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The lower crust of SE Belarus: petrological, geophysical and geochemical constraints from xenoliths

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Abstract

Highly altered alkaline lamprophyre tuffs of Devonian age from SE Belarus contain lower crustal xenoliths that vary from feldspar-rich mafic granulites to rarer feldspar-poor eclogitic granulites, together with coarse-grained hornblendites. Textural evidence, mineral compositions and *T*/*P* data (700–800°C/0.9–1.1 GPa) of the garnet granulite xenoliths indicate that equilibration occurred in the lower crust and therefore we consider that these xenoliths represent samples of the in situ lower crust. They are highly altered, but the main primary minerals are almandine garnet, diopside and andesine plagioclase. K-feldspar, scapolite, biotite, apatite, ilmenite, Ti-magnetite and rutile are accessory minerals. A subsequent metasomatic event, post-dating the granulite facies metamorphism, has introduced widespread pargasitic amphibole. The granulites are basic in composition, are enriched in the most incompatible trace elements and have large negative Nb anomalies (La_N/Nb_N range from 3 to 12), inferring a subduction-related origin. REE patterns show variable LREE enrichment (La/Yb = 6–27) and some positive Eu anomalies, indicating plagioclase accumulation. Trace element and REE patterns for one sample suggest a melt composition. The granulites may represent a gabbroic underplate of unknown age and may be petrogenetically related to the Osnitsk–Mikashevichi plutonic belt.

Hornblendite xenoliths are relatively unaltered and consist of pargasite or pargasite plus magnetite and apatite. Amphibole compositions are very homogeneous and resemble those in the granulite xenoliths. The $^{40}Ar/^{39}Ar$ plateau age is 381 ± 2 Ma. LREE-enriched patterns, low $^{87}Sr/^{86}Sr$ ratios and textural constraints suggest that these hornblendites precipitated from a magma formed from a small degree partial melt of a garnet-bearing mantle. They may be petrogenetically related to the host lamprophyres, possibly as a cumulate from an earlier flux of magma and related to the metasomatic event that affected the granulites.

Whole rock Vp values calculated from modal mineralogies of the granulite xenoliths range from 6.7 to 7.5 km/s, assuming equilibration at 800°C/1.0 GPa. They are compatible with seismic velocity measurements of the lower crust beneath SE Belarus, which reach 7.2 km/s. Xenoliths with the least modal plagioclase and to some extent the most modal clinopyroxene and garnet give the highest whole rock Vp values. Calculated Vp values for hornblendite xenoliths (~6.75 km/s at 800°C/1.0 GPa) suggest that identification of hornblendite in the lower crust from Vp values alone would be difficult.

Keywords: xenoliths; granulite; garnet; hornblendite; Belarus

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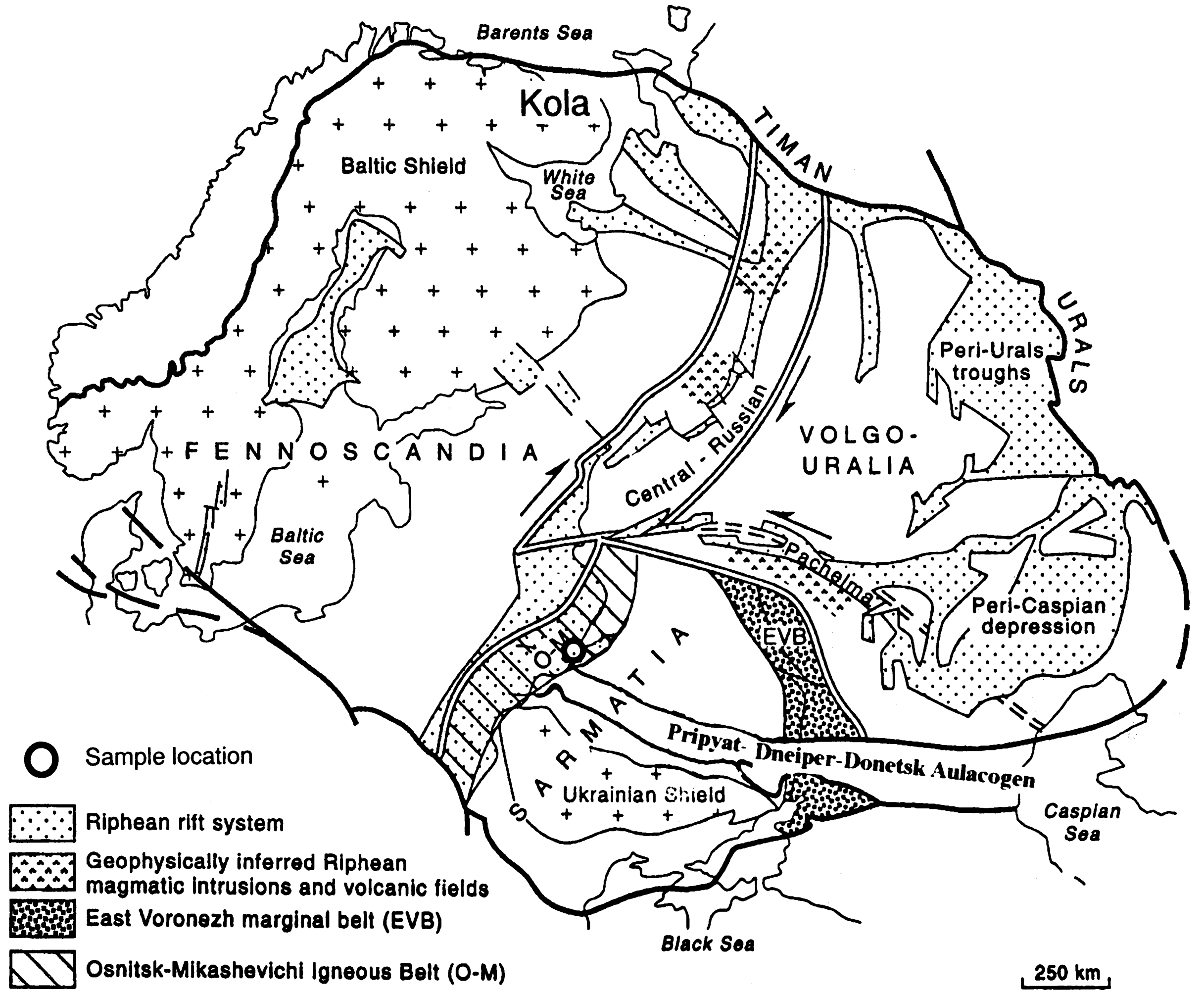

Fig. 1. Geological sketch map of the Baltic Shield and East European Platform (after Bogdanova et al., 1996) showing the location of the Zhlobin diatremes, the source of the Belarus xenoliths.

1. Introduction

1.1. Xenoliths from Archaean and Phanerozoic terrains

Our understanding of the nature and origin of the lower continental crust has been greatly improved by recent advances in the acquisition of high quality geochemical and geophysical data for xenolithic material derived from the deep crust (see reviews by Rudnick (1992) and Downes (1993)). In this paper we present new data for deep crustal xenoliths and their host magmas from SE Belarus, which forms part of the Archaean/Proterozoic East European craton. We show that the lower crust in this region is predominantly composed of mafic garnet granulites which may represent a basaltic underplating event(s) of unknown age and that it is compositionally quite distinct from mafic granulites from Belarus metamorphic terranes. The granulites have been metasomatised by mantle-derived fluids or melts which have precipitated extensive hornblende. In order to develop an interpretation of the lower crustal structure beneath SE Belarus, modal estimates of minerals and

Table 1
Electron microprobe analysis of constituent minerals in Belarus xenoliths (wt%) (abbreviations: GG = garnet granulite; H = hornblendite; or = K-feldspar; plg = plagioclase)

	Amphibole													
	By4x GG	By5x GG	By8x GG	By9x GG	By6x H	Bel1 H	Bel2 H	Bel3 H	Bel7 H	Bel10 GG	Bel8 GG	Bel9 GG	Bel4 GG	Bel6 GG
SiO_2	41.21	41.8	40.62	42.13	42.52	41.56	43.26	41.77	41.64	40.44	43.54	43.05	41.27	42.70
TiO_2	1.61	1.02	2.33	0.73	2.26	2.60	2.30	2.43	2.57	3.04	1.55	1.86	2.70	0.86
Al_2O_3	15.34	12.23	12.89	13.31	11.04	11.42	11.21	11.91	11.60	14.00	13.53	12.84	13.59	14.77
FeO	17.86	18.05	17.46	13.79	12.31	14.59	11.48	14.01	15.87	12.19	6.84	7.09	14.37	10.36
MnO	0.00	0.00	0.00	0.14	0.12	0.18	0.07	0.07	0.03	0.00	0.00	0.15	0.00	0.00
MgO	9.06	9.64	9.00	11.85	13.45	11.17	13.46	11.52	10.31	12.05	15.69	16.07	10.97	13.93
CaO	10.02	9.77	9.93	9.40	10.48	10.80	11.20	11.02	10.36	11.81	12.02	12.01	9.96	10.38
Na_2O	3.40	3.60	4.02	4.09	3.72	3.81	3.76	3.81	4.21	2.95	2.92	2.96	3.49	4.15
K_2O	1.29	1.65	1.31	1.07	1.29	1.40	1.42	1.55	1.32	2.08	2.26	1.79	1.63	0.56
Total	99.79	97.76	97.56	96.51	97.19	97.53	98.16	98.09	97.91	98.56	98.35	97.82	97.98	97.71

	Garnet								
	By4x GG	By5x GG	By9x GG	Bel4 GG	Bel5 GG	Bel6 GG	Bel8 GG	Bel9 GG	Bel10 GG
SiO_2	39.09	38.96	39.14	38.90	39.27	40.09	40.57	39.55	38.42
TiO_2	0.07	0.00	0.18	0.06	0.00	0.10	0.05	0.23	0.14
Al_2O_3	21.62	21.62	21.86	21.59	21.98	22.24	22.49	21.98	21.51
Cr_2O_3	0.00	0.00	0.00	0.00	0.00	0.00	0.32	0.00	0.00
FeO	24.26	24.30	21.60	24.60	22.71	19.34	17.74	21.62	23.40
MnO	0.38	0.63	0.56	0.57	0.50	0.39	0.43	0.00	0.46
MgO	8.27	7.86	9.92	7.80	9.00	11.23	12.86	9.09	7.38
CaO	6.21	6.53	6.08	6.32	6.24	5.81	5.54	7.19	8.17
Total	99.90	99.91	99.34	99.84	99.70	99.20	100.00	99.66	99.48
Alm	50.8	50.7	44.7	52.1	48.1	39.0	36.8	46.0	49.3
And	1.4	2.09	1.9	0.0	0.0	0.0	0.0	0.0	0.0
Gross	15.5	15.8	14.7	17.2	16.9	15.6	13.8	19.5	22.0
Pyr	31.4	30.0	37.6	29.5	34.0	42.4	47.6	34.5	27.7
Spess	0.8	1.4	1.2	1.2	1.1	0.8	0.9	0.0	1.0

	Biotite	
	By5x GG	By6x GG
SiO_2	35.79	37.26
TiO_2	0.46	3.78
Al_2O_3	16.91	13.93
FeO	20.60	12.53
MnO	0.00	0.15
MgO	11.73	15.87
CaO	0.00	0.15
Na_2O	1.26	1.12
K_2O	8.73	8.72
Total	95.48	93.53
$Mg^{\#}$	0.50	0.69

	Pyroxene			
	By5x GG	Bel4 GG	Bel8 GG	Bel9 GG
SiO_2	50.98	50.94	50.27	51.33
TiO_2	0.46	0.35	0.74	0.98
Al_2O_3	4.07	4.13	6.47	9.11
FeO	9.23	9.11	6.17	5.63
MnO	0.00	0.00	0.00	0.00
MgO	12.61	12.19	12.85	12.22
CaO	21.48	21.51	21.02	17.81
Na_2O	1.17	0.71	2.02	2.86
Total	100.00	98.94	99.54	99.94

	Scapolite
	Bel8 GG
SiO_2	45.40
Al_2O_3	26.00
CaO	16.21
Na_2O	4.03
K_2O	0.16
MgO	0.69
SO_3	5.84
Total	98.33

	Monazite
	By5x GG
P_2O_5	31.92
CaO	5.00
Nd_2O_3	6.46
Ce_2O_3	31.45
La_2O_3	22.10
SiO_2	0.60
Total	97.53

Table 1 (*continued*)

	Feldspar																
	By4x plg GG	By5x plg GG	By9x plg1 GG	By9x plg2 GG	By4x or GG	By5x or GG	By9x or GG	Bel4 or GG	Bel4 plg GG	Bel5 or GG	Bel5 plg GG	Bel8 or GG	Bel8 Plg GG	Bel9 Plg1 GG	Bel9 plg2 GG	Bel10 plg GG	Bel10 Or GG
SiO_2	58.47	58.58	56.99	60.59	64.72	64.70	62.61	63.86	57.43	64.12	56.46	62.7	59.47	61.09	57.65	55.58	63.06
TiO_2	0.00	0.00	0.00	0.00	0.67	0.00	0.00	0.00	0.00	0.00	0.00	0.68	0.00	0.00	0.00	0.00	0.18
Al_2O_3	25.61	25.81	26.86	23.32	17.81	18.17	17.50	18.12	26.12	18.37	27.27	20.18	25.25	23.84	26.18	28.02	17.81
FeO	0.00	0.00	0.12	0.00	0.00	0.31	0.00	0.43	0.00	0.00	0.00	0.37	0.00	0.00	0.00	0.00	0.24
MnO	0.00	0.00	0.00	0.00	0.00	0.00	0.39	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.56	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.40
CaO	7.54	7.73	8.54	4.89	0.00	0.00	0.00	0.14	7.64	0.00	9.15	0.36	6.25	2.07	8.17	10.03	0.11
Na_2O	6.75	6.35	6.55	8.10	0.00	0.18	0.17	0.32	7.30	0.23	6.53	2.54	6.46	5.72	7.07	6.01	0.58
K_2O	0.49	0.88	0.23	0.75	15.98	15.57	15.47	14.96	0.91	16.88	0.43	11.48	2.53	5.95	0.40	0.18	15.24
Total	98.86	99.35	99.29	97.65	99.18	98.93	96.14	98.39	99.4	99.6	99.84	98.31	99.96	98.67	99.47	99.82	97.62
Ab	60.1	56.7	57.4	71.7	0.0	1.7	1.6	3.0	60.0	2.1	55.0	24.7	55.8	53.1	59.7	51.5	5.4
An	37.1	38.1	41.3	23.9	0.0	0.0	0.0	0.8	35.0	0.3	42.6	1.9	29.8	10.6	38.1	47.5	0.6
Or	2.9	5.2	1.3	4.4	100.0	98.3	98.4	96.2	5.0	97.6	2.4	73.4	14.4	36.3	2.2	1.0	94.0

their laboratory derived compressional (Vp) wave velocities have been used to calculate possible bulk rock Vp values. The calculated values agree well with observed lower crustal seismic velocities in the region.

1.2. Geological setting

During Devonian times the East European platform was affected by widespread magmatism, related to a series of rising mantle plumes and rift systems (Wilson and Lyashkevich, 1996). The Pripyat–Dneiper–Donets (PDD) rift, located in the south-western segment of the East European Craton (Fig. 1), was the site of voluminous Devonian magmatic activity (Kusznir et al., 1996). Several xenolith-bearing diatremes associated with this magmatism occur on the north-east shoulder of the PDD rift, close to the Loev–Braguin saddle (Fig. 1) in Belarus. They cut through the 1.97–2.02 Ga old Osnitsk–Mikashevichi Igneous Belt (Shcherbak, 1991) and therefore their entrained xenoliths could be representative of the lower crust of the Sarmatian segment of the East European Platform. The age of the diatremes is well constrained stratigraphically to the Late Frasnian (~370 Ma). Our samples were taken from drillings of sub-surface diatremes in the Zhlobin field, situated on the banks of the Dnieper river, between Rogachev and Zhlobin in SE Belarus. In this study the host magmatic rocks and their entrained xenoliths will be compared with selected magmatic rocks from the PDD rift and xenoliths from the northern Baltic Shield. We will show that xenoliths representing the lower crust of Belarus are similar to lower crustal xenoliths from the Kola peninsula, NW Russia.

1.3. Analytical techniques

Electron microprobe analyses of mineral compositions (Table 1) were obtained using a Jeol 733 Superprobe (Birkbeck College) with an Oxford Instruments ISIS energy dispersion system. Analytical conditions were 15 kV accelerating voltage, a spot diameter of 1–2 μm and a counting time of 100 s. Host magmatic rocks and 6 xenoliths of suitable size were analysed by X-ray fluorescence (University of Edinburgh), using glass discs for major elements and pressed powder blocks for trace elements (Table 2). Rare earth element (REE) data were obtained from Royal Holloway, University of London, using a Phillips PV8060 ICP-AES simultaneous/sequential spectrometer (Table 3). REE and other trace element data were obtained on hornblende separates from hornblendite xenoliths by ion probe at the University of Pavia, Italy (Table 4). Sr and Nd isotope ratios were obtained from Royal Holloway, University of London, using a VG354 five collector mass

Table 2
XRF major and trace element compositions of host magmatic rocks, hornblendite and granulite xenoliths from Belarus (abbreviations: H = hornblendite xenoliths; GG = garnet granulite; * signifies below detection limit)

	By10 host	By11 host	By12 host	By13 host	By14 host	By15 host	Bel2 H	Bel3 H	Bel7 H	By5x GG	By9x GG	Bel8 GG
SiO_2	39.73	39.26	35.29	44.95	29.33	42.57	41.93	40.66	40.55	50.56	50.94	45.43
Al_2O_3	8.53	8.82	7.56	10.04	13.46	12.30	11.04	11.62	10.85	20.48	17.37	16.04
Fe_2O_3	12.07	14.52	12.45	12.26	12.76	13.59	12.34	16.90	16.78	8.06	7.57	10.51
MgO	14.34	17.73	17.53	12.87	5.84	11.43	14.63	12.41	12.93	4.55	7.66	12.18
CaO	9.71	5.74	10.05	4.34	15.16	8.05	10.59	10.04	8.48	6.09	5.25	10.32
Na_2O	1.22	1.40	1.41	1.17	1.76	2.63	3.15	3.05	2.84	3.33	3.04	1.17
K_2O	3.307	2.086	1.382	5.365	2.56	1.724	1.502	1.503	1.445	2.756	4.617	1.277
TiO_2	1.417	1.539	1.618	1.715	2.16	1.838	2.116	2.316	2.247	1.188	0.382	0.455
MnO	0.197	0.25	0.189	0.098	0.76	0.237	0.123	0.145	0.152	0.104	0.109	0.188
P_2O_5	0.446	0.478	0.766	0.676	0.86	0.845	0.02	0.018	0.324	0.529	0.053	0.100
LOI	9.44	8.37	12.19	6.37	15.12	5.15	2.06	1.03	2.30	1.58	2.26	1.01
Total	100.41	100.19	100.45	99.85	99.77	100.37	99.50	99.69	98.90	99.22	99.25	98.67
Nb	30.6	34.2	69.5	59.2	81.0	82.7	11.3	11.4	14.6	10.0	5.5	2.1
Zr	86.4	92.6	150.4	123.1	173.0	171.7	44.1	37.1	56.7	112.2	43.0	64.1
Y	12.9	13.9	22.3	16.5	27.0	23.3	7.8	8.1	8.1	13.6	8.3	35.5
Sr	582.1	219.0	648.0	467.0	1074	1099.9	290.5	323.0	359.0	831.1	308.7	838.1
Rb	42.4	35.7	31.3	64.6	36.7	27.9	10.6	8.2	10.4	14.3	32.2	25.9
Th	4.1	5.2	5.9	6.4	8.3	9.1	3.7	2.5	3.4	*	*	*
Pb	5.9	4.9	4.0	4.4	13.5	10.2	22.6	*	9.5	13.6	6.4	6.2
Zn	131.2	176.1	98.2	80.7	260.3	197.3	68.7	90.6	93.7	48.6	49.6	52.7
Cu	246.1	255.9	214.2	252.3	294	220.7	1018.0	286.4	502.5	78.0	31.2	13.6
Ni	155.8	154.4	236.7	50.1	47.0	53.3	248.2	65.9	51.3	32.7	41.1	167.7
Cr	379.7	359.5	264.5	105.6	125	93.0	158.9	31.1	25.8	32.3	108.3	1678.3
V	260.1	269.2	256.7	298.9	373	289.3	353.3	426.6	405.9	192.3	152.5	171.9
Ba	1258.8	441.2	864.3	849.2	888.0	1185.3	202.0	225.6	3669.7	825.0	497.8	3002
Sc	30.4	28.8	19.1	25.4	21.7	18.3	56.7	35.9	31.0	18.5	33.8	49.1
La	26.0	36.7	90.4	59.9	93.0	81.2	*	*	*	27.8	*	25.0
Nd	27.9	32.7	64.9	42.6	77.5	50.1	11.1	6.0	*	23.9	3.2	29.7
Ce	54.2	63.3	157.0	96.0	171.0	126.1	20.0	9.1	*	61.9	14.1	54.9

spectrometer (Table 5). Results from SRM 987 ranged from 0.710227 ± 14 to 0.710253 ± 9, and an internal laboratory Aldrich standard gave a value of 0.51142 ± 4, equivalent to a value of 0.511856 for the La Jolla standard. An ^{40}Ar–^{39}Ar plateau age was determined on hornblende separated from a hornblendite xenolith, using a step heating method, at the Institute of Earth Sciences, University of Montpellier. Pressure estimates were calculated using garnet–clinopyroxene–plagioclase–quartz equilibria (Newton and Perkins, 1982; Powell and Holland, 1988) and temperature estimates were made using Fe^{2+}/Mg equilibria in garnet–clinopyroxene (Powell, 1985). The absence of quartz from the xenoliths means that pressures are maximum estimates. Direct measurements of compressional wave velocities (Vp) cannot be made because of the small size of the Belarus samples (Christensen and Fountain, 1975; Kern and Schenk, 1985). Therefore we have calculated Vp values for eight of the least altered granulite xenoliths using modal mineral proportions obtained from a 1000 point count (Table 6) and single crystal Voigt–Reuss–Hill averages of compressional wave velocities from published sources (Christensen, 1989; Furlong and Fountain, 1986) that correspond most closely to the mineral compositions in these samples (Table 7). Temperature and pressure derivatives (Table 7) for each mineral phase were used to correct Vp values to an equilibration temperature of 800°C and pressure of 1.0 Gpa. Results are given in Table 8.

Table 3
REE abundances in ppm in host magmatic rocks and granulite xenoliths from Belarus (analysis by ICP, University of London)

	By10	By11	By12	By13	By14	By15	Bel8	By5x	By9x
La	27.33	29.01	67.26	48.85	71.20	70.62	18.9	29.7	5.0
Ce	57.73	61.79	133.55	96.8	150.37	136.78	49.59	57.67	9.2
Pr	6.93	7.26	14.68	10.92	15.64	15.12	7.37	6.85	1.03
Nd	25.81	27.18	49.86	37.78	54.5	52.07	31.1	28.0	4.6
Sm	4.35	4.49	7.71	5.90	8.34	8.16	5.78	4.65	0.96
Eu	1.27	1.26	2.20	1.67	2.60	2.40	1.82	2.03	0.55
Gd	3.57	3.76	6.17	4.77	7.40	6.44	5.8	4.03	1.23
Dy	2.80	2.78	5.07	3.73	5.22	5.06	5.58	2.61	1.23
Ho	0.51	0.51	0.98	0.69	1.01	0.96	1.16	0.49	0.27
Er	1.22	1.23	2.49	1.68	2.18	2.29	3.3	1.24	0.55
Yb	0.99	0.93	2.31	1.47	1.90	1.92	2.76	1.12	0.81
Lu	0.14	0.13	0.35	0.22	0.31	0.28	0.43	0.18	0.13
La/Yb	27.61	31.19	29.12	33.23	37.47	36.78	6.85	26.52	6.17

2. Results

2.1. Petrology, mineralogy and mineral chemistry

Little work has been published on the Devonian diatremes from SE Belarus, although Wilson and Lyashkevich (1996) reported geochemical and isotopic analyses of two alkali ultramafic rocks, described as kimberlite-like, from this region. The xenoliths contained within the Belarussian diatremes have not been previously analysed by modern geochemical techniques. The drillcore samples consist of: (a) grey/green highly altered, brecciated and veined lapilli tuffs, (b) magmatic clasts of alkaline ultramafic rocks (By10–15) which are also highly altered; these often contain lower crustal granulite and/or hornblendite xenoliths; (c) post-magmatically altered garnet granulite xenoliths, ranging in size from 1–3 cm; and (d) hornblendite and hornblende–apatite xenoliths (2–4 cm), which are somewhat fresher than most of the other rocks.

Table 4
REE and trace element abundance in amphibole separates from hornblendite xenoliths from Belarus (analysis by SIMS ion probe, Pavia) (* signifies below detection limit)

	By6x	Bel2	Bel3	Bel7
	H	H	H	H
K	*	11107	13097	10696
Sc	53	53	35	40
Ti	12000	12146	14709	14443
V	355	329	386	416
Cr	168	97	45	55
Rb	*	13	17	15
Sr	256	297	359	275
Y	6	6	6	6
Zr	39	36	33	42
Nb	8	9	10	9
Ba	178	189	270	215
La	3.47	4.50	4.00	2.83
Ce	9.78	14.40	11.80	9.20
Nd	8.03	9.90	8.70	8.20
Sm	2.00	2.27	2.06	2.06
Eu	0.69	0.68	0.72	0.67
Gd	1.76	1.60	1.70	1.84
Dy	1.43	1.40	1.42	1.39
Er	0.59	0.56	0.58	0.66
Yb	0.46	0.52	0.54	0.77
La/Yb	58.3	8.7	7.4	3.6
La/Nb	0.4	0.5	0.4	0.3

The host magmatic rocks (By10–15) are fine grained and contain calcite pseudomorphs after melilite or plagioclase. Also present are elongate apatite crystals (0.5 mm) with Sr-rich rims (13.6% SrO) and Sr-bearing barite. Skeletal diopside crystals have been partially pseudomorphed by chlorite and calcite. Fresh acicular crystals of aegerine often form epitaxial overgrowths on the diopside laths. Highly altered biotite crystals are rare. The groundmass is composed of micro-crystalline chlorite, calcite, Ti-rich magnetite and rare chalcopyrite and barite. The least altered sample (By11) contains abundant megacrysts of diopside with coronas of chlorite, rare anhedral crystals of sphene, together with magnetite, K-feldspar and barite. From their

Table 5
Sr and Nd isotopic data for host magmatic rocks, hornblendite and granulite xenoliths from Belarus (abbreviations: H = hornblendite (separates); GG = granulites; WR = whole rock (hornblendite); i = initial ratio, 370 Ma for granulites and host and 381 Ma for hornblendites)

	$^{87}Sr/^{86}Sr$	$(^{87}Sr/^{86}Sr)_i$	ϵSr_i	$^{143}Nd/^{144}Nd$	$(^{143}Nd/^{144}Nd)_i$	ϵNd_i
By11(host)	0.70905 ± 1	0.70657	35.66			
By12(host)	0.70559 ± 1	0.70486	11.37			
By14(host)	0.70469 ± 1	0.70420	1.61	0.512321 ± 5	0.512097	−1.27
By6x(H)	0.70347 ± 1			0.512447 ± 5		
Bel2(H)	0.70361 ± 1	0.70293	−15.90	0.512562 ± 5	0.512216	1.34
Bel3(H)	0.70379 ± 1	0.70305	−14.13	0.512422 ± 5	0.512065	−1.61
Bel7(H)	0.70368 ± 1	0.70283	−17.30	0.512376 ± 5	0.541197	−2.94
Bel2(WR)	0.70390 ± 1	0.70333	−10.23			
Bel3(WR)	0.70400 ± 1	0.70360	−6.32			
Bel7(WR)	0.70515 ± 1	0.70470	9.24			
By5x(GG)	0.70444 ± 1	0.70418	1.67	0.511418 ± 5	0.511175	−19.26
By9x(GG)	0.70628 ± 1	0.70464	8.32	0.511929 ± 3	0.511623	−10.51
Bel8(GG)	0.70458 ± 1	0.70411	0.73	0.511732 ± 4	0.511460	−13.70

textural appearance, By14 and By15 appear to be the same rock, but By14 is much more altered than By15, which may explain its unusual major element chemistry.

Granulite xenoliths from the Belarus diatremes have fine to medium grained granoblastic textures showing characteristic triple junction equilibrium fabrics (Fig. 2). The mineralogy is consistently: ga + plg ± cpx ± amp ± opx ± ap ± scap ± rt ± bi ± K-fsp. Estimated modal proportions are given in Table 6. Both plagioclase-rich garnet granulites and plagioclase-poor (plag < 5%) eclogitic granulites are present. The pink subhedral–anhedral garnets occur in variable amounts (6–36%). All garnets have pressure release cracks and rims which are altered to cryptocrystalline kelyphite. Inclusions of rounded apatite and magnetite are present in several garnets; one apatite inclusion in By5x contained very small monazite crystals as well as small exsolution needles of rutile. The garnets are unzoned almandine-rich solid solutions with a representative compositional range of 37–52 wt% almandine, 28–48 wt% pyrope, 14–22 wt% grossular, 0–2 wt% andradite and 0–1.4 wt% spessartine (Table 1, Fig. 3). Unzoned anhedral–subhedral plagioclase feldspars are mostly andesine with a compositional range of An_{48}–An_{35}, Ab_{60}–Ab_{51}, Or_5–Or_1. However, samples By9x and Bel9 contain rare oligoclase (An_{24}, Ab_{72}, Or_4) and a rare ternary feldspar (An_{11}, Ab_{53}, Or_{36}), respectively. Only micro-antiperthites were found in Bel8. Modal proportions of plagioclase are highly variable, ranging from 1–49%. K-feldspar (Or_{96}–Or_{100}) occurs as small

Table 6
Estimated (1000 point count) modal mineralogy for granulite xenoliths from Belarus (mineral abbreviations: Ga = garnet Plg = plagioclase. Cpx = clinopvroxene, Kf = K-feldspar, Op = opaques, Scap = scapolite, Amp = amphibole, Ap = apatite, Bi = biotite)

	Modal mineralogy									
Sample	Ga	Plg	Cpx	Kf	Op	Scap	Amp	Ap	Bi	Alt
Bel10	28.1	1.0		5.0	4.0		29.8	0.2		31.9
By4x	25.2	9.1		4.5	3.0		18.2	5.6		34.4
Bel8	36.4	4.2	34.4			16.3	8.2	0.5	tr	
Bel9	10.2	22.0	18.2		5.8		20.4	3.0		20.4
Bel4	34.4	20.3		11.3	2.8		11.6	1.2		18.4
Bel5	28.2	20.0		12.6	4.0			2.0	tr	33.2
By9x	5.5	35.5		11.9	tr		35.2		4.0	7.9
By5x	16.4	48.7	1.4	10.7	tr		8.0	Tr	tr	14.8

Table 7
Experimental compressional wave velocities and their pressure and temperature derivatives for minerals commonly found in Belarus granulite xenoliths, used for calculating whole rock compressional wave velocities (Vp) (compressional wave velocities (Vp) and their *P* and *T* derivatives without superscript were taken from Furlong and Fountain (1986) and references therein). Plagioclase was assumed to be An_{50}. Vp for garnet corresponds to the formula $3(Fe_{63}Mg_{29}Ca_8)O{\cdot}Al_2O_3{\cdot}3SiO_2$. Where Vp values are not available (e.g. scapolite), they were calculated from published elastic moduli (Bass, 1995; Christensen, 1989). Temperature and pressure derivatives for plagioclase, K-feldspar and scapolite were assumed to be the same as those of quartz, derivatives for hornblende and biotite were equated to those of hypersthene, and rutile derivatives were used for the opaques. Derivatives for garnet of composition $3(Fe_{76}Mg_{21}Ca_3)O{\cdot}Al_2O_3{\cdot}3SiO_2$ were used. Similar assumptions have been used by Furlong and Fountain (1986) and Hynes and Snyder (1995). Apatite was not adjusted for temperature and pressure, as it is present in only small modal proportions

	Vp (km/s)	dVp/dP (km/s/GPa)	− dVp/dT (km/s/K)
Garnet	8.17	7.84×10^{-2a}	3.93×10^{-4a}
Diopside	7.85	2.04×10^{-1}	6.33×10^{-4}
Plagioclase	6.166	13.67×10^{-2}	0.16×10^{-3}
Scapolite	5.63^b	13.67×10^{-2}	0.16×10^{-3}
Amphibole	7.04^a	2.04×10^{-1}	6.33×10^{-4}
Opaque	7.4	7.6×10^{-2}	0.943×10^{-3}
K-feldspar	5.556	13.67×10^{-1}	0.16×10^{-3}
Apatite	6.7	0	0
Biotite	5.26^a	2.04×10^{-1}	6.33×10^{-4}

[a] Values from Christensen (1989).

[b] Value calculated from elastic constants from Bass (1995).

Table 8
Calculated values of seismic compressional wave velocities (km/s) for selected granulite xenoliths from Belarus assuming equilibration at 800°C and 1.0 GPa. In the first column amphibole was included in the modal estimate and in the second column amphibole was recalculated as diopside. Alteration products were assumed to have been derived from primary diopside and calculated accordingly (abbreviations: cpx = clinopyroxene; amp = amphibole)

	cpx + amp (800°C/1.0 GPa)	amp = cpx (800°C/1.0 GPa)
Bel8	7.22	7.27
Bel10	7.26	7.46
Bel9	7.00	7.13
By4x	7.17	7.29
Bel4	7.03	7.11
By9x	6.45	6.69
By5x	6.65	6.70
Bel5		6.88

anhedral interstitial crystals (Table 1, Fig. 4). Pyroxenes are all unzoned diopside; their modal proportions vary from 1–34%. However, primary pyroxenes are absent from most samples. In By5x they occur as small (0.5–1.5 mm) anhedral–subhedral crystals, which have been pseudomorphed by amphibole, and as much larger, relatively fresh, subhedral crystals in the other samples. Brown amphibole occurs as crystal aggregates possibly pseudomorphing primary pyroxene. Using IMA nomenclature (Leake, 1978), ferroan pargasite and ferroan pargasitic hornblende are the main compositions (Table 1) with $Mg^{\#}$ ranging from 0.5–0.7, although Bel8 and Bel9 contain pargasitic hornblende ($Mg^{\#} = 0.85$). Where more than one amphibole composition occurs in a sample, no distinguishing petrographic features were found. Table 1 shows the compositions of accessory minerals, including rare platy biotite found ($Mg^{\#} = 0.5$) in By5x and altered phlogopite ($Mg^{\#} = 0.88$) in Bel8. Scapolite (mizzonite) occurs only in Bel8. Minor phases include magnetite, apatite often in close contact with anhedral rutile, and secondary calcite. Temperature determinations give values ranging from 700–800°C and maximum pressure estimates are 0.9–1.1 GPa. Without exception the granulites have been overprinted by subsequent retrogressive metamorphic events, with mineral assemblages of chlorite ± amphibole ± white mica ± sphene. The coexistence of mineralogies from more than one metamorphic facies in the xenolith suite reveals a complex and extended metamorphic history. The retention of small amounts of plagioclase in the eclogitic granulites may be explained kinetically.

Hornblendite xenoliths are coarse to medium grained (0.5–10 mm) melanocratic ultramafic rocks, consisting either of pure amphibole or of amphibole with magnetite (up to 10%) and apatite (up to 20%). A retrograde metamorphic event has variably altered these rocks, introducing minerals such as chlorite, zeolites, calcite, barite, chalcopyrite and pyrite. Secondary biotite ($Mg^{\#} = 0.69$) and K-feldspar were also found (Table 1). Texturally most hornblendite xenoliths show no evidence of stress, having near decussate fabrics, although Bel2 and Bel3 are atypical having granoblastic and proto-mylonitic textures, respectively. Compositionally the amphiboles are

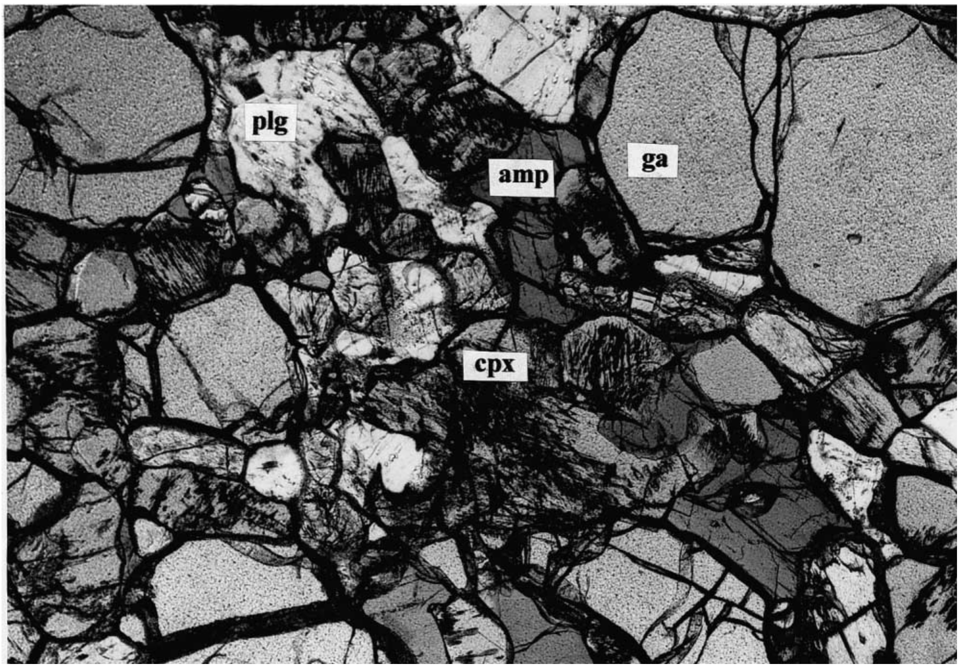

Fig. 2. Typical garnet granulite xenolith from the Belarus suite (Bel9). Garnet (Ga), diopside (cpx) and plagioclase (Plg) show an equilibrium fabric. Metasomatic amphibole (amp) is also present in this rock.

very homogeneous, being predominantly ferroan pargasite or ferroan pargasitic hornblende (Table 1), similar to those found in the garnet granulite xenoliths.

2.2. Geochemistry

The host magmatic rocks have high losses on ignition (5–15 wt%), which may be due to primary mineralogy, or more likely, alteration. With the exception of By14, which is the most highly altered rock, the SiO_2 content ranges from 35–45 wt%, while the MgO content is 11–17 wt%, indicating that the hosts were silica-undersaturated mafic magmas (Table 2). All samples are alkalic (Fig. 5), although the high degree of alteration and mobility of K and Na may have affected original alkali concentrations. Sample By14 has an anomalous major element composition (29 wt% SiO_2 and 5.8 wt% MgO). Its high wt% CaO (15%) correlates to its high content of secondary calcite and the highest loss on ignition. Trace element data for the host magmatic rocks show a considerable range (Table 2). Ni and Cr concentrations vary from 237–50 ppm and 380–93 ppm, respectively. High Ba and Sr concentrations may reflect barite and calcite alteration. Incompatible trace element abundances, normalised to primitive mantle, show an enrichment in all elements (Fig. 6a) with generally greater concentrations in the most incompatible elements. By11, By15 and By12 also show a trough at K. The higher K and Sr concentrations in By10 and By13 may be the result of more extensive alteration. Zr, Ti and Y concentrations are low. REE patterns are parallel and show strong LREE enrichment (Table 3, Fig. 7a), with La/Yb ratios from 28–38. The trace element concentrations of By14 are very similar to those of By15. $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ initial ratios ($t = 370$ Ma) for By14 are 0.70420 and 0.512097, respectively, and fall in the enriched field of the $\epsilon Sr-\epsilon Nd$ diagram (Fig. 8a). $^{87}Sr/^{86}Sr$ initial ratios for By11 and By12 are 0.70657 and 0.70486 respectively. These high values are thought to be due to alteration.

The highly altered nature of the granulite xenoliths may have significantly affected major and trace element concentrations, especially the more mobile

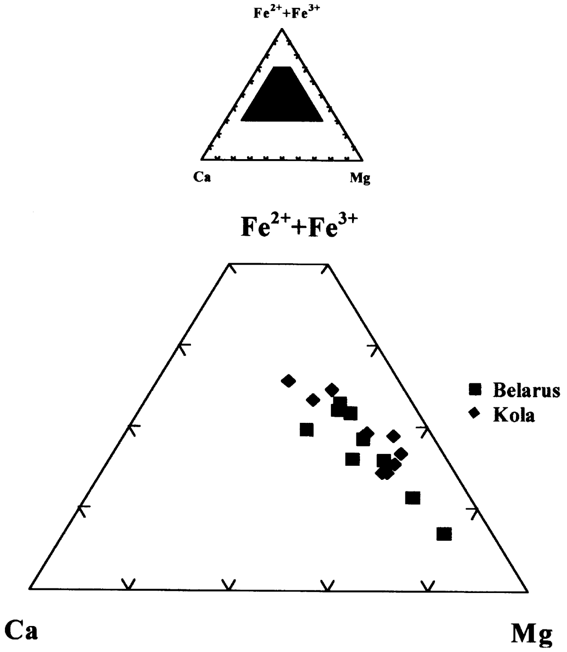

Fig. 3. Ternary plot of representative compositions of garnets from granulite xenoliths from Belarus (this paper) and Kola (Kempton et al., 1995).

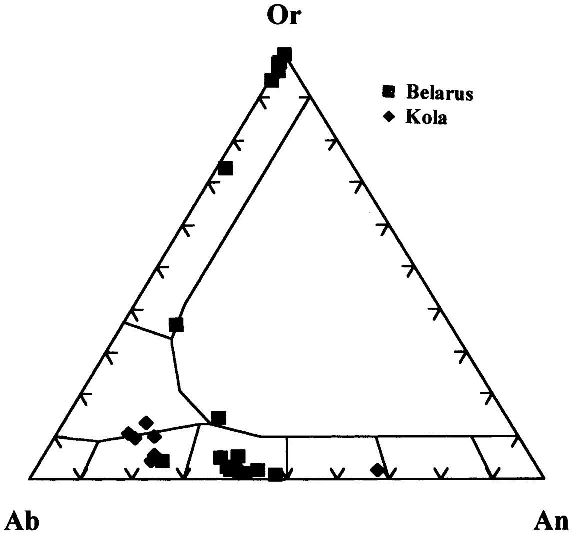

Fig. 4. Ternary plot of representative plagioclase compositions from granulite xenoliths from Belarus (this paper) and Kola (Kempton et al., 1995).

elements. In addition, the presence of metasomatic amphibole would undoubtedly affect whole rock chemistry and possibly mineral compositions, and so their interpretation requires caution. Garnet granulite and eclogitic granulite xenoliths from Belarus are basic in composition (Table 2), with 45–51 wt% SiO_2 and 5–12 wt% MgO. Eclogitic granulite Bel8 has higher MgO and lower SiO_2 contents than garnet granulites By5x and By9x. K_2O exceeds Na_2O in two of the three samples, with a maximum K_2O/Na_2O of 1.5. However, alteration may have affected alkali metal concentrations. P_2O_5 concentrations vary greatly from 0.05 to 0.53 wt%, Bel5x and By9x being significantly depleted in P, whereas all other major element concentrations are similar. Low Ni concentrations in By5x and By9x (33–41 ppm) contrast with the higher value of 168 ppm in Bel8. Normative compositions (Table 9) show Bel8 and By5x to be sub-alkaline. By9x is weakly alkaline in composition, but this is likely to be a consequence of alteration and not a primary characteristic. Trace element data (Table 2), normalised to primitive mantle, show the granulites to be variably enriched in the most incompatible elements, with very low concentrations of the less incompatible elements P, Zr, Ti and Y. All samples show prominent negative Nb anomalies ($La_N/Nb_N = 2.9–12.3$) and are variably enriched in Ba and Sr (Fig. 6b). They are LREE enriched, with La/Yb varying from 6 to 27 (Table 3, Fig. 7b). By5x and Bel8 have similar REE concentrations but By9x shows much lower total concentrations. REE patterns for By5x and By9x are similar and have significant positive Eu anomalies ($Eu/Eu^* = 1.4$ and 1.6, respectively), which coincide well with their Sr enrichment (Fig. 6b). Measured $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios for the granulites are low, ranging from 0.70444–0.70628 to 0.511418–0.511929, respectively (Table 5).

The three least altered hornblendites (Bel2, Bel3 and Bel7) have major element compositions that plot in the basanite field of the total alkali–SiO_2 diagram (Cox et al., 1979). Their major elements show little variation with 41–42 wt% SiO_2, 12–15 wt% MgO and 9–11 wt% CaO. Trace element concentrations from amphibole separates are also relatively homogeneous and, when normalised to primitive mantle, show enrichment in the most incompatible elements (Fig. 6c). Whole rock trace element

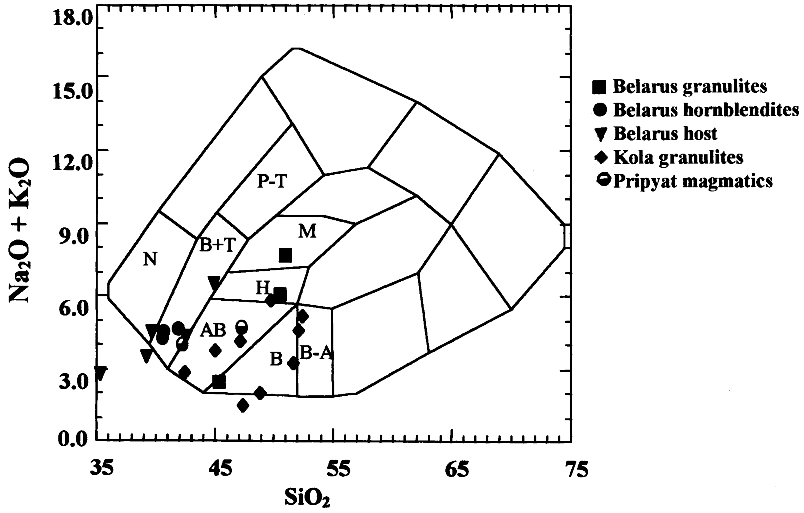

Fig. 5. $K_2O + Na_2O$ vs SiO_2 plot (Cox et al., 1979) for Belarus host magmatic rocks, hornblendite xenoliths from Belarus (this paper) and granulite xenoliths from Belarus (this paper), Kola xenoliths (Kempton et al., 1995) and Pripyat magmatics (Wilson and Lyashkevich, 1996). Abbreviations: B = basalt, T = tephrite, N = nephelinite, P = phonolite, AB = alkali basalt, B-A = basaltic andesite, H = hawaiite.

concentrations are very similar for the two least altered samples (Bel2 and Bel3), but Bel7 has notably higher concentrations of Ba (Fig. 6d) than Bel2 and Bel3, probably due to secondary barite. Both Bel2 and Bel3 are strongly depleted in P. The higher P value in Bel7 is reflected in its much higher modal apatite. However, transition metal concentrations vary significantly, e.g. 51–248 ppm Ni and 26–159 ppm Cr. Fig. 7c shows that hornblendite xenoliths are slightly LREE enriched (La/Yb = 3.6–8.7) with a maximum at Nd, except for Bel2. Hornblende mineral separates from the hornblendite xenoliths have low measured $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios (Table 5) that plot within the depleted mantle field when age corrected to 381 Ma (Fig. 8a). Similar isotopic values were obtained from whole rock samples (Table 5, Fig. 8a), with the exception of Bel7 which gave a slightly higher $^{87}Sr/^{86}Sr$ value due to its more extensive alteration.

2.3. Calculated whole rock Vp values

Calculated bulk rock Vp values (Table 8) for the Belarus xenoliths at 800°C and 1.0 GPa vary between 6.7 and 7.5 km/s. The higher Vp values (7.3–7.5 km/s) were obtained from eclogitic granulites, which are significantly richer in garnet and diopside and poorer in plagioclase. An inverse linear relationship is found between modal plagioclase and bulk rock Vp (Fig. 9).

Many sources of error exist in the calculation of whole rock Vp values. These include (a) an assumption that the thin sections are representative of both the sample and the body of lower crust from which they were derived (see Downes, 1993; Rudnick and Fountain, 1995), (b) that the 1000 point-count accurately represents the volumetric abundance of minerals in the samples, (c) that above 0.2 GPa complete annealing of microcracks occurs (Kern and Schenk, 1985), hence the effective pore space within these rocks under granulite facies conditions is zero, and (d) linear relationships exists between Vp and temperature and between Vp and pressure. The values used are in many cases approximations and assumptions, and for simplicity assume that averaging due to the randomness of orientation of anisotropic mineral phases produces an isotropic sample. This may be a large source of error, as metamorphism often produces preferred orientation in minerals which can markedly affect bulk rock Vp (Christensen, 1965; Kern, 1978). However, Christensen and Mooney (1995) have shown that velocity anisotropy is quite limited in mafic garnet granulites

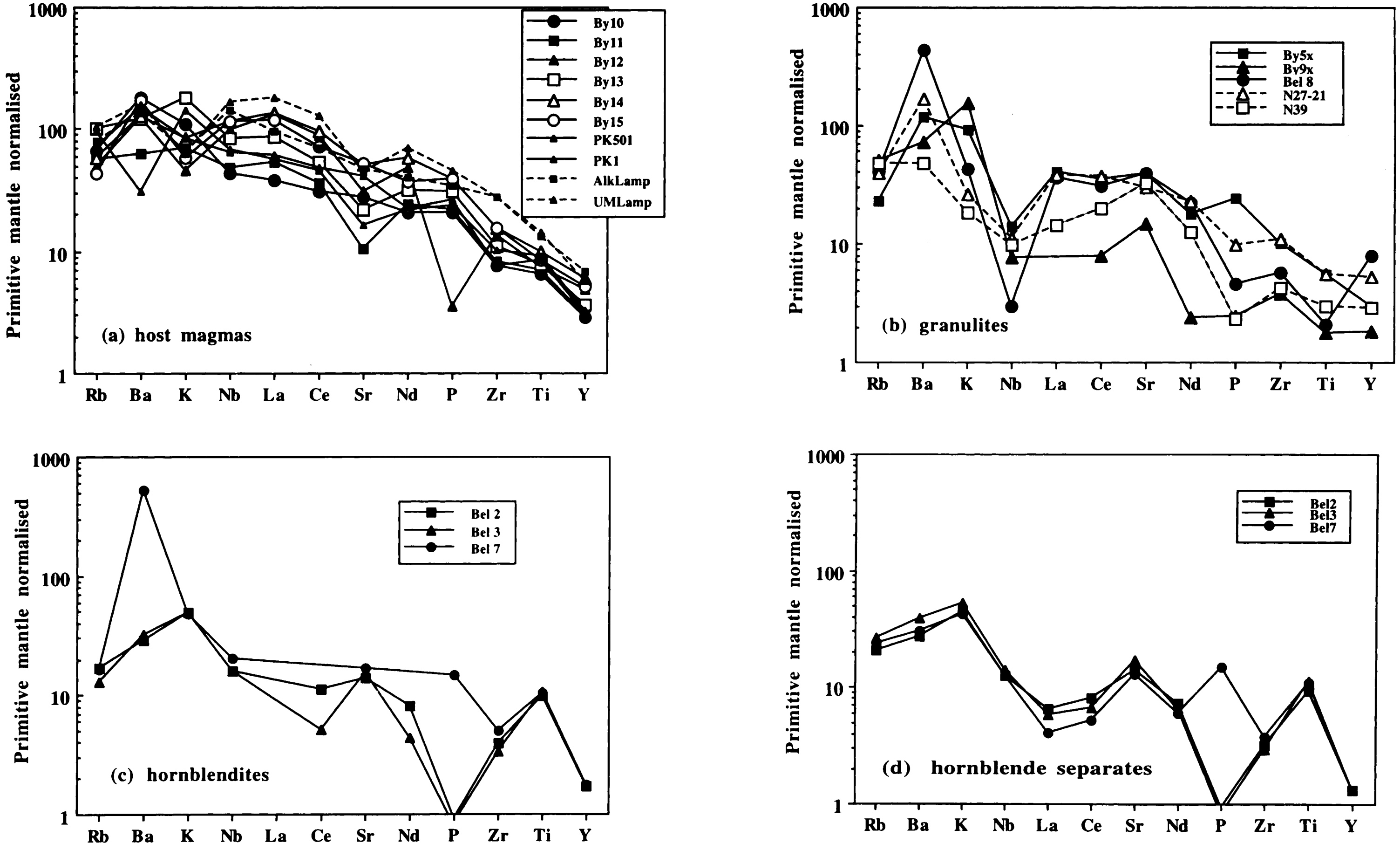

Fig. 6. Trace element plot of (a) whole rock host magmatic rocks from Belarus, kimberlite-like rocks PK501 and PK1 from Pripyat segment (Wilson and Lyashkevich, 1996) and average alkaline lamprophyre (Rock, 1991); (b) whole rock granulite xenoliths from Belarus (this paper) and Kola (Kempton et al., 1995); (c) amphibole separates from hornblendite xenoliths from Belarus (this paper) and Kola (Beard, A., unpubl. data); (d) whole rock hornblendite samples from Belarus (this paper). Normalised to primitive mantle (Sun and McDonough, 1989).

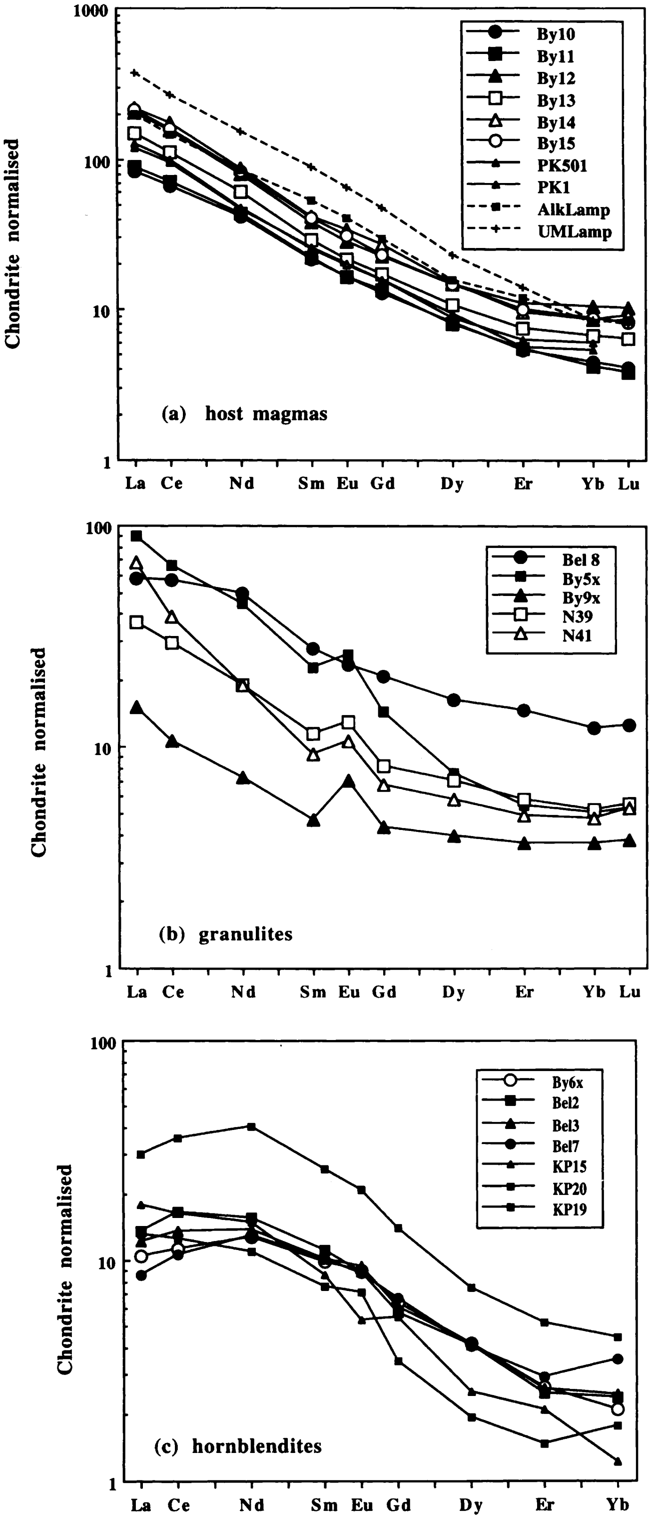

Fig. 7. REE plots of: (a) whole rock host alkaline lamprophyres from Belarus and kimberlite-like rocks samples PK501 and PK1 from the Pripyat segment (Wilson and Lyashkevich, 1996); (b) whole rock granulite xenoliths from Belarus (this paper); (c) amphiboles from Belarus (this paper) and Kola (Beard, A., unpubl. data). Normalisation coefficients from Nakamura (1974).

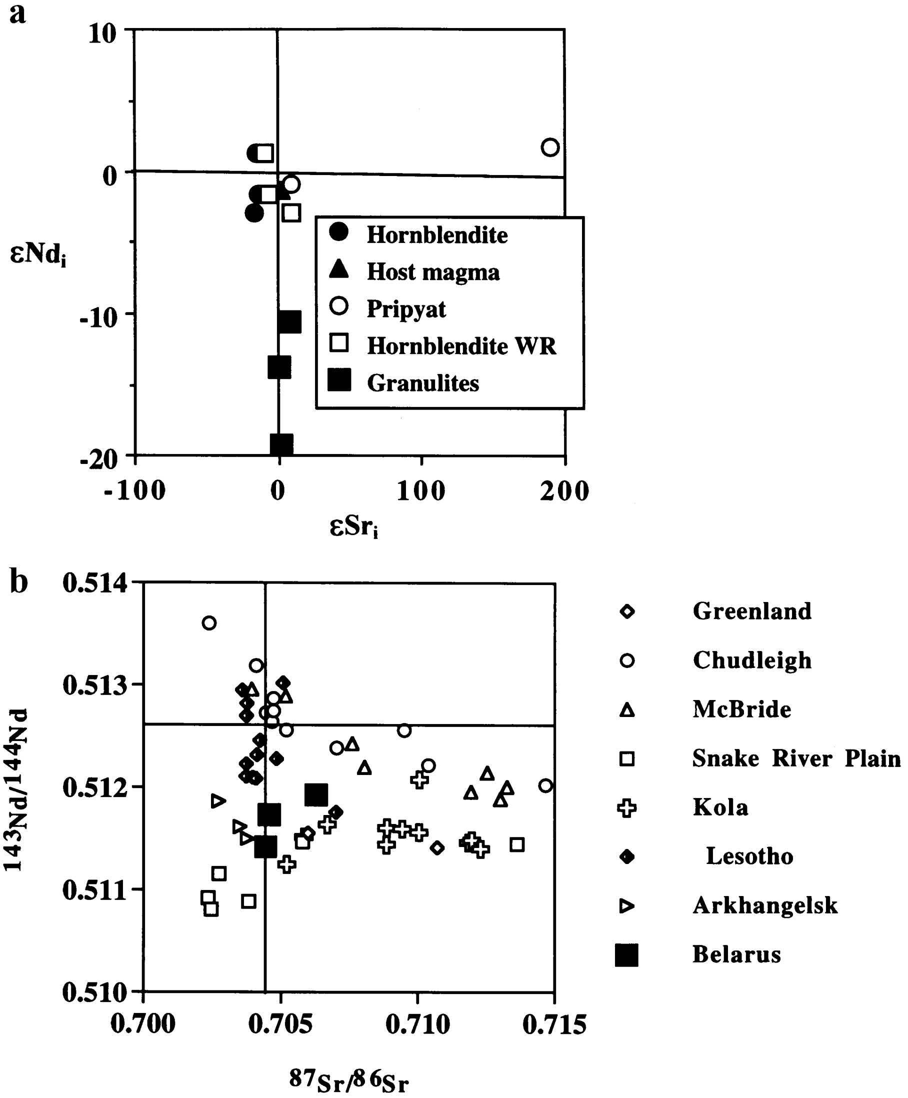

Fig. 8. (a) ϵNd versus ϵSr for granulite ($t = 370$ Ma) and hornblendite ($t = 381$ Ma) xenoliths and host magmatic rocks from Belarus ($t = 370$ Ma) (this paper) and for comparison kimberlite-like samples from Pripyat ($t = 370$ Ma) (Wilson and Lyashkevich, 1996); (b) present-day Sr and Nd isotopic values for Belarus xenoliths compared with selected lower crustal xenoliths from elsewhere in the world. Data from Kempton et al., 1995 (Kola); Huang et al., 1995 (Lesotho); Leeman et al., 1985 (Snake River Plain); Rudnick et al., 1986 (Chudleigh); Rudnick, 1990 (McBride); Garritt, D., pers. comm. (Greenland).

(4.7% at 35 km) and they as well as Kern and Schenk (1985) show that anisotropy decreases with increasing pressure. Rudnick and Jackson (1995) have calculated up to 2.6% anisotropy for similar rocks. Laboratory measured compressional wave velocities at ambient *T*/*P* for lower crustal garnet granulite xenoliths from the Kola peninsula suggest an anisotropy of 5.5% or less (J.V. Korhonen, pers. comm.). No evidence for mineral anisotropy was found in the Belarussian granulite xenoliths and the general proliferation of equant granoblastic fabrics throughout the suite allows us to assume isotropic conditions. The inclusion of the varying amounts of diopside in the xenoliths at the

Table 9
Normative mineralogy of granulite xenoliths from Belarus (mineral abbreviations: Q = quartz, C = corundum, Ab = albite, An = anorthite, Ne = nepheline, Di = diopside, Hy = hypersthene, Ol = olivine, Mt = magnetite, Il = ilmenite, Ap = apatite)

	By9x	By5x	Bel8
Q	0	0	0
C	0	2.27	0
Ab	12.50	29.05	10.23
An	20.86	27.59	35.88
Ne	7.69	0	0
Di	4.69	0	13.15
Hy	0	14.71	6.00
Ol	19.44	3.83	23.18
Mt	1.82	2.10	2.51
Il	0.75	2.33	0.89
Ap	0.12	1.19	0.23

expense of amphibole increases the bulk rock Vp values by a very small amount (0.24–0.05 km/s).

Most high quality data for laboratory measured bulk rock Vp values have measurement uncertainties of up to ±0.5% (Rudnick and Fountain, 1995; Christensen and Mooney, 1995). We consider that the inherent sources of errors in measuring modal mineralogy could exceed this and that the uncertainty in the % anisotropy in the Belarus lower crust may alter our values significantly. The temperature estimates for the Belarus xenoliths are thought to have errors of ±25°, this represents a Vp difference of ±0.01 km/s. It is difficult to estimate the total error in such an investigation; however, we suggest an error of around ±0.5 km/s.

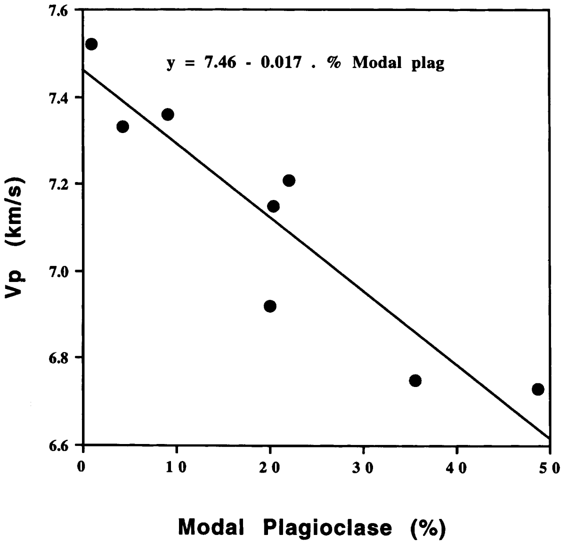

Fig. 9. Plot of estimated modal plagioclase versus calculated whole rock Vp values for garnet granulite and eclogitic granulite xenoliths from Belarus. The line of best fit through the data has also been plotted.

2.4. Age determination

A hornblende separate from the hornblendite xenolith (By6x) from Belarus yielded a plateau age of 381 ± 2 Ma, similar to the stratigraphic age of eruption of the host magmas (~370 Ma). It may represent a crystallisation age for the hornblendite bodies in the crust, below the Ar blocking temperature of 685 ± 53°C in hornblende (Berger and York, 1981) or a post-entrainment cooling age. The small size and high degree of alteration of the granulite xenoliths precluded age determinations.

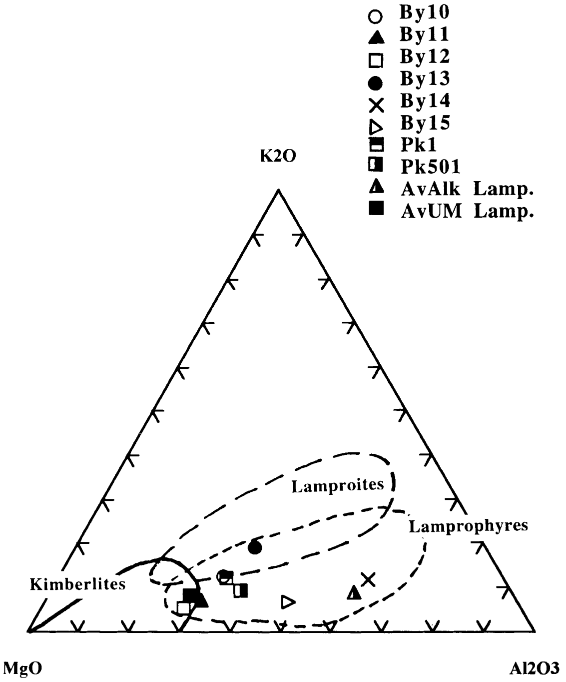

Fig. 10. Whole rock major element discrimination diagram for kimberlite, lamprophyre and lamproite (Bergman, 1987) showing host rocks from Belarus (this paper) and kimberlite- like rocks from Pripyat (Wilson and Lyashkevich, 1996) and average ultramafic lamprophyre (Rock, 1991).

3. Discussion

3.1. Classification and origin of host magmatic rocks

The highly altered nature of the host rocks makes classification difficult, but they are clearly alkaline mafic volcanics with lamprophyre/nephelinite/basanite affinities. The high Ca concentration correlates well with loss on ignition and may be due to alteration. Major element data (Fig. 10) show that the host rocks fall within the lamprophyre field, close to average ultramafic lamprophyre (Rock, 1991), although one overlaps with the lamproite field and another with the kimberlite field. However, trace element and REE data do not suggest lamproite or kimberlite affinities for these samples. Unlike their major element compositions, trace element and REE abundances indicate an alkaline rather than ultramafic affinity for these lamprophyres (Figs. 6a and 7a). Thus, despite their alteration, we suggest that these rocks are alkaline lamprophyres.

Ni and Cr concentrations indicate that both primitive and more evolved rocks are present. Enrichment in the most incompatible elements (Table 2, Fig. 6a) may suggest a small degree partial melt from the mantle or a melt extracted from an enriched mantle source. Trace element plots for rocks described as possible Devonian kimberlites (PK501 and PK1) from the Pripyat segment (Wilson and Lyashkevich, 1996) lie within the range of trace element patterns for the host magmatic rocks (Fig. 6a). The K trough in several Belarus and Pripyat magmatic rocks indicates the presence of a residual K-bearing phase, such as phlogopite, in their source (Rogers et al., 1992). It is unlikely that fractional crystallisation of a K-rich phase would have formed such prominent troughs. LREE enrichment (Fig. 7a), together with the low Ti, Zr and Y concentrations, suggests a low degree partial melt from a garnet-bearing source. The parallel form of the REE patterns and similar La/Nb ratios (0.85–1.1) suggest that they are comagmatic. The very similar $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ initial ratios of By14 compared with sample PK501 from Pripyat (Wilson and Lyashkevich, 1996) fall close to Bulk Earth and strongly suggest derivation from the same mantle source (Fig. 8a).

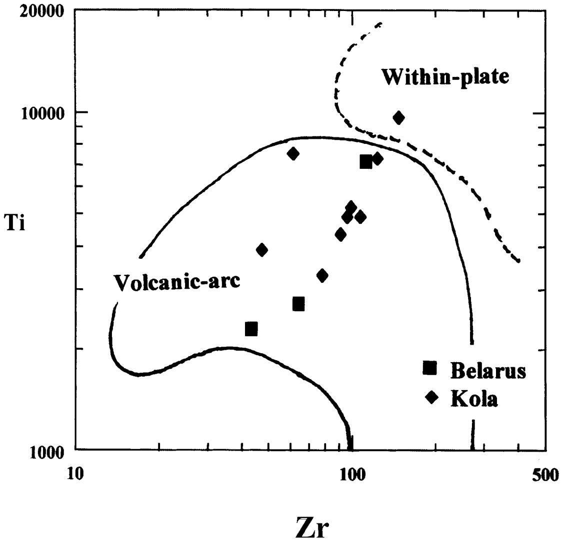

Fig. 11. Ti–Zr trace element discrimination diagram for basalts (Pearce et al., 1981).

3.2. Origin of granulite xenoliths

The granoblastic polygonal textures found in the Belarus garnet granulites and eclogitic granulites show that these rocks have equilibrated in the absence of deviatoric stress, under granulite-facies metamorphic conditions. This is supported by *T*/*P* determinations (700–800°C, 0.9–1.1 GPa). Bel8, the least altered and most primitive sample, shows relatively smooth trace element and REE patterns (Figs. 6b and 7b); its composition may thus be indicative of a melt composition, as opposed to a cumulate. By5x and By9x show more irregular trace element patterns, with prominent Sr peaks (Fig. 6b). These coincide with the large positive Eu anomalies (Fig. 7b) and suggest that these rocks are plagioclase cumulates. However, variations in concentrations and patterns observed for trace element and REE may also be a result of metamorphic mineral segregation. Low transition metal concentrations in By5x and By9x may indicate either a fractionation or accumulation event or metamorphic segregation. The higher Ni concentration in Bel8 more closely resembles a slightly evolved magmatic composition. Trace element discrimination diagram (Fig. 11) suggest volcanic-arc signatures for the granulites; this and the pronounced Nb troughs (Fig. 6b) points to a

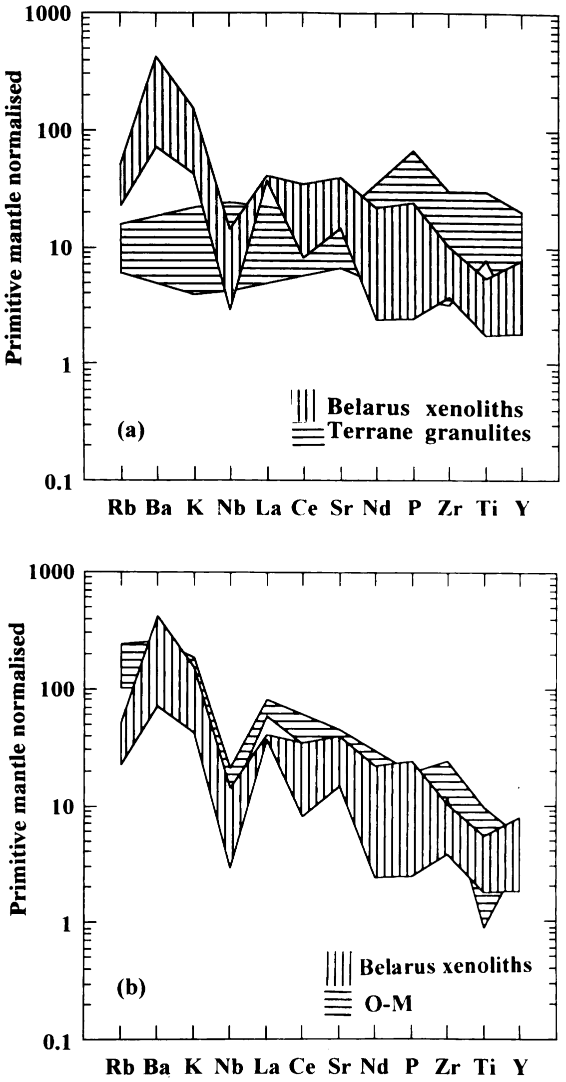

Fig. 12. Trace element plots for granulite xenoliths (this paper) compared to rocks from (a) granulite-facies terranes in Belarus (L. Taran, pers. comm.) and (b) Osnitsk–Mikashevichi Igneous belt (O–M) (L. Taran, pers. comm.).

subduction-related origin. The Belarus granulite xenoliths have quite distinct trace element signatures (Fig. 12a) from the Belarus granulite terrane rocks (L. Taran, pers. comm.). In particular, the terrane granulites have lower abundances in the large ion lithophiles (LIL) Rb, Ba and K, and higher abundances of FeO and TiO_2. However, when trace element patterns of the Belarus granulite xenoliths are compared to rocks from the Osnitsk–Mikashevichi Igneous Belt (L. Taran, pers. comm.), similarities do exist (Fig. 12b). Most notable features are the LIL enrichment and the Nb troughs. These similarities may indicate a related petrogenesis. Isotope systematics for the Belarus granulite xenoliths (Table 5; Fig. 8b) show similar values to other lower crustal xenoliths sampled through Archaean/Proterozoic crust, in particular those from Kola (Kempton et al., 1995) and Greenland (Garritt, pers. comm.). T_{DM} model ages were calculated for granulite xenoliths using present-day $^{143}Nd/^{144}Nd$ of 0.513114 and $^{147}Sm/^{144}Nd$ of 0.222 (Michard et al., 1985). The values obtained were 1.9–2.1 Ga, which are very similar to ages of rocks from the Osnitsk–Mikashevichi Igneous Belt (Shcherbak, 1991). These ages are Proterozoic and in this sense are similar to model ages calculated for mafic lower crustal xenoliths from Kola (Kempton et al., 1993) and as with the Kola xenoliths suggest that their protolith was extracted from a mantle source during Proterozoic times. Certainly these rocks are much older than the host magmatism.

Xenoliths derived from thick Archaean/Proterozoic crust are quite different from those derived from thin Phanerozoic crust (Downes, 1993). Typically, xenoliths derived from Phanerozoic crust show a wider range of compositions including felsic metaigneous and rare metasedimentary lithologies. They tend to be garnet-free, two-pyroxene granulites, which yield relatively low pressures. Higher pressure garnet-bearing granulites are rare from Phanerozoic crust, whereas these are common rock types in the Belarus suite. The abundance of metasomatic amphibole in Belarus xenoliths also contrast strongly with xenoliths from Phanerozoic crust.

The Belarus granulite xenoliths show few petrographic and geochemical similarities to the assumed Archaean-derived Lace kimberlite xenolith suite, which have rather unusual compositions (Dawson and Smith, 1987) or to the Snake River Plain xenolith suite, which are predominately two pyroxene paragneisses with felsic to intermediate compositions and equilibration pressures of only 5 kbars, (Leeman et al., 1985). However, many petrological and mineralogical similarities exist between Belarus and xenoliths from beneath the Archaean terrains of Greenland (Garrit et al., in prep.), Montana (Collerson et al., 1988), West Africa (Toft et al., 1989) and Yakutia (Shatsky et al., 1990). Common mineral assemblages for granulites from these areas are ga + cpx + plg ± qz ± opx ± scap ± amp ± op ± rt in varying concentrations, and commonly garnet granulites grade into eclogitic granulites and eclogites (Toft et al., 1989). The Proterozoic Lesotho xenolith suite is also similar in mineralogy to the Belarus xenoliths (Rogers, 1977; Griffin et al., 1979; Rogers and Hawkesworth, 1982). However, the West African and Montana suites contain a wider range of compositions than the Belarus xenolith suite. Above all, the Belarus xenoliths strongly resemble the granulite xenolith suites from the Kola peninsula, NW Russia, reported by Kempton et al. (1995) and the Arkhanglesk kimberlite province (Markwick and Downes, 2000). The similarities between these lower crustal Archaean/Proterozoic xenolith suites is well exemplified by the following comparison between the Belarus and Kola xenolith suites.

Garnet granulites and eclogitic granulites from Kola are also dominated by granoblastic fabrics, and as with the Belarus xenoliths this suggests equilibration under granulite facies conditions. *T*/*P* determinations (Kempton et al., 1995) are also consistent with granulite facies metamorphic conditions. Kola xenoliths are compositionally very similar to the eclogitic granulites from Belarus. Garnets from Belarus and Kola xenoliths have similar unzoned almandine-rich compositions and fall in the field for garnet granulites (Mottana, 1986). Thicker kelyphitic rims in Kola garnets may be due to differing protolith bulk compositions or to a higher equilibration pressure for the Kola xenoliths. The relatively greater modal abundance of rutile in the Kola xenoliths may also support this. Quartz is absent from Belarus granulite xenoliths but is often present in Kola granulite xenoliths. Pyroxenes in both suites have similar diopsidic compositions. Kola granulite xenoliths show a wider range of feldspar composition than the Belarus granulite xenoliths (Fig. 4), but this may be due to the more limited sampling of the Belarus suite. Both suites have been metasomatised, the Belarus suite more so than the Kola suite, producing amphibole as pseudomorphs after pyroxene in both suites and rarer phlogopite-rich rocks in the Kola suite (Kempton et al., 1995). The amphiboles in both suites are pargasites with similar ranges in $Mg^{\#}$ (0.45–0.85) and are typical of amphiboles found in xenoliths that have equilibrated under granulite facies conditions (Rudnick, 1992). However, the amphiboles in the Belarus granulites rarely exhibit an equilibrium texture, possibly crystallising later under amphibolite facies conditions. The formation of small crystals of amphibole as pseudomorphs after clinopyroxene suggests that amphibolite-facies conditions were not maintained for long before the entrainment of the granulites. Both suites contain rocks with abundant scapolite, indicating a high P_{CO_2} volatile content in the lower crust.

Trace element abundances in the Belarussian granulite xenoliths have similar mantle-normalised patterns to those from Kola (Kempton et al., 1995), showing LILE and LREE enrichment, and Nb troughs, but with a narrower range of concentrations (Fig. 6b). In particular, samples N27-21 and N39 show very similar patterns to By5x, Bel8 and By9x, respectively. The LREE enrichment displayed by the Belarus granulite xenoliths (Fig. 7b) is also seen in Kola granulite samples. However, in general total REE concentrations in Kola samples are greater. Only two Kola granulite xenoliths (N39 and N41) show positive Eu anomalies (Fig. 7b) and may have a similar origin to By5x and By9x.

The very close similarities in petrography, mineral chemistry, whole rock major and trace element concentrations between Belarus and Kola granulite xenoliths strongly suggests a similar magmatic origin and subsequent complex metamorphic history. Compositional, mineralogical, petrological and geophysical data strongly suggest that the Belarus rocks have formed as a mafic underplate, as has been suggested for the Kola xenoliths (Kempton et al., 1995). The Nb troughs in both suites may infer a similar subduction-related origin. However, the smaller Nb troughs exhibited by Kola xenoliths could also suggest an origin related to plume activity (e.g. Sharkov et al., 1997; Cadman et al., 1995). It is possible that the most recent additions of underplate material, whilst crystallising from a mafic melt under granulite facies conditions, would necessarily release aqueous fluids. These would, due to density contrasts, rise into older granulites and possibly under lower temperature/pressure conditions form metasomatic fluids which could be responsible for introducing both amphibole and mica under amphibolite facies

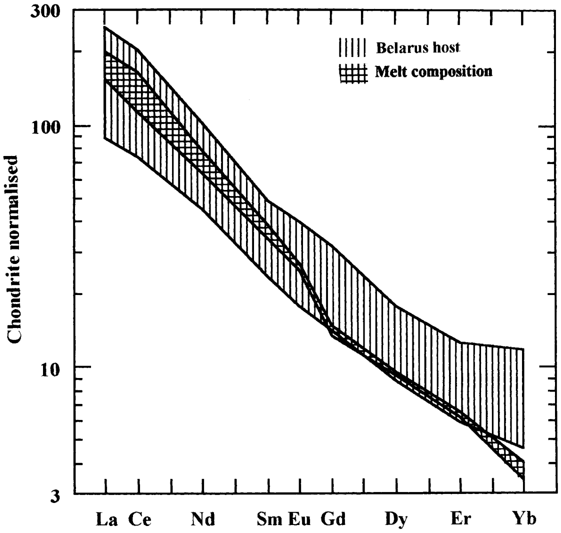

Fig. 13. Modelled REE patterns for a basaltic melt in equilibrium with amphiboles from hornblendite xenoliths from Belarus (this paper) and REE plots for Belarus host magmatic rocks (this paper), re-calculated for volatile-free compositions.

conditions. However, the similarity in compositions between the amphiboles in the granulites and the hornblendites, and the relatively short equilibration times of the amphiboles in the granulites, makes it highly probable that the amphiboles are related to the formation of the hornblendites and also to the host magmas. Whatever the origins of the amphibole in the granulites, it is certain that the metasomatism post-dates granulite facies metamorphism.

3.3. Origin of hornblendite xenoliths

Hornblendite xenoliths from Belarus mostly have decusate/granoblastic fabrics which suggest that they crystallised under a homogeneous stress-field environment. The rarer proto-mylonitic fabrics may have been created just prior to entrainment and transportation in the host magmas. The textures resemble those of hornblendite xenoliths from the Kola suite (A. Beard, pers. comm.).

Major element compositions of the amphiboles in the hornblendite xenoliths are compatible with lower crustal and/or mantle origins (Rudnick, 1992). Trace element patterns (Fig. 6c) for Belarus hornblendite xenoliths are similar to each other and to some Kola hornblendite xenoliths (Beard et al., in prep.). Amphiboles from Belarus hornblendite xenoliths and similar xenoliths from Kola (A. Beard, pers. comm.) are all LREE-enriched (Fig. 7c). REE patterns of Belarus hornblendites closely match some of the Kola hornblendites and the relative concentrations of REE are also similar. Subtle differences between the hornblendites could be the result of a variation in source mineralogy and subsequent chemical evolution. It seems very likely from mineral chemistry, LREE enrichment and the low measured $^{87}Sr/^{86}Sr$ ratios, that the amphibole precipitated from a LREE-enriched mantle-derived mafic magma. Figs. 5 and 8a may suggest a petrogenetic relationship between the hornblendite xenoliths and the host magmas, the xenoliths having crystallised from an earlier flux of lamprophyric melt injected into the crust. Isotope data (Fig. 8a) may support this premise as ϵNd and ϵSr values for the host rock are similar to those of the hornblende separates and whole rock hornblendite xenoliths. Using K_D values from Zack et al. (1996), REE concentrations for a melt existing in equilibrium with the amphiboles from hornblendite xenoliths were calculated (Table 4, Fig. 13). The patterns show greater LREE enrichment for this melt, which may indicate that it originated from a smaller degree of partial melting than did the host magmatic rocks. However, notably the range of LREE concentrations of the model melt lies within that of the host magmatic rocks when modelled using internally consistent data from Zack et al. (1996) but this is not so for the heavier REE. La/Nb ratios for the hornblendite xenoliths are remarkably similar (0.3–0.5). The range of values for the host magmatic rocks are approximately twice that for the hornblendite xenoliths. Modelling precipitation of hornblende from the host lamprophyres using K_D La (amp/liq) = 0.068 and K_D Nd (amp/liq) = 0.20 (Zack et al., 1996), gives values which may support the premise that a petrogenetic link exists between the hornblendite xenoliths and host magmatic rocks. Attempts to model parent melt compositions from the hornblende data using K_D values from Nicholls and Harris (1980) and Arth (1976) are less successful, partly due to the limited K_D data available.

3.4. Interpreting whole rock Vp calculations

Geochemical and petrological analysis shows that many of the Belarus xenoliths are mafic garnet granulites which have equilibrated at temperatures and pressures of at least 800°C and 1.0 GPa respectively. The range of estimated bulk rock Vp values (Table 8) for the Belarus xenoliths is 6.7–7.5 km/s and support the premise that they have equilibrated under lower crustal conditions.

Holbrook et al. (1992) reported that the continental lower crust has *T*/*P* corrected Vp values of 6.4–7.5 km/s and is essentially bimodal in its distribution of bulk rock Vp values with a lower group of 6.7–6.8 km/s and a higher group of 7.0–7.2 km/s. More recently Christensen and Mooney (1995) have shown that compressional wave velocities gradually increase with depth in the crust and that at depths of 40 km a bimodal distribution of Vp values is seen with peaks at 6.8–6.9 and 7.2 km/s. Based on a compilation of laboratory measured compressional wave velocities made on rocks from metamorphic terrains, the lower velocity range of Holbrook et al. (1992) has been suggested to contain lithologies such as anorthosites, mafic granulites and intermediate granulites, whilst the higher range may represent granulite facies metapelites or pyroxenites. Rudnick and Fountain (1995) suggested slightly different mean Vp values for some categories of rock type. Their characterisation of 'type sections' of continental crust also vary. However, they considered mafic granulites, mafic amphibolite-facies rocks, anorthosites and high-grade metapelites to be the most probable rock types within the high Vp range (≥6.9 km/s), whereas the most likely lithologies for the intermediate Vp range (6.5–6.9 km/s) include intermediate granulites and metapelites or mixed mafic and felsic granulites. The very extensive work of Christensen and Mooney (1995) strongly suggests that Proterozoic lower crust has a mafic composition and towards its base grades into garnet granulite, this being the dominant rock type directly above the Mohorovicic discontinuity. It is expected that the calculated bulk rock Vp values will be higher than any corresponding laboratory measured values. This is due to the presence of mineral alteration and in particular grain boundary alteration, which has been shown to give lower (5–12%) measured Vp values when compared to calculated values (Jackson et al., 1990; Rudnick and Jackson, 1995).

A relationship between compressional wave velocities and whole rock chemistry has been previously reported by Birch (1961). More recently a relationship of Vp to mineralogy has been demonstrated by Rudnick and Jackson (1995). The higher bulk rock Vp values (7.3–7.5 km/s) derived from plagioclase-poor, garnet + clinopyroxene-rich eclogitic granulites and the observed linear relationship between modal plagioclase and whole rock Vp values (Fig. 9) may give a relatively quick and accurate method of predicting in situ seismic velocities for similar lower crustal rocks. Anisotropic fabrics would necessarily increase the bulk rock Vp values. The ranges of compressional wave velocities obtained from this method lie mostly within the intermediate to higher ranges given by Rudnick and Fountain (1995). They are also consistent within the lower crustal Vp range of Holbrook et al. (1992) and support the premise of Christensen and Mooney (1995) that the deepest Proterozoic lower crust is composed of mafic garnet granulites. For the hornblendite xenoliths, which have calculated compressional wave velocities of approximately 6.75 km/s at 800°C/1.0 GPa and isotropic fabrics, it is unlikely that compressional wave velocities alone could distinguish them from other in situ lower crustal lithologies, and that Vp/Vs ratios would be required (Ward and Warner, 1991; Musacchio et al., 1997).

Deep seismic profiles of the Belarussian crust (Giese et al., 1997) give lower crustal compressional wave velocities of 6.6–7.2 km/s. Seismic profiles across the Pripyat Trough section of the Osnitsk–Mikashevichi Igneous Belt (EUROBRIDGE'95, EUROBRIDGE Seismic working group, 2001) show, for a depth consistent with our calculations, a Vp range of 6.8–7.4 km/s. The mantle is defined at 50 km depth by a Vp jump from 7.2 to 8.3 km/s (Giese et al., 1997) and 7.4 to 8.0 km/s (EUROBRIDGE'95, EUROBRIDGE Seismic working group, 2001). Thus, the calculated whole rock Vp values for the Belarus xenoliths are consistent with the deep seismic measurements of the lower crust in this area. Durrheim and Mooney (1994) showed that at the base of Proterozoic crust exists a high velocity layer (7.0–7.6 km/s), which has been attributed to extensive basaltic underplating. In the lower crust of East

Poland, a 5–6 km layer of high velocity (~7.2 km/s) crust of unknown composition overlies the mantle (Grad and Ryka, 1996). This relatively thin high velocity layer may contain rocks of similar composition and metamorphic facies to those from the Belarus xenolith suite. Seismic studies of the Archaean Kola Peninsula through to the Proterozoic East European Platform (Beloussov and Pavlenkova, 1991) and of the Karelian section of the Baltic Shield (Hjelt et al., 1996) identify a reflective high velocity ($Vp > 7$ km/s) layer in the lower crust. On the basis of our calculations both garnet granulite and eclogitic granulite lithologies may be likely rock types in this layer. As a consequence of the variation of modal plagioclase in these rock types, the acoustic impedance contrasts may be large enough to account for some of the observed lower crustal reflectivity.

4. Conclusions

From the petrological, geochemical and geophysical evidence discussed above, a number of conclusions can be drawn for the xenoliths and host magmatic rocks of the Zhlobin area of SE Belarus.

(1) The host rocks, which are thought to be alkaline lamprophyres, have entrained a limited range of rocks derived from the lower crust including the predominant mafic garnet granulite xenoliths, rarer mafic eclogitic granulite xenoliths and hornblendite xenoliths. Intermediate and felsic metaigneous and metasedimentary rocks have not been found. LREE-enriched patterns, K troughs, low concentrations of Ti, Zr, Y and low $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ initial ratios suggests that the host lamprophyres were derived from a small degree partial melt of a K-bearing garnetiferous mantle source.

(2) A Hornblendite xenolith yielded an Ar–Ar age of 381 ± 2 Ma. Trace element and isotope evidence indicates that they are probably related to the host magmas and may represent cumulates from an earlier flux of alkaline lamprophyre magma.

(3) *T*/*P* data and calculated whole rock Vp values determined from the granulite xenoliths are consistent with the expected equilibration conditions and measured seismic wave velocities in the lower crust of the Sarmatian segment of the East European Craton.

(4) The granulites have probably formed as a subduction-related mafic underplate and crystallised under granulite facies metamorphic conditions. Introduction of metasomatic pargasitic hornblende occurred in most samples shortly before entrainment as xenoliths.

(5) This study has highlighted the similarity of petrography and geochemistry that exists between mafic lower crustal xenoliths from Belarus and Kola (Kempton et al., 1995). Similar comparisons with other xenolith suites which have been sampled through Archaean/Proterozoic crust, are beginning to suggest that the mineralogy and geochemistry of the lower crust in these areas as sampled via xenoliths is to some extent related to thickness and age of the crust.

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