Mantle transect of the Karelian Craton from margin to core based on P-T data from garnet and clinopyroxene xenocrysts in kimberlites

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Abstract

Peridotitic garnet and clinopyroxene xenocrysts from the 1.2 Ga Lentiira–Kuhmo and the 760 Ma Kuusamo kimberlite fields, emplaced within the Central Karelian Craton, have been studied using major and trace element geochemistry to obtain information on the stratigraphy, compositional variability, and evolutionary history of the underlying lithospheric mantle. An earlier study on the 600 Ma Kaavi–Kuopio kimberlite field showed that near the SW margin of the craton the 230-km-thick subcontinental lithospheric mantle (SCLM) exhibits a well-developed 3-layer structure indicative of episodic construction. New Ni in garnet thermometry on samples farther into the craton, from Kuhmo, Lentiira, and southern and northern Kuusamo, gives temperature ranges of 800–1400, 700–1450, 700–1450 and 600–1300 °C, respectively, and when extrapolated to the geotherm determined for the Kaavi–Kuopio area, indicates sampling intervals of c. 100–240, 80–250, 80–250 and 65–210 km, respectively. The results demonstrate that the SCLM of the Karelian Craton reaches its greatest thickness in the Lentiira–Kuhmo and southern Kuusamo area and thins towards the North and South. The mantle stratigraphy of the craton core, compared to the craton edge, shows less compositional variation, with only two distinguishable horizons corresponding generally to the two deepest layers at Kaavi–Kuopio. Additionally, based on the Mg# of pyropes, the level of depletion of peridotites comprising the mantle lithosphere in the central craton is significantly higher than at the craton edge represented by Kaavi–Kuopio. Coupled with the rarity of mantle-derived chrome diopside, the implication is that this portion of the mantle either underwent high levels of partial melting to produce very refractory residua or experienced less melt-modification than average SCLM subsequent to formation. The craton core provides a wide diamond window up to 110 km thick compared to c. 40 km at the craton edge.

Keywords: kimberlite, lithosphere, mantle, xenocrysts, pyrope, Finland

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1 Introduction

The Archean eastern and northern parts of the Fennoscandian Shield comprise the Karelian–Kola–Kuloi-megacraton, which hosts numerous kimberlitic rocks representing various ages and mineralogical types

(Fig. 1). From oldest to youngest the known kimberlite provinces in the region are: 1. Kemozero, located on the northern shore of Lake Onega and dated at 1760 Ma (Lobkova et al., 2006). 2. Lentiira–Kuh-

Figure 1. Generalized geological map of Fennoscandia. The Archean/Proterozoic boundary marks the subsurface extent of the Archean craton as determined by Nd isotopes. The black diamonds represent diamond-bearing kimberlites and lamproites, the diamond mines in the Arkhangelsk peninsula are also indicated. The lithospheric mantle cross section across the Karelian Craton presented in this study is indicated by a dashed line.

mo–Kostamuksha, covering both sides of the Finnish–Russian border, contains 1200 Ma dykes with characteristics of both olivine lamproites and Group II kimberlites (O'Brien et al., 2007; Antonov & Ulianov, 2008). 3. Kaavi–Kuopio, where the bulk of the Finnish Group I kimberlites were emplaced about 600 Ma ago (O'Brien & Tyni, 1999; O'Brien et al., 2005). 4. Kuusamo, where both Group I and Group II kimberlites have recently been discovered. Two of the Group I kimberlites have been dated and give the same 760 Ma age within error. For this paper the Kuusamo kimberlites have been divided into northern (N) and southern (S) Kuusamo fields. 5. Arkhangelsk and Terskii, which lie farther to the north and east, contain kimberlites of 360 to 380 Ma in age (Mahotkin et al., 2000). These kimberlites represent members of the Devonian Kola Alkaline Province. Arkhangelsk kimberlites are divided geographically and minera-

logically into Eastern mica-poor (Group I) and Western micaceous (Group II) subtypes (e.g. Beard et al., 2000). The Western group hosts Europe's first diamond mine, Lomonosova, which covers six individual kimberlite pipes. In the near future mining activity will also start in the Eastern Group when production at the Grip pipe commences.

1.1 Summary of results from Kaavi-Kuopio

Previous studies of Kaavi–Kuopio kimberlite xenoliths and xenocrysts have demonstrated that the subcontinental lithospheric mantle (SCLM) sampled by these magmas at the Archean–Paleoproterozoic boundary is distinctly stratified. The mantle xenolith suite demonstrates that a shallow zone (<900 °C) of garnet–spinel harzburgite is underlain by a zone of more fertile garnet peridotites (180–240 km, Peltonen et al., 1999). Systematic studies of mantle xenocrysts have provided additional information on the mantle section (Lehtonen et al., 2004; Lehtonen, 2005a). Three distinct layers (A, B and C) can be identified based on garnet xenocryst compositions and Ni-derived temperatures, when extrapolated to the local geotherm (Kukkonen & Peltonen, 1999; Kukkonen et al., 2003): (a) A shallow, 60–110 km, garnet–spinel peridotite layer; (b) A variably depleted peridotitic horizon from 110 to 180 km containing diamond-indicative subcalcic harzburgitic garnet, (c) A deep layer, >180 km, composed largely of fertile peridotites.

Layer A is distinguished by enigmatic CCGE garnets (named after clinopyroxene–chromite–garnet–equilibrium by Kopylova et al., 2000) that are depleted with respect to Ti, Y, Zr and REE but fertile in their major element compositions, as demonstrated by elevated Ca contents. Layer A peridotites have “ultradepleted” arc mantle -type compositions, and have been metasomatised by radiogenic $^{187}\text{Os}/^{188}\text{Os}$, presumably as a result of metasomatism by slab-derived fluids (Peltonen & Brüggmann, 2006). Layer A does not exist at the core of the craton and is thought to represent predominantly Proterozoic arc complex lithosphere. Nevertheless, it has maximum T_{RD} ages

of ~2.6 Ga, suggesting at least some Archean material exists within the layer (Peltonen & Brügmann, 2006). Layer A has a sharp lower contact against Layer B, interpreted as a shear zone along which Layer A was thrust beneath the Karelian Craton margin crust during ~1.9 Ga continental collision (Peltonen & Brügmann, 2006).

Layer B is considered to represent the least modified SCLM in this cross-section, stabilized during the Paleoproterozoic. Xenoliths from this horizon are characterized by unradiogenic Os isotopic compositions. $^{187}\text{Os}/^{188}\text{Os}$ shows a good correlation with indices of partial melting implying an age of ~3.3 Ga for komatiite extraction (Peltonen & Brügmann, 2006), a likely marker of extensive high pressure melting that is postulated to have left these and similar depleted cratonic peridotites worldwide as residues (Boyd, 1989; Griffin et al., 2009). Significantly, this age corresponds closely with the oldest formation ages of the overlying crust (Mutanen & Huhma, 2003). Subcalcic harzburgitic garnets (G10) with depleted contents of Ti, Y, Zr and HREE exist only in Layer B. In contrast, the lherzolitic garnets (G9) are mostly Ti-rich and have REE_N patterns typical of Ca-saturated mantle garnets (e.g., Shimizu, 1975). This was likely caused by a metasomatic re-enrichment event as a result of interaction with silicate melts (cf. Griffin et al., 1989b; Stachel et al., 1998) close in composition to megacryst-forming magmas (Burgess & Harte, 1999, 2004). Abundant Ti-rich megacrystal garnets attest to the presence and polybaric fractionation of such magmas. However, the existence of harzburgitic and some Ti-poor lherzolitic pyropes requires that some portions of this Archean mantle remained unaffected by melt interaction.

Layer C, at depths greater than 180 km, contains no G10 garnets, and is dominated by Ti-rich peridotite pyropes (G11). Significantly however, there exist a few Ti-poor, trace-element depleted G9 pyropes from this layer that possibly represent remnants of Layer B that once extended to greater depths. The osmium isotopic composition of the Layer C xenoliths is more radiogenic compared to Layer B, yielding only Proterozoic T_{RD} ages (Peltonen & Brügmann, 2006), and is

interpreted to represent a melt-metasomatised equivalent to Layer B that was injected by melts c. 2.0 Ga during break-up of the Archean craton.

1.2 The aim of this study

The aim of this study is to extend the dataset on the SCLM transect from the edge of the Karelian Craton at Kaavi–Kuopio to the Central Karelian Craton by studying mantle-derived materials from Kuhmo, Lentiira, southern Kuusamo (S-Kuusamo) and northern Kuusamo (N-Kuusamo) kimberlites. The working methods are similar to those used in the Kaavi–Kuopio study (Lehtonen, 2005b). However, because only a relatively small volume of hard-rock kimberlite is available from the central craton kimberlites, and because mantle xenoliths appear to be less common in some of these rocks, this work must rely solely on mantle xenocryst data. For comparison purposes, Kaavi–Kuopio data are presented throughout the paper side by side with the new datasets. Summary diagrams of some of the Kuhmo data presented in this study have been published elsewhere (O'Brien et al., 2003).

2 Samples

Xenocryst samples from the Lentiira-Kuhmo area are from the Seitaperä (Dyke 16) and Rantala kimberlites and from heavy mineral concentrates taken from till samples. The Kuusamo samples represent xenocrysts from four southern Kuusamo kimberlites, Kalettomannupuro, Kattaisenvaara, Käsmä 45 and Käsmä 47, and one northern Kuusamo kimberlite. Rantala and Lentiira samples were made available to this study by Kopane Diamond Developments Plc, southern Kuusamo samples by Sunrise Diamonds Plc, and northern Kuusamo samples by Mantle Diamonds Ltd. and Kopane Diamond Developments.

In the case of kimberlite rock samples, xenocryst grains (0.25–2.0 mm) were liberated by lightly crushing the kimberlite material, followed by heavy medium separation, and hand picking. The processing of till samples is described in Lehtonen et al. (2005). Hundreds of garnet and clinopyroxene (cpx) xenoc-

Table 1. Representative electron microprobe and LA-ICP-MS analyses of garnet xenocrysts from the Kuhmo-Lentiira kimberlites.

Sample ID	Kimberlite	Garnet type*	EPMA analyses										LA-ICP-MS analyses		
			SiO ₂ wt. %	TiO ₂ wt. %	Al ₂ O ₃ wt. %	Cr ₂ O ₃ wt. %	FeO wt. %	MnO wt. %	MgO wt. %	CaO wt. %	Na ₂ O wt. %	Total wt. %	Sc ppm	Ni ppm	Ga ppm
7901.02	SEITAPERÄ	G1	41.52	0.89	22.17	0.27	9.48	0.24	20.38	4.36	0.07	99.44	60	66	16.2
7896.02	SEITAPERÄ	G9A	40.88	0.03	19.83	5.34	6.91	0.36	20.21	5.61	0.00	99.17	116	31	2.1
7921.02	SEITAPERÄ	G9A	41.19	0.17	20.60	3.64	7.39	0.36	20.61	4.92	0.03	98.94	98	38	6.1
10715.02	SEITAPERÄ	G9A	41.45	0.22	20.58	3.81	7.18	0.36	21.03	4.85	0.06	99.64	99	41	7.5
9220.02	SEITAPERÄ	G9A	40.72	0.22	20.79	3.88	7.42	0.35	20.99	4.96	0.03	99.40	87	47	7.8
9240.02	SEITAPERÄ	G9A	40.05	0.20	19.42	5.16	6.62	0.32	20.81	5.35	0.03	98.06	118	77	6.1
9238.02	SEITAPERÄ	G9A	40.77	0.35	19.21	5.21	6.78	0.29	20.98	5.39	0.02	99.08	105	80	5.8
7825.02	SEITAPERÄ	G9A	41.48	0.20	19.88	4.70	5.73	0.27	21.89	4.96	0.00	99.13	110	81	7.3
7816.02	SEITAPERÄ	G9A	41.67	0.24	19.83	4.81	5.79	0.23	21.91	5.01	0.02	99.56	102	81	7.2
9241.02	SEITAPERÄ	G9A	41.05	0.22	19.56	5.16	6.59	0.33	21.05	5.39	0.03	99.45	113	83	6.2
7814.02	SEITAPERÄ	G9A	41.68	0.24	19.87	4.72	5.85	0.24	22.06	5.00	0.03	99.74	105	86	7.1
7827.02	SEITAPERÄ	G9A	41.28	0.11	18.23	6.47	5.94	0.23	20.92	5.66	0.01	98.89	119	91	5.8
7828.02	SEITAPERÄ	G9A	42.00	0.19	19.95	4.71	5.73	0.26	22.07	5.02	0.02	99.99	104	94	6.7
7930.02	SEITAPERÄ	G9A	41.17	0.20	18.53	5.83	5.39	0.23	21.37	5.59	0.02	98.38	109	99	5.1
9221.02	SEITAPERÄ	G9B	41.10	0.21	20.38	4.39	7.38	0.42	20.81	4.95	0.07	99.79	89	21	4.1
10717.02	SEITAPERÄ	G9B	41.25	0.17	20.34	4.00	7.30	0.37	21.03	5.02	0.00	99.59	99	40	7.7
7889.02	SEITAPERÄ	G10A	41.93	0.01	21.15	3.57	6.42	0.31	22.75	3.21	0.00	99.39	80	35	2.2
7893.02	SEITAPERÄ	G10A	42.13	0.00	21.18	3.61	6.40	0.29	22.95	3.20	0.00	99.77	78	39	2.2
9227.02	SEITAPERÄ	G10A	41.12	0.00	18.60	7.12	6.50	0.39	23.38	1.98	0.00	99.11	114	43	1.0
10700.02	SEITAPERÄ	G10A	41.24	0.10	17.26	8.20	4.85	0.22	22.66	4.72	0.00	99.34	125	96	2.8
10694.02	SEITAPERÄ	G11	41.79	0.33	20.57	3.41	5.92	0.28	22.21	4.69	0.04	99.28	95	33	2.9
7892.02	SEITAPERÄ	G11	42.02	0.44	20.82	3.40	6.07	0.25	22.05	4.69	0.06	99.84	89	83	7.1
9237.02	SEITAPERÄ	G11	41.65	0.37	20.63	3.41	6.08	0.24	22.17	4.80	0.04	99.43	78	86	7.5
9247.02	SEITAPERÄ	G11	41.38	0.36	20.53	3.34	5.98	0.24	22.16	4.86	0.03	98.95	90	89	8.5
7922.02	SEITAPERÄ	G11	41.63	0.40	21.11	2.52	6.18	0.25	22.09	4.80	0.03	99.05	85	98	6.3
7824.02	SEITAPERÄ	G11	41.78	0.37	19.07	5.12	5.98	0.25	21.54	5.19	0.04	99.36	114	111	7.4
10771.02	RANTALA	G1	40.51	1.34	21.02	0.08	11.71	0.35	18.61	5.10	0.08	98.88	103	30	23.1
10728.02	RANTALA	G10A	41.74	0.01	19.65	5.78	5.92	0.35	24.34	1.46	0.02	99.38	271	29	0.6
10746.02	RANTALA	G10A	41.28	0.06	18.24	6.78	5.17	0.28	22.78	4.28	0.00	98.93	133	91	2.2
10725.02	RANTALA	G10A	41.67	0.07	18.39	7.18	5.29	0.25	22.66	4.32	0.00	99.92	140	103	3.4
10732.02	RANTALA	G11	41.38	0.52	18.28	6.50	5.72	0.31	21.54	5.48	0.04	99.88	112	80	6.1
10734.02	RANTALA	G11	41.19	0.50	18.28	6.36	5.75	0.25	21.46	5.43	0.04	99.41	108	84	5.6
10763.02	RANTALA	G11	41.13	0.90	19.39	4.39	5.66	0.25	22.21	4.77	0.09	98.88	99	95	7.3
8778.00	LENTIIRA	G1	41.80	0.67	21.05	2.43	6.93	0.25	21.94	5.08	0.05	100.28	89	130	6.8
7541.00	LENTIIRA	G9A	41.55	0.07	18.96	6.26	6.01	0.30	21.15	5.83	0.00	100.16	116	69	3.5
9915.00	LENTIIRA	G9A	41.69	0.14	20.01	5.15	6.33	0.34	21.54	5.18	0.03	100.45	87	29	5.2
9921.00	LENTIIRA	G9A	41.01	0.23	19.39	5.52	6.43	0.32	21.39	5.32	0.05	99.69	101	50	5.5
7590.00	LENTIIRA	G9B	41.75	0.05	20.99	3.89	7.53	0.46	20.67	5.14	0.00	100.51	123	21	2.3
9898.00	LENTIIRA	G10A	41.65	0.05	17.83	7.56	5.09	0.21	23.24	3.55	0.00	99.25	174	116	1.4
9901.00	LENTIIRA	G10A	40.62	0.22	15.85	10.27	5.74	0.29	22.39	3.77	0.05	99.27	180	85	1.4
7540.00	LENTIIRA	G10B	40.50	0.06	18.26	7.33	6.82	0.38	20.90	4.99	0.00	99.28	123	49	3.5
7603.00	LENTIIRA	G11	41.72	0.49	21.62	2.58	6.80	0.30	21.79	4.32	0.06	99.70	80	49	8.4
9922.00	LENTIIRA	G11	40.28	1.01	17.56	6.50	6.50	0.23	20.55	5.79	0.03	98.47	111	98	6.1

*) Garnet type after Grütter et al., 2004 **) RGP96 = Ryan et al., 1996

n.a. = not analyzed

n.d. = not detected

Y ppm	Zr ppm	La ppm	Ce ppm	Pr ppm	Nd ppm	Sm ppm	Eu ppm	Gd ppm	Dy ppm	Er ppm	Yb ppm	Lu ppm	TNi (C°) RGP96**
17.7	110.3	0.03	0.57	0.20	1.80	1.18	0.50	1.68	2.82	2.04	1.94	0.27	1125
0.3	3.8	0.31	1.79	0.36	1.46	0.04	0.02	0.12	0.20	0.10	0.44	0.13	894
10.9	8.1	0.14	0.31	0.10	0.68	0.41	0.19	0.84	1.36	1.21	1.52	0.26	945
11.4	9.4	0.54	1.16	0.20	1.20	0.56	0.15	0.67	1.96	1.23	1.48	0.26	973
10.5	8.0	0.97	2.31	0.40	1.96	0.73	0.21	0.87	1.66	1.15	1.48	0.20	1008
2.9	39.4	0.17	0.63	0.29	2.44	0.84	0.27	0.65	0.63	0.48	0.30	n.d	1188
3.0	42.3	0.17	0.51	0.77	1.99	0.65	0.21	0.54	0.79	0.39	0.36	n.d	1203
7.0	8.1	0.03	0.93	0.30	2.34	0.59	0.13	0.70	0.94	0.96	0.75	0.17	1206
6.8	9.9	0.14	1.10	0.27	2.13	0.65	0.28	0.44	1.30	1.08	0.43	0.17	1209
3.0	36.4	0.23	0.66	0.28	2.38	0.52	0.28	0.48	0.51	0.45	0.15	0.00	1216
7.4	7.3	0.20	1.17	0.17	1.79	0.30	0.15	0.46	1.02	0.89	0.86	0.16	1230
3.2	8.3	0.65	2.49	0.35	2.66	0.84	0.20	0.35	0.65	0.52	0.56	0.16	1255
8.2	18.6	0.41	n.d	0.42	2.78	1.04	0.25	0.77	1.43	0.94	1.11	0.15	1270
8.9	14.1	0.23	1.14	0.27	2.56	0.85	0.34	1.07	1.62	0.99	1.14	0.29	1296
10.6	14.1	0.02	0.11	0.04	0.63	0.46	0.28	0.84	1.60	1.19	1.50	0.22	805
11.1	7.8	0.18	0.44	0.11	1.16	0.55	0.25	0.46	1.63	1.34	1.30	0.30	964
0.6	1.1	0.55	5.51	1.27	3.63	0.17	0.05	0.01	0.01	0.03	0.30	0.16	926
0.8	1.4	0.73	6.64	1.26	4.30	0.03	0.01	0.07	0.12	0.13	0.58	0.19	955
0.3	3.2	0.18	1.23	0.40	2.30	0.27	0.14	0.13	0.17	0.20	0.19	0.04	981
2.4	2.8	0.09	1.32	0.35	2.13	0.07	0.14	0.45	0.48	0.17	0.51	0.20	1279
1.0	8.9	0.30	3.02	0.54	2.86	0.59	0.15	0.10	0.19	0.16	0.16	0.18	914
14.8	32.5	0.04	0.51	0.14	1.39	0.66	0.36	1.13	2.14	1.57	2.12	0.33	1219
11.9	30.4	0.10	0.60	0.14	1.37	0.45	0.27	1.20	1.87	1.72	1.66	0.18	1230
14.3	35.9	0.16	0.49	0.18	1.60	0.80	0.42	1.37	2.57	2.07	1.86	0.29	1246
15.5	25.3	0.19	0.60	0.21	1.66	0.54	0.40	1.51	2.08	1.45	2.21	0.39	1290
12.5	30.4	0.21	1.65	0.43	3.31	1.21	0.46	1.82	2.21	1.47	1.59	0.29	1350
42.7	370.3	0.05	0.63	0.26	2.61	2.04	1.03	4.19	6.59	5.13	6.06	0.83	888
1.7	45.3	n.d	1.19	0.65	6.90	1.80	0.15	1.02	0.86	0.10	n.d.	0.08	874
2.6	55.6	0.12	1.26	0.41	4.20	3.16	0.44	1.18	0.58	0.17	0.24	0.03	1257
2.9	67.3	0.39	1.13	0.44	5.07	2.39	0.70	1.93	0.63	0.22	0.36	0.06	1314
9.3	67.8	0.05	1.07	0.40	3.18	1.35	0.43	1.62	1.54	0.92	0.98	0.18	1203
9.1	62.6	0.05	1.11	0.30	3.82	1.18	0.47	1.16	1.46	0.94	1.40	0.22	1223
10.4	122.7	0.15	0.71	0.26	2.59	1.48	0.63	2.08	2.66	0.92	0.67	0.13	1274
18.7	72.7	0.14	9.59	0.23	1.69	1.12	0.61	1.75	3.24	2.24	2.28	0.41	1434
0.9	14.7	0.15	1.16	0.41	3.97	0.18	0.05	0.27	0.05	0.15	0.32	0.09	1145
9.8	9.4	n.d	0.07	n.d.	0.7	0.45	0.31	0.71	1.25	1.10	1.25	0.20	873
8.8	53.5	n.d	0.13	0.07	0.8	0.85	0.36	1.20	1.49	1.08	1.24	0.20	1029
3.0	10.7	0.03	0.34	0.02	1.07	0.94	0.39	1.34	1.04	0.36	0.18	n.d.	797
2.3	6.5	0.48	3.70	0.84	3.79	0.14	0.09	0.48	0.38	0.44	0.45	0.12	1373
0.8	73.7	0.07	1.21	0.34	3.0	1.17	0.35	0.96	0.41	0.14	0.21	n.d	1228
2.5	2.6	n.d	1.07	0.31	2.37	0.13	-0.00	0.39	0.32	0.32	0.66	0.06	1025
17.7	44.3	0.06	0.20	n.d.	0.86	0.48	0.43	1.58	2.73	1.96	2.28	0.33	1024
10.3	149.9	0.18	1.24	0.32	2.6	1.88	0.75	2.54	2.64	0.89	0.77	0.15	1292

Table 2. Representative electron microprobe analyses of garnet xenocrysts from the Kuusamo kimberlites.

Sample ID	Kimberlite	Garnet type*	EPMA analyses				
			SiO ₂ wt.%	TiO ₂ wt.%	Al ₂ O ₃ wt.%	Cr ₂ O ₃ wt.%	FeO wt.%
7014.06	Käsmä 45	G11	41.95	1.03	17.90	4.60	7.19
7021.06	Käsmä 45	G11	40.29	1.01	17.70	5.98	7.15
7018.06	Käsmä 45	G9A	41.89	0.09	20.39	3.87	6.35
13914.05	Käsmä 47	G9A	41.40	0.04	15.83	9.70	5.93
13916.05	Käsmä 47	G10A	41.15	0.05	16.09	9.33	5.86
13968.05	Käsmä 47	G11	42.07	0.95	18.79	4.48	6.51
9802.04	Kalettomanpuro	G1	41.95	0.52	22.90	0.50	8.09
9808.04	Kalettomanpuro	G1	41.66	0.47	22.76	0.51	8.52
9785.04	Kalettomanpuro	G9A	41.52	0.15	19.52	5.29	6.79
9786.04	Kalettomanpuro	G9A	41.82	0.19	19.22	5.76	5.87
9798.04	Kalettomanpuro	G9B	42.15	0.02	21.31	3.41	5.90
9765.04	Kalettomanpuro	G9A	41.44	0.19	19.39	5.62	5.85
9769.04	Kalettomanpuro	G9A	41.68	0.29	18.91	6.38	5.96
9758.04	Kalettomanpuro	G9B	41.15	0.08	17.34	8.23	5.75
9771.04	Kalettomanpuro	G10A	42.18	0.01	20.96	3.85	5.49
9759.04	Kalettomanpuro	G10A	42.29	0.02	21.19	3.79	5.53
9755.04	Kalettomanpuro	G10A	42.02	0.15	18.80	6.67	5.84
9766.04	Kalettomanpuro	G10A	42.28	0.02	21.97	2.96	5.42
9761.04	Kalettomanpuro	G11	42.07	0.34	20.15	4.62	6.06
9789.04	Kalettomanpuro	G11	42.01	0.68	20.44	3.31	7.68
12413.04	Kattaisenvaara	G9A	40.91	0.06	19.53	5.33	6.42
12372.04	Kattaisenvaara	G9B	41.19	0.02	19.89	5.08	6.36
12365.04	Kattaisenvaara	G9B	40.72	0.12	17.71	7.69	7.17
12410.04	Kattaisenvaara	G9A	40.51	0.02	19.18	5.48	6.22
12408.04	Kattaisenvaara	G9A	40.99	0.24	19.70	4.63	6.28
12407.04	Kattaisenvaara	G11	40.90	0.67	19.25	4.60	6.35
12309.04	Kattaisenvaara	G11	41.05	0.82	16.83	7.43	6.62
9018-POH-02	N. Kuusamo	G1	41.82	0.56	21.37	2.40	7.19
9001-PIN-02	N. Kuusamo	G1	40.74	0.61	21.74	1.79	7.66
D478	N. Kuusamo	G1	41.58	0.77	22.68	0.28	7.73
9069-POH-02	N. Kuusamo	G10A	40.25	0.00	18.81	6.64	6.23
D478	N. Kuusamo	G10A	40.62	0.09	17.31	8.60	6.10
D478	N. Kuusamo	G10A	40.88	0.12	17.24	8.93	5.94
9196-KES-02	N. Kuusamo	G10A	40.61	0.07	18.06	7.93	5.99
D478	N. Kuusamo	G10A	41.77	0.00	20.16	5.26	5.77
D478	N. Kuusamo	G10A	41.84	0.02	20.22	5.31	6.01
9034-KES-02	N. Kuusamo	G10A	40.76	0.24	17.37	8.30	6.33
9059-POH-02	N. Kuusamo	G11	40.79	0.52	19.29	4.93	6.64
9295-KES-03	N. Kuusamo	G11	41.28	0.64	21.96	2.21	6.97
9083-KES-02	N. Kuusamo	G11	41.36	0.51	19.63	5.06	6.04
9001-PIN-02	N. Kuusamo	G11	41.36	0.44	19.69	4.79	6.69
9043-POH-02	N. Kuusamo	G9A	40.69	0.33	18.89	6.00	6.49
D478	N. Kuusamo	G9A	41.32	0.26	20.55	4.25	6.24
D478	N. Kuusamo	G9A	40.40	0.20	17.55	8.07	6.28
D478	N. Kuusamo	G9A	41.70	0.21	21.09	3.68	5.88
D478	N. Kuusamo	G9A	41.07	0.18	21.11	3.34	6.11
9039-POH-02	N. Kuusamo	G9A	40.87	0.25	20.55	3.65	6.93
035-KUU-00	N. Kuusamo	G9A	41.42	0.19	20.89	3.83	6.05
9066-KES-02	N. Kuusamo	G9B	39.99	0.22	17.62	8.11	6.95

*) Garnet type after Grütter et al., 2004 **) RGP96 = Ryan et al., 1996

					trace data			
MnO wt.%	MgO wt.%	CaO wt.%	Na2O wt.%	Total wt.%	Ti ppm	Mn ppm	Ni ppm	TNi (C ^o) RGP96**
0.32	20.86	5.62	0.10	99.63	5680	2259	86	1230
0.28	20.93	5.77	0.13	99.30	5615	2262	105	1323
0.26	21.98	5.16	0.00	100.05	295	1881	125	1412
0.09	20.94	5.80	0.02	99.80	215	2116	118	1379
0.05	21.09	5.67	0.02	99.35	215	2099	117	1377
0.16	21.75	5.24	0.09	100.17	5536	1893	119	1386
0.27	21.10	4.41	0.04	99.87	4490	2340	51	1037
0.34	21.34	4.15	0.06	99.87	2421	2446	66	1127
0.36	21.16	4.70	0.04	99.61	648	2764	45	998
0.28	21.53	5.24	0.02	99.98	965	2127	60	1092
0.27	21.69	5.18	0.00	100.00	85	1964	64	1115
0.27	21.41	5.26	0.01	99.53	939	2138	69	1143
0.25	20.97	5.81	0.02	100.27	1500	1983	74	1170
0.29	20.00	6.41	0.01	99.34	346	2175	76	1181
0.26	23.01	3.92	0.02	99.75	79	1887	61	1098
0.26	23.20	3.91	0.01	100.23	79	1872	67	1132
0.26	21.79	4.91	0.02	100.50	736	2053	70	1149
0.24	24.04	3.20	0.01	100.18	92	1713	76	1181
0.25	21.56	5.04	0.04	100.22	1794	2009	74	1170
0.26	20.94	5.22	0.03	100.61	3569	1874	105	1323
0.33	21.10	5.46	0.03	99.26	714	3258	12	691
0.34	20.98	5.67	0.01	99.63	72	3105	19	780
0.44	18.55	7.10	0.00	99.59	738	3441	29	872
0.30	21.05	5.13	0.02	97.97	174	2366	41	970
0.27	21.88	5.00	0.03	99.18	1411	2103	84	1222
0.25	21.55	5.20	0.05	98.93	3848	1920	105	1321
0.28	20.51	6.22	0.06	99.93	5420	2136	110	1343
0.24	21.61	4.22	0.07	99.64	722	2489	53	1050
0.26	21.46	4.29	0.07	98.94	1544	2202	75	1176
0.29	22.33	4.15	0.08	100.31	3817	1918	91	1257
0.32	23.17	2.36	0.01	98.25	723	2395	54	1056
0.29	22.18	4.28	0.01	99.83	399	2126	64	1115
0.26	22.31	4.14	0.01	100.16	606	2070	67	1132
0.18	23.31	2.68	0.01	99.22	2361	1964	70	1149
0.23	24.94	1.53	0.02	100.02	104	1877	74	1170
0.28	25.08	1.53	0.01	100.71	101	1876	80	1202
0.35	20.96	4.57	0.03	99.18	2937	2068	95	1276
0.28	21.16	5.16	0.09	99.21	938	3061	26	850
0.29	22.19	4.48	0.09	100.61	449	2783	54	1056
0.29	21.84	4.77	0.05	99.75	3632	2112	73	1165
0.28	20.91	5.26	0.04	99.72	3172	2049	88	1242
0.26	20.88	5.45	0.04	99.35	2324	2584	39	956
0.28	22.09	4.77	0.05	100.17	1362	2030	61	1098
0.34	20.89	5.74	0.03	99.93	1121	2153	69	1143
0.26	22.59	4.55	0.05	100.32	994	1909	71	1154
0.29	22.57	4.51	0.05	99.70	970	1903	73	1165
0.24	21.23	5.08	0.05	99.14	1647	2050	78	1192
0.28	21.79	4.83	0.03	99.37	4427	2153	93	1267
0.40	19.66	5.83	0.03	99.12	11	2311	71	1154

rysts were recovered from the kimberlite and till concentrates and mounted for analysis. In this study only unsorted garnet concentrates were used in order to gain unbiased populations for each of the kimberlites.

3 Analytical techniques

Garnet major element compositions were determined by Cameca SX50 and SX100 electron microprobes (EMP) at the Geological Survey of Finland (GTK), applying an acceleration voltage of 25 kV, probe current of 48 nA and beam diameter of 1 μm . The parameters for clinopyroxene analyses were 15 kV, 30 nA and 5 μm , respectively. Selected garnet xenocrysts from Lentiira–Kuhmo were analyzed for trace elements by LA-ICP-MS at the University of Cape Town, using the methodology and equipment described in Grégoire et al. (2003). Typical theoretical detection limits are in the range of 10–20 ppb for REE, Zr and Y, and 2 ppm for Ti and Ni. Trace Ni data by EMP were obtained on pyrope grains employing 500 nA probe current, 600s counting times on peak plus background positions and were reduced by the CSIRO TRACE program for the SX50 (Robinson & Graham, 1992). Cross-checking of the trace methods shows that Ni analyses in pyrope by EMP can achieve similar precision to those of LA ICP-MS down to a level of ca. 10 ppm.

The garnets were classified based on their major element composition into harzburgitic (G10), lherzolitic (G9), wehrlitic (G12), high-Ti peridotitic (G11), low-Cr megacrystal (G1), eclogitic (G3, G4), pyroxenitic (G4, G5), and crustal varieties according to Grütter et al. (2004). For equilibration temperatures the Ni thermometer (Griffin et al., 1989a) was applied using the calibration of Ryan et al. (1996). Equilibration pressures and temperatures of peridotitic clinopyroxene were calculated using the single-grain cpx thermobarometer of Nimis & Taylor (2000).

4 Results

Major and trace element analyses for a large portion of the Kuhmo, Lentiira and Kuusamo samples

used in this study are available as an Open File Report (Lehtonen et al., 2008) at the Geological Survey of Finland website (www.gtk.fi). Selected analyses are presented in Tables 1 and 2. Trace element data obtained by LA-ICP-MS is available only from Kuhmo and Lentiira samples.

4.1 Garnet major element geochemistry

CaO and Cr_2O_3 contents of mantle garnets from Kuopio, Kaavi, Kuhmo, Lentiira, S-Kuusamo and N-Kuusamo kimberlites are shown in Figure 2. It is evident that lherzolitic garnets (G9) predominate over other garnet varieties in samples from all kimberlite fields and that only a minority classifies as subcalcic harzburgitic grains (G10). Relatively Cr-poor, but Ti-rich, pyropes of megacryst composition (G1) are also common at all localities. Pyrope populations from the edge of the craton (Kaavi–Kuopio) are considerably poorer in harzburgitic pyropes compared to the craton core samples represented by Lentiira–Kuhmo. However, there is a prominent eclogitic component (G3/G4) at Kaavi–Kuopio (grains in the lower right of Fig. 2) that is nearly absent from the Lentiira–Kuhmo mantle and in terms of diamond potential, may compensate for the paucity of G10 source rocks. Fewer grains are available from the N- and S-Kuusamo locations, but both areas show a relatively high proportion of the harzburgitic to lherzolitic pyropes, suggesting a depleted mantle more similar to that at Lentiira–Kuhmo.

4.2 Clinopyroxene thermobarometry

Single-grain thermobarometry from clinopyroxene xenocrysts and clinopyroxene–garnet aggregates from Kaavi–Kuopio kimberlites fits reasonably well with a regional geotherm of 36 mW/m² calculated using heat flow constraints, xenolith modes and geophysical properties (Kukkonen & Peltonen, 1999; Kukkonen et al., 2003). Figure 3 shows the correspondence between the xenolith and xenocryst data and includes reference fields for the Slave Craton from Grütter (2009).

Figure 2. Cr_2O_3 vs. CaO of Kuopio, Kaavi, Kuhmo, Lentiira, and southern and northern Kuusamo kimberlite-derived garnet xenocrysts. Crustal garnets are excluded from this diagram using the classification of Schulze (2003). The harzburgite, lherzolite and non-peridotite fields are redrawn after Gurney (1984) and the wehrlite field is separated according to Sobolev et al. (1973). Arrow pointing NE in the Kaavi-Kuopio diagram marks the “CCGE” garnet trend, i.e. chromite-clinopyroxene-garnet equilibrium, recognized in spinel-garnet peridotites from the Jericho kimberlite by Kopylova et al. (2000). Layer A explained in text, section 1.1.

Constructing single-clinopyroxene geotherms (after Nimis and Taylor, 2000) for Kuhmo, Lentiira and S-Kuusamo areas has proven much more difficult than for Kaavi–Kuopio because of the rarity of high temperature (>1100 °C) clinopyroxene grains. Their virtual absence is somewhat enigmatic since high temperature lherzolitic garnets are well represented at both localities, suggesting clinopyroxene-bearing source rocks in the deeper parts of the SCLM. However, PT data on clinopyroxenes from xenoliths extracted from ca. 1.2 Ga Group II-olivine lamproite dike rocks from Kostomuksha that are mineralogical and compositional analogues and time correlative to the Lentiira–Kuhmo rocks, provide evidence that the lithospheric mantle at this time, in this part of the world had a cold geotherm similar to that at Kaavi–Kuopio (Antonov & Ulianov, 2008).

Although not abundant, our N-Kuusamo samples did contain sufficient mantle-derived clinopyroxene grains from which the local geotherm could be defined. In Figure 3 the data are plotted against the Kaavi–Kuopio data, and it can be seen that there is only a minimal difference between the two single-cpx-based geotherms and that they deviate approximately the same amount from the more accurately defined Kaavi–Kuopio xenolith geotherm of Kukkonen & Peltonen (1999). Consequently, we have used the xenolith-based geotherm for all Central Karelian data, including Kuhmo, Lentiira and S-Kuusamo.

4.3 Garnet thermometry and trace element geochemistry

4.3.1 T_{Ni} histograms

T_{Ni} histograms for Kaavi, Kuopio, Kuhmo, Lentiira, S- and N-Kuusamo garnet xenocrysts are presented in Figure 4a. Megacryst garnets (G1) with $\text{Cr}_2\text{O}_3 > 1.5$ wt.% and $\text{MgO} > 18$ wt.% are included in the diagrams for illustrative purposes, based on the reasoning that they reach the same contents of Cr_2O_3 and MgO as do the G9 pyropes, implying equilibration with olivine similar in composition to that in the lherzolites. The megacryst temperatures give rea-

sonable results, however the coexisting olivine may have lower Ni than normal peridotite and in that case the megacryst temperatures would be underestimated.

The Kaavi–Kuopio data show a bimodal distribution including a strong low temperature peak at 700 to 850°C (Layer A) consisting dominantly of CCGE pyropes, that are classified either as wehrlitic (G12) or lherzolitic (G9) varieties according to Grütter et al. (2004), and a stronger abundance peak at 1000 to 1150°C (Layer B) consisting of lherzolitic (G9), harzburgitic (G10), high-Ti peridotitic (G11), wehrlitic (G12) and megacryst (G1) varieties. 1150 °C separates mantle containing strongly depleted pyropes (G9 and G10) based on their Zr contents from more enriched mantle, as seen in Figure 5

In the Kuhmo and Lentiira terrains the sampling of garnet-bearing mantle was between 800–1400 and 700–1450 °C, respectively (Fig. 4a). The harzburgitic component (G10) is distributed almost throughout both mantle sections, although in Lentiira samples there is an obvious concentration near the base of the unmodified lithosphere from 1250 to 1400°C and an absence of G10s at temperatures <850 °C. Kaavi–Kuopio and Lentiira–Kuhmo share the common feature that the highest temperature at which G10 garnets exist also represents the highest temperature at which any low Zr pyropes (G9 and G10) occur (Fig. 5).

The T_{Ni} histogram of the S-Kuusamo garnet xenocrysts is dominated by lherzolitic grains at temperatures below 1150 °C and by high-Ti peridotitic grains at higher temperatures. The range of sampling is roughly the same as at Lentiira, with some very high temperature grains (>1400 °C). G10 garnets are clearly concentrated around 1100 °C. High-Cr megacrystal grains (G1) are very rare compared to Lentiira–Kuhmo and Kaavi–Kuopio. The N-Kuusamo histogram is narrower than in S-Kuusamo and the major horizon of sampling lies between 1050–1250 °C. This is also the horizon that contains abundant G10 and high-Cr G1 garnets, rare G9 varieties also exist at higher and lower temperatures. There are no grains with temperatures exceeding 1350 °C in N-Kuusamo whereas in S-Kuusamo they are relatively

Figure 3. P-T calculated for cpx-garnet pairs from Kaavi-Kuopio and cpx xenocrysts from Kaavi-Kuopio and northern Kuusamo kimberlites using the thermobarometer by Nimis & Taylor (2000). 36 mW/m² geotherm from Kukkonen & Peltonen (1999) calculated for the Karelian craton at 600 Ma and reference fields for the Slave Craton from Grütter (2009) are also marked. Depth in km is converted from pressure according to Kukkonen et al. (2003). The diamond-graphite transition is redrawn after Kennedy & Kennedy (1976).

common, indicating that the lithosphere is thinning towards the North.

Whether T_{Ni} above 1400 °C represent pyrope grains in lithospheric mantle equilibrated on a geotherm can be called into question. In the overall dataset, the bulk of the grains giving high temperatures are G11, high Ti pyropes, representing fertile or refertilized mantle, consistent with heating by melt percolation and implying temperatures above the conductive geotherm. This could be one explanation for the highest temperatures calculated. However, at Kuhmo, Lentiira and South Kuusamo these same high temperatures are recorded from G10 pyropes that show no evidence of melt metasomatism (see Section

4.3.3) indicating that such high temperatures can indeed exist in the lowermost sections of the non-conducting lithospheric mantle along a cool geotherm.

4.3.2 Ti and Mg# vs. temperature and depth

Variations in pyrope TiO₂ contents provide a useful means to map the relative fertility of the mantle sample (e.g. Griffin & Ryan, 1995). Mg# of garnet also reflects mantle bulk composition but is not as straightforward to interpret because it is also strongly affected by temperature dependent Mg-Fe exchange with olivine, orthopyroxene, clinopyroxene and Mg-chromite (Stachel et al., 2003). In addition, Ca has

Figure 4. Distribution of T_{Ni} (A), TiO_2 vs. T_{Ni} (B) and $Mg\#$ vs. T_{Ni} (C) for Kuopio, Kaavi, Kuhmo, Lentiira, and southern and northern Kuusamo pyrope xenocrysts. Garnet classification according to Grütter et al. (2004), T_{Ni} calculated using the calibration of Ryan et al. (1996) and depths by extrapolating the T_{Ni} temperatures of pyropes to the Kaavi-Kuopio geotherm determined by Kukkonen et al. (2003). Megacryst garnets (G1) with $Cr_2O_3 > 1.5$ wt.% and $MgO > 18$ wt.% are shown for illustrative purposes, see section 4.3.1 for discussion. Layers A, B and C also explained in text.

a significant crystal chemical effect on the Mg-Fe partitioning between garnet and olivine (O'Neill & Wood, 1979; Stachel et al., 2003).

The two breaks, at 850 and 1150 °C, described previously in the Kaavi-Kuopio mantle stratigraphy, are also apparent in the TiO_2 contents and $Mg\#$ of garnets (Figs. 4b and c). The lower temperature boundary at c. 850 °C or c. 110 km is marked by a break from low

Ti CCGE pyropes (G12, G9) with $Mg\#$ around 0.80 in Layer A to pyropes that exhibit a wide range in Ti and $Mg\#$ in Layer B. The boundary at 1150 °C or c. 175 km is the limit below which only a very few TiO_2 -depleted pyropes occur, G10s are absent and pyrope $Mg\#$ values are less variable (Layer C).

In the upper parts of these mantle sections at Lentiira and Kuhmo, TiO_2 contents of pyropes vary con-

Figure 5. Zr contents of pyropes vs. T_{Ni} (Ryan et al., 1996) from Kaavi-Kuopio and Lentiira-Kuhmo. Garnet classification after Grütter et al. (2004).

siderably less than those at Kaavi–Kuopio. However, there is a general increase in pyrope TiO_2 with depth, especially at Lentiira (Fig. 4b). The large population of G11 high-Ti pyrope grains (TiO_2 0.5–1.0 wt.%), starting at 1150 °C or c. 175 km depth, represents a break in the lithosphere below which the mantle has been more affected by melt-metasomatism. Also evident in the Lentiira and Kuhmo pyrope data is a

strong Mg# versus depth correlation (Fig. 4c) that in both cases fit along well-defined linear trends evident in all of the pyrope data from these areas, irrespective of pyrope chemistry. This strong linearity suggests these portions of mantle have been have not seen any thermal perturbations close in time to kimberlite/lamproite magmatism and likely have been in equilibrium for long periods of time.

The Kuusamo garnet data resemble those from Lentiira–Kuhmo in regard to Ti contents and Mg#. As at Lentiira–Kuhmo, strong compositional variations in pyropes are seen only at temperatures higher than 1000 °C, corresponding to 140 km in depth, in both the S- and N-Kuusamo datasets. The TiO₂ contents of garnets are relatively constant (<0.15 wt.%) from 700 until 1000 °C, at higher temperatures Ti-enriched garnets appear. The most obvious difference compared to Lentiira–Kuhmo is the relatively narrow temperature range over which harzburgitic grains have been sampled, from roughly 1100–1300 °C. The thickness of the lithosphere decreases from over 250 km in S-Kuusamo to 220 km in N-Kuusamo.

4.3.3 Zr and Y contents

Systematics of the Zr–Y concentrations in Lentiira–Kuhmo pyropes are compared to the field for similar data from Kaavi–Kuopio in Figure 6. The Kaavi–Kuopio compositional field shows a roughly linear increase of Zr and Y from depleted mantle compositions (Griffin & Ryan, 1995) of harzburgites (Zr < 25 ppm, Y < 10 ppm), through lherzolites and finally to enriched megacryst and G11 pyrope compositions (up to 60 ppm Zr and 30 ppm Y). The increase

in Zr and Y is coupled to increasing Ti, and this overall trend to higher Ti, Zr and Y has been interpreted to indicate an increasing effect by melt metasomatism in the Kaavi–Kuopio mantle (Lehtonen, 2005a).

Two obvious differences from the case described above are shown by the Lentiira–Kuhmo data. First, the overall level of melt metasomatism is far weaker at Lentiira–Kuhmo in terms of extreme enrichments in Zr and Y. Second, a completely different style of metasomatism, one of Zr enrichment without concomitant Y (Fig. 6) or Ti (not shown) enrichment is exhibited by a range of pyrope types from G10 to G9 to G11. This latter style of metasomatism is only poorly developed in the mantle section at <1150 °C, but is clear in many grains at higher temperatures. The fact that it affected such a range of pyrope types without converting them all to G11 is a clear distinction from the melt metasomatism trend, and instead points to some other metasomatic agent.

4.4 Rare-earth element contents of peridotitic garnet xenocrysts

4.4.1 Kaavi–Kuopio

C1-chondrite normalized REE_N profiles of all garnet xenocrysts classified as megacrysts and Ti-rich peridotitic pyropes (G11), along with the majority of the

Figure 6. Y vs. Zr for the Lentiira-Kuhmo pyrope xenocrysts. The two temperature horizons (above and below 1150 °C) are discussed in text. Kaavi-Kuopio data from Lehtonen (2005A) are shown for comparison as a shaded field. Dashed line defines depleted compositions (from Griffin and Ryan, 1995).

lherzolitic pyropes (Layers B and C), have so called N-type ("normal") REE profiles showing strong LREE depletion relative to MREE and HREE with steady enrichment from Sm_N to Yb_N (Fig. 7). This type of profile is typical of Ca-saturated mantle garnets (e.g. Shimizu, 1975). A subset of the lherzolitic and nearly all of the Layer B harzburgitic grains have sinusoidal (S-type) REE_N patterns, characteristic of subcalcic harzburgitic garnets (e.g. Shimizu, 1975). The S-type REE_N profiles show a weaker LREE depletion, a maximum at Pr or Nd and a trough in the MREE bottoming at Dy, Ho or Er.

Two ratios, $(\text{Nd/Ce})_N$ and $(\text{Nd/Dy})_N$ allow evaluation of the steepness and direction of the LREE pattern slope and the degree of sinuosity of the overall profile in garnet, respectively, and these are plotted against each other in Figure 8. A $(\text{Nd/Dy})_N < 1$ refers to N-type REE patterns, and with increasing values of $(\text{Nd/Dy})_N$ above 1, the level of sinuosity in the REE profile increases. For Kaavi–Kuopio, nearly all of the G10 grains have $(\text{Nd/Dy})_N > 1$ and, importantly, all of the Ti, Y and Zr depleted G9 pyropes (Fig. 7) have $(\text{Nd/Dy})_N > 10$, showing even greater profile sinuosity than the G10 grains. Nevertheless, the bulk of the G9 grains and essentially all of the G11 and megacryst composition grains have $(\text{Nd/Dy})_N < 1$. The vast majority of the Kaavi–Kuopio pyropes have $(\text{Nd/Ce})_N$ ranging from 2–9 (centered around the value of 5, Fig. 8), implying a range from nearly flat to increasingly negative LREE slopes.

4.4.2 Lentiira–Kuhmo

As in Kaavi–Kuopio, G11 and megacrystic garnet xenocrysts are overwhelmingly N-type based on their REE_N signatures (Fig. 7). However, the G9 lherzolitic pyrope REE_N profiles from Lentiira–Kuhmo contrast markedly from those at Kaavi–Kuopio. First, less than half of G9 pyropes from Lentiira–Kuhmo have $(\text{Nd/Dy})_N < 1$, i.e. N-type REE_N patterns. Second, the remaining G9 REE_N patterns are moderately to strongly sinusoidal, similar to those in lherzolitic inclusions in diamonds reported by Stachel et al. (2004), and overlapping completely with the Len-

tiira–Kuhmo G10 grains in $(\text{Nd/Dy})_N$ (Fig. 8). Third, although the data are relatively few, it appears that the Lentiira–Kuhmo G10 grain REE profiles show a systematic flattening of LREE slope with increasing sinuosity. Fourth, for the remaining grains at Lentiira–Kuhmo, the LREE slopes are uniformly flatter relative to Kaavi–Kuopio pyropes, with $(\text{Nd/Ce})_N$ ranging from 1 to 5 for 90 % of the grains. Importantly this is true even for the G11 pyrope populations, the most melt-metasomatised pyrope variety (Fig. 8).

5 Discussion

5.1 Thickness of lithospheric mantle

Based on the data presented here, the lithospheric mantle thickens from about 220 km at Kuopio to nearly 250 km in the Lentiira–Kuhmo–S. Kuusamo section and then thins again to about 220 km at the N. Kuusamo locality. The latter is in the direction of the Kandalaksha graben where the mantle lithosphere may have already been considerably thinned during the Middle Riphean (c. 1200 Ma) (Baluev et al., 2000). Of course, the source of error involved in these depth estimations are significant and include: (a) Ni measurements in pyrope at a level of only ppm, (b) uncertainties in the pyrope Ni thermometer calibration, (c) extrapolation to a single geotherm for all the kimberlite regions for magmatism spanning a range in age of 600 Ma, (d) deepest (highest T) pyropes not representing the true asthenosphere–lithosphere boundary due to variable sampling by the kimberlites. Despite these simplifications and difficulties, geophysical evidence such as teleseismically derived tomographic cross sections for the central Fennoscandian shield published by Sandoval et al. (2004) are surprisingly consistent with our interpretation above. This gives us confidence that the overall view of lithospheric mantle thicknesses presented in this paper is realistic and relatively accurate.

5.2 Pyrope composition and distribution

Not only is the lithospheric mantle thinner at the craton edges (Kuopio–Kaavi and N. Kuusamo), rel-

Kaavi-Kuopio **Lentiira-Kuhmo**

Figure 7. C1-Chondrite normalized (McDonough & Sun, 1995) REE profiles of mantle-derived garnet xenocrysts from Kaavi-Kuopio and Lentiira-Kuhmo kimberlites. The samples are subdivided according to rock type (Grütter et al., 2004) and temperature horizon.

Figure 8. Nd/Ce (N) vs. Nd/Dy (N) for garnet xenocrysts from Kaavi-Kuopio and Lentiira-Kuhmo. (N = normalized to C1-chondrite composition after McDonough & Sun, 1995). The diagram distinguishes between S-type (sinuoidal) and N-type garnet REE_N patterns. The reference line at $(\text{Nd/Ce})_N = 5$ helps highlight the distinction in LREE slope in all garnet types from Kaavi-Kuopio compared to Lentiira-Kuhmo.

Figure 9. Distribution of Mg# and Cr# vs. depth for Lentiira pyrope xenocrysts. The strong linearity in the Mg# plot is not visible in the Cr# plot, implying that the Mg# correlation is a function of temperature, not bulk rock composition.

ative to the core (Lentiira–Kuhmo–S. Kuusamo) but also the distribution of some of the mantle rock-types shows similarities at the craton edges. For example G10 pyropes are limited to a 40–50 km window at the craton edges (Fig. 4), but occur throughout a thick section of up to 100 km, in the craton core. Based on this thick section of mantle containing harzburgite, it might be expected that craton core samples would also contain a higher ratio of G10 to G9 pyropes but this is not the case for N. Kuusamo which is particularly G10-rich (Fig. 2).

Another similarity between the north and south craton edge sections is the lack of any correlation between Mg# and depth, which, in contrast, is displayed nicely in the central craton sections (Fig. 4c). The overall magnesium number (Mg#) of the central craton mantle lithosphere is high relative to the mantle at Kaavi-Kuopio, and seems to increase uniformly

with depth. Two possibilities exist for this correlation; either it is strictly due to temperature effects on the gt/ol Fe/Mg K_D (O'Neill & Wood, 1979; O'Neill, 1980) or an effect of an increase in the degree of partial melting with increasing depth. To test these two alternatives, Mg# and Cr# versus depth for Lentiira are plotted in Figure 9. Although Mg# increases uniformly with temperature, Cr# does not, corroborating a temperature control on pyrope Mg#. Since the Ni thermometry methodology of Ryan et al. (1996) employed here is calibrated against the gt/ol Fe/Mg K_D mentioned above, the strong correlation of Mg# with T_{Ni} is expected in a well-behaved, closed system. The strong correlation in the craton core data contrast with those at the craton mantle edges where there have been considerably more disturbances due to open system behavior and an influx of higher volumes of metasomatic components.

5.3 Metasomatic effects – Melt infiltration

Several lines of evidence suggest that the lithospheric mantle at the craton edge has undergone considerably more melt metasomatism than the craton core mantle.

5.3.1 Contrasting Mg# & Ti profiles

As described above, a range in Mg# of pyropes taken from a mantle section is expected, but a wide range in Mg# at a specific depth requires that the pyropes sampled from that depth are derived from peridotites with a range in fertility, i.e. a range in the amount of basalt component that remains to be extracted or that has been added (e.g. Griffin et al., 2009). Returning to Figure 4c, the garnets from Kuhmo and Lentiira show significantly less variation in Mg# than those from the craton edges at any given depth. This holds true even for the megacryst suite (G1) pyropes. Although enigmatic, the origin of the low Cr megacryst suite is normally considered as one of crystallization from asthenosphere derived melts that have undergone significant interaction with the overlying lithospheric mantle (e.g. Hops et al., 1992). We suggest that the small spread in craton-core megacryst Mg# reflects a system buffered by a dominant peridotite endmember. In contrast, in the craton edge systems, melt was locally the dominant component, and produced a population of megacrysts with a broad range of Mg#, from 85 (peridotite dominated) to 78 (melt dominated).

The Ti contents of pyropes at Kuhmo and Lentiira show a different relationship compared with that of Mg# (Fig. 4b and 4c). This is probably because of the dominance of the melt component in a melt + peridotite mixture in terms of Ti mass-balance [hundreds of ppm Ti in Karelian craton peridotites (Peltonen et al., 1999) versus 1–3 wt.% TiO₂ in Finnish kimberlites (O'Brien & Tyni, 1999)]. Consequently even very small amounts of melt infiltration will have a large effect on pyrope Ti contents (see Burgess & Harte, 1999). In Figure 4b even the craton core pyropes show a range of TiO₂ at any given depth, al-

though the effect is mostly limited to the base of the SCLM where lherzolitic pyropes (G9) has been converted to high Ti peridotitic G11 pyropes. In fact the G11 pyropes from all of the mantle sections closely resemble megacryst pyropes in terms of Ti concentration (Fig. 4b), strongly suggesting that the G11 grains all represent peridotitic pyropes that recrystallized due to interaction with infiltrating megacryst-producing magmas. This is further corroborated by the fact that the G11 pyropes uniformly have N-type REE profiles indistinguishable from those of the megacryst pyropes (Fig. 7) and are uniformly enriched in Y and Zr relative to the G9 and G10 pyropes (Figs. 5 and 6).

Although we do not have trace element data for the Kuusamo pyropes, Ti and Mg# strongly suggest that the mantle at the Kuusamo localities is quite similar to that at Lentiira-Kuhmo, albeit with a more disrupted Mg# profile, suggesting a somewhat larger component of melt metasomatism. Although the N-Kuusamo lithospheric mantle is shallower than at the rest of the craton core sections, it nevertheless contains the same population of high Mg# and low Ti G10 pyropes that we infer to be Archean in age.

5.3.2 Evidence for Remnants of Layer B in Layer C

The base of the lithosphere at Kaavi-Kuopio is sufficiently distinct from that at the core of the craton that the entire lower lithospheric mantle (Layer C) at Kaavi-Kuopio might be envisioned to have an entirely separate and unique origin. For example, given its location, it could be interpreted to represent Proterozoic mantle that was emplaced tectonically during collision of the primitive arc at 1.9 Ga (Peltonen et al., 1999). However, there are a few lherzolitic grains from layer C, both in the Kuopio and Kaavi mantle cross-sections, which argue instead for a different paragenesis of layer C. These rare but important grains have coupled low Ti, Y and Zr along with REE profiles that match very closely those of the harzburgitic grains. Moreover, these few grains mimic very closely in terms of Ti, Y, Zr and REE profiles the G9 grains that occur in the deeper levels of the craton core section at Lentiira-Kuhmo (Fig. 7). We suggest

that these few deep G9 grains from Layer C are in fact remnants of the depleted peridotites that once existed at that depth at Kaavi–Kuopio, analogous to those from the same depth at Lentiira–Kuhmo. Melt metasomatism, as described earlier, has converted nearly all of the Layer C rocks to relatively fertile, but primitive mantle rock types. During this metasomatic process, all G10 and most G9 grains in layer C were converted to G11. It is also likely that the bulk of the peridotitic diamonds that may have existed in this lower layer were destroyed due to the oxidizing effects of the addition of melt (cf. Luth et al., 1990).

5.4 Metasomatic effects – Fluid infiltration

Fluid metasomatism has been invoked to explain the sinusoidal REE_N patterns of harzburgitic (G10) pyropes, an observation that appears to be a worldwide phenomenon (e.g., Stachel et al., 2004), and the G10s from the Kaavi–Kuopio and Lentiira–Kuhmo mantle sections are no exceptions (Figs. 7 and 8). The worldwide link between G10 garnets and diamonds suggests that the formation of these minerals in mantle peridotite requires infiltration of CHO-rich fluids, enriched also in incompatible trace elements (Stachel et al., 2004). Malkovets et al. (2007) has further refined this model wherein diamond is precipitated by the oxidation of asthenospheric melt-derived, methane-rich fluids that, at the same time, produce subcalcic high-Cr, S-type REE_N garnet (G10) as they penetrate along fractures into pristine, highly depleted harzburgitic lithosphere. As a continuation of this process, the melts themselves invade the lithosphere, causing melt metasomatism that refertilizes harzburgite to lherzolite by introduction of Ca, Al and Fe, and progressively altering the REE_N signatures of pyrope to N-type.

There are several observations that seem to support this model. Trace element analyses of diamonds have indeed revealed the presence of strongly light REE-enriched fluids (Rege et al., 2006) that could have caused the S-type REE_N signatures diagnostic of G10's. Refertilization of harzburgites to lherzolites has also been evidenced by compositional zonation

of garnets within mantle xenoliths (e.g. Griffin et al., 1989b, 1999; Burgess & Harte, 1999).

The following differences in the characteristics of metasomatism seen at Lentiira–Kuhmo compared to Kuopio–Kaavi require a larger role for fluid metasomatism in the former: (a) elevated Zr in some grains without concomitant increases in Ti and Y or changes in major element composition (Griffin & Ryan, 1995), (b) a larger percentage of sinusoidal REE_N profile G9 and G10 grains, all of which are Ti, Y and Zr depleted, (c) relatively minor disturbances to major element compositions (e.g., Mg#), and (d) a broader distribution of harzburgite throughout the mantle section. Certainly the Kaavi–Kuopio and Lentiira–Kuhmo lithospheric mantle peridotites have undergone both fluid and melt metasomatism, as evidenced by the occurrence of sinusoidal G10 and high-Ti peridotite G11 grains at both localities, but the relative amount of fluid metasomatism is considerably less in the Kaavi–Kuopio pyrope signature. However, following the model of Malkovets et al. (2007), it is possible that this same level of fluid metasomatism previously existed in the Kaavi–Kuopio mantle section. If the fluid metasomatism-produced harzburgites provided channelways for the ingress of melts further into the lithospheric mantle, then the melts may have preferentially destroyed these harzburgite zones, erasing the record of the older fluid metasomatism event that remains prevalent in the Lentiira–Kuhmo mantle section.

A corollary of this model is that the distribution of diamonds and garnets in the lithosphere may be laterally heterogeneous on relatively small scales. In regard to this study, it could mean that similar compositional layers of SCLM recognized in various areas are not necessarily continuous but rather local features produced by metasomatism. However, teleseismic data, discussed briefly earlier (Sandoval et al., 2004), imply the existence of a continuous depleted thick mantle root throughout the area through which our cross section from Kuopio to N. Kuusamo is drawn. Taking all of these factors into account, Figure 10 summarizes our present understanding of the SCLM in this region. The diagram includes the

Figure 10. Simplified geological cross-section of the Archean lithosphere across the 600-km-long transect (see Fig. 1 for locations). The Kaavi-Kuopio SCLM stratigraphy implies a peridotitic diamond window between 140 km and 180 km, the lower temperature stability limit of diamond (Kennedy & Kennedy, 1976) and the deepest level where G10 pyropes are found. A well-sampled component of highly diamondiferous eclogite xenoliths (Peltonen et al., 2002) in addition to a high number of diamond-indicative Group I eclogitic garnet xenocrysts (Lehtonen, 2005B) add significantly to the diamond potential of this mantle section in contrast to the eclogite-poor central Craton. Nevertheless, in Lentiira-Kuhmo the peridotitic diamond window determined by the diamond-graphite stability curve and the presence of G10 garnets is extremely thick, exceeding 100 km, starting around 140 km and ending near 250 km. In Kuusamo the main diamond-bearing peridotitic horizon is between 140 and 200 km, with possibly some diamondiferous material even at depths comparable to Lentiira-Kuhmo. Modified after Peltonen et al. (2008).

main mantle layers and connects the Archean layer B under Kaavi–Kuopio with the Archean section under Lentiira–Kuhmo–Kuusamo. In a broad sense this is correct, even though, as discussed above, in detail even the G10-bearing portions of the Archean craton mantle lithosphere have been significantly more melt-modified at the craton edges.

6 Conclusions

The mantle-derived xenolith and xenocryst records from Kaavi, Kuopio, Kuhmo, Lentiira and S- and N-Kuusamo kimberlites indicate that the subcontinental lithospheric mantle of the Karelian craton shows considerable variation from margin to core. The Karelian Craton reaches its greatest thickness in the

central portion, around the Lentiira–Kuhmo–S. Kuusamo terrain (>250 km), and thins towards the SW and NE edges, as indicated by mantle cross-sections at Kaavi–Kuopio (230 km) and northern Kuusamo (220 km). The shallow and deep layers of the 3-layer SCLM underlying Kaavi–Kuopio have been strongly affected by the Paleoproterozoic Svecofennian orogeny as demonstrated by isotopic evidence from mantle xenoliths (Peltonen & Brüggmann, 2006), the scatter of pyrope Mg# within depth profiles, the complete conversion of harzburgite-bearing mantle to lherzolite in the deepest layer (Layer C) and the overall fertility of the Archean mantle section (Layer B) that nonetheless contains remnants of extremely depleted zones. The lithospheric mantle in the Lentiira–Kuhmo area, in contrast, represents the most pristine,

highly depleted, and harzburgite-rich portion of the mantle section with lherzolite and harzburgite relatively uniformly distributed throughout. The Kuusamo region is located relatively close (300 km) to the Belomorian graben area, which separates the Karelian and Kola cratons. Considerable melt metasomatic transformation and concomitant thermal erosion of the SCLM has taken place in the graben area, resulting in thinning of the lithosphere, an effect that can be seen in the deepest mantle layer at N.-Kuusamo. Nevertheless the Kuusamo mantle is more similar and only slightly more disrupted than the mantle at Lentiira–Kuhmo, and is significantly more pristine than the mantle at than the Kaavi–Kuopio craton edge mantle.

Pyrope trace element data consistently show stronger evidence of fluid metasomatism in the peridotites from Lentiira–Kuhmo than at Kaavi–Kuopio. This evidence includes: (a) elevated Zr in some grains at low Ti and Y; (b) a preponderance of G9 and G10 grains with sinusoidal REE profiles, all of which are Ti, Y and Zr depleted; (c) uniformly increasing Mg# with depth implying a closed, equilibrated and therefore long-lived system; (d) harzburgite sources throughout the mantle section. The evidence provided here suggests that the Lentiira–Kuhmo lithospheric mantle may represent some of the most pristine Archean mantle lithosphere anywhere on the Earth.

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