Petrology of mafic lavas within the Onega plateau, central Karelia: evidence for 2.0 Ga plume-related continental crustal growth in the Baltic Shield

Article *in* Contributions to Mineralogy and Petrology · January 1998

DOI: 10.1007/s004100050355 · Source: OAI

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Contrib Mineral Petrol (1998) 130: 134–153

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Petrology of mafic lavas within the Onega plateau, central Karelia: evidence for 2.0 Ga plume-related continental crustal growth in the Baltic Shield

Received: 17 June 1997 / Accepted: 24 September 1997

Abstract The Onega plateau constitutes part of a vast continental flood basalt province in the SE Baltic Shield. It consists of Jatulian-Ludikovian submarine volcanic, volcaniclastic and sedimentary sequences attaining in places 4.5 km in thickness. The parental magmas of the lavas contained ~10% MgO and were derived from melts generated in the garnet stability field at depths 80–100 km. The Sm-Nd mineral and Pb-Pb whole-rock isochron ages of 1975 ± 24 and 1980 ± 57 Ma for the upper part of the plateau and a SHRIMP U-Pb zircon age of 1976 ± 9 Ma for its lower part imply the formation of the entire sequence within a short time span. These ages coincide with those of picrites in the Pechenga-Imandra belt (the Kola Peninsula) and komatiites and basalts in the Karasjok-Kittilä belt (Norway and Finnmark). Together with lithostratigraphic, chemical and isotope evidence, these ages suggest the derivation of the three provinces from a single large (~2000 km in diameter) mantle plume. These plume-generated magmas covered ~600,000 km^2 of the Baltic Shield and represent a major contribution of juvenile material to the existing continental crust at 2.0 Ga. The uppermost Onega plateau lavas have high $(Nb/Th)_N = 1.4–2.4$, $(Nb/La)_N = 1.1–1.3$, positive $\varepsilon Nd(T)$ of $+3.2$ and unradiogenic Pb-isotope composition ($\mu_1 = 8.57$), comparable with those of modern oceanic plume-derived magmas (oceanic flood basalt and ocean island basalt). These parameters are regarded as source characteristics. The lower sequences have $(Nb/Th)_N = 0.58–1.2$, $(Nb/La)_N = 0.52–0.88$ and $\varepsilon Nd(T) = -2.6$. They have experienced mixing with 10–30% of continental crust and resemble contaminated lavas from other continental flood basalt provinces. The estimated Nb/U ratios of 53 ± 4 in the uncontaminated rocks are similar to those found in the modern mantle (~47) suggesting that by 2.0 Ga a volume of continental crust similar to the present-day value already existed.

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Editorial responsibility: J. Hoefs

Introduction

Two major processes are responsible for the addition of new juvenile material to the continental crust. First, it forms *continuously* at island arcs through the subduction of oceanic plates into the mantle (Taylor and McLennan 1985). The second process involves *periodic* delivery of new material by mantle plumes. These generate large amounts of basaltic and komatiitic melts, which erupt very rapidly and create vast volcanic plateaux in the oceans and on the continents. Oceanic plateaux are accreted laterally to the continents by the processes of plate tectonics, thus becoming new fragments of continental crust (Ben-Avraham et al. 1981). Mantle plumes rising beneath the continents create continental flood basalt provinces and thus contribute to the vertical growth of continental crust (McKenzie and Bickle 1988).

Sometimes outpourings of lava onto the continental crust are volumetrically in excess of any contemporary volcanic processes. For instance, the abundances of magmatic products of the Permo-Triassic Siberian Traps, the largest Phanaerozoic flood basalt province, exceeded 1.5×10^6 km^3. These lavas were erupted over a period of ~600,000 years and now cover an area of $\sim 3.4 \times 10^5$ km^2 (Czamanske et al. 1992; Renne and Basu 1991).

Most known continental flood basalt provinces range in age between 15–250 Ma. However, they have several Proterozoic equivalents such as the 1.1 Ga Keweenawan basalts of the Lake Superior district (Lightfoot et al. 1991), and the 1.3 Ga Coppermine River basalts in Northwest Territories (Dupuy et al. 1992; Griselin et al. 1997). Rarer Archaean examples include the 2.7 Ga Fortescue basalts in Australia and Ventersdorp volcanics of South Africa (Nelson et al. 1992). Here we report data for 2.0 Ga basalts and related rocks of the Onega plateau in the SE Baltic Shield, which represent a new temporal group in the history of continental flood volcanism. The main objectives of the study include: (1) detailed field studies of the lavas in order to deduce the emplacement environment; (2) geochemical and isotope-geochronological studies of whole-rock samples and mineral separates from volcanic and plutonic rocks in order to deduce the timing of their formation, the petrogenetic mechanisms responsible for the generation and evolution of the parental magmas and to correlate these rocks with the other Jatulian-Ludikovian sequences all over the Baltic Shield; (3) evaluation of contributions from different sources to the petrogenesis of the rocks; (4) place some constraints on the timing of continental crust formation in the region.

Regional geology

The Baltic Shield exhibits a large number of temporal groups of mafic rocks related to different tectonic settings and has long been regarded as a type region for Palaeoproterozoic plate tectonics (Gaál and Gorbatschev 1987). The shield is divided into three main domains, the Caledonian, Svecofennian, and Archaean Domain. The latter includes the Karelian, Belomorian and Kola Peninsula Provinces (Gorbatschev and Bogdanova 1993). The Karelian Province is a typical Archaean gneiss-greenstone terrain. It occupies a total area of ~350,000 km^2 in the SE Baltic Shield (Fig. 1) and is covered by Palaeoproterozoic epicontinental basins and volcano-sedimentary belts. The Palaeoproterozoic rocks are generally divided into several informal stratigraphic units: the Sumian-Sariolian, Jatulian, and Kalevian (Laajoki 1986; Gáal and Gorbatschev 1987). In Russian Karelia, the Jatulian is overlain by the Ludikovian and Vepsian rocks (Sokolov 1987).

The Jatulian and Ludikovian rocks show a wide areal distribution all over the Baltic Shield (black areas within the inset map in Fig. 1). They form a number of epicontinental basins representing spatially separated relicts of a once continuous cover deposited over the Sumian-Sariolian rocks and the Archaean basement. Among the others these relicts include the Karasjok-Kittilä volcanic belt and its satellites with abundant ultramafic lavas and volcaniclastic lithologies (Often 1985; Saverikko 1985; Barnes and Often 1990), basaltic lavas and tuffs, shallow marine quartzitic sandstones, arkoses and stromatolitic dolomites in Norway and northern Finland (Pharaoh and Pearce 1984; Pharaoh et al. 1987) and the Pechenga-Imandra belt in the Kola Peninsula containing picrites and basalts (Hanski and Smolkin 1989; Hanski 1992). In the Karelian Province, the Jatulian-Ludikovian sequences include marine associations of dolomites, black shales, sandstones, volcanic ash and basaltic lavas of the Onega plateau and several smaller basins extending to the NW. These sequences attain thicknesses of 4500 m and cover a total area of > 50,000 km^2 at the present erosion level (Svetov 1979; Sokolov 1987). Uneroded sequences were probably thicker and areally more extensive. The presence of huge amounts of lava erupted led several workers (e.g. Eskola 1963; Pharaoh and Pearce 1984) to propose the existence of the so-called "Jatulian continent", which was postulated by these authors to occupy at that time the Karelian, Belomorian and Kola Provinces altogether.

The available geochronological data for the Jatulian of the Baltic Shield are rather controversial. There are quite a few dates for 2.2 Ga mafic layered sills cutting Jatulian quartzites all over the Karelian province, i.e. in northern Karelia, Kainuu, Kuusamo, Peräpohja and central Lapland (e.g. Huhma et al. 1990; Nykänen et al. 1994; Vuollo and Piirainen 1992). This means that a rather large part of the Jatulian metasediments and some of the mafic lavas are older than 2.2 Ga. Younger (~2.1 Ga) dike generations have also been recognised (Vuollo et al. 1992), and these can also be used to constrain the age of supracrustal rocks. However, direct datings of Jatulian volcanic rocks are so far too limited. Huhma et al. (1990) reported a Sm-Nd isochron age of 2090 ± 70 Ma for basaltic metavolcanics from the Peräpohja area. A felsic lava has been dated in the Kuopio area, yielding a precise U-Pb zircon age of 2062 ± 2 Ma. In addition, a basic feeding dike was dated at 2115 ± 6 Ma in the Kiihtelysvaara area, Finnish northern Karelia (Pekkarinen and Lukkarinen 1991). The Pechenga-Imandra-Varzuga belt in the Kola Peninsula is thought to have formed over a large time interval between 2300–1980 Ma (Balashov et al. 1993; Smolkin 1997). On the other hand, reliable isotope data are available only for the upper part of the belt within the Pechenga synform (1977 ± 52 Ma, Sm-Nd mineral isochron, 1988 ± 39 Ma, Pb-Pb whole-rock isochron, and 1970 ± 5 Ma, U-Pb single-fraction zircon data (Hanski et al. 1990; Hanski 1992). It is very probable that at least the upper 3000 m thick mafic-ultramafic part of the sequence (the Pilgujärvi Formation) accumulated within a relatively short period of time at ~1.98 Ga. In the Kittilä greenstone belt, clinopyroxene separates and whole-rock samples from mafic lavas define a Sm-Nd isochron with an age of 1990 ± 35 Ma and an initial εNd value of +3.7 ±0.2 (Hanski et al. 1997). Krill et al. (1985) published a Sm-Nd isochron age of 2085 ±85 Ma for a komatiite rock suite from the Karasjok greenstone belt, which adjoins the Kittilä greenstone belt in the north. This suite contains both LREE (light rare earth element) depleted and LREE-enriched lavas, and the meaning of this age is somewhat problematic because of the combination of rocks having probably different initial Nd-isotope composition. It is very likely that these komatiites erupted simultaneously with the emplacement of the mafic lavas within the Kittilä belt.

Onega plateau

The Onega plateau proper represents the largest intracratonic basin within the Karelian Province and occupies a total area of ~35,000 km^2 (Fig. 1). The Jatulian and Ludikovian volcanic rocks within the Onega plateau, which are the subject of this study, were considered by previous workers as representing continental flood basalts, on the basis of their extensive areal distribution and large thicknesses (Svetov 1979; Sokolov 1987). In the southern and northern outskirts of the village of Hirvas, some 90 km to the NW of Petrozavodsk (Fig. 1), the base of the Jatulian sequences is represented by a single 27 m thick lava sheet. It is overlain by a layer of rhythmically bedded pebbly quartz conglomerates, gravellites and shallow-water sandstones some 10 m thick. This grades upwards into a sequence of mafic tuffs and pillow and massive lava flows with well preserved pahoehoe structures, which are indicative of subaerial emplacement. The lava pile is intruded by abundant mafic sills (e.g. the Koikarsky sill). The total thickness of the

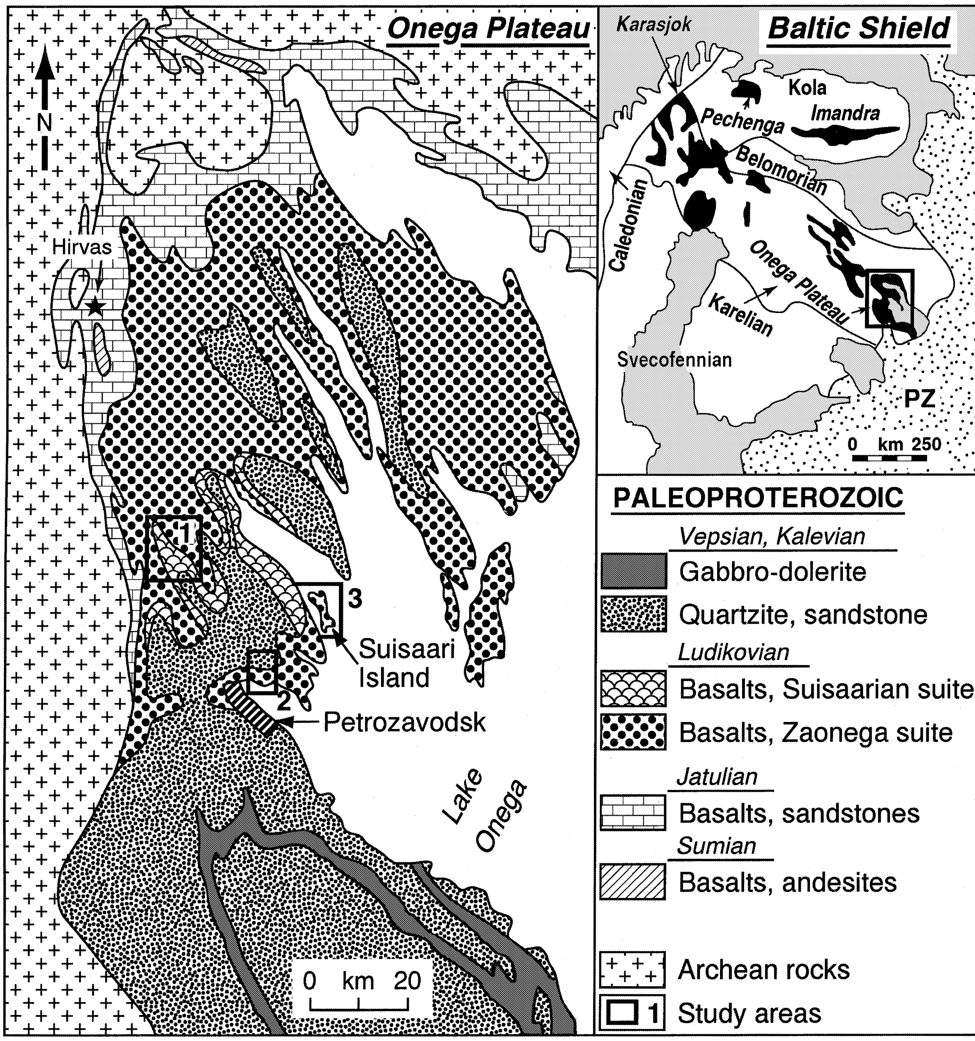

Fig. 1 Geological sketch map of the Onega plateau. *Black areas* in the *inset* map reflect the distribution of the Jatulian and Ludikovian rocks

Jatulian sequences in the area attains ~700 m. For SHRIMP U-Pb zircon studies, sample #9118 was collected from a coarse-grained leucocratic lens at the top of the single lowermost sheet. It consists of an upper part with unevenly distributed amygdaloidal layers and a massive lower part.

The Ludikovian rocks, which were separated from the Jatulian by Sokolov (1987), constitute the lower Zaonega suite and the upper Suisaarian suite. The Zaonega suite has a thickness of up to 1800 m and is composed of submarine volcanic, volcaniclastic and sedimentary sequences including clay-carbonate rocks, dolomites, shungite-bearing basalt tuffs and tuffites, massive, pillow and variolitic basalts. The Suisaarian suite is up to 700 m thick and is made up of volcanic rocks accumulated at shallow-water depths (basalt tuffs and tuff-conglomerates, hyaloclastites, pillow and variolitic basalts). The rocks were metamorphosed under the prehnite-pumpellyite to greenschist facies conditions. The two suites were sampled within three reference areas: (1) Angozero; (2) Solomennoe; (3) Suisaari Island (Figs. 1–3).

The Angozero area

This area is located ~40 km NW of Petrozavodsk and is composed of the Zaonega and Suisaarian suite rocks. The former are represented by Cpx (clinopyroxene)-Pl (plagioclase)-phyric basalts and tuffs. These are intruded by the Konchozero peridotite sill. The Suisaarian suite is divided into four volcanic units; each of them shows regular facies changes outwards from volcanic centres. The lowermost unit comprises tuffites and tuffs intercalated with Cpx-Ol (olivine)-porphyritic basalt lava flows. The second unit consists of Pl-phyric basaltic lavas and tuffs. In the third unit, Cpx-Ol-porphyritic basalts are predominant, while Cpx-Pl-phyric basalts are the main constituent of the uppermost unit. The Cpx-Ol-porphyritic basalts from the lower unit were found to contain wallrock xenoliths, mainly metamorphosed schists. The total thickness of the suite in the area exceeds 420 m. Our studies include the Cpx-Ol-porphyritic basalts of the first (sample 89103) and the third (samples 89104, 89105, 105/1) units and the Konchozero sill, which ranges in thickness from several tens of metres in the flanks up to 200 m in the centre (Figs. 2 and 3). Samples from the sill represent a section from the lower (samples 9111 and 9112) and middle peridotites (sample 9112/3) to the gabbro from the upper chill (samples 9112/1 and 9112/2). Detailed petrographic and mineralogical information for the sill and the lavas can be found in Puchtel et al. (1995).

The Solomennoe area

This is located in the northern part of the Petrozavodsk Bay of Lake Onega (Fig. 1). The volcanic sequence is composed entirely of the Zaonega suite rocks and is divided into five lava units. The lowermost unit is ~50 m thick. It comprises four to five basalt lava flows interbedded with shungite-bearing metasediments. The

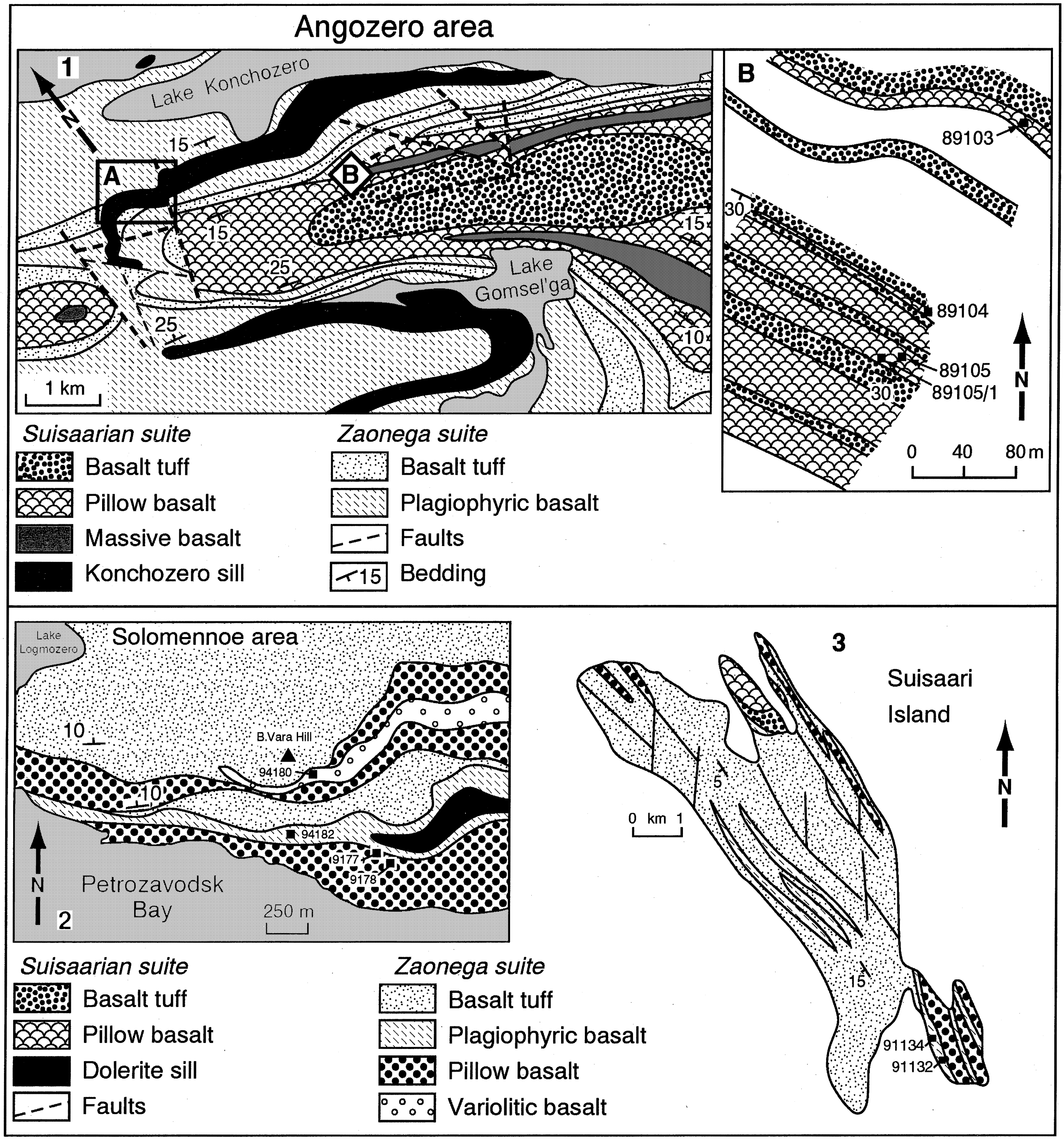

Fig. 2 Sketch maps of the study areas

second unit is up to 30 m thick and consists of two to three lava flows including Pl-phyric basaltic andesite. Upper parts of the lava flows are pillowed; single pillows attain 2 m in diameter and contain small rounded quartz amygdules at the top and elongated pencil-like ones at the bottom oriented subperpendicular to the flow boundaries. Some pillows include central drainage tubes filled with quartz and/or calcite. One sample was collected from the plagioclase-phyric lavas (94182) and the other two from the pillow-textured amygdaloidal lavas (9177 and 9178). The third unit is ~40 m thick and consists of interlayered horizons of agglomerate tuffs and siliceous mudstone. The fourth unit is composed of variolitic pillow basalt represented by sample 94180. The uppermost unit comprises shallow-water agglutinate and agglomerate basalt tuffs and lava breccias. The total documented thickness of the sequence in the area above the Onega Lake level is ~400 m.

The Island of Suisaari

This is composed of the Zaonega and Suisaarian suite volcanic rocks, but the former are predominant. They are represented by

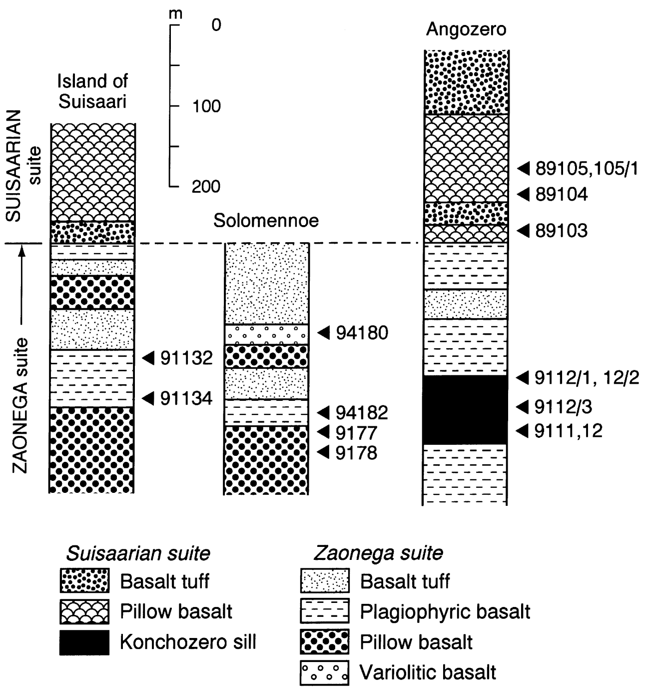

Fig. 3 Stratigraphic profiles through the three study areas

intercalated massive, pillow and amygdaloidal basalt lavas, hyaloclastites and tuffs. The SE part of the island is made up essentially of Pl-phyric basalts, which can be correlated with those in the Solomennoe area. Two samples (91132 and 91134) were collected from these basalts. The Suisaarian suite is restricted to the northern part of the island. The base is composed of tuffite, which gives way to Ol-Cpx-porphyritic basalt lavas and tuffs up to 200 m thick. These rocks can be correlated with Ol-Cpx porphyritic basalts from the Angozero area. The total thickness of the volcanic pile is around 1000 m.

Analytical techniques

Major element abundances were determined by X-ray fluorescence on fused glass discs using a Philips PW-1404 spectrometer at the University of Mainz. The trace elements Cr, V, Co, and Ni were analysed on pressed powder pellets on the same XRF machine. The REE and the rest of the trace elements were determined on an upgraded PlasmaQuad PQl ICP-MS (inductively coupled plasma-mass-spectrometer) at the University of Kiel following the method outlined by Garbe-Schönberg (1993). The REE concentrations in the clinopyroxene separates and in the Angozero rocks were determined at the Institute of Ore Deposit Geology (IGEM) in Moscow by ID-TIMS (thermal ionisation mass spectrometry) using the method described by Zhuravlev et al. (1989). Accuracies are estimated as follows: major elements – 1%; trace elements – 5% for the XRF and the ICP-MS analyses. For the ID-TIMS measurements of REE, the accuracy was 2% or better.

The Sm-Nd-isotope investigations were carried out at the Max-Planck-Institut (MPI) für Chemie in Mainz. Both Nd and Sm were run on a MAT-261 Finnigan mass-spectrometer under static mode. The effects of fractionation during Nd runs were eliminated by normalizing to $^{146}Nd/^{144}Nd = 0.7219$. The $^{143}Nd/^{144}Nd$ isotope ratio measurements of the La Jolla standard during the period of data collection yielded a value of 0.511854 ± 18 ($2\sigma_{pop.}$; $n = 30$). The reported analytical uncertainties used in the regression calculations were 0.2% for $^{147}Sm/^{144}Nd$ and for $^{143}Nd/^{144}Nd$, the $2\sigma_{pop.}$ value for the La Jolla standard corresponding to its external reproducibility.

The Pb-isotope data were obtained at the MPI für Chemie using a MAT-261 Finnigan mass-spectrometer and following the method described by Dupré and Arndt (1990). The total blank was ~100 pg and is negligible compared with the Pb concentrations in the rocks (0.8–2 ppm). Error input was determined by mass fractionation of 0.04% per atomic mass unit. To calculate the age, $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ data were regressed using the York (1969) procedure, the ISOPLOT program (Ludwig 1992) and assuming an error correlation coefficient of 0.9. All errors on age and initial isotopic ratios are quoted at 2σ.

Single zircon grains from the picrobasalt sample 89103 were analysed at the MPI für Chemie, using the evaporation method of Kober (1986). Laboratory procedures as well as comparisons with conventional and ion-microprobe zircon dating are published elsewhere (Kröner and Todt 1988). Several single zircon grains from the Jatulian basalt sample #9118 were analysed by SHRIMP (sensitive high resolution ion microprobe) at the Curtin University in Perth applying the analytical procedures similar to those described by Compston et al. (1984). The Sri Lankan gem zircon CZ-3 used as a standard has been described by Pidgeon et al. (1994). The age of this zircon, determined by conventional analysis, is 564 Ma and the concentration of U is 530–560 ppm.

Results

Clinopyroxene compositions

The REE data for clinopyroxenes from the peridotites and the basalt sample 89104 are listed in Table 1 and plotted on Fig. 4. The pyroxenes have similar REE patterns and nearly identical total levels of rare earth element abundances.

Rock compositions and element mobility

As with all metamorphosed rocks, the geochemical data must be treated with caution because of the potential problem of element mobility. In general, establishing the behaviour of elements and their mobility during secondary alteration processes is more straightforward in komatiites than in basalts because of a greater number of possible liquidus minerals in basaltic liquids and the substantial influence of *P-T* conditions upon the composition of a fractionating mineral assemblage. In the

Table 1 Rare earth element data for clinopyroxene. The concentrations were determined at IGEM in Moscow by ID-TIMS using the method described by Zhuravlev et al. (1989)

Sample	9112	9112/3	89104	Average
La	1.13	1.40	1.19	1.24
Ce	4.26	4.76	4.45	4.49
Nd	5.33	5.57	5.40	5.43
Sm	1.99	2.06	1.93	1.99
Eu	0.633	0.693	0.621	0.65
Gd	2.42	2.50	2.32	2.41
Dy	2.28	2.37	2.25	2.30
Er	1.15	1.20	1.14	1.16
Yb	0.863	0.909	0.887	0.89
$(La/Sm)_N$	0.358	0.428	0.388	0.392
$(Gd/Yb)_N$	2.27	2.23	2.12	2.20

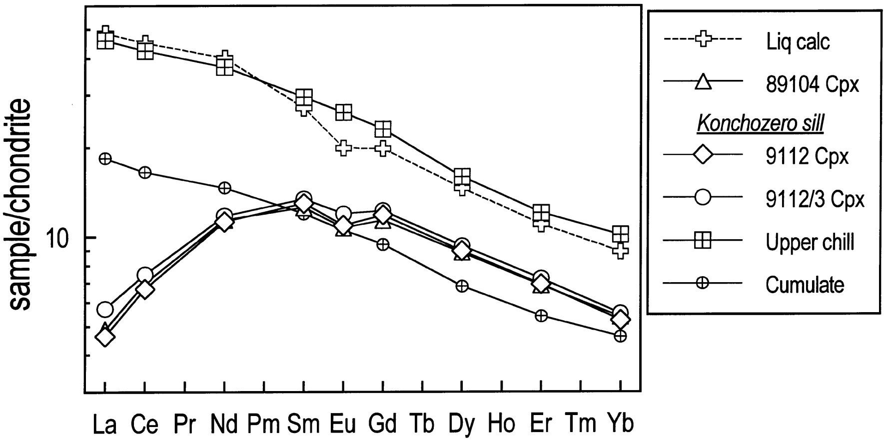

Fig. 4 Chondrite-normalised (Evensen et al. 1978) REE abundances in clinopyroxenes from the Konchozero sill peridotites and picrobasalt sample 89104

present case, however, we were able to solve this problem owing to the relatively low degrees of secondary alteration of the rocks studied, which resulted in either complete preservation of a primary liquidus phase (clinopyroxene) or its pseudomorphic replacement (olivine), thereby allowing the reconstruction of the magmatic mineral association. From the available data we conclude that the major and trace elements define trends consistent with the primary mineralogy, except for the alkalis, Rb, Sr, and Ba. The CaO and SiO_2 may also be disturbed due to calcite and quartz formation during alteration, and this mobility could be responsible for the scatter of CaO in the Angozero lava samples.

The Angozero lavas have elevated MgO concentrations (14–21%); Mg-numbers [mg# = Mg/(Mg + Fe)] range between 0.65–0.75. Using the TAS diagram (Le Bas and Streckeisen 1991) they can be classified as basalts on the basis of the SiO_2 contents. The Zaonega suite lavas from the Solomennoe and Suisaari areas have MgO contents lower than ~7% (mg# = 0.39–0.53) and correspond to basalt, basaltic andesite and dacite. These samples show a positive slope in the diagrams CaO, Sc, and TiO_2 versus MgO, which can be accounted for by clinopyroxene removal.

Chemical compositions of the rocks are listed in Table 2 and plotted on Figs. 5 and 6. The analytical data for the Angozero lavas and the Konchozero sill rocks form two separate linear trends that coincide only for Al_2O_3 and Pb; for most other components, the trends are distinct (Fig. 5). In the Angozero lavas, the data define a trend with a positive slope in the diagram MgO versus CaO. The observed change in Ca/Al ratios is only partly consistent with clinopyroxene fractionation; the low CaO in the lavas compared with common basaltic rocks can be explained by some loss of CaO during secondary alteration. For the other components but Sc, the basalts display relatively steep trends which intersect the MgO axis between 25 and 30%. Although these trends could be interpreted as the results of fractionation or accumulation of olivine plus clinopyroxene, the large variation in the intercepts and the changes in trace element ratios (see below) suggest that the process was more complicated. It seems likely that the compositions of these rocks were also affected by variations in conditions of mantle melting and perhaps mantle source compositions.

For the Konchozero sill there are generally two groups of samples that define two-point trends. Although it is difficult to reach firm conclusions with such a limited data set, we can use these trends to infer the types and compositions of minerals that crystallised. It is significant that the intercepts on the MgO axis for elements excluded from clinopyroxene (e.g. Al, Ti, Nb, Th, Pb) are close to 40% MgO, whereas CaO and Sc give much higher intercepts. We calculate that the fractionating assemblage consists of olivine and clinopyroxene in the proportions 80:20. On the basis of these considerations we try to infer the composition of fractionating olivine and the composition of the parental magma to the Konchozero sill. For our calculations we use only the elements that are incompatible in clinopyroxene. To the extent to which these calculations give consistent results, it can be inferred that the crystallizing olivine had 43–44% MgO. This corresponds to a composition Fo_{82-83}. If a partition coefficient $D^{ol\text{-}liq}$ (Mg-Fe) of 0.31 is adopted (Beattie et al. 1991) and it is assumed that Fe^{3+} was less than 10% of the total iron in the magma, then the liquid phase of the sill can be shown to contain between 8 and 10% MgO. The upper chilled margin of the sill contains 9% MgO and can be regarded as having the composition close to that of the liquid from which the sill had crystallised. These estimates are much lower than the amount of MgO in the upper chill of basalt flows and therefore it must be assumed that these magmas contained excess olivine + clinopyroxene, as is consistent with their petrography.

In order to test this hypothesis, we calculate the composition of the melt using the REE contents in the cumulate clinopyroxenes and the published partition coefficients between tholeiitic melt and clinopyroxene (Green 1994). The calculated liquid composition is similar to the composition of the sample 9112/2 from the upper chilled margin of the sill (Fig. 4). Thus on the basis of the two independent lines of evidence, we

Table 2 Major and trace element data. Analyses recalculated on an anhydrous basis (*LOI* loss on ignition)

Sample	Zaonega suite						Suisaarian suite: Angozero								
	Solomennoe				Suisaari Island		Lavas				Konchozero sill				
	94180	9177	9178	94182	91132	91134	89103	89104	89105	89105/1	9111	9112	9112/3	9112/1	9112/2
SiO_2	48.9	51.5	67.6	55.5	53.1	53.7	46.0	48.4	46.4	47.4	43.9	44.0	43.9	50.3	49.3
TiO_2	1.76	1.37	1.46	1.31	1.52	1.60	1.20	1.59	2.06	2.02	0.84	0.80	0.84	1.95	1.91
Al_2O_3	13.1	17.5	11.2	16.2	16.0	15.8	7.7	10.5	10.5	11.7	4.82	4.79	4.68	14.1	14.0
Fe_2O_3	14.2	11.2	8.16	10.4	10.6	10.1	14.0	13.6	14.6	15.0	14.0	13.8	14.3	9.71	10.1
MnO	0.18	0.17	0.15	0.15	0.17	0.17	0.18	0.19	0.19	0.18	0.18	0.18	0.19	0.20	0.20
MgO	7.55	6.34	2.62	4.40	5.28	4.46	20.9	16.0	15.5	14.1	29.6	29.5	29.3	8.98	8.40
CaO	12.5	7.96	4.70	7.17	9.72	8.92	9.70	7.90	9.24	7.37	6.35	6.64	6.42	11.2	12.0
Na_2O	1.04	2.49	2.93	4.49	2.29	1.83	0.20	1.39	0.99	1.71	0.15	0.14	0.21	3.19	3.46
K_2O	0.69	1.39	1.07	0.15	1.15	3.28	0.02	0.30	0.30	0.28	0.07	0.08	0.07	0.23	0.30
P_2O_5	0.13	0.11	0.12	0.14	0.13	0.15	0.13	0.14	0.17	0.16	0.05	0.05	0.05	0.15	0.18
LOI	3.02	4.66	1.22	3.13	2.03	1.46	5.28	4.69	4.05	4.29	7.42	7.29	7.95	3.54	3.37
Total	100.11	99.91	99.73	100.04	99.89	99.97	100.24	100.01	100.12	100.09	100.69	100.75	100.80	99.37	100.71
Cr	169	43	42	37	74	63	3006	1518	1384	1111	2158	2142	2317	464	456
V	284	232	215	207	222	232	219	299	322	326	150	144	162	327	316
Co	60	37	35	46	40	32	123	86	83	76	125	120	128	43	43
Ni	131	30	22	33	65	57	1274	811	705	564	1715	1660	1637	225	193
Zr	131	130	138	121	156	168	78	112	140	140	49	49	50	117	120
Hf	3.30	3.20	3.64	3.12	3.99	4.27	2.06	2.78	3.91	3.57	1.31	1.31	1.33	2.83	2.90
Nb	13.6	11.2	10.7	10.9	12.6	13.1	7.50	11.7	15.4	16.4	5.22	5.26	5.53	13.0	13.3
Ta	0.750	0.593	0.597	0.576	0.791	0.792	0.439	0.709	0.866	0.899	0.320	0.343	0.361	0.737	0.749
Sc	32.3	29.4	19.9	25.0	22.1	22.2	29.1	29.9	33.7	33.7	21.0	22.3	22.4	30.3	32.5
Y	19.7	18.1	16.9	18.3	17.6	18.4	13.0	17.6	23.3	23.5	9.12	9.31	9.14	21.7	21.5
Sr	337	448	137	244	530	914	15	80	65	93	17	17	20	259	271
Rb	16.9	40.2	21.7	3.4	15.1	52.1	0.55	6.1	6.3	5.9	5.3	4.3	5.4	21.7	5.4
Ba	318	312	194	43	306	837	26	116	93	100	8.1	2.1	10	83	87
Pb	3.92	5.20	5.02	4.66	4.91	4.41	1.23	1.70	1.52	1.74	0.776	0.751	0.796	1.97	2.04
U	0.406	0.407	0.521	0.413	0.466	0.482	0.232	0.203	0.263	0.264	0.099	0.107	0.098	0.217	0.220
Th	1.52	2.03	2.42	1.98	2.13	2.25	0.693	0.757	0.982	0.896	0.356	0.290	0.355	0.818	0.830
La	15.4	15.9	20.6	16.6	20.5	21.8	7.14	9.41	13.1	12.3	4.37	4.51	4.68	10.8	11.2
Ce	33.5	34.2	43.4	35.2	43.4	45.4	17.2	22.9	31.2	29.8	10.3	10.5	11.0	27.2	27.0
Nd	18.9	17.9	20.9	18.4	22.1	23.2	10.6	14.4	19.0	18.5	6.64	6.92	7.13	17.9	17.7
Sm	4.35	4.02	4.38	3.97	4.73	5.03	2.73	3.82	4.95	4.82	1.71	1.84	1.87	4.58	4.54
Eu	1.34	0.997	1.14	1.20	1.45	1.55	0.720	1.15	1.66	1.61	0.542	0.611	0.587	1.50	1.52
Gd	4.26	3.80	4.12	3.81	4.47	4.70	2.81	3.94	5.08	4.95	1.85	1.93	2.00	4.74	4.70
Dy	3.67	3.25	3.49	3.24	3.73	3.90	2.54	3.45	4.39	4.34	1.65	1.73	1.76	4.18	4.04
Er	1.89	1.74	1.74	1.68	1.84	1.92	1.29	1.76	2.24	2.22	0.856	0.897	0.905	2.06	2.00
Yb	1.61	1.53	1.58	1.54	1.62	1.71	1.08	1.44	1.89	1.85	0.712	0.761	0.768	1.72	1.68
Nb/U[a]								51.1	51.6	60.2	48.3	59.7	51.2	52.5	52.6

[a] Calculated from Th-U-Pb relationships

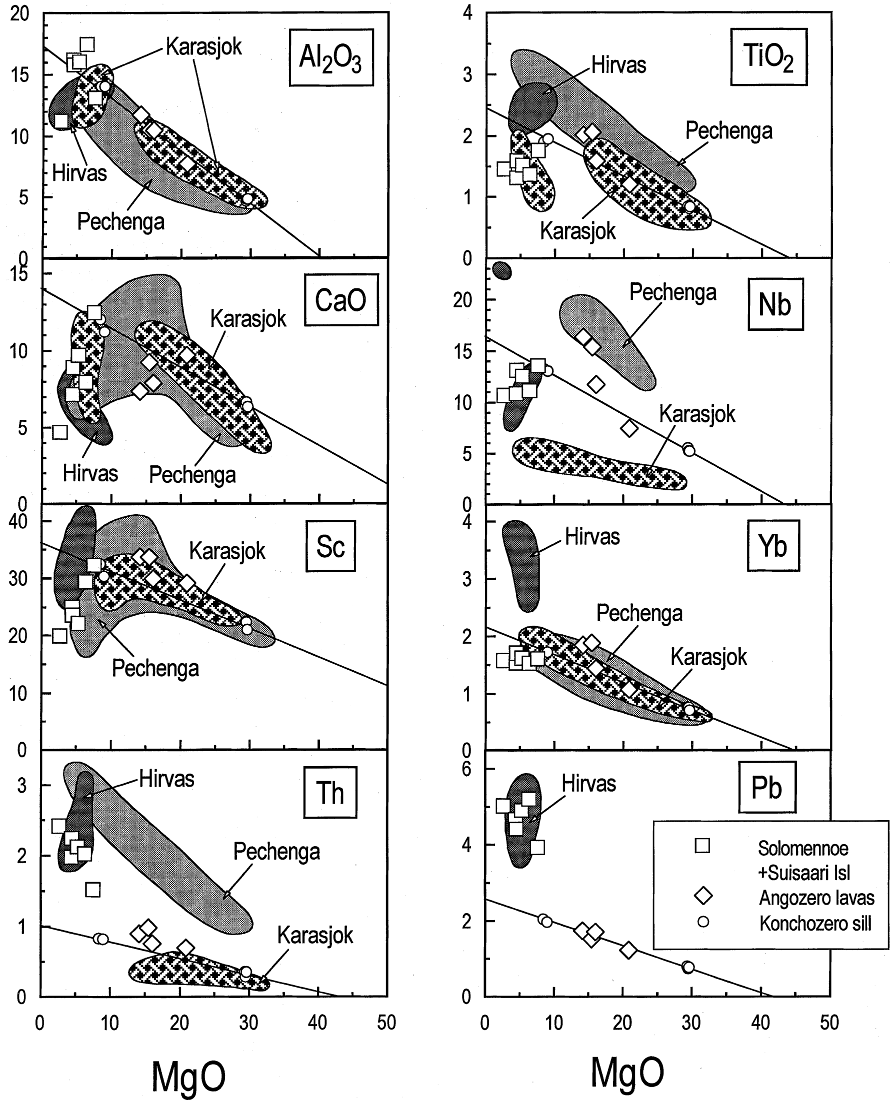

Fig. 5 Variation diagrams of selected major and trace elements. Data for the Pechenga belt are from Hanski (1992) and Hanski and Smolkin (1995) and those for the Karasjok belt from Barnes and Often (1990) and Krill et al. (1985). The olivine control lines are drawn for the Konchozero sill only

suggest that the liquid parental to the Konchozero sill contained 8–10% MgO and had a composition similar to that of the chilled sample 9112/2. Although there is no guarantee that this magma was related genetically to the volcanic rocks, for our purposes we will assume that this liquid was also parental to the whole volcano-plutonic association.

The Suisaarian volcanic and hypabyssal rocks from the Angozero area have substantially lower Th and Pb concentrations compared with those found in the Zaonega suite rocks (Fig. 5). In this respect the former are closer to the Karasjok komatiites, while the latter resemble the Pechenga and Hirvas lavas. The significance of these differences is discussed below.

Shown in Fig. 6 are primitive mantle-normalised (Hofmann 1988) abundances of major and trace elements. These patterns reveal clear differences between the rocks from the two suites. The Angozero rocks are characterized by moderately fractionated profiles $[(La/Yb)_N = 4.0–4.7]$, nearly chondritic Ti/Zr ratios of 90–100, moderate enrichment in LREE $[(La/Sm)_N = 1.5–1.7]$ and relative depletion in Al, Y, Sc and HREE $[(Gd/Yb)_N = 2.0–2.3]$. They show pronounced maxima at Nb and Ta $[(Nb/La)_N = 1.1–1.3; (Nb/Th)_N = 1.4–2.4]$. The rocks are characterized by negative Pb-anomalies $[(Ce/Pb)_N = 1.4–2.2]$. These patterns are similar to those found in oceanic flood basalts (OFB) and ocean island basalts (OIB) (Fig. 6). Samples from the Zaonega suite display much greater enrichment in the more-incompatible elements $[(La/Yb)_N = 7.0–8.8; (La/Sm)_N = 2.2–3.0]$, but show the same degrees of depletion in Al, Y, Sc and HREE $[(Gd/Yb)_N = 2.0–2.2]$. They are enriched in Zr (Ti/Zr = 60–80), Th and U, show negative Nb and Ta anomalies $[(Nb/La)_N = 0.52–0.88; (Nb/Th)_N = 0.58–1.2]$, the patterns resembling those found in the continental crust (Fig. 6).

Isotope-geochronological data

The Sm-Nd isotope data are presented in Table 3 and plotted on Fig. 7. The data for the bulk samples of the

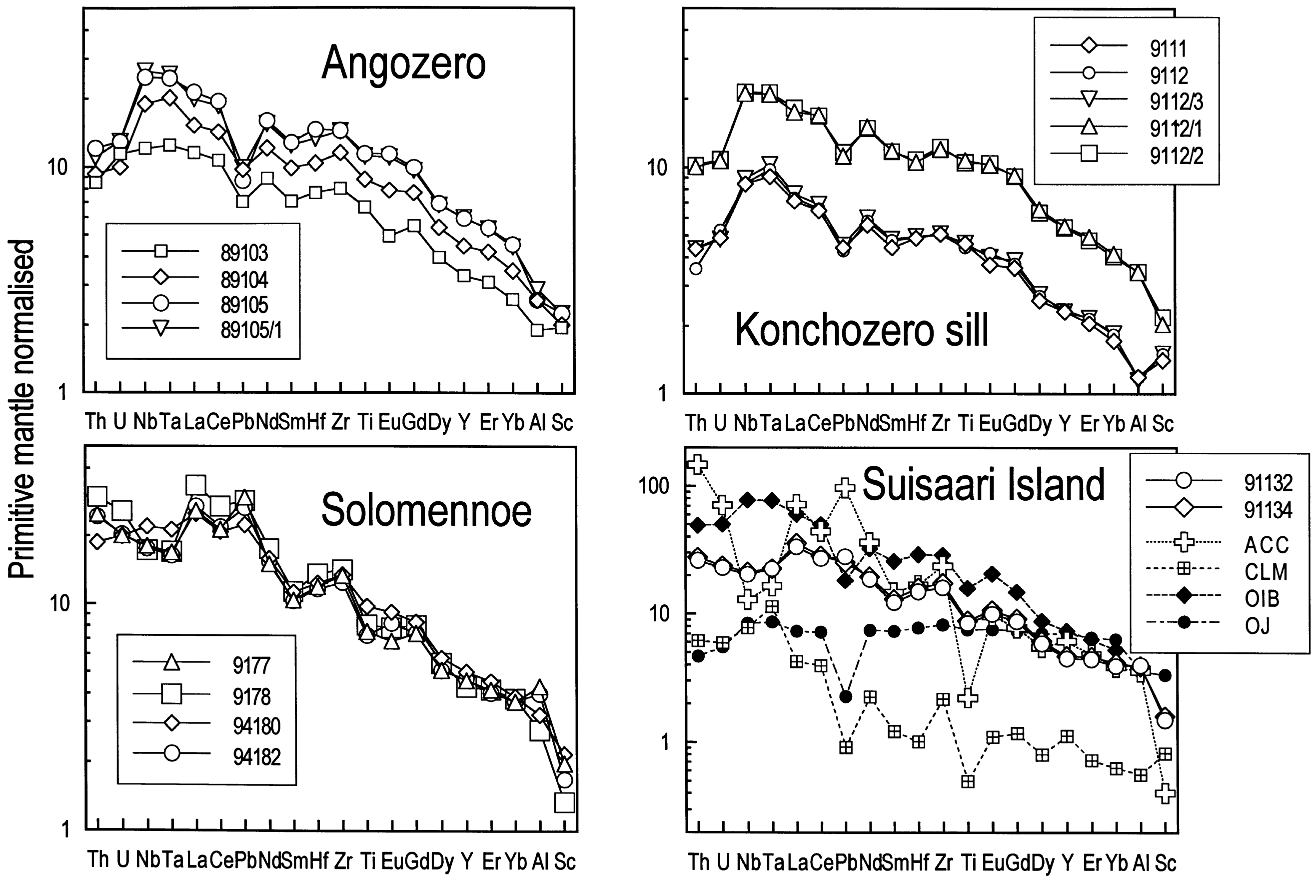

Fig. 6 Abundances of selected major and trace elements normalised to primitive mantle values of Hofmann (1988). [*ACC* Archaean continental crust (Rudnick and Fountain 1995), *CLM* continental lithospheric mantle (McDonough 1990), *OIB* average ocean island basalt (Sun and McDonough 1989), *OJ* average composition of basalt from the Ontong Java plateau (compiled using the data of Mahoney et al., 1993, and Neal et al., in press)]

Konchozero sill peridotites and gabbros together with the clinopyroxene separates define an isochron with a slope corresponding to an age of 1975 ± 24 Ma and εNd(T) of +3.2 ± 0.1. All whole-rock samples from both suites define a linear array [MSWD (mean square weighted deviates) = 10] with a steeper slope corresponding to an age of 3420 Ma, some 1.5 Ga older than the isochron age. Again, there is a distinct correlation between the stratigraphy and isotope compositions. The samples mostly form two data clusters, one from the Solomennoe and Suisaari Island area with low Sm/Nd ratios and εNd(T) values ranging between 0.60 and −2.9, and the other from the Angozero area with higher Sm/Nd and initial εNd values of +2.0 to +3.3.

Lead isotopes were measured on leached samples because of the possibility of uranium mobility during interaction with sea water. Only the Angozero rocks were studied. The Pb-isotope data are reported in Table 4 and plotted on Fig. 8. Residues from the basalt samples, Konchozero sill rocks and mineral separates define an isochron with an age of 1980 ± 57 Ma. The μ_1, or time-integrated $^{238}U/^{204}Pb$ of the source of the rocks is 8.57 ± 0.03. This value is close to the μ_1 of 8.62 obtained by Hanski (1992) for the Pechenga picrites.

In the $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ plot, the data also define a linear trend with a somewhat larger scatter (MSWD = 30). These data allow us to estimate the $^{232}Th/^{238}U$ ratio in the volcanic rocks. The estimated Th/U ratio is 3.4 ± 0.3, in good agreement with the average Th/U ratio of 3.5 ± 0.4 calculated for these rocks from the trace element data.

Several zircons were separated from the picrobasalt sample 89103 (Angozero) for single grain evaporation. They are slightly elongated brownish crystals containing zoned subhedral cores. Most of them were too small to yield persistent and strong ion beams in the mass spectrometer. One grain, some 120 μm long and oval shaped, yielded a $^{207}Pb/^{206}Pb$ age of 2693 ± 3 Ma. The shape and structure of the crystal indicate that it crystallised from a felsic, not mafic magma. Taking into account that this zircon is some 700 Ma older than the host rock, we conclude that it represents a xenocrystic grain assimilated together with the wallrocks during the magma ascent.

Zircons separated from the Jatulian sample #9118 for SHRIMP studies show simple habit determined by (100) prism and (101) pyramid faces. They are inhomogeneously coloured from light to dark brown, large irregular embayed grains morphologically resembling zircons that crystallise in mafic intrusive rocks (Poldervaart 1956). About thirty grains were mounted in epoxy resin,

Table 3 Nd-isotope data (*WR* whole rock, *Cpx* clinopyroxene). All Nd-isotope ratios were bias-corrected to the La Jolla standard $^{143}Nd/^{144}Nd$ = 0.511860. The initial εNd values were calculated using the present-day parameters of the chondrite uniform reservoir (CHUR) adopted from Jacobsen and Wasserburg (1980) to be the following: $^{147}Sm/^{144}Nd$ = 0.1967, $^{143}Nd/^{144}Nd$ = 0.512638

Sample	Sm, ppm	Nd, ppm	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	εNd (1975)
Konchozero sill					
9111 WR	1.661	6.446	0.15573	0.512272 ± 13	3.26
9111 Cpx	1.809	4.886	0.22387	0.513151 ± 14	
9112 WR	1.657	6.411	0.15626	0.512271 ± 10	3.11
9112 Cpx	2.017	5.418	0.22507	0.513159 ± 18	
9112/3 WR	1.700	6.604	0.15562	0.512255 ± 13	2.96
9112/3 Cpx	2.015	5.451	0.22344	0.513149 ± 12	
9112/2 WR	4.232	16.71	0.15306	0.512240 ± 11	3.31
9112/2 Cpx	2.127	6.007	0.21402	0.513036 ± 9	
9112/1 WR	4.356	17.07	0.15427	0.512247 ± 7	3.14
Angozero					
89103	2.670	10.05	0.16052	0.512269 ± 6	1.98
89104	3.687	14.05	0.15870	0.512295 ± 7	2.96
89105	4.862	18.81	0.15629	0.512255 ± 5	2.79
89105/1	4.868	18.48	0.15928	0.512267 ± 5	2.26
Solomennoe					
94180	4.629	19.99	0.14002	0.511932 ± 4	0.60
9177	3.854	17.67	0.13183	0.511645 ± 4	−2.94
9178	4.347	20.62	0.12744	0.511592 ± 6	−2.86
94182	4.751	21.99	0.13057	0.511630 ± 4	−2.92
Suisaari Island					
91132	4.724	22.00	0.12980	0.511653 ± 6	−2.27
91134	4.614	21.69	0.12858	0.511655 ± 5	−1.92

polished until they were sectioned approximately in half and analysed using SHRIMP. However, only five grains contained relatively clear transparent areas, which gave concordant U-Pb results. These results are presented in Table 5 and plotted on Fig. 9. Analyses of dark opaque zircon show various degrees of discordance. Five concordant points give the average of 1976 ± 9 Ma.

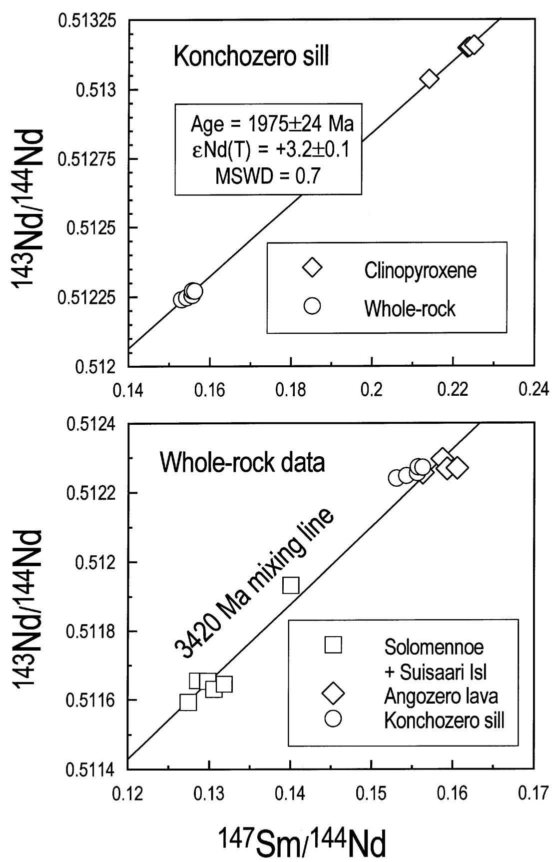

Fig. 7 Sm-Nd evolution diagrams

Discussion

Mantle melting and differentiation

The geochemical features of the basalts and the Konchozero sill provide some constraints on the composition and conditions of generation of their parental magmas. The chemical similarities between the Angozero lavas and the gabbro-peridotites are particularly

Table 4 Pb-isotope data (*Pl* plagioclase). The analyses were corrected for isotopic fractionation determined by analysis of the standard SRM-982 to be 1.00135 ± 0.00036 (2σ) per atomic mass unit. The μ_1 value was calculated using a single-stage model, assuming 4.50 Ga as the age of the Earth and Canyon Diablo values (Tatsumoto et al. 1973) for the starting isotopic composition

Sample	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	μ_1
89104	17.222	15.360	36.814	8.55
89105	21.568	15.890	41.171	8.56
89105/1	21.312	15.854	40.607	8.55
9111	18.414	15.516	38.504	8.58
9112	18.422	15.494	38.466	8.53
9112/1	18.378	15.522	38.091	8.60
9112/2	18.899	15.586	38.347	8.60
9112/3	22.342	15.997	42.551	8.58
9111 Cpx	19.310	15.640	38.797	8.61
9112 Cpx	17.151	15.346	37.353	8.54
9112/3 Cpx	17.173	15.358	37.275	8.56
9112/2 Pl	14.389	15.031	34.357	8.59

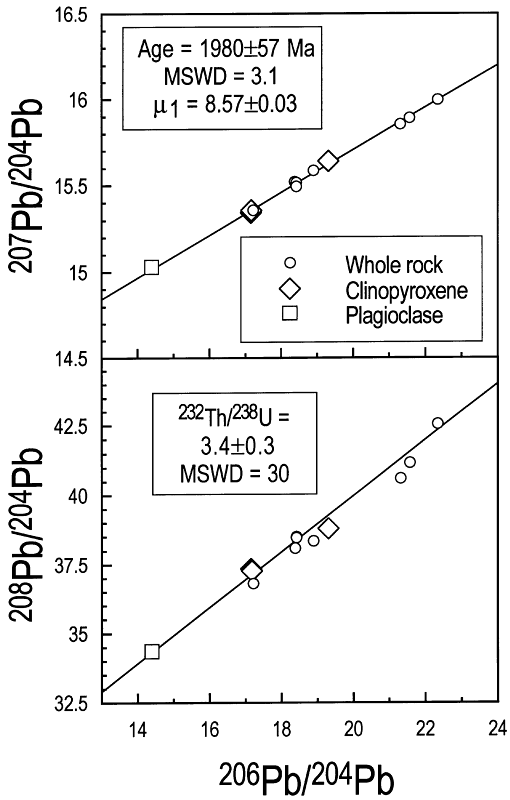

Fig. 8 Pb-Pb diagrams

evident in Figs. 5 and 6 and justify their classification as a single volcano-plutonic association. As shown above, the parental magma of the Onega lavas and the sill is inferred to have had 8–10% MgO (similar to the composition of the chilled samples 9112/1-12/2). However, it is not likely that these samples represent the composition of the primary magma. As is the case for many continental flood basalt provinces, the erupting magmas usually represent moderately evolved liquids that resulted from long-term fractionation both *en route* to the surface and in subcrustal lithospheric magma chambers (e.g. Wooden et al. 1993). The pronounced Al-, Y-, Sc-, and HREE-depletions suggest generation of the initial melt in the garnet stability field at depths 80–100 km (Gudfinnsson and Presnall 1996), and thus deeper than the base of the thinned continental lithosphere (Menzies 1990). At these depths, primary mantle melts are magnesian even at low degrees (5–10%) of partial melting (e.g. Herzberg 1992, 1995). Additionally, if the degree of melting is low, the trace element abundances in these melts will be high. We suggest that this was the case for the Onega plateau lavas. The lack of upward decline in Gd/Yb ratios in the Onega rocks, normally observed in other continental flood basalt provinces and interpreted as a result of ascent of the site of melting, suggests that the primary magmas separated from their source region while still in the garnet stability field. This may imply a large thickness of the continental lithosphere in the region. After the magmas separated from the plume source

Table 5 Th-U-Pb SHRIMP data for zircons from basalt sample #9118. The reported uncertainties correspond to 1σ

Label	U ppm	Th ppm	Th/U	$^{204}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$	$^{208}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{208}Pb/^{232}Th$	Age $^{207}Pb/^{206}Pb$	% Concordance
p^1	1938	3549	1.83	0.00021	0.12127 ± 49	0.5082 ± 17	0.3745 ± 49	6.26 ± 9	0.1039 ± 15	1975 ± 7	104
p^4	1968	5843	2.97	0.00024	0.12109 ± 48	0.8121 ± 22	0.3683 ± 49	6.15 ± 9	0.1007 ± 14	1972 ± 7	102
p^5	1806	5260	2.91	0.00055	0.12053 ± 68	0.8193 ± 27	0.3500 ± 46	5.82 ± 9	0.0985 ± 14	1964 ± 10	98
p^7	950	1662	1.75	0.00011	0.12189 ± 47	0.4905 ± 17	0.3641 ± 48	6.12 ± 9	0.1020 ± 14	1984 ± 7	101
p^9	887	4109	4.63	0.00008	0.12150 ± 59	1.2771 ± 42	0.3680 ± 49	6.17 ± 9	0.1014 ± 14	1978 ± 9	102

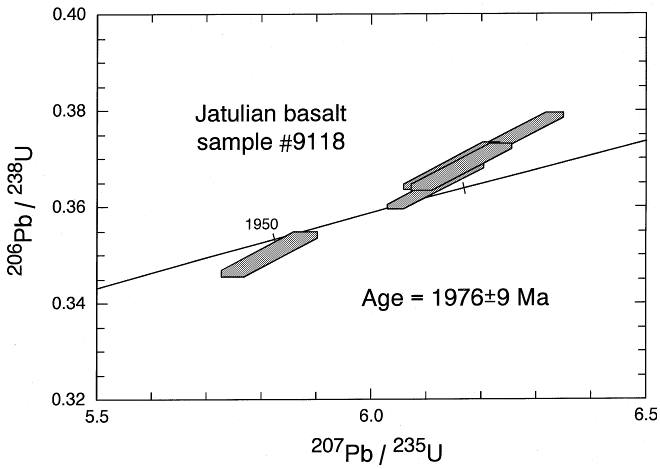

Fig. 9 Concordia diagram showing ion microprobe single-grain analyses for zircons from Jatulian basalt sample #9118

region, they passed through the continental lithospheric mantle and crust. As is evident from petrography and chemistry of the rocks, their differentiation *en route* to the surface was controlled by two major liquidus phases, clinopyroxene and olivine. According to the experimental data on komatiites and basalts, at 1 atm pressure clinopyroxene starts to crystallise from a liquid containing 7.5–8.5% MgO (Kinzler and Grove 1985; Thy 1995). Moreover, some authors argue that clinopyroxene joins olivine on the liquidus at low pressure in even more magnesian magmas (e.g. represented by MORB glasses with 10–12% MgO, Grove and Bryan 1983), and increasing pressure extends the clinopyroxene stability field towards the more magnesian compositions (O'Donnell and Presnall 1980; Elthon and Scarfe 1984). This evidence suggests that the Onega parental magmas probably differentiated in subcrustal magma chambers and then the evolved Ol + Cpx mush was erupted either as thin picrobasalt flows, which solidified before any substantial differentiation had occurred, or were emplaced as shallow-level sills. One of these, the Konchozero sill, represents a large and slowly cooled body, in which differentiation probably continued after emplacement, when it mostly crystallised olivine.

Isotope-geochronological data and the timing of formation of the volcanic sequence

The Sm-Nd mineral and Pb-Pb whole-rock isochron ages of 1975 ± 24 and 1980 ± 57 Ma correspond to the time of emplacement of the sill and to the eruption of the Ludikovian lavas. The age of these rocks coincides also with the Sm-Nd, Pb-Pb and U-Pb zircon ages of picrobasalts from the Pechenga-Imandra belt and komatiites and basalts from the Karasjok-Kittilä belt (listed earlier). This implies contemporaneous outpouring of large volumes of mafic-ultramafic magmas over the vast territory in the Baltic Shield and together with evidence for their formation deep in the asthenospheric mantle suggest to us that the Karasjok-Pechenga-Onega magmas are the melting products of a mantle plume. Below we will provide more arguments to support this model.

The SHRIMP U-Pb zircon age of 1976 ± 9 Ma for the Jatulian lava sheet is somewhat contradictory with respect to the previously published age data and general opinion about the age of the Jatulian rocks (see review in "Regional geology"). Several possible interpretations of this age may be proposed. Firstly, this unit may not correlate with the Jatulian, but is part of the Ludikovian sequence. This is possible, though the Hirvas locality is regarded as stratotype area of the Jatulian within the Onega plateau (Svetov 1979) and is lithologically similar to the Jatulian elsewhere in the shield. Secondly, the sampled body may represent a hypabyssal sill rather than a lava. As mentioned above, there exist a number of hypabyssal mafic rocks in the Hirvas area; in addition, Vuollo et al. (1992) published a U-Pb zircon age of 1965 ± 10 Ma for a diabase dike generation cutting Jatulian metasediments in northern Karelia. The studied rock comes from a single body overlain by terrigenous sediments and it is certainly possible that it represents a sill. Nevertheless, this "sill" is chemically and isotopically very similar to the lavas overlying the metasediments; it is most probable that they had a related origin and that the entire sequence was emplaced simultaneously. We would suggest that the Jatulian sequence studied is correlative with the uppermost part of the Jatulian rocks developed over the Karelian province, and this part is contemporaneous with the Ludikovian lavas and sills. However, in order to reconcile these inconsistencies, direct reliable isotope datings of the Jatulian rocks within their reference areas are highly desirable.

The bulk-rock data show an excessive slope in a Sm-Nd isotope plot (Fig. 7). The rocks from the two suites most likely had distinct initial Nd-isotope compositions. The negative εNd(T) values in the Zaonega lavas suggest that their parental magmas were derived, at least in part, from a source region with a long-term LREE-enrichment. This kind of source could have been continental lithospheric mantle and/or ancient felsic crust. The xenoliths of underlying schists and xenocrystic zircons in the basalts are direct evidence for crustal contamination. This xenolithic material was probably inherited from felsic wallrocks assimilated during ascent of the basaltic melts. The isotope data on xenocrystic zircons suggest that the host crustal rocks were at least 2.7 Ga old. This is an age commonly found in the underlying granitoids (Kulikov et al. 1990).

Crustal contamination

Assimilation of even small amounts of felsic crustal rocks results in a sharp increase in the abundances of Ba, Pb, U, Th, and LREE, but should have little effect on the concentrations of Ta, Nb, Y, Ti and HREE. This produces negative Ta-, Nb-, and Ti-anomalies in contaminated rocks. In contrast, the average subcontinental lithospheric mantle (Fig. 6) is enriched in Nb and Ta relative to Th and La. Consequently, mixing with material from the continental lithospheric mantle will not produce such anomalies in mantle melts. The effect of crustal contamination is clearly seen in magnesian rocks, which have low levels of highly incompatible elements. Such magmas, being characterized by high liquidus temperatures and low viscosities, are very susceptible to contamination (e.g. Huppert and Sparks 1985). Contamination, however, will have much smaller effect on flood basalts, which have one to two orders of magnitude higher concentrations than komatiites and substantially lower liquidus temperatures. Nevertheless, as can be seen in Fig. 10, distinct positive correlations exist, for example, between Nb/Th and Nb/La and negative correlations are observed between Nb/La or εNd and La/Sm. These trends can be explained by mixing of a mantle-derived magma with a component of the upper Archaean continental crust having low εNd(1975), Nb/Th, and Nb/La ratios.

The upper crust in the adjacent Vodla and Belomorian Blocks, which form the basement for the Onega plateau, consists of TTG-gneisses and mafic and felsic granulites with ages of 3.1–2.6 Ga (Kulikov et al. 1990; Bogdanova and Bibikova 1993; Lobach-Zhuchenko et al. 1993; Timmerman and Daly 1995; Bibikova et al. 1996). The average Nd-isotopic parameters of these rocks ($^{143}Nd/^{144}Nd = 0.51093$, $^{147}Sm/^{144}Nd = 0.1011$, $n = 49$) were compiled from the data of these authors. Trace element data for these rocks are not available, and so, in order to evaluate the extent of contamination of the primary basaltic magmas, we used a trace element composition similar to that of the average mafic and felsic Archaean granulite reported by Rudnick and Fountain (1995).

The composition of the asthenospheric magmas that gave rise to the Onega plateau lavas can be estimated using the compositions of plume-derived rocks in oceanic settings, because there is no reason to suspect that distinct types of mantle plume ascend beneath the oceans and the continents. Shown in Fig. 11 are $(Nb/Th)_N$ versus $(Nb/La)_N$ data for basalts from the Ontong Java oceanic plateau, oceanic island basalts, and several

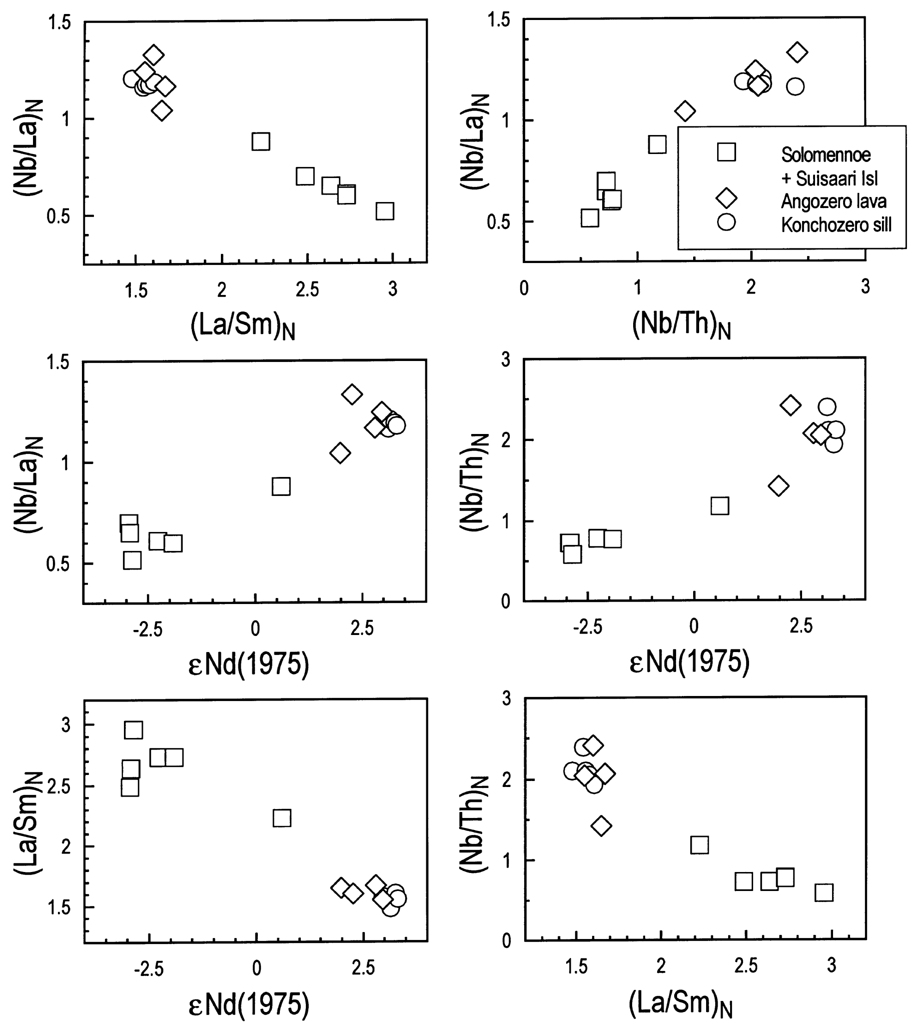

Fig. 10 Diagrams for selected elemental ratios

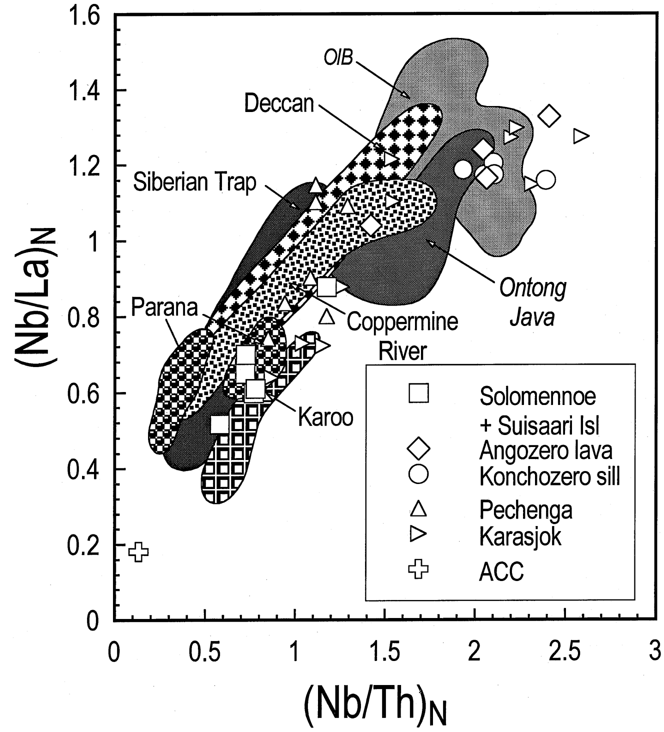

Fig. 11 $(Nb/Th)_N$ versus $(La/Nb)_N$ data for the Onega plateau as compared with the Pechenga and Karasjok lavas, the Ontong Java plateau basalts, oceanic island basalts (*OIB*), and the largest continental flood basalt provinces. The data sources used here and in Fig. 13 are available from the senior author upon request

continental flood basalt provinces including the Siberian Traps, Deccan Traps, Coppermine River, Karoo and Parana plateau. The Angozero rocks plot within the fields of the Ontong Java and OIB lavas and are probably the best estimates of the plume composition. In contrast, the data for the Zaonega lavas straddle the trend towards the composition of the continental crust, and plot within the fields of crustally contaminated continental flood basalts. We chose sample 9112/2 from the upper chilled margin of the Konchozero sill as the parental endmember in the contamination model. The Nd-isotope compositions of the crustal endmember and the Onega plateau rocks were recalculated for the time of the mixing event (1975 Ma). The results of the mixing calculations are shown in Fig. 12 and the trace element and Nd-isotope data support the binary mixing model. The estimated degrees of contamination are consistent for different trace element and isotope ratios and decrease up through the sequence from 30 to 10% in the Zaonega suite samples to 4% in one of the Angozero samples, in which several xenocrystic zircons were found; the uppermost Angozero lavas and the sill are regarded as representing uncontaminated mantle compositions. This decline in the degree of contamination up the volcanic sequences is a common feature of most continental flood basalt provinces (e.g. Wooden et al. 1993) and is probably attributable to the fact that the later portions of magma on their way through the crust used the same channels, which by that time were already covered with basalt.

Composition of the mantle sources and tectonic setting of the Onega plateau

Using geological and geochemical data we can place some constraints on the composition of the mantle source of the Onega lavas, which we interpret to represent a relict of a once areally extensive province of continental flood basalts. In the Karelian Province, Jatulian and Ludikovian volcanic rocks cover an area of $\sim$50,000 km^2 at the present erosion level, and in places attain a thickness of 4500 m. In Norway and adjacent parts of northern Sweden and Finland (e.g. Karasjok-Kittilä belt), as well as in the Pechenga-Imandra belt of the Kola Province, at least the upper parts of the Jatulian and Ludikovian successions can be correlated with those of the Onega plateau. Moreover, these rocks have similar ages and reveal substantial overlap in terms of chemical and isotope compositions (Figs. 6 and 10). If they were parts of a single flood basalt province, as proposed by Eskola (1963), this province would have covered a total area of $\sim$600,000 km^2, comparable with the distribution of Phanaerozoic continental flood basalts (Coffin and Eldholm 1994).

The Onega plateau contains a wide range of basalt types. The Angozero rocks have high Nb/Th, Nb/La, and Ce/Pb ratios similar to those found in recent oceanic plateau basalts and modern OIB (Fig. 11). We interpret compositions of the Angozero rocks as being representative of their mantle source. The positive εNd(1975) values of $+3.2 \pm 0.1$ imply long-term LREE-depletion in the source although the magmas are enriched in these elements. As pointed out above, the Onega magmas could have been the result of relatively low degrees of partial melting that could explain their LREE-enrichment. However, both the Pechenga and the Karasjok magmas also show enrichment in LREE. Because the Karasjok magmas are komatiitic in composition, and probably formed by higher degrees of melting of perhaps 10 to 20% (Barnes and Often 1990), we suggest that the mantle sources for all magmas, which were related to the 1.98 Ga event in the Baltic Shield, have been enriched in incompatible elements shortly before the magmatic event.

The εNd(T) values in the uncontaminated Onega rocks are lower than the range of $+4.0$ to $+5.3$ determined for depleted MORB from the contemporaneous Purtuniq ophiolite (Hegner and Bavier 1991), but plot on the PREMA (prevalent mantle) evolution line postulated by Stein and Hofmann (1994), which leads to the composition of plume-derived magmas from the recent Pacific oceanic plateaux ($+3.3$ at 1975 Ma). There is, however, no direct match between the Onega rocks and modern oceanic volcanics in terms of trace element and REE concentrations (Fig. 13). The former are distinguished from OIB by lower abundances of moderately

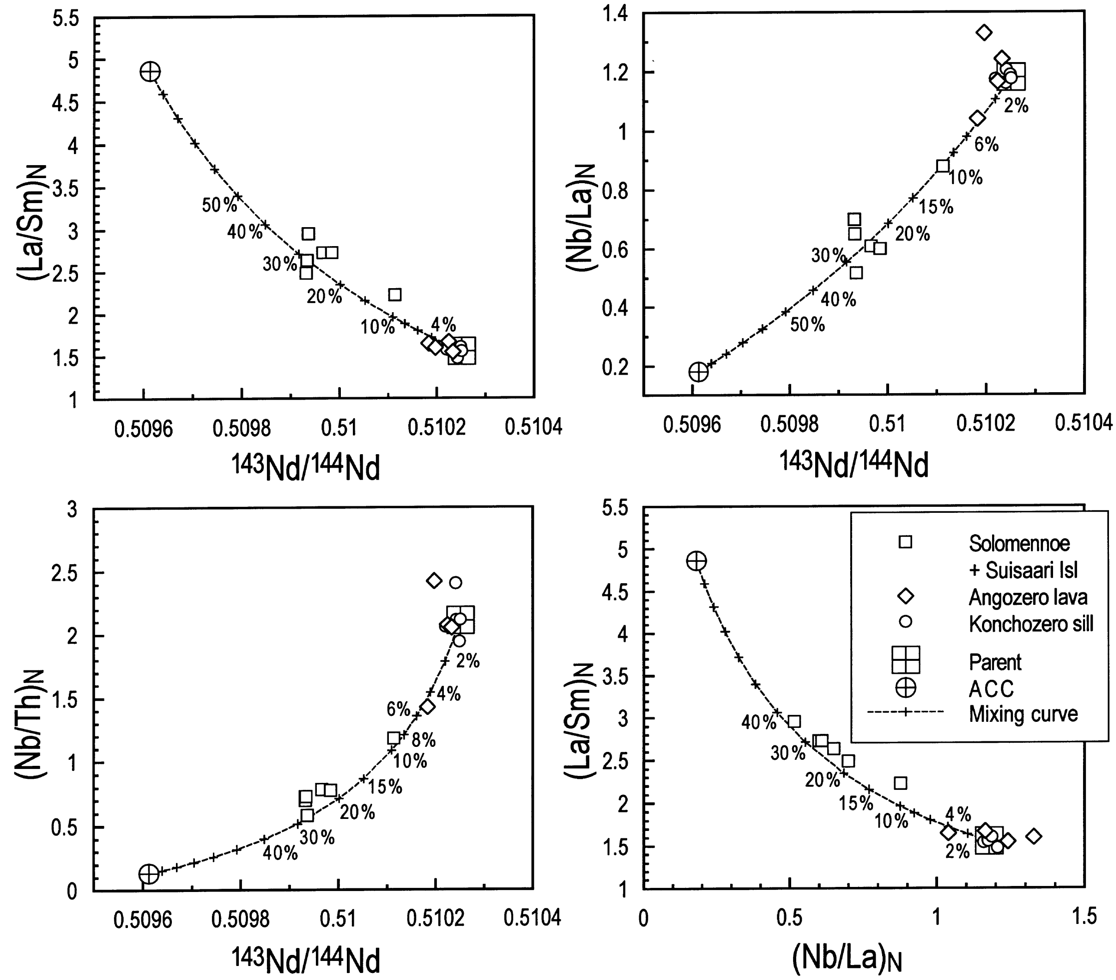

Fig. 12 Diagrams illustrating the postulated mixing model between the primary Onega plateau magma and Archaean continental crust (*ACC*)

and highly incompatible elements and subdued fractionation of HREE, and from oceanic flood basalts by higher concentrations of incompatible elements and by HREE-fractionation (most OFB have flat patterns). A better match is found when the Onega rocks are compared with continental flood basalts (CFB). The Angozero basalts plot together with the Gudchikhinsky picrites from the Siberian province, and share similar characteristics with non-contaminated flood basalts from other areas, whereas the Zaonega lavas are similar to contaminated Siberian Trap volcanic rocks (Fig. 13). Moreover, when the entire sequence is considered and the geological context is taken into account, a comparison with CFB is most appropriate.

Using the arguments introduced by Morgan (1971) and Duncan (1978) and further developed by Campbell and Griffiths (1990) and McKenzie and Bickle (1988), we suggest that the formation of the studied flood basalt province was controlled by the interaction between a mantle plume and continental crust. A plume model has been tentatively proposed by Hanski (1992) and Walker et al. (1997) for the formation of the Pechenga mafic-ultramafic sequences and for the origin of the Karasjok-Kittilä komatiites by Barnes and Often (1990). These komatiites are pillow lavas and volcaniclastic lithologies erupted into shallow water, and are interpreted to represent primary liquid composition (Barnes and Often 1990).

Using the method of Nisbet et al. (1993) and Abbott et al. (1994) we estimate that average liquidus temperature of the Karasjok magma was 1520 ± 20 °C. In fact, the Karasjok komatiites represent the most magnesian and high-temperature magmas so far reported for the Proterozoic. These high liquidus temperatures imply that the source of the Karasjok komatiites was much hotter than the ambient mantle (1470 °C at 2.0 Ga, Richter 1988). The liquidus temperature is directly related to the potential mantle temperature (McKenzie and Bickle 1988), and we calculate the latter to be 1730 ± 30 °C. Therefore, the ascending material that gave rise to the Karasjok komatiites was some 250 °C hotter than the surrounding mantle; this can be accounted for by their origin in a mantle plume. The Karasjok komatiites share the same Al- and HREE-depletion characteristics with the Onega lavas and Pechenga picrites suggesting that their melting also

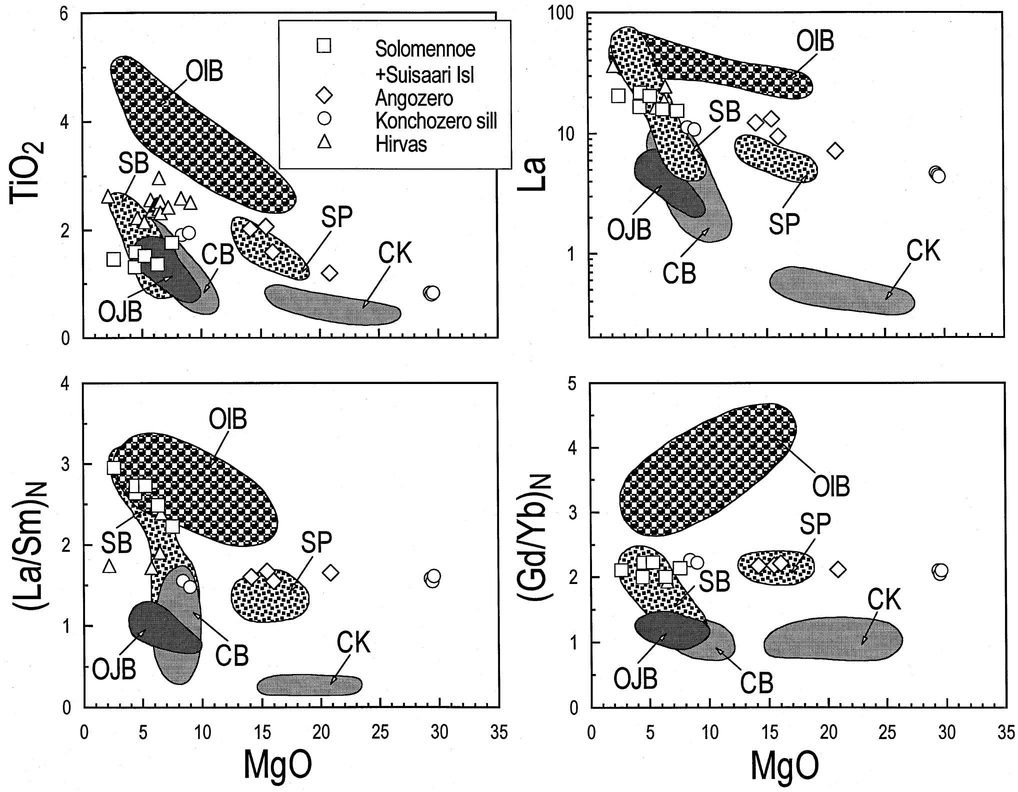

Fig. 13 TiO_2, La, $(La/Sm)_N$, and $(Gd/Yb)_N$ versus MgO data for the Onega plateau rocks as compared with the oceanic flood basalts (*OJB* Ontong Java basalts, *CB*, *CK* Caribbean basalts and komatiites/picrites), oceanic island basalts (*OIB*) and continental flood basalts (*SB*, *SP* Siberian Trap basalts and picrites). Hirvas – unpublished data of the authors

occurred in the garnet stability field (Barnes and Often 1990). Using the method of McKenzie and Bickle (1988) and Abbott (1996) we estimate that the depth of melting initiation of these komatiites was 240 ± 40 km. Similar calculations for the Angozero parental magma give a depth of initiation of ~100 km. We argue that the Karasjok komatiites were formed in a hot plume tail, while the Onega plateau and Pechenga-Imandra basalts are probably the products of melting in a cooler plume head similar to the mechanism proposed by Campbell et al. (1989). Taking into account the present areal distribution of these belts, we estimate that this fossil plume head might have had a diameter of ~2000 km, comparable to the head size of the largest extant plumes such as the Iceland plume (Gill et al. 1995).

On the southwestern margin of the postulated Baltic continental flood basalt province there is a 1.95 Ga ophiolite belt running from the Outokumpu district in northern Karelia as far as the vicinity of Jormua in the Kainuu schist belt (Peltonen et al. 1996). This ophiolite is interpreted as part of an ocean basin, and its emplacement was possibly triggered by a reorganization of plate motion related to the 1.98 Ga plume. The time interval involved (~30 Ma) is comparable to modern areas showing interaction between rifting and plume activity like the Red Sea-Gulf of Aden, even though we are aware that not all plumes cause rifting and not all rifts require plumes to initiate them.

Flooded Precambrian continents: fact or fiction?

From the petrography and field relations it is evident that the rocks from the middle and upper stratigraphic levels within the Onega, Karasjok-Kittilä and Pechenga-Imandra successions were erupted in a submarine environment. If it is accepted that some of these magmas were contaminated with, and thus erupted onto continental crust, this raises the question of why the continent in this part of the world was submerged during the Palaeoproterozoic. In fact, this is a general problem that relates to many Archaean greenstone sequences. In the Kambalda region of Western Australia, for example, the entire volcanic sequence is clearly submarine, but the magmas appear to have travelled through continental crust on their way to the surface (Arndt and Jenner 1986; Compston et al. 1986). Post-3.2 Ga volcanic sequences of the Pilbara, also in Western Australia (Nelson et al. 1992), the komatiitic and tholeiitic lavas of the Belingwe Belt in Zimbabwe (Bickle et al. 1994), and komatiitic basalts of the Vetreny belt in the Baltic Shield (Puchtel et al. 1997a) also appear to have erupted under water in basins within older continental crust. There is thus an interesting contrast between the Phanaerozoic situation, for example, in the Deccan, Parana or Siberian continental flood basalt provinces where most lavas erupted subaerially (Saunders et al. 1992), and the Precambrian settings. Furthermore, in the well-documented Tertiary North Atlantic province, the arrival of a mantle plume head beneath the continental lithosphere caused the uplift of the continent some 20 Ma before the subaerial eruption of picrites and basalts (Skogseid et al. 1992; Nadin and Kusznir 1995). With thermal relaxation and the opening of the North Atlantic the continent then subsided beneath the ocean to form the dipping reflector sequence (Saunders et al. in press). The North Atlantic situation fits with modern plume theory: the plume arrives, uplifts the continent, erupts basalts, and then with thermal relaxation, the continent submerges. What happened at Onega and in Archaean greenstone sequences, and why were large portions of the continents submerged in the Precambrian during the eruption of lavas? If the total amount of water in the oceans was not significantly greater than at present (and there is no reason to believe otherwise), this suggests that the capacity of ocean basins was less, leading to the submergence of at least the rifted parts of the continents. There are several possible reasons why this might be so. Many of the relevant arguments have already been aired during the discussion of continental freeboard (Galer 1991; Galer and Mezger in press). There is an interplay between the volume of the continents and the average depths of the oceans. A large proportion of flooded continental crust implies either that the volume of continental crust was large, or that the average ocean depth was less, perhaps because the oceanic ridges were more extensive, more active, and shallower or because a greater portion of the ocean basins was filled by volcanic plateaux. In the case of the Onega province, a large volume of continental crust is quite credible: the province formed after the late Archaean, a period accepted by most authors as a period of rapid crustal growth. For the mid to late Archaean greenstone belts the situation is more delicate. Either the continental crust was already extensive, or a smaller volume of continental crust was compensated by far more extensive shallow oceanic crust. Below we treat this problem in some more detail.

Nb-Th-U-Pb data: evidence for the early formation of the continental crust

The initial isotope composition of the lead in the Angozero rocks can be estimated using simplified lead-evolution models assuming 4.50 Ga as an "effective" age of the Earth (actually, the age of core formation) advocated by some authors (Allègre et al. 1995; Galer and Goldstein 1996). This view is consistent with the recognition that the Earth accretion probably lasted several tens of million years. A single-stage model then yields a μ_1 of 8.57 ± 0.03. This value implies an unradiogenic Pb composition of the Angozero rocks. It is even lower than those in the Pacific Ontong Java (8.63), Manihiki (8.62), and Caribbean (8.71) oceanic plateaux, which originate far away from any sizable continental masses. We suggest therefore that no older felsic crust was involved in the petrogenesis of the Angozero rocks and their Th-U-Pb relationships reflect those in their mantle sources.

Although U is thought to be more mobile than Th because of its much greater solubility under oxidizing conditions, the Th/U ratio estimated from the slope of the linear correlation in the $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram (3.4 ± 0.3) is indistinguishable from that calculated from the trace element data (3.5 ± 0.4). This suggests an essentially immobile behaviour of U in the Angozero rocks. Nevertheless, in order to be safe, we calculate "real" U concentrations from the measured Th abundances and Pb-isotope data. Further on, using estimated U concentrations we calculate Nb/U ratios in the uncontaminated Angozero rocks. These values range between 48 and 60 with an average of 53 ± 4 (Table 1).

Hofmann et al. (1986) and Hofmann (1997a) proposed that Nb/U systematics in ancient mafic-ultramafic lavas can be used to deduce the history of continental crustal growth. This is attributed to the fact that the Nb/U ratio has changed from the primitive value of ~30 before continent formation to ~47 in the present-day mantle (Hofmann 1997b) due to the removal of the continental crust. However, it remains unclear whether it has grown continuously ever since (Moorbath 1997) or was formed in the early Archaean and then was maintained in a steady-state through continuous recycling back into the mantle (Armstrong 1981). The high Nb/U ratios (~47) recently detected in lavas from the Archaean Kambalda and Kostomuksha provinces (Sylvester et al. 1997; Puchtel et al. in press) provide an argument in support to the latter model. Moreover, the Onega plateau lavas show Nb/U ratios that are even higher than the average value of ~47 observed in modern MORB and OIB. This implies that the source regions of the Onega plateau lavas have experienced the degree of continental material extraction comparable to that found in the modern mantle. This suggests in turn that by 2.0 Ga and possibly as early as 2.8 Ga the volume of continental crust similar to the present-day value already existed. However, as stated by Hofmann (1997a), great caution must be exercised while extending this conclusion and accepting it as a general constraint on the evolution of the terrestrial continental mass. Further detailed studies of even older rocks from various continents are of utmost importance in order to clear up this issue.

Concluding remarks

1. Mafic lavas and the Konchozero peridotite sill constitute the upper part of the Onega plateau, a fragment of the major flood basalt province in the SE Baltic Shield.
2. The parental magmas contained 8–10% MgO and were derived from melts formed in the garnet stability field at depths 80–100 km in a mantle plume. Their

differentiation occurred in subcrustal magma chambers and was controlled by two major liquidus phases, clinopyroxene and olivine. The basalt was then erupted as lava flows and emplaced as sills.

3. The Sm-Nd mineral and Pb-Pb whole-rock isochron ages of 1975 ± 24 and 1980 ± 57 Ma for the Angozero lavas and the sill and the SHRIMP U-Pb zircon age of 1976 ± 9 Ma for the Jatulian basalt reflect the time of emplacement of the Onega plateau lavas and suggest that the formation of the entire sequence occurred within a short time span. These ages coincide with those of picrites and komatiites from the upper part of the Pechenga-Imandra basin in the Kola Peninsula and the Karasjok-Kittilä belt in Lapland. The data suggest that the arrival of a mantle plume head was accompanied by simultaneous catastrophic eruption of vast volumes of basaltic, picritic and komatiitic magmas in the SE Baltic Shield. These probably covered ca. 600,000 km^2 and represent a major contribution of juvenile material to the existing continental crust 2.0 Ga ago.

4. The uppermost Suisaarian suite rocks (Angozero area) have high Nb/Th, Nb/La and Ce/Pb ratios, high positive εNd(T) values, and unradiogenic Pb-isotope characteristics. They represent primitive compositions comparable with those of modern plume-derived lavas such as oceanic island and oceanic plateau volcanic rocks. These parameters are regarded as source characteristics. The lowermost Zaonega suite volcanic rocks, on the other hand, have low Ce/Pb, Nb/Th and Nb/La ratios and negative εNd(T) due to contamination with 10 to 30% felsic crust and thus have much in common with contaminated flood basalts from several CFB provinces.

5. The Nb/U ratios in the uncontaminated Angozero rocks are similar to those found in the modern mantle. This provides evidence that by 2.0 Ga the volume of continental crust similar to the present-day value already existed.

6. The Onega plateau represents a sequence of flood basalts erupted through and onto submerged continental crust. In this respect it represents a type of activity common in the Precambrian but rare in the Phanaerozoic.

Acknowledgements We thank V. Kevlich, N. Ashikhmina and V. Pavlichenko for obtaining mineral separates and handpicking and Jon Snow for helpful comments. Very detailed and constructive reviews by Andrew Kerr and Eero Hanski are appreciated. This study was supported by an Alexander von Humboldt and Max-Planck Institut für Chemie fellowships to I. Puchtel and by the Russian Foundation for Fundamental Research.

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