

Sr–Nd–Pb isotopic compositions of the Kovdor phoscorite–carbonatite complex, Kola Peninsula, NW Russia

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Abstract

The Sr, Nd and Pb isotopic compositions for the Kovdor phoscorite–carbonatite complex (PCC), Kola Peninsula, NW Russia, have been determined to characterize the mantle sources involved and to evaluate the relative contributions of a plume and subcontinental lithospheric mantle in the formation of the complex. The Kovdor PCC is a part of the Kovdor ultramafic–alkaline–carbonatite massif, and consists of six intrusions. The initial isotopic ratios of the analyzed samples, calculated at 380 Ma, display limited variations: ϵ_{Nd} , +2.0 to +4.7; $^{87}\text{Sr}/^{86}\text{Sr}$, 0.70319 to 0.70361 (ϵ_{Sr} , –12.2 to –6.2); $^{206}\text{Pb}/^{204}\text{Pb}$, 18.38 to 18.74; $^{207}\text{Pb}/^{204}\text{Pb}$, 15.45 to 15.50; $^{208}\text{Pb}/^{204}\text{Pb}$, 37.98 to 39.28. The Nd and Sr isotope data of the Kovdor PCC generally fit the patterns of the other phoscorites and carbonatites from the Kola Alkaline Province (KAP), but some data are slightly shifted from the mixing line defined as the Kola Carbonatite Line, having more radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. However, the less radiogenic Nd isotopic compositions and negative $\Delta 7/4$ values of Pb isotopes of the analyzed samples exclude crustal contamination, but imply the involvement of a metasomatized lithospheric mantle source. Isotopic variations indicate mixing of at least three distinct mantle components: FOZO-like primitive plume component, EMI-like enriched component and DMM-like depleted component. The isotopic nature of the EMI- and DMM-like mantle component observed in the Kovdor samples is considered to be inherited from metasomatized subcontinental lithospheric mantle. This supports the previous models invoking plume–lithosphere interaction to explain the origin of the Devonian alkaline carbonatite magmatism in the KAP.

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1. Introduction

Carbonatite magmas are now generally thought to have been derived from the mantle, although their origin is still

controversial. Whether the source of the parental melts is lithospheric or the asthenospheric, and whether they represent primary magmas or differentiated products by separation of immiscible melt fractions or by crystal fractionation from carbonated alkali silicate melts (Bell and Tilton, 2001 and references therein) is still debated. Even though this debate is far from settled, numerous recent studies from carbonatites occurring in extensive

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continental rift systems (e.g., East Africa and Kola Peninsula) have suggested important contributions of mantle plumes and plume–lithosphere interaction in their genesis (e.g., Tilton et al., 1998; Simonetti et al., 1998; Tolstikhin et al., 1999, 2002; Bell and Tilton, 2001).

In the case of the Devonian Kola carbonatites, Kramm (1993) firstly proposed two mantle end-members for the sources involved in their genesis using Sr–Nd isotope data and defined a mixing line, called the Devonian ‘Kola Carbonatite Line (KCL)’; one end-member lies within the enriched quadrant, similar to the EMI component, and the other in the depleted quadrant, similar to the DM component. However, other studies (e.g., for Turiy, Dunworth and Bell, 2001; for Vuoriyarvi and Khibina, Zaitsev et al., 2002) have proposed that at least three or four mantle components are necessary to explain the evolution of ultramafic–alkaline–carbonatitic complexes in the Kola Alkaline Province (KAP). Dunworth and Bell (2001) have proposed that the isotopically depleted end-member of KCL represents a plume (‘Devonian equivalent FOZO’) component derived from the lower mantle rather than DMM. This hypothesis is supported by recent studies of noble gases (He, Ne and Xe) in some Kola alkaline carbonatite massifs, which has shown that their isotopic signatures are similar to those of OIBs and that they contain primordial noble gases from the lower mantle (Tolstikhin et al., 2002).

FOZO is a hypothetical mantle end-member, and is thought to be derived from the lower and/or degassed mantle. It has moderately depleted Sr and Nd isotopic signatures, radiogenic Pb isotopes and elevated $^3\text{He}/^4\text{He}$ ratios (Hart et al., 1992; Hauri et al., 1994). Mantle materials with the isotopic nature of FOZO have now been identified in many plumes worldwide: e.g., Oligocene Afar basalts (Orihashi et al., 2001; Bertrand et al., 2003), Trinidad and Martin Vaz basalts, South Atlantic (Siebel et al., 2000) and Pitcairn basalts, Pacific Ocean (Woodhead and Dewey, 1993). The FOZO component is thus considered to be a possible candidate capable of generating plume-derived carbonatite melts.

In a conventional Nd–Sr isotope correlation diagram, the compositional field of FOZO lies in the mixing array between DMM and EMI and cannot be easily recognized as a distinct mantle end-member. The Pb isotope data are thus a prerequisite for the better understanding of contributions of various mantle components to the genesis of mantle-derived magmas. However, systematic Pb isotope data from the KAP rocks have never been published so far.

In this paper we firstly report Pb isotopic compositions from the Kovdor phoscorite–carbonatite com-

plexes (PCC), Kola Peninsula. The Pb isotopes combined with the Nd and Sr isotope data from the Kovdor PCC are used to constrain the chemical and isotopic nature of their mantle source regions. We evaluate the role of the FOZO-like plume and other mantle components and will discuss plume–lithosphere interaction as a possible mechanism for the genesis of the isotopically heterogeneous KAP rocks.

2. Geological background

The Kola Peninsula, the northeastern segment of the Precambrian Baltic Shield, is characterized by complicated block structures, which were reactivated during several tectono-magmatic episodes. During the Devonian, numerous ultramafic, alkaline and carbonatitic intrusions were emplaced over an area of more than 100,000 km² which extends from eastern Finland to the eastern Kola Peninsula, and comprise the KAP (Fig. 1). Most of the intrusions seem to have been emplaced along old suture or reactivated graben structures. Two major structural lineaments appear to control the emplacement of these intrusions: the NE–SW trending Kontozero Graben and the NW–SE trending Kandalaksha Graben (Dudkin and Mitrofanov, 1994).

More than 20 Devonian alkaline–ultramafic and alkaline complexes occur in the KAP, and fourteen of which contain carbonatites (Bulakh et al., 2004). Various age determinations have shown that the majority of the complexes was formed within the narrow range of 380–360 Ma (Kramm and Sintern, 2004), corresponding to the latest Middle to Upper Devonian.

The Kovdor massif is a complex, multiphase and concentric intrusion (Fig. 2). It occurs along an extension of the NW–SE trending Kandalaksha Graben, and crops out over an area of approximately 40 km². It intrudes Archean gneisses and granitic gneisses of the Belomorian group. The main rock types are from the oldest to youngest: ultramafic cumulate rocks, melilitolites (turjaite, ultramelilitolites), foidolites (melteigite–ijolite–urtites), phoscorite–carbonatite complex (PCC) and nepheline syenite. The ultramafic rocks include olivinites (>85 vol. % olivine) and pyroxenites (>85 vol.% clinopyroxene), and constitute central core of the massif. The melilitolites form an incomplete ring around the ultramafic core. The main rock type of the melilitolites is turjaite consisting of melilite (25–50 vol.%), nepheline (15–30 vol.%), phlogopite (20–25 vol.%) and diopside (5–25 vol.%). The ultramelilitolites (>85 vol.% melilite and <5 vol.% nepheline) form lenticular bodies in the turjaites. The foidolites are very heterogeneous in terms of texture, grain size and mineralogy. They are mainly

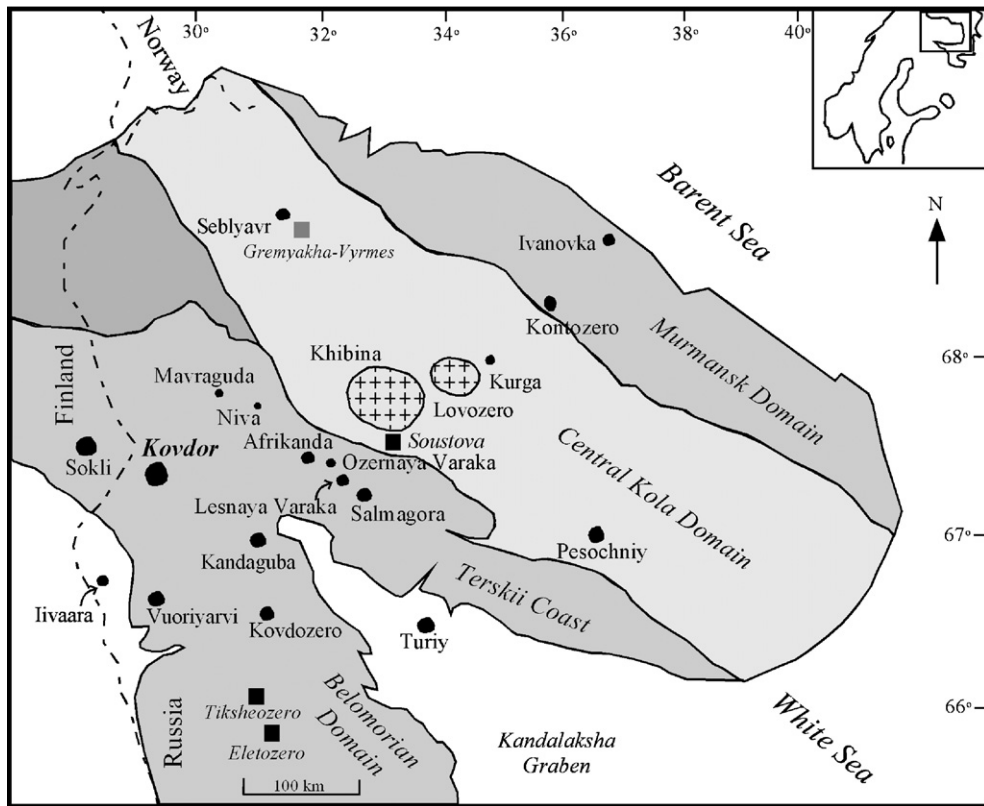


Fig. 1. Map of the Kola Alkaline Province showing the location of the Proterozoic and Paleozoic alkaline intrusions (modified from Bell et al., 1996). Oval shapes filled with cross (Khibina and Lovozero), Paleozoic peralkaline (agpaitic) nepheline syenite complexes; irregular black shapes, Paleozoic alkaline ultramafic complexes; grey square, Proterozoic gabbroic complexes.

composed of clinopyroxene (30–65 vol.%), nepheline (10–50 vol.%), titanite (5–10 vol.%) with minor apatite, magnetite, perovskite, biotite and calcite. The nepheline syenites occur as veins or dikes in the central part of the massif. These various ultramafic and alkaline silicate rocks are considered to have been formed from several batches of a carbonated olivine melanephelinite parental magma by a mechanism involving fractional crystallization, magma mixing and/or contamination (Verhulst et al., 2000).

In the south-western part of the complex, around the contact between pyroxenite and ijolite units, a stock of phoscorite and carbonatite, the main target of this study, is emplaced. Phoscorites are essentially composed of apatite, magnetite and forsterite or diopside (\pm phlogopite or tetraferriphlogopite). They are usually associated with carbonatites and share the same mineral assemblage and composition (Krasnova et al., 2004). Enrichment in magnetite and silicates visually distinguishes phoscorites from carbonatites.

Six successive stages (Ia, Ib, IIa, IIb, IIIa and IIIb) of phoscorite emplacement are distinguished. In the last

four stages, the phoscorites are associated with the carbonatites. Precise mineral age dating (Amelin and Zaitsev, 2002) indicates that all constituents of the Kovdor PCC were emplaced within a very short period at \sim 378 Ma. The rock types and successions of the Kovdor PCC are summarized in Table 1. The phoscorites of the Kovdor PCC are very heterogeneous in terms of relative abundance of rock-forming minerals and are conventionally named by Russian geologists using abbreviations composed of capitalized first letters of their prevailing minerals; e.g., Forsterite–Magnetite phoscorite refers to FM phoscorite and indicates that the modal amount of forsterite (F) is larger than that of Magnetite (M). One of the characteristic features in the Kovdor PCC is the occurrence of various kinds of breccias which clearly separate the successive intrusive stages.

3. Analytical methods

The Sr and Nd isotope analyses were performed on twelve representative carbonatites and associated

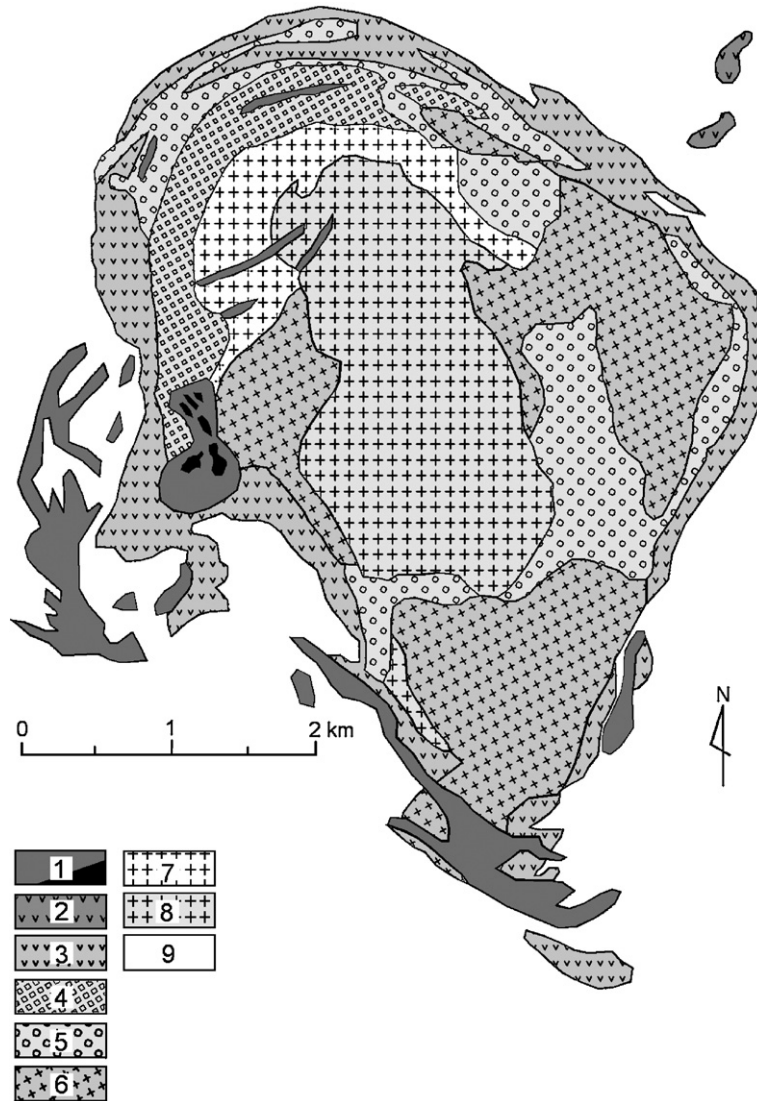


Fig. 2. Geological map of the Kovdor massif in the Kola Alkaline Province (after Ternovoy et al., 1969). 1, Carbonatites and phoscorites; 2, nepheline syenite; 3, ijolite–melteigite; 4, monticellite-bearing melilitolites; 5, melilitolites; 6, pyroxenite; 7, phlogopite–diopside–olivine rocks; 8, dunite/peridotite (olivinite); 9, Archean basement.

phoscorites from the Kovdor PCC. Among them, six samples (two C1, three C2 and one MCF phoscorites (IIaP1)) which have relatively low U/Pb and Th/Pb ratios were selected for Pb isotope analyses. All samples were analyzed with unleached whole rock powders and completely decomposed in a mixture of HF–HClO₄ for Sr–Nd isotopic analyses and in a mixture of HF–HNO₃ for Pb isotopic analyses. Isotope ratios were measured using a Finnigan MAT-262 thermal ionization mass spectrometer at the Laboratory for Radiogenic Isotope Geochemistry, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China. Rb, Sr, Sm and Nd concentrations were determined by the

isotope dilution method. Chemical separation of Rb, Sr, Sm and Nd was carried out by the cation exchange technique. Total blank levels for the whole procedure were below 200 pg for Rb and Sr, and 50 pg for Nd and Sm. Measured ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd ratios were normalized to ⁸⁶Sr/⁸⁸Sr=0.1194 and ¹⁴⁶Nd/¹⁴⁴Nd=0.7219, respectively. Replicate analyses of NIST SRM-987 and La Jolla Nd gave ⁸⁷Sr/⁸⁶Sr=0.710208±0.000009 (N=10) and ¹⁴³Nd/¹⁴⁴Nd=0.511826±0.000008 (N=12), respectively. Pb was eluted and purified in HBr media using standard anion exchange technique. Total procedure blanks were less than 50 pg. Pb isotope data were corrected for fractionation of 0.1%

Table 1

Succession of the Kovdor phoscorite and carbonatite intrusions (after Krasnova and Kopylova, 1988)

Stage		Phoscorite	Carbonatite and other 'vein-like' rocks
III	IIIb	Magnetite–dolomite–richterite with ilmenite (MDR) (P4) Breccia	Calcite–dolomite carbonatite with labuntsovite (CD6) Dolomite carbonatite with kovdorskite (D5) Calcite–dolomite carbonatite with ilmenite (CD4)
	IIIa	Magnetite–dolomite–forsterite with ilmenite (MDF), magnetite–dolomite–calcite (MDC) (P3) Breccia	Dolomite carbonatite (D3)
II	IIB	Magnetite–calcite–forsterite with tetraferriphlogopite (MCF(tphl)–rare metal (RM)) (P2) Breccia	Calcite carbonatite with tetraferriphlogopite (C2)
	Ia	Magnetite–calcite–forsterite with green phlogopite (MCF), magnetite–calcite (MC) (P1) Breccia	Calcite carbonatite (C1)
I	Ib	Magnetite–apatite–forsterite (MAF) Breccia	Apatite (A)
	Ia	Forsterite (F), forsterite–apatite (FA)	

per atomic mass unit based on repeated analyses of NIST SRM-981 Pb standard. Pb, U and Th concentrations were determined using an ICP-MS at the Korea Polar Research Institute (KOPRI). Detailed analytical procedures for ICP-MS were presented by Hur et al. (2003). Raw data obtained were calculated using the Isoplot program (Ludwig, 2001), giving $2\sigma_m$ error. Details of analytical techniques on chemical separation and measurement are described in Chen et al. (2002). The initial Sr, Nd and Pb isotopic ratios at 380 Ma were calculated using the decay constants of 1.42×10^{-11} year $^{-1}$ for ^{87}Rb , 6.54×10^{-12} year $^{-1}$ for ^{147}Sm , 1.55×10^{-11} year $^{-1}$ for U^{238} , 9.85×10^{-10} year $^{-1}$ for

U^{235} and 4.95×10^{-11} year $^{-1}$ for Th^{232} (Steiger and Jäger, 1977).

4. Results and discussion

The measured Sr, Nd and Pb isotopic ratios with age (380 Ma) corrected values are given in Tables 2 and 3. As an index for isotopic compositions that corresponds to the mantle sources, we will use the present-day mantle end-member components (DMM, HIMU, EMI and FOZO) defined from oceanic volcanic rocks by Hart et al. (1992). All analyzed samples are characterized by high Sr contents; 2556–9649 ppm in the carbonatites

Table 2

Rb–Sr and Sm–Nd isotope data, calculated ϵ -values and model ages for the Kovdor phoscorites and carbonatites

Sample	Rock	Rb (ppm)	Sr (ppm)	$^{87}\text{Sr}/^{86}\text{Sr}$	$(^{87}\text{Sr}/^{86}\text{Sr})_t$	$\epsilon_{\text{Sr}(t)}$	Sm (ppm)	Nd (ppm)	$^{143}\text{Nd}/^{144}\text{Nd}$	$(^{143}\text{Nd}/^{144}\text{Nd})_t$	$\epsilon_{\text{Nd}(t)}$	T_{DM}
KV1	IaC1	2.8	2618	0.703426±11	0.703409	−9.1	11.0	77.2	0.512514±5	0.512300	3.0	0.8
KV3	IaC1	38.3	2556	0.703718±11	0.703484	−8.1	10.5	72.9	0.512511±13	0.512294	2.8	0.8
KV5	IaP1	0.3	972	0.703246±16	0.703241	−11.5	3.4	22.9	0.512591±11	0.512370	4.3	0.7
KV7	IaP1	2.2	684	0.703241±7	0.703192	−12.2	2.0	14.1	0.512604±9	0.512385	4.6	0.7
KV9	IIBc2	1.0	4414	0.703461±9	0.703458	−8.5	12.2	84.7	0.512467±7	0.512250	2.0	0.8
KV18	IIBc2	2.8	3181	0.703415±17	0.703401	−9.2	8.5	59.6	0.512564±9	0.512349	3.9	0.7
KV19	IIBc2	21.2	4792	0.703507±9	0.703483	−8.7	12.5	85.6	0.512507±4	0.512287	2.7	0.8
KV35	IIBc2	7.1	9649	0.703498±14	0.703487	−8.0	20.4	149.4	0.512592±10	0.512387	4.6	0.7
KV37	IIBc2	2.0	4541	0.703551±15	0.703544	−7.2	10.4	50.0	0.512622±13	0.512308	3.1	0.9
KV14	IIBp2	2.5	1058	0.703503±9	0.703467	−8.3	8.5	58.3	0.512609±7	0.512389	4.7	0.7
KV15	IIBp2	34.9	678	0.704268±6	0.703462	−8.4	4.0	30.5	0.512541±6	0.512346	3.9	0.7
KV39	IIBp2	3.6	851	0.703681±7	0.703614	−6.2	2.8	17.0	0.512565±7	0.512314	3.2	0.8

$(^{87}\text{Sr}/^{86}\text{Sr})_t$, $\epsilon_{\text{Sr}(t)}$, $(^{143}\text{Nd}/^{144}\text{Nd})_t$ and $\epsilon_{\text{Nd}(t)}$ at 380 Ma were calculated using the following parameters: $^{87}\text{Sr}/^{86}\text{Sr}_{\text{UR}(0)}=0.7045$, $^{87}\text{Rb}/^{86}\text{Sr}_{\text{UR}(0)}=0.0827$, $\lambda^{87}\text{Rb}=1.42 \times 10^{-11}$ year $^{-1}$, $^{143}\text{Nd}/^{144}\text{Nd}_{\text{CHUR}(0)}=0.512638$, $^{147}\text{Sm}/^{143}\text{Nd}_{\text{CHUR}(0)}=0.1966$, $\lambda^{147}\text{Sm}=6.54 \times 10^{-12}$ year $^{-1}$. Model ages with reference to the depleted mantle (T_{DM}) were calculated by the following parameters: $^{143}\text{Nd}/^{144}\text{Nd}_{\text{DM}(0)}=0.513153$, $^{147}\text{Sm}/^{144}\text{Nd}_{\text{DM}(0)}=0.2136$ (Liew and McCulloch, 1985).

Abbreviations: IaC1, Ia stage C1 calcite carbonatite; IIBc2, IIB stage C2 calcite carbonatite; IaP1, Ia stage P1 phoscorite; IIBp2, IIB stage P2 phoscorite.

Table 3
Pb isotope data for phosphorites and carbonatites from the Kovdor complex

Sample no.	Rock type	Concentration			Measured ratio			μ	κ	Initial ratio (at 380 Ma)		
		U (ppm)	Th (ppm)	Pb (ppm)	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$			$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
KV1	IlaC1	0.04	0.13	1.41	18.754	15.482	38.373	1.812	6.271	18.645	15.476	38.254
KV3	IlaC1	0.03	0.55	1.74	18.481	15.454	38.765	1.028	20.755	18.419	15.451	38.371
KV7	IlaP1	0.24	0.84	1.21	19.138	15.533	38.859	12.576	46.473	18.380	15.492	37.977
KV9	IibC2	0.12	1.22	2.23	18.674	15.479	38.971	3.415	36.264	18.468	15.468	38.283
KV18	IibC2	0.01	0.20	1.93	18.668	15.495	38.261	0.329	6.769	18.648	15.494	38.132
KV19	IibC2	0.33	2.95	1.62	19.566	15.542	41.675	13.764	126.417	18.737	15.498	39.276

Abbreviations: IlaC1, Ila stage C1 calcite carbonatite; IibC2, Iib stage C2 calcite carbonatite; IlaP1, Ila stage P1 phosphorite.

and 678–1058 ppm in the phosphorites. Sm, Nd, U, Th and Pb concentrations are low in both carbonatites and phosphorites.

4.1. Sr, Nd and Pb isotopic compositions

The Sr and Nd isotope data obtained from this study are combined with the previously published data for the Kola carbonatitic rocks and depicted on a conventional Nd–Sr isotope correlation diagram (Fig. 3). The initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the analyzed samples, calculated at 380 Ma, display a limited variation ranging from 0.70319 to 0.70361 and 0.51225 to 0.51239, respectively (Table 2). These

compositions are similar to the results reported by Verhulst et al. (2000) for whole rock carbonatites and phosphorites from the Kovdor complex. Because crustal contamination is unlikely to influence the Sr and Nd isotopic compositions of carbonatites, these ratios are considered to reflect the isotopic compositions of their mantle sources. The Sr and Nd isotopic signatures of the Kovdor carbonatites are not distinguished from those of the associated phosphorites (Fig. 3). This feature has been also recognized in the Sokli, Turiy, Seblyavr and Vuoriyarvi PCCs (Kramm et al., 1993; Dunworth and Bell, 2001; Brassinnes et al., 2003). Most data obtained from PCCs (Kovdor, Sokli, Turiy, Seblyavr and Vuoriyarvi) in the KAP fall into the

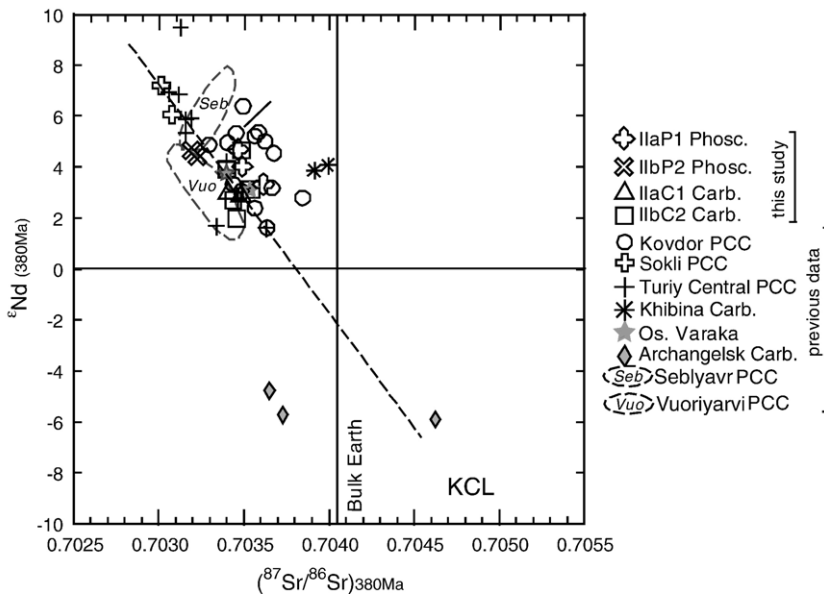


Fig. 3. $\epsilon_{\text{Nd}}(380\text{ Ma})$ versus initial $^{87}\text{Sr}/^{86}\text{Sr}(380\text{ Ma})$ isotope plot for the Kovdor phosphorites and carbonatites. The data are compared to those of other Kola phosphorites and carbonatites. Previous data sources: Sokli, Kramm et al. (1993); Khibina, Kramm and Kogarko (1994); Kovdor, Verhulst et al. (2000); Archangelsk, Mahotkin et al. (2000); Turiy, Dunworth and Bell (2001); Sebljavr and Vuorijarvi, Brassinnes et al. (2003). All epsilon values are calculated at 380 Ma. KCL, Kola Carbonatite Line from Dunworth and Bell (2001). Carb., carbonatite; Phosc., phosphorite; PCC, phosphorite–carbonatite complex.

depleted quadrant, but are less depleted compared to the MORB-like DMM component, and generally cluster along the mixing path defined as the KCL. In contrast, carbonatites from Khibina and Archangelsk, which are not associated with phoscorites but nepheline syenites and kimberlites, respectively, show substantially different isotopic characteristics from those of the Kola PCCs. They are more radiogenic compared to the Kola PCCs.

In the diagrams of $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, the analyzed samples, with one exception, plot below and above the Northern Hemisphere Reference Line (NHRL) of Hart (1984), respectively (Fig. 4). Thus, the samples are characterized by negative $\Delta 7/4$ and positive $\Delta 8/4$ values of Pb isotopes. The initial and present-day isotopic compositions plot to the right of and below the 4.55 Ga geochron and the Pb evolution curve of Stacey and Kramers (1975) and thus appear to have not been contaminated by a high U/Pb upper crustal component (Fig. 4a). It is noteworthy that the present-day and the initial Pb isotopic ratios define a linear array in Fig. 4a, but on the other hand, a more scattered pattern is apparent in Fig.

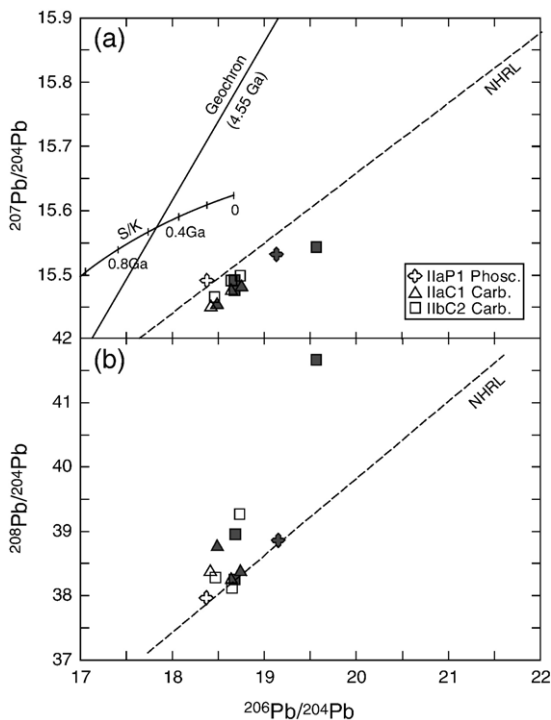


Fig. 4. Plots of (a) $^{207}\text{Pb}/^{204}\text{Pb}$ and (b) $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ for phoscorites and carbonatites from the Kovor massif. Closed and open symbols are measured and initial isotopic compositions, respectively. NHRL, Northern Hemisphere Reference Line from Hart et al. (1992); S/K, The Stacey and Kramers (1975) Pb evolution curve.

4b. In both cases, the initial isotopic ratios of the samples show a tighter cluster than the measured values. This implies that the U–Th/Pb system of the Kovdor PCC was not significantly disturbed since its emplacement. Minor effects of alteration cannot be ruled out and such an effect may be responsible for the data (one IlaC2 sample) slightly different than the main isotopic composition. However, the analyzed samples display almost identical initial $^{206}\text{Pb}/^{204}\text{Pb}$ ratios.

In contrast to Sr and Nd isotopes, Pb isotopes are more sensitive to contamination by isotopically distinctive crustal Pb. It is thus important to evaluate the effects of crustal contamination. The low $^{207}\text{Pb}/^{206}\text{Pb}$ ratios of the Kovdor PCC (negative $\Delta 7/4$ values) preclude significant contributions of upper crustal material. Alternatively, the lower continental crust, which is regarded as a major reservoir of unradiogenic Pb, can be considered as a contaminant. Such unradiogenic Pb is thought to be the result of U depletion, together with other large-ion lithophile elements (LILE), during granulite-facies metamorphism (Rudnick and Presper, 1990; Cohen et al., 1991), or due to primary low U/Pb ratios in areas of significant underplating with gabbroic cumulates (van Calsteren et al., 1993). In both cases, lower crustal rocks have low Rb/Sr as well as low U/Pb and, with time, contain unradiogenic Pb and also relatively unradiogenic Sr isotopic signatures. However, no correlation between Sr and Pb isotope ratios, and between U/Pb and Pb isotopes in the Kovdor PCC (Tables 2 and 3) further suggests the inherited lead isotopes. Additionally, the available O and C isotope compositions of the Kovdor PCC (Zaitsev and Bell, 1995) support that they have not been significantly affected by crustal contamination. The positive $\Delta 8/4$ and negative $\Delta 7/4$ values of Pb isotopes of the Kovdor PCC could indicate that the Th/U and U/Pb ratios may have been increased by small melt fractions or fluids relatively recently, reflecting recent metasomatism of mantle sources.

We therefore conclude for the data shown in Fig. 4: (1) the negative $\Delta 7/4$ values suggest an ancient evolution of the Pb isotopes in a low- μ environment ($\mu = ^{238}\text{U}/^{204}\text{Pb}$); (2) on the other hand, the relatively high $^{206}\text{Pb}/^{204}\text{Pb}$ for a given $^{207}\text{Pb}/^{204}\text{Pb}$ ratios record a more recent evolution in a high- μ environment; (3) the scattered data in a plot of $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ (Fig. 4b) are indicative of a variable time-integrated Th/U ratios.

4.2. Contributing mantle source(s)

The initial Pb isotopic ratios (to the right of the geochron, Fig. 4a), ε_{Sr} (–12.2 to –6.2) and ε_{Nd} (+2.0 to

+4.7) values indicate that the Kovdor PCC has originated from source regions with a time-integrated enhancement of the U/Pb ratio, somewhat lower Rb/Sr ratio with respect to Bulk Earth and a time-integrated depletion in LILEs relative to CHUR.

Fig. 5 shows the correlations between Sr and Pb isotopes for the Kovdor PCC, and compares them with other young carbonatites in the world (<120 Ma, Tilton et al., 1998; Bell and Tilton, 2001). The reference carbonatites represent samples from three continents plus two additional oceanic settings (Tilton et al., 1998). Tilton et al. (1998) have noted that the reference carbonatites have similar isotopic patterns in spite of very diverse locations. Additionally, Bell and Tilton (2001) pointed out that the Chilwa (southern Tanzania) and Napak (East African rift group) carbonatites show isotopic similarities to the reference carbonatite group, and their isotopic compositions suggest a certain contribution of the FOZO component to the sources. The Chilwa and Napak carbonatites are considered here as the same group with the reference

carbonatites. To evaluate the relative contributions of specific mantle end-members in generating the Kovdor rocks, we have to estimate the isotope ratios of mantle end-members (DMM, HIMU, EMI and FOZO) at 380 Ma with reasonable accuracy. However, the absence of seafloor volcanics older than 190 Ma prevents direct analysis of mantle reservoirs beyond that age. Furthermore, it is difficult to compare directly the isotope data of young carbonatites worldwide with those of the Kovdor PCC due to the large age differences. We will use therefore our measured Sr and Pb isotope data for a comparison with a database from young carbonatites. The isotopic compositions of young carbonatites used in our diagrams are also corrected to present-day values (Bell and Tilton, 2001). There is no significant difference between the measured values and the initial Sr and Pb ratios because of the extremely low Rb/Sr and low U/Pb and Th/Pb ratios of the analyzed samples. We excluded Nd isotope compositions in this comparison because of considerable differences between the present-day $^{143}\text{Nd}/^{144}\text{Nd}$

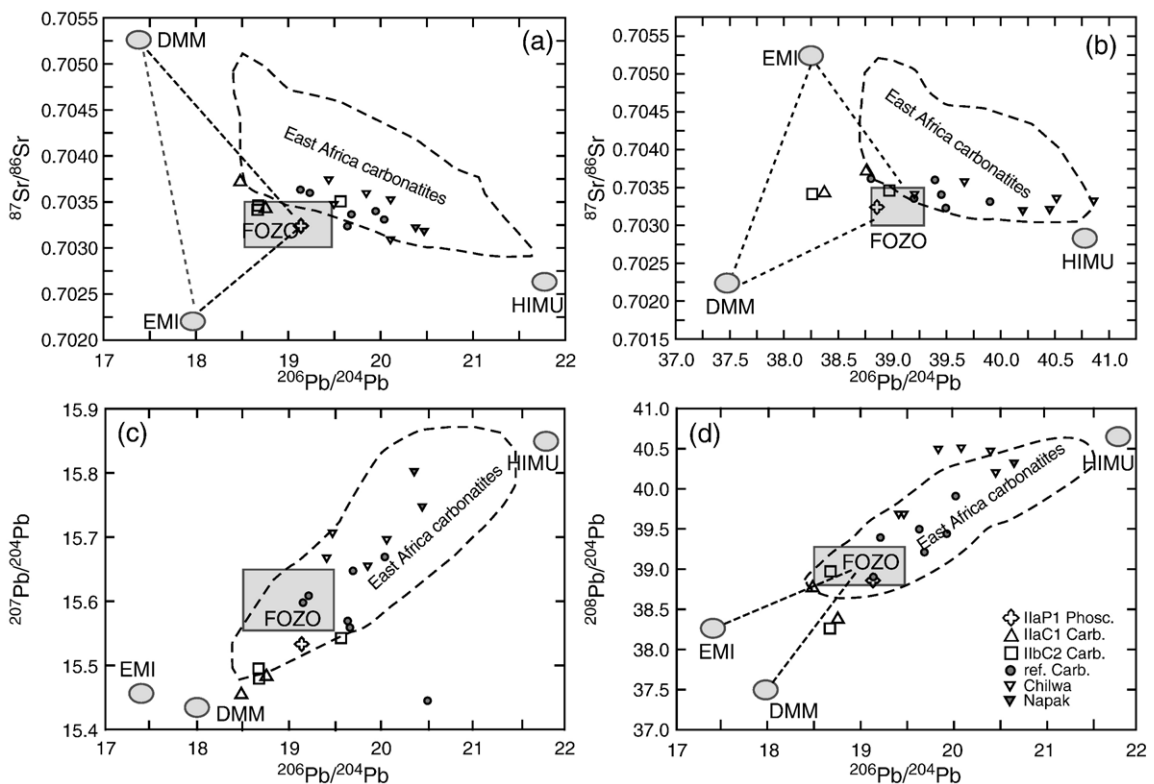


Fig. 5. Correlation diagrams of (a) $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, (b) $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{208}\text{Pb}/^{204}\text{Pb}$, (c) $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ and (d) $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ for the Kovdor PCC. Isotopic compositions of young (<120 Ma) carbonatite complexes worldwide are added to the diagrams. Data sources for carbonatites: Chilwa and Napak, East Africa (Simonetti and Bell, 1994; Bell and Tilton, 2001); reference carbonatites (Nelson et al., 1988; Tilton et al., 1998; Bell and Tilton, 2001). EMI, HIMU and DMM are approximations of mantle end-members taken from Hart et al. (1992). Values for FOZO ('Focal Zone') taken from Hauri et al. (1994).

and the initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in our analyzed samples due to the relatively high Sm/Nd ratios and low Sm and Nd concentrations. This comparison provides an idea for the evaluation of the minimum number of isotopically distinct mantle sources involved and for the characterization of the source composition of the Kovdor PCC melts.

The Sr and Nd isotopic signatures of many Kola carbonatites investigated have commonly been intermediate between the depleted mantle (DMM) and the enriched mantle (EMI) component, and thus have been usually explained by simple binary mixing of the two components. However, our combined Sr and Pb isotope systematics suggests another. The Kovdor phoscorites and carbonatites have slightly higher $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios compared to those of DMM and EMI, indicating involvement of a more radiogenic Pb component in the sources (Fig. 5a and b). According to the trends observed in Fig. 5, the additional radiogenic Pb component is likely to have been derived from FOZO which is commonly assumed to represent lower mantle material. The influence of a HIMU-like source appears minor on the basis of these diagrams. Moreover, generally HIMU islands such as St. Helena, Tubuaii, and Mangai are known to have negative $\Delta 8/4$ values of Pb isotopes. Thus, the high $^{208}\text{Pb}/^{204}\text{Pb}$ values at a given $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of the Kovdor samples (Fig. 5d) do not support the possibility of a HIMU-like component contribution to the sources.

The isotope compositions of the reference carbonatite group are comparable to those of the Kovdor PCC (Fig. 5). The Pb isotopic values of the reference group show a slightly more radiogenic nature compared to that of FOZO. In contrast, the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of the Kovdor samples well correspond to that of the FOZO component, whereas the $^{207}\text{Pb}/^{204}\text{Pb}$ ratios are slightly lower than that of FOZO. Even if the Pb isotopes from the Kovdor PCC and the reference group are slightly different from the compositional field of FOZO, their overall compositions show an obvious contribution of the FOZO component to the sources of the Kovdor PCC and reference group. The higher $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ isotopic ratios of the reference group can be explained by an additional involvement of a HIMU-like mantle component to their sources.

Although the most pronounced isotopic signature shown in the samples analyzed is the FOZO-like mantle component, it is unlikely that the Kovdor PCC has been derived from a simple melting of a deep FOZO-like mantle plume alone. The contribution of a depleted component is also necessary to account for the lower $^{207}\text{Pb}/^{204}\text{Pb}$ isotope ratios compared to that of the FOZO

end-member. In fact, they plot closer to the field of DMM than to FOZO in a $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 5c). Correlations between ϵ_{Nd} and initial Sr ratios and between Sr and Pb isotopes (Figs. 3 and 5) show that the Kovdor samples are not as depleted as the DMM component, and do not seem to define a simple binary mixing array between the FOZO and DMM, or EMI end-members. Instead, the isotopic compositions of the Kovdor samples suggest that the source reservoir might involve three mantle components (DMM, FOZO and EMI).

4.3. Plume–lithosphere interaction

Recently, many petrogenetic models proposed for the generation of carbonatites place great emphasis on the role of a plume. Actually, much scientific evidence advocates that many carbonatites and related alkaline silicate rocks are genetically related to plume activities, e.g., the presence of carbonatites associated with flood basalts (Decan, Paraná) or related to known plumes (Canary Island, the Cape Verdes) and the primitive nature of noble gas isotopes in some carbonatites (Gerlach et al., 1988; Bell and Tilton, 2001; Tolstikhin et al., 2002; Bell and Rukhlov, 2004). However, the actual role of plumes in the genesis of carbonatite magma is still contentious; does it only provide a heat source and cause melting of the overlying lithospheric mantle or is it also a major contributing melt source?

Various empirical lines of evidence, backed by theoretical considerations, support the view that mantle plumes are radially zoned in composition and temperature (Campbell, 1998; Bijwaard and Spakman, 1999; Breddam et al., 2000). During the ascent of a rising plume from the core–mantle boundary, the plume head is considered to entrain surrounding upper or lower mantle materials. The classic model of Griffiths and Campbell (1990) predicts that the plume head will be dominated by lower mantle material. On the other hand, Stuart et al. (2000) presumed that the plume head would be dominated by upper mantle material if a plume sourced at the core–mantle boundary stalls at the 660 km thermal boundary layer. Both cases predict that the plume head, where partial melting dominantly occurs by adiabatic decompression, may incorporate several different parts of the mantle. With this context, Bell and Tilton (2001) suggested that both HIMU and EMI sources are stored within the deep mantle, and that the East African carbonatites have been derived from a heterogeneous mantle plume covering the complete isotopic spectrum between HIMU and EMI signatures.

This plume inhomogeneity model effectively explains large variations of isotope data shown in ultramafic, alkaline and carbonatitic complexes of the KAP. The widespread voluminous alkaline carbonatite magmatism during the very short period (360–380 Ma) seems to be a convincing evidence for the impact of a mantle plume beneath the Kola Peninsula during the late Devonian. The isotope data obtained from this study along with the noble gas isotope data from several carbonatite complexes including Kovdor PCC in the KAP (Tolstikhin et al., 2002) also support a certain contribution of plume material in the generation of the Kola carbonatites. From the noble gas concentrations (He, Ne and Ar) and isotopic systematics, Tolstikhin et al. (2002) considered that the KAP magmatism was related to a lower mantle plume material with an initial $^4\text{He}/^3\text{He}$ of about 3×10^4 . However, the involvement of lithospheric mantle during the migration of degassed plume material rising from a deep mantle source to the shallow level across the solidus of carbonated peridotite cannot be ruled out. It has been suggested in many cases that the large variations in isotope data of carbonatites or alkaline provinces are related to the interaction between upwelling plume and continental lithosphere (e.g., Stewart and Rogers, 1996; Simonetti et al., 1998; Pik et al., 1999). Another important factor in order to estimate the relative importance of plume mantle versus lithospheric mantle is the thickness of the lithosphere which is considered to primarily control the amount of upwelling and hence the degree of melting by decompression within the plume head, and contamination of magmas by subcontinental lithospheric mantle (Gibson et al., 1995; White and McKenzie, 1995). The present-day lithospheric thickness beneath the Kola Peninsula is estimated to be at least 200 km (Artemieva, 2003) and is considered to have been similar in Devonian times (Mahotkin et al., 2000). Instead of continental flood basalts, the presence of mafic potassic rocks such as diamondiferous kimberlites and ultramafic lamprophyres in Terskii coast (Beard et al., 2000) also supports the existence of a thick cratonic lithosphere beneath the Kola region. This thick cratonic lithosphere raises the possibility of the interaction of magmas derived from a plume en route to the surface with lithospheric mantle.

As we demonstrated, the Sr–Nd–Pb isotope systematics from the Kovdor PCC is better explained by the mixing of three mantle end-members (DMM-, EMI- and FOZO-like components). The imprints of DMM- and EMI-like isotopic signatures are likely to be inherited from metasomatized subcontinental litho-

sphere which is commonly considered to have both enriched and depleted isotopic signatures. The depleted upper mantle is considered to have been processed by metasomatic melts (10% to 30%) originating by extraction of ~0.3% to 0.5% from the asthenospheric mantle (Taintan and McKenzie, 1994; McKenzie and O’Nions, 1995). These repeated metasomatic processes for a long time interval allow enriched isotopic signatures to be generated in subcontinental lithospheric mantle (Richardson et al., 1984).

The Pb isotopic signatures from the Kovdor PCC provide another evidence for an involvement of the metasomatic lithosphere in their sources. Tolstikhin et al. (2002) also suggested model age of about 450 to 750 Ma for the upper mantle metasomatism in the KAP based on the previous Rb–Sr age results (Kramm et al., 1993), and insisted contributions of three components into the sources of parental melts of the Kola rocks: plume material, metasomatized upper mantle and air-saturated ground water. However, it is very difficult to estimate the exact proportions of the isotopically distinct mantle components involved to the sources because of the poor information on the compositions of mantle end-members during the Devonian.

5. Concluding remarks

The Sr and Nd isotopic ratios of the Kovdor PCC are compatible with the previous results and fit well the patterns of the other phoscorites and carbonatites in the KAP. The Pb isotope data on the Kovdor PCC, firstly obtained from this study, provide an important insight into the interaction between a plume derived from deep mantle and the metasomatized lithospheric mantle during the magma generation and ascent. The most striking feature of the Sr–Nd–Pb correlations is that the Kovdor data plot close to the field of FOZO. Although the most pronounced isotopic signature observed is the FOZO-like component, the data array cannot be explained by a simple melting of a FOZO-like mantle plume alone. The Pb isotopic systematics (negative $\Delta 7/4$ and positive $\Delta 8/4$ values of Pb isotopes) of the Kovdor PCC clearly suggests the contribution of metasomatized lithospheric mantle material to the source region. The Sr–Nd–Pb isotopic compositions of the Kovdor PCC indicate a mixing of three distinct mantle reservoirs: FOZO-like primitive plume component, EMI-like enriched component and MORB-like DMM component. It is thus concluded that the primary magma of the Kovdor PCC was formed from the interaction between the previously metasomatized lithospheric mantle and a lower mantle plume.

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