0.16	0.15	0.18	0.15	0.14	0.10	0.12	0.11	0.10	0.22	0.25		0.21	0.47
Zr/Y	2.21	2.43	2.27	2.53	2.60	2.91	2.82	3.26	1.71	2.75	2.08	2.44	2.50	4.61	5.64	2.84	3.32	3.65
Nb/La	2.04	2.74			1.55	0.85			2.38	0.99		0.84	1.67					0.66

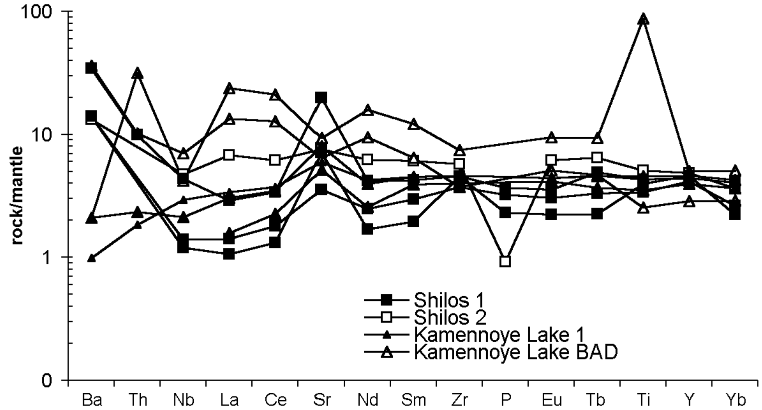

Figure 11. Spidergram for basalts of the northern margin of the Vodlozero domain.

island arc (as with rhyolites of the basalt–andesite–dacite series) and (2) through melting of oceanic tholeiitic basalts of the lower succession followed by fractional crystallization (as with the adakite series) [*Puchtel et al.,* 1999]. However, the bimodality of the volcanic assemblage is rather suggestive of its rift affinity. The Shilos dacites and rhyolites have been proposed to result from melting of enriched basalt under plume impact [*Lobach-Zhuchenko et al.,* 1999]. Continental margin rift affinity for the Shilos and Kamennye Ozera belts is evidenced by the following geologic data. The greenstone belts are located among ancient tonalites and granodiorites, dated at 3130 Ma at the Vyg River [*Lobach-Zhuchenko et al.,* 1993] and at 2980±12 Ma south of the Kamennye Ozera structure [*Lobach-Zhuchenko et al.,* 1999]. West of the Shilos belt, among the early tonalites, basaltic and rhyolitic dikes compositionally similar to the belt volcanites were mapped.

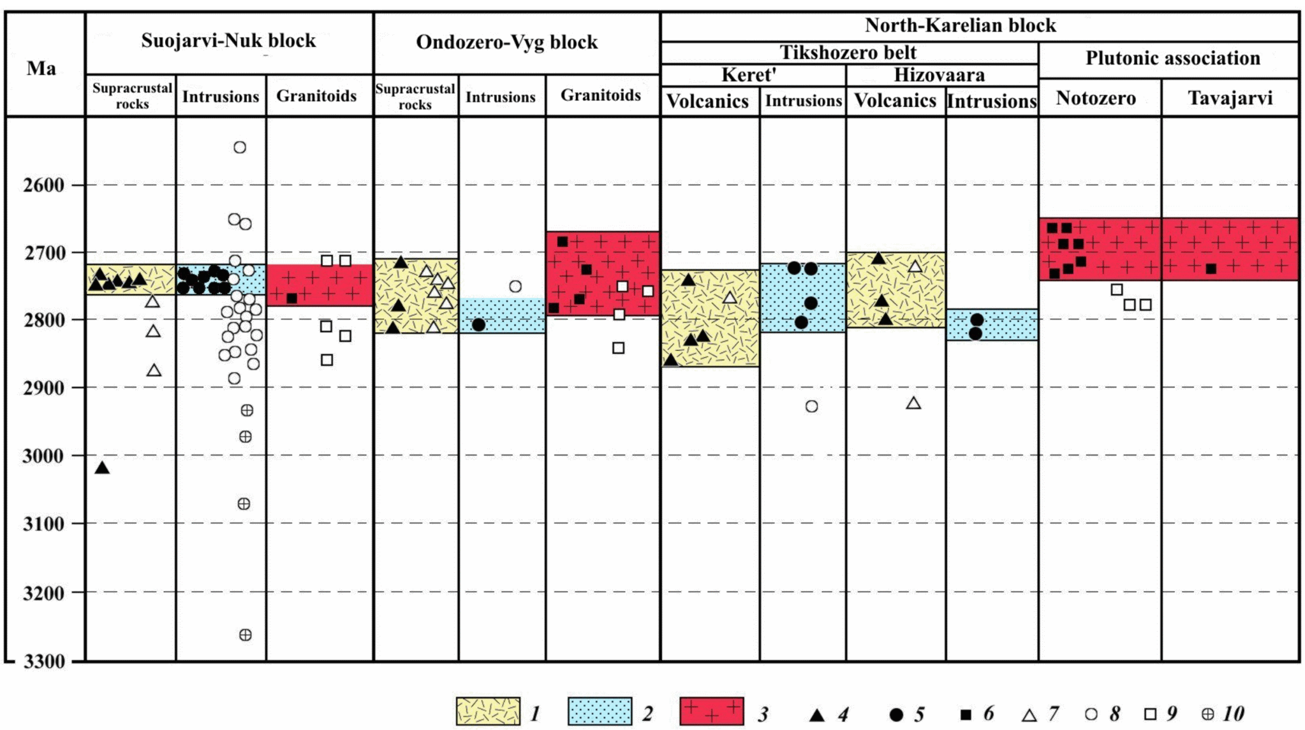

Figure 12. Sketch showing correlation of Archean petrogenesis in the central Karelian domain [*Chekulaev et al.,* 2002]. 1 – supracrustal rocks, 2 – intrusive rocks confined to greenstone belts, 3 – areally extensive granitoids. U–Pb zircon ages for (4) supracrustal rocks, (5) intrusive rocks associated with greenstone belts, (6) areally extensive plutonic rocks. T_{Nd}(DM) model ages for (7) supracrustal rocks, (8) intrusive rocks associated with greenstone belts, (9) areally extensive granitoids. 10 – model ages for rocks marking the domain's boundaries.

Dikes composed of amphibolites chemically similar to volcanites of the belts have also been mapped in the interior of the Vodlozero domain [*Lobach-Zhuchenko et al.,* 1999]. Their age, equaling 2860±90 Ma, was established at the Vodla River [*Sergeev et al.,* 1990].

Coevally to volcanism in rift-related structures of the northern margin of the Vodlozero domain, vigorous magmatism took place throughout the domain (Figures 7, 3 and 4). In the western accretionary structure, the second magmatic phase occurred, involving acid and intermediate volcanism, as well as mafic, dioritic, tonalitic, and trondhjemitic intrusions, and, for the first time in the shield's geologic history, two-feldspar granite intrusions [*Kovalenko,* 2000]. In the domain's interior, more ancient granitoids and gneisses were intruded by granodiorites and tonalites with Nd isotope earmarks pointing to contamination by a more ancient crustal material (Figure 3). Magmatism was accompanied by heavy deformations and amphibolite facies metamorphism [*Lobach-Zhuchenko et al.,* 2002]. Based on geologic setting, spatial distribution, and isotope geochemical characteristics of the rocks just mentioned, generation of mantle and crustal melts during this phase is best explained assuming that a rising mantle plume resulted in high-T mafic melts being emplaced in the crust, to induce crustal melting.

Therefore, magmatic and tectonic processes that occurred during the above three phases between 3.2–2.83 Ga gave rise, on the southeast of the shield, to the first relatively stable and large (over 60,000 km^2) feature of the Fennoscandian shield—the Vodlozero domain.

Within the western Karelian domain, at its eastern margin and in its central part, in the end of this time interval rift-related structures came into being, their related bimodal volcanism (with a considerable predominance of mafic and ultramafic volcanites) giving rise to the Kostomuksha and Tipasjarvi–Kuhmo–Suomussalmi greenstone belts. Generation of the lower, mafic section of the Kostomuksha belt is timed at 2843±39 Ma [*Puchtel et al.,* 1997]. Crustal extension zones, expressed by rift-related greenstone belts, were complemented by high-grade metamorphism and strong isoclinal folding in-between the belts. These endogenic processes occurred between 2858–2837 Ma, marking the second phase of deformations and metamorphism in the tonalitic gneiss assemblage [*Luukkonen,* 1985].

The Kola province provided the stage for initiation of rift-related greenstone belts (e.g., the Kolmozero–Voronya belt) and sequential formation of their related komatiite–tholeiite (2.92–2.87 Ga) and basalt–andesite–dacite (2.88–2.79 Ga) series. These belts in their present day form, just as the marginal belt of the Vodlozero domain, apparently result from tectonic juxtaposition of rock assemblages generated in a variety of geodynamic settings. Thus, petrologic and geochemical characteristics of the komatiite magmatism of this phase point to its origin in the most high-T and "primitive" (in terms of its isotope geochemical composition) axial portion of a mantle plume [*Vrevsky,* 2000], whereas the basalt–andesite–dacite series of the greenstone belts show affinity to modern island arc systems. In the central Kola Peninsula, greenstone belts came into being coevally to structural and metamorphic reworking of supracrustal rocks of the Kola Group, which involved zoned granulite–amphibolite metamorphism and tonalite–trondhjemite plutonism, dated at 2902–2835 Ma [*Mitrofanov et al.,* 1997]. The rocks of a similar plutonic suite of the Murmansk massif yielded a series of Nd model ages in the range 2.86–2.9 Ga, constraining the lower age of formation of this crustal block [*Timmerman and Daly,* 1995].

During this phase of evolution of the study area, the space between the Kola sialic segment and the western Karelian and Vodlozero terranes was likely occupied by ocean. In particular, the "mafic zones" of the Belomorian foldbelt, composed of amphibolites after volcanic rocks and (partly) probably after gabbroic dikes and large harzburgitic lenses, clearly have features in common with proto-ophiolites [*Lobach-Zhuchenko et al.,* 1998; *Stepanov,* 1981; *Stepanov and Slabunov,* 1989]. Dacites occurring in the mafic zone are dated at 2878±13 Ma [*Bibikova et al.,* 1999].

At ca. 2.85 Ga, the zone separating the Kola and Karelian segments accommodated a large sedimentary basin, where the Chupa Group graywackes were deposited, and volcanism took place. Dacites alternating with the Chupa aluminous gneisses are dated at 2850±30 Ma, and the first metamorphic event involving the Chupa gneisses is timed at 2820±20 Ma [*Bibikova et al.,* 2001], which implies that this event was roughly coeval to metamorphism affecting the Kola Group rocks. Detrital zircon cores from the Chupa gneisses in the vicinity of Lake Pulangskoe have been dated between 2900–3000 Ma [*Bibikiva et al.,* 2001], which defines the lower depositional age. The age of acid volcanism in the Chupa Group is coeval to the formation for the Keretozero Group of the Keretsky greenstone belt, situated at the boundary between the Karelian segment and Belomorian belt [*Slabunov,* 1993] (Figure 6). Volcanites of the Keretsky greenstone belt have been dated at 2877±45 Ma (U–Pb zircon age from andesitic metatuff) and at 2829±30 Ma (neck facies andesite) [*Bibikova et al.,* 1999]. Composition of volcanites of the Keretsky belt is indicative of their generation in an island arc type setting [*Slabunov,* 1993].

Therefore, by 2.8 Ga, the principal Archean domains of the Baltic shield (Vodlozero, western Karelian, Kola, and Belomorian) had been formed. At ca. 2800 Ma, the area separating these domains was the locus of formation of island arcs that propagated in a southwesterly direction, to form the northern Karelian system of greenstone belts (Figure 8; the north Karelian segment), and in a southerly direction (the Suojarvi–Nyukozero and Ondozero–Vygozero segments of the central Karelian domain; Figure 8). Geochronologic and isotope data listed in Figure 8 clearly indicate a younger (2.85–2.70 Ga) age for the continental crust of this region, where the western Karelian, Vodlozero, and Belomorian–Kola ancient terranes are joined together.

Conclusions

Archean continental crust of the Fennoscandian shield was generated both by progressive addition of sialic crust over time (chiefly at convergent boundaries of ancient plates) and through reworking of ancient fragments with input of juvenile material derived from rising mantle plumes. The latter process was accompanied by redistribution of material in the crust, with rocks of granitoid composition being transferred to shallower crustal levels. Assumedly, these processes were controlled by two principal mechanisms. The formation of accretionary and collisional orogens and their coeval magmatism at active margins of ancient plates were driven by subduction (in the context of the plate tectonic mechanism). Generation of rift-related structures and associated magmatism apparently resulted from rising mantle plumes. Brittle deformations involving rifting and related to the ascent of mantle plumes did not lead to break-up of young continental crust. The main outcome of this mechanism was a voluminous input of high-T magma in the lithosphere, to form within-crust layers of "asthenosphere," where granite melts were produced.

It remains open to discussion whether the episodes of mafic magmatism at 2.99, 2.96–2.94, and 2.92 (2.91) Ga were induced by different pulses of the same deep-seated mantle plume (a first-order plume, or superplume) or whether this magmatism was driven by two different plumes. The 2.92 Ga episode of mafic magmatism might be due to a mantle plume rising immediately after the subduction event at 3.0–2.94 Ga; in this particular case, the plume may have originated from a shallower depth, at the lower mantle/upper mantle boundary. This inference is suggested by the fact that mafic magmatism with an age of 2.92 Ga at the northern margin of the Vodlozero domain immediately postdated the cessation of subduction processes at the domain's western margin. The ascent of another mantle plume may have been triggered by a mantle slab sinking to the D' discontinuity (the lower mantle/upper mantle boundary).

The higher geothermal gradient in the Archean must have been due to rising mantle plumes. The interplay of two mechanisms, plume and plate tectonics, in the Archean was responsible for the diversity of rock assemblages on the Fennoscandian shield. Either of the mechanisms operated during ca. 50–80 m.y.

Acknowledgments. This work was supported by the Russian Foundation for Basic Research (project nos. 02-05-65052 and 01-05-64930).

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(Received 16 January 2003)