The north-eastern Polish anorthosite massifs: petrological, geochemical and isotopic evidence for a crustal derivation

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ABSTRACT

Deeply buried 1.5 Ga Polish anorthosites, accessible only by bore holes, reveal diagnostic features of some massif-type anorthosites (polybarism, jotunitic parent magma), diapirically emplaced in the mid crust together with the rapakivi granites of the EW-trending Mazury complex, intruded along a major crustal discontinuity. Geochemical modelling and isotope data corroborate recent experimental work on the basaltic system in dry conditions: the source rock of the parental magma is a gabbronorite, necessarily lying in the lower crust. Since no Archaean crust is known in the region, high initial $^{188}Os/^{187}Os$ ratios for sulphide-oxide isochrons and negative ε_{Nd} values are best accounted for by melting a $\sim$ 2.0 Ga mafic crust.

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Introduction

Diapiric emplacement is a now widely accepted process for most massive anorthosites (see References in Ashwal, 1993). Evidence of polybaric crystallization is demonstrated by the occurrence of high alumina orthopyroxene megacrysts (HAOM), shown experimentally to have equilibrated at lower crustal pressure (11–13 kbar) (Longhi et al., 1993), and brought to their final level of emplacement at depths corresponding to 3–5 kbar by diapirism (Berg, 1977; Barnichon et al., 1999).

Experimental data have also shown that a range of compositions from high alumina basalt (Fram and Longhi, 1992) to jotunite (Fe–Ti–P-rich hypersthene monzodiorite) (Vander Auwera et al., 1998b) can be possible parents to the noritic series of anorthosites and related rocks of the anorthosite, mangerite, charnockite and (rapakivi) granite (AMCG) suite. Recent data in the basaltic system in dry conditions at pressures up to 13 kbar (Longhi et al., 1999) have constrained the composition of the source rocks of these parent magmas: a plagioclase + orthopyroxene + clinopyroxene-bearing source is required, and thus mafic granulites or gabbronorites from layered complexes are possible source

rocks. It has also been recognized in the Nain plutonic series (Emslie et al., 1994; Hamilton et al., 1998), the Laramie Complex (Scoates and Chamberlain, 1997) or the Grenville Province (Higgins and van Breemen, 1992) that the anorthosite massifs coincide with structural weaknesses in the lithosphere which have channelled and facilitated the diapiric emplacement (Scoates and Chamberlain, 1997; Duchesne et al., 1999).

The anorthosite massifs of NE Poland are part of the Mazury Complex which was emplaced along an E–W-trending lineament. The anorthosites show HAOM, characteristic of a polybaric evolution, and melts of the jotunite kindreds. These petrological diagnostic features point to a crustal origin, a conclusion corroborated by Sm–Nd and Re–Os isotopic data.

Geological setting and petrology

Two anorthosite massifs, the Suwalki and the Sejny massifs (Fig. 1), belong to the eastern part of the Mazury Complex (N. Poland), which is made up of a variety of (rapakivi-like) granites, emplaced around 1.5 Ga (Claesson et al., 2001) in the crystalline basement of the East European Craton along an E–W fault zone (Kubicki and Ryka, 1982). They are overlain by a 580–1200-m-thick Phanerozoic sedimentary cover and have been explored by geophysics and drilling (Ryka and Podemski, 1998).

The 250-km^2 Suwalki massif shows in cross-section a domical structure

whose centre is made up of anorthosites, with some norites and Fe-Ti layered deposits, capped by gabbronorites and diorites. An Re–Os isochron age of 1559 $\pm$ 37 Ma has been obtained on sulphide and oxide minerals from the deposits (Stein et al., 1998; Morgan et al., 2000). Andesine plagioclase is the main rock-forming mineral, followed by orthopyroxene, clinopyroxene, Ti-magnetite and ilmenite. Apatite, spinel and titanite are accessories (Juskowiak, 1998) as well as sulphides (Wiszniewska, 1998). The smaller Sejny massif also comprises gabbroidic rocks, syenites and granitoids.

New data on the geochemistry of the anorthosites and related rocks

Petrological and geochemical characteristics of new samples coming from drill cores of the two massifs are reported here. Major and trace elements on whole rocks and on plagioclases are presented in Tables 1 and 2. An electron microprobe was also used for mineral analysis, particularly of zoned grains.

Anorthosites and norites have compositions typical of cumulate rocks. Anorthosites are essentially made up of plagioclase (An45–55) and accordingly their REE distributions (Fig. 2a) show strong positive anomalies (Eu/Eu* up to 8.5) with large [La/Yb]$_N$ ratios (30–60). The contents in elements incompatible with plagioclase, such as Zr, Nb and transition elements, are low and mostly controlled by the

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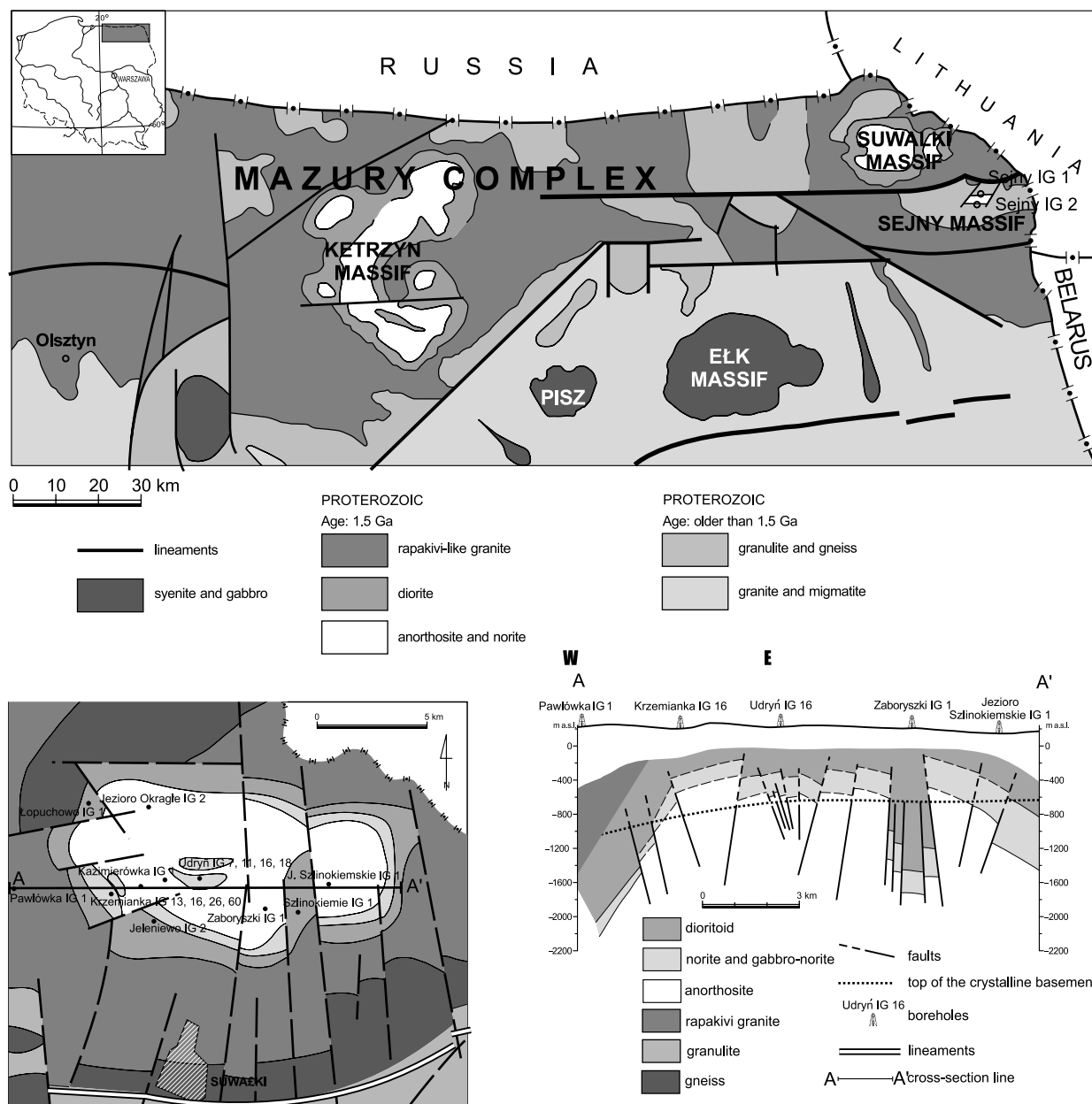
Fig. 1 The Mazury Complex and the anorthosite massifs. Top: schematic geological map of the Mazury Complex essentially made up of rapakivi and rapakivi-like granites, anorthosites and related rocks. The complex extends westwards some 100 km to the Baltic sea, and eastwards *c*. 30 km in Lithuania (Veisiejai Complex) and Belarus (Kapciamiestis massif) (Skridlaite *et al.* unpubl. obs.). Lower left: geological map of the Suwalki and Sejny anorthosites in NE Poland. Same legend as in lower right figure. The bore holes from which the samples studied here have been collected are indicated. Lower right: cross-section through the Suwalki anorthosite dome (after Juskowiak, 1998). Emplacement of bore holes indicated. Note the thickness of the sedimentary cover (from 600 to 900 m).

mafic mineral content. REE span a relatively large interval of concentrations, which can be explained partly by fractional crystallization and partly by varying amounts of apatite crystallized from trapped liquid (sample P14 has the highest REE and P_2O_5 contents). Norite REE distributions (Fig. 2b) display smaller positive anomalies (Eu/Eu* $\approx$ 2) and [La/Yb]$_N$ $\approx$ 8 in agreement with a larger pyroxene content of the norite.

Besides these data, several other petrological and geochemical characteristics permit comparison with other anorthosite occurrences and particularly with the Rogaland anorthosites, Norway (Duchesne, 1987; Duchesne *et al.*, 1999). Several features point to similar petrogenetic processes and mechanisms of emplacement which suggest that both occurrences are derived from a crustal source.

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Table 1 Major (%) and trace (p.p.m) element composition in anorthosites, norites, and jotunites

	Anorth P-12	Anorth P-13	Anorth P-18	Anorth P-14	Anorth P-21	Norite P-10	Norite P-11	Norite P-08	Norite P-07	Jotunite JW9718	Jotunite JW974
SiO_2	52.99	53.91	53.88	52.10	54.36	50.61	50.20	52.16	51.92	46.74	42.53
TiO_2	0.23	0.11	0.20	0.12	0.26	0.36	0.27	0.33	0.39	2.26	3.45
Al_2O_3	26.91	27.46	27.33	28.40	26.64	24.94	23.48	19.81	18.72	16.33	15.19
Fe_2O_3	0.34	0.16	0.29	0.25	0.36	0.79	0.93	1.54	1.78	3.14	3.62
FeO	1.30	0.62	1.11	0.95	1.39	3.02	3.57	5.90	6.83	12.04	13.67
MnO	0.02	0.01	0.01	0.02	0.02	0.06	0.10	0.15	0.18	0.26	0.26
MgO	0.58	0.16	0.22	0.52	0.28	1.97	3.79	5.57	6.48	4.46	5.40
CaO	10.56	10.99	10.85	11.99	10.49	11.33	10.66	8.42	7.86	9.31	9.81
Na_2O	4.10	3.97	3.81	3.39	4.08	2.98	2.65	2.69	2.44	2.50	2.76
K_2O	0.61	0.67	0.73	0.54	0.86	0.50	0.48	0.67	0.69	0.52	0.50
P_2O_5	0.01	0.01	0.01	0.13	0.03	0.04	0.01	0.01	0.04	0.66	1.23
LOI	1.37	1.34	0.51	1.03	0.66	2.48	2.39	2.58	1.47	0.34	0.18
Total	99.16	99.49	99.07	99.54	99.58	99.41	98.93	100.49	99.56	99.89	100.12
U	0.02	0.01	0.03	0.19	0.12	0.05	0.03	0.14	0.38	0.09	0.13
Th	0.04	0.11	0.11	0.34	0.96	0.67	0.47	0.77	1.73	0.49	1.06
Zr	3.8	1.9	4.7	2.2	29	27	13	20	40	96	79
Hf	0.4	0.3	0.4	0.4	1.0	1.0	0.6	0.8	1.5		1.7
Zr/Hf	9	5	12	6	28	27	21	24	27		47
Nb	0.29	0.53	0.63	1.26	2.61	2.38	1.27	2.14	3.80	10.20	10.71
Ta	0.09	0.06	0.07	0.17	0.14	0.09	0.09	0.14	0.28	0.45	0.52
Nb/Ta	3	9	9	8	19	26	14	15	14	23	21
Rb	5	4	4	6	5	7	6	7	13	7	7
Sr	797	866	846	706	800	692	624	332	477	684	608
Ba	286	301	379	214	515	242	178	168	282	510	319
K/Rb	1015	1401	1496	762	1532	586	697	750	427	662	570
K/Ba	18	19	16	21	14	17	22	33	20	8	13
V	57	17	33	23	36	80	58	62	92	485	522
Cr	71	80	160	71	131	220	154	239	367		
Zn	28	2	13	3	2	43	33	71	112	198	244
Co	7	2	4	4	5	16	20	24	42	67	66
Cu	15	4	10	17	12	38	14	52	53		
Ga	20	20	21	25	22	20	17	12	20	23	28
Pb	3	7	4	14	7	4	2	5	7	6	8
Y	1	1	2	7	3	6	5	4	8	25	43
La	4.3	4.6	5.8	20.7	15.0	7.3	5.2	5.2	10.0	34	48
Ce	6.8	8.0	10.3	37.1	25.2	13.8	10.5	9.9	19.6	72	112
Pr	0.7	0.9	1.1	4.0	2.7	1.6	1.2	1.1	2.1	10.1	16.8
Nd	2.7	3.1	3.7	15.8	9.9	6.3	4.5	4.6	8.6	42	71
Sm	0.5	0.3	0.7	2.6	1.6	1.1	0.8	0.7	1.5	7.6	14.7
Eu	0.8	0.8	1.0	1.8	1.5	0.8	0.6	0.5	0.8	2.4	3.7
Gd	0.32	0.25	0.56	2.17	1.10	1.14	0.74	0.66	1.32	6.61	12.74
Tb	0.04			0.30	0.14	0.21	0.11	0.12	0.20	0.85	1.66
Dy	0.23	0.16	0.31	1.45	0.60	0.94	0.64	0.59	1.25	4.60	8.18
Ho	0.05	0.04	0.07	0.31	0.14	0.21	0.16	0.15	0.31	0.89	1.61
Er	0.10	0.08	0.14	0.63	0.26	0.52	0.39	0.37	0.79	2.44	3.77
Tm				0.08			0.07	0.06	0.13	0.32	0.51
Yb	0.08	0.07	0.11	0.46	0.15	0.50	0.40	0.37	0.79	2.30	2.82
Lu				0.04	0.02	0.08	0.07		0.10	0.33	0.37
La/Yb n	35	45	35	29	63	9	8	9	8	10	11
Eu/Eu*	6.32	8.59	4.73	2.32	3.50	2.06	2.27	2.06	1.76	1.04	0.83
Mg/(Mg + Fe)(atom)	0.39	0.27	0.22	0.44	0.23	0.49	0.61	0.58	0.58	0.35	0.36
FeOt/MgO + FeOt	0.73	0.83	0.86	0.69	0.86	0.65	0.54	0.57	0.57	0.77	0.76
Peral (mol%)										0.76	0.66
Agp (mol%)										0.29	0.33
Sat T zircon °C										650	599
Sat T apatite °C										823	848
CIPW normative composition (wt%)											
An	60	62	61	67	60	68	69	64	65	60	54
Feldspar	91	92	90	91	91	81	76	67	63	56	54
Mafic	4	2	3	4	4	12	18	27	31	42	45

Sample location: P-12: Krzemianka, P-13: Kazimierowka; P-18, P-21: Udryn; P-14: Lopuchowo; P-10, P-11, P-8, P-7, JW9718: Sejny; JW974: Suwalki.

Analytical methods: Major elements by XRF on Li-borate discs; trace elements by ICP-MS (Vander Auwera et al., 1998a) or by XRF on pressed powders.

FeO calculated on the basis of $Fe_2O_3 = 0.19Fe_2O_3t$ (following Kress and Carmichael (1991) at 3 kbar and 1160 °C).

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Table 2 Major and trace element composition of plagioclase

	U16	Krz60	JezO2	IP-G	IP-B
in %					
CaO	10.99	10.80	11.31	10.80	10.32
K_2O	0.54	0.51	0.56	0.57	0.56
in p.p.m.					
Fe	3287	5455	7203	4755	5874
Ti	247	326	583	342	456
Ba	268	265	302	285	261
Sr	884	840	797	885	880
Rb	4.9	4.8	5.2	3.6	3.6
Pb	2.2	2.2	3.3	6.2	2.3
La	7.2	6.2	6.1	2.9	2.6
Ce	9.5	7.1	8.0	4.4	4.0
Pr				0.57	0.48
Nd	4.4	2.8	3.1	1.7	1.5
Sm	0.65	0.33	0.44	0.33	0.33
Eu	0.63	0.69	0.76	0.58	0.56
Gd	0.62	0.25	0.40	0.30	0.33
Tb	0.09	0.03	0.05		0.06
Dy	0.42	0.16	0.25	0.22	0.19
Ho	0.08	0.03	0.05	0.03	
Er	0.24	0.10	0.14	0.09	0.10
Yb	0.19	0.08	0.13	0.08	0.10
Lu				0.01	0.01
An (wt%)	55	54	56	54	51

U16, Krz60, JezO2 were separated by physical methods from rocks of Udryn, Krzemianka and Jezioro Okragle drill cores, respectively. IP-G, IP-B are iridescent plagioclase megacrysts from Suwalki.

High-alumina orthopyroxene megacrysts. Megacrysts of orthopyroxene, up to 10 cm long, are found in coarse-grained anorthosites (An62–66). They show evidence of deformation by kinking and locally by granulation (polygonization). Inversely zoned plagioclase grains are common in the deformed parts of the megacryst, and lamellae of plagioclase can locally be observed in the (100) planes of the host pyroxene (Fig. 3). The host orthopyroxene is En68 and has a relatively low Al_2O_3 content of 2–3%. The plagioclase granules vary from An76 in the centre to An94 at the rim (inverse zoning), in contrast with the plagioclase of the anorthosite which is less calcic (An66). Similar textures are observed in most massif-type anorthosites (see Ashwal, 1993), and particularly in the Rogaland anorthosites (Maquil and Duchesne, 1984). Following Emslie (1975), the plagioclase lamellae and granules are produced by an exsolution process during decreasing temperature and pressure, and are locally strain-activated. The original Al_2O_3 content of the orthopyroxene megacrysts is strongly pressure dependent (Longhi *et al.*, 1993). In the present case, due to extensive external granule exsolution, no reintegration of the exsolved products in the host mineral to reconstruct the primary composition of the megacrysts is possible. Therefore, the depth of formation of the megacrysts is still to be ascertained but its value is certainly higher than the depth of final emplacement.

Iridescent plagioclase megacrysts. Fragments up to 10 cm in size of granulated iridescent plagioclase with blue, green and purple colours are commonly found embedded in an anorthosite matrix with the characteristic violet colour. Evidence of deformation of the megacrysts are curved twin planes and polygonized (granulated) margins. The plagioclase crystals are crowded with minute needles of rutile, Fe–Ti oxides, and antiperthite exsolutions. They also locally contain association of very small pyroxene and oxide grains that possibly result from the crystallization of melt inclusions. Compositions of iridescent plagioclase are given in Table 2. They show high Sr contents (*c.* 900 p.p.m.) and An content *c.* 52–55%. Such compositions are typical of massif-type anorthosite plagioclase megacrysts and

contrast with plagioclase from layered intrusions, which, as pointed out by Emslie (1985), have Sr contents less than 500 p.p.m. Similar megacrysts are found associated with high-alumina orthopyroxene megacrysts in Rogaland massive anorthosites (Duchesne, 1987). Their REE contents are compared to the present megacrysts in Fig. 2(d). The overall contents fall in the same range of values, but with higher positive Eu anomalies and La/Yb ratios than in the Suwalki anorthosites. Both differences can be explained by mixing small amounts (2–3%) of jotunitic melt (see below) with the Rogaland plagioclase with the lowest REE content. Due to the low REE contents of the plagioclase, small inclusions of relatively REE-rich material, such as the melt inclusions observed here, strongly influence the bulk composition of the megacrysts.

Jotunitic chilled melts. Fine-grained Fe–Ti–P-rich jotunites have been sampled and analysed (Table 1). Petrographically, they show apatite crystals with high aspect ratios, which are disseminated together with small rounded Fe–Ti oxide grains in all silicate minerals, a texture characteristic of chilled rocks (Demaiffe and Hertogen, 1981; Wiebe, 1984; Duchesne and Hertogen, 1988; Vander Auwera *et al.*, 1998). Two samples slightly different in compositions are presented here: JW97-18 is made up of an unzoned plagioclase (An63), low-Ca pyroxene (mg# = 59), high-Ca pyroxene (mg# = 74), apatite, Fe–Ti oxides and accessory biotite; JW97-4 contains the same minerals, with plagioclase An54 and slightly more evolved pyroxenes (mg# = 56 and 70, respectively). The chemical compositions of these two jotunitic melts are reported in Table 1. Both rocks are characteristically high in Fe (up to 19.8% Fe_2O_3t), Ti (up to 3.45% TiO_2) and P (up to 1.2% P_2O_5). The REE distributions (Fig. 2d) are moderately differentiated ($[La/Yb]_N \approx 10$) and do not show significant Eu anomalies. These geochemical characteristics are similar to Rogaland jotunites as shown in Fig. 2(d). On variation diagrams (Fig. 4) displaying the jotunite liquid line of descent (Vander Auwera *et al.*, 1998b), JW97-18 plots close to the primitive jotunites defined in Rogaland; JW97-4, although somewhat less
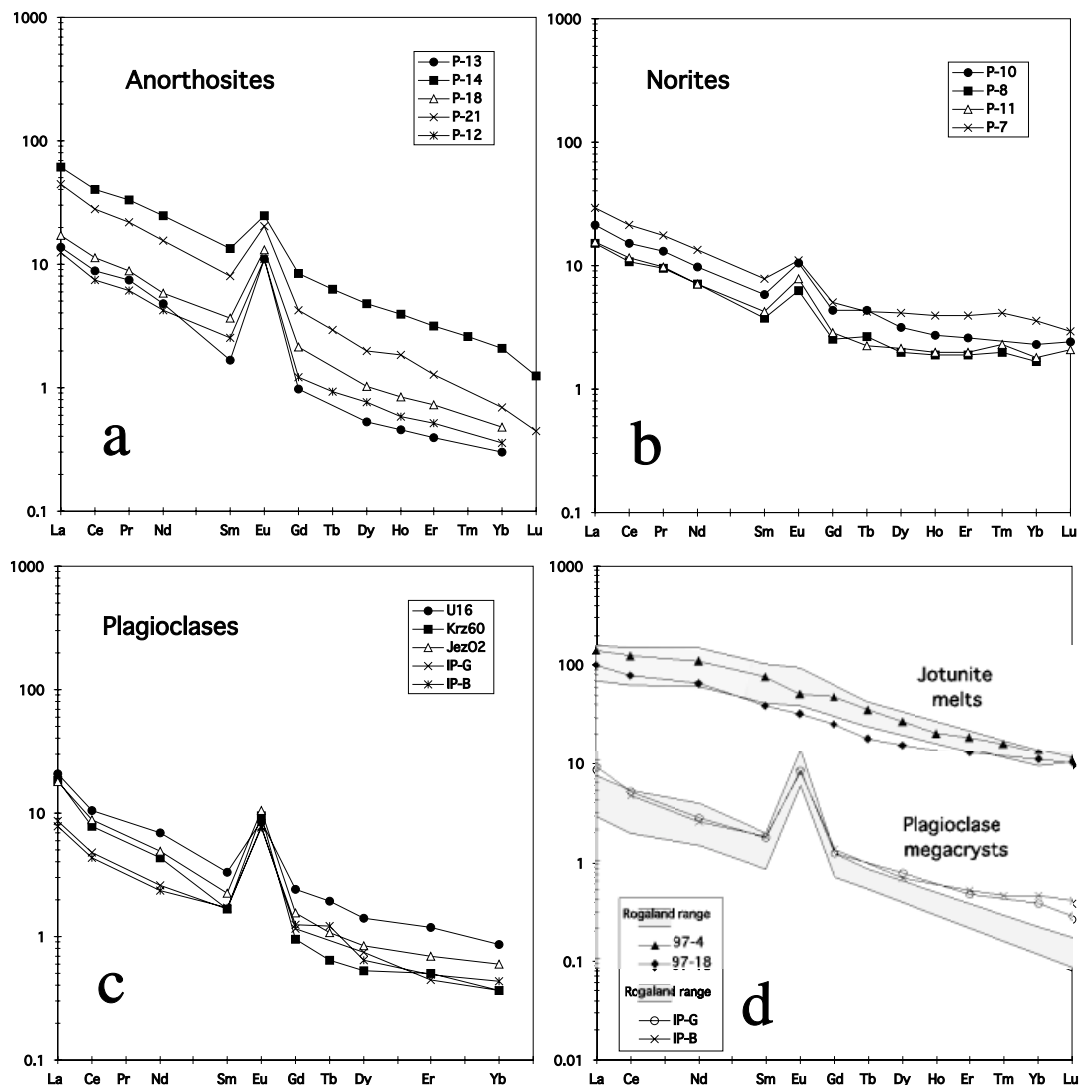
Fig. 2 Chondrite-normalized REE distributions (data from Tables 1 and 2) in (a) anorthosites, (b) norites, (c) iridescent plagioclase megacrysts and plagioclase separated from norites, and (d) REE in iridescent plagioclase megacrysts and jotunites compared to composition of similar materials from Rogaland.

evolved, has many common features with high-alumina basalts from Laramie and Maloin (Kolker *et al.*, 1991).

Jotunites as parental magma

The first steps of crystallization of the two fine-grained jotunitic samples, JW97-4 and JW97-18, have been modelled using the MELTS algorithm (Ghiorso and Sack, 1995). The fO_2 and pressure were set at FMQ and 3 kbar, respectively, as these are plausible conditions for anorthosite crystallization (e.g. Ashwal, 1993; Vander Auwera and Longhi, 1994). Results are shown in Table 3. In the

first two steps, the calculated compositions of liquidus plagioclase and near liquidus low-Ca pyroxene (in JW97-4: first An68, then An60 and En65, and in JW97-18: first An70, then An59 and En58) are in the range of the observed compositions. In both modelling, whitlockite crystallizes instead of apatite because anhydrous crystallization has been imposed. Whitlockite crystallization is somewhat delayed (1160 °C) in JW97-18, as is the case in primitive jotunites, but it appears at the liquidus (1200 °C) of JW97-4, suggesting that some apatite accumulated in these samples. A Ti-rich phase also crystallizes in both

sequences and a low-Ca pyroxene has a relatively long interval of crystallization, in agreement with the widespread occurrence of orthopyroxene in the anorthosite. The good agreement between observed and calculated phase compositions as well as the similar order of appearance of crystallizing phases corroborate the hypothesis that these samples represent parent magmas.

The REE contents of liquids in equilibrium with plagioclase have been calculated using appropriate partition coefficients (Vander Auwera *et al.*, 1998b) and REE data acquired on mineral separates (Table 2). Except

Fig. 3 Highly strained (kinked and polygonized) zone of a high-alumina orthopyroxene megacryst (HAOM). Grains of inversely zoned plagioclase (250–1000 μm) associated with Fe–Ti oxide minerals are aligned along kink planes grossly perpendicular to the 100 plane of the HAOM. In the lower part of the field, plagioclase exsolutions, locally connected to plagioclase granules, are still visible. Crossed polars. Width of view: 3.7 mm.

Table 3 Modelling the first crystallization steps of jotunitic liquids through the MELTS algorithm

T (°C)	Crystallizing phases
Sample JW97.4	An68Ab32 – Whitlockite
1200	
1170	An60Ab39Or1 – Whitlockite – Pigeonite (Wo9En65Fs26)
1130	An46Ab51Or3 – Whitlockite – Pigeonite (Wo14En55Fs31) – Augite (Wo38En43Fs19) – Spinel (Mgt33Sp22Uvsp44)
Sample JW97.18	
1210	An70Ab29Or1
1160	An59Ab40Or1 – Whitlockite – Pigeonite (Wo9En58Fs33)
1110	An50Ab47Or3 – Whitlockite – Augite (Wo37En35Fs28) – Spinel (Mgt39Sp14Uvsp47)

for Eu, which is not considered because it is also controlled by the fO_2, the calculated liquids have REE patterns in the range of those of the jotunitic samples (Fig. 5). This observation gives further support to a jotunitic parent magma.

Jotunites as crustal derivates

It has been shown by Longhi *et al.* (1999) on the basis of experimental data in the basaltic system from 1 atm to 13 kbar in dry conditions that jotunite melts can evolve by fractional crystallization at pressure < 4 kbar, with or without contamination by acidic material, to produce the AMCG suite. The petrological and

geochemical characteristics of the NE Polish anorthosites also belong to the AMCG suite. This conclusion imposes severe constraints on the origin of the parent melts. In the pressure range at which high-alumina orthopyroxene megacrysts are in equilibrium with plagioclase, that is at the depth of the deep-seated magma, these melts stand on thermal highs of the phase diagram (Longhi *et al.*, 1999). They thus cannot be derived by fractionation of any basaltic or intermediate liquids. Nor can they be produced by mixing of mantle-derived products with acidic crustal material. The only possible source is a gabbronorite (i.e. a two-pyroxene and plagioclase-bearing rock). Since such a lithology cannot be found in the mantle (except close to Moho), it must be considered as crustal.

Sm–Nd and Re–Os evidence for a crustal origin

Sm–Nd isotope data for anorthosite, norite and jotunite (including JW97-4 and 97-18) from the NE Polish anorthosites are shown in Table 4 and in an ε_{Nd}-time evolution diagram in Fig. 6. The anorthosite samples have low Sm and Nd concentrations of 0.3–0.7 and 2–5 p.p.m., respectively, and low $^{147}Sm/^{144}Nd$ ratios of 0.08–0.10, reflecting the composition of the plagioclase, while corresponding values for jotunite and norite are 7–22 p.p.m., 40–126 p.p.m and 0.11–0.12. Depleted mantle model ages (DePaolo, 1983) are between 1.9 and 2.3 Ga. ε_{Nd} values calculated at 1.5 Ga range from −2 to −5.

The Sm–Nd results clearly demonstrate that the anorthosite massifs are not composed entirely of materials directly derived from a 1.5-Ga mantle source. The source is either older continental crust, or a mixture between mantle and older crustal materials. In the latter case, this crustal material must be older than the obtained model ages, in all likelihood Archaean. However, there are no indications of the existence of Archaean crust in this region. On the contrary, a regional survey of the Precambrian crystalline basement of the East European (Claesson *et al.*, 2001), and a reconnaissance Nd isotope investigation of various rock types from all major lithotectonic units in the Precambrian crystalline basement of north-eastern Poland (Claesson and Ryka, 1999) indicate that this entire crustal region is Palaeoproterozoic with Nd model ages of between 1.9 and 2.3 Ga. Our preferred interpretation of our Nd model ages is therefore that the NE Polish anorthosites are derived from Palaeoproterozoic continental crust, and that the Nd model ages approximate the age of this crust.

The Nd model ages reported here form a distinct group, but do nevertheless scatter significantly. Some scatter must be ascribed to the Sm/Nd fractionation caused by the melting which formed the Suwalki magmas, and following crystalliza-
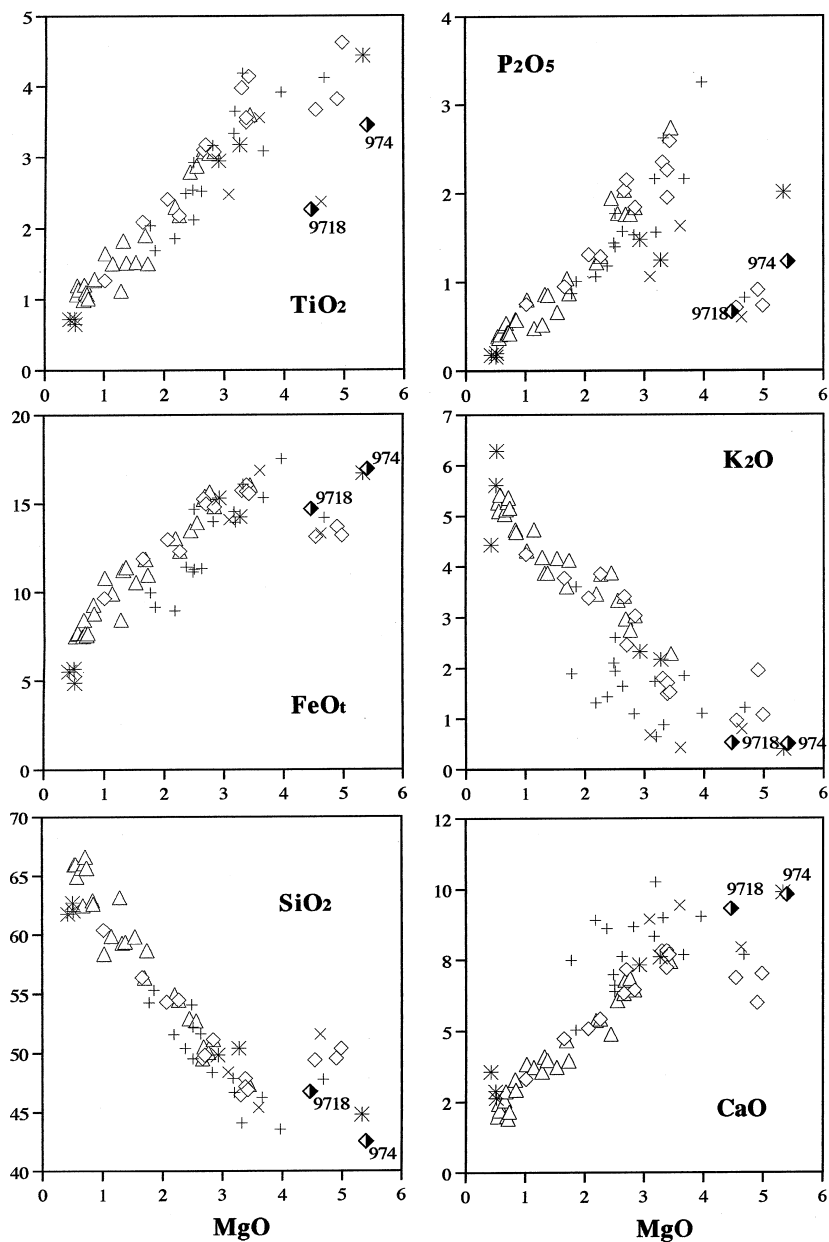
Fig. 4 Variation diagrams comparing the Polish chilled rocks to the liquid line of descent of the Rogaland jotunites and other occurrences. Legend: half-filled diamond: Suwalki–Sejny chilled rocks; +: Grenville occurrences (Owens *et al.*, 1993); ×: Nain (Wiebe, 1979; Emslie *et al.*, 1994); *: Maloin (Kolker *et al.*, 1991; Mitchell *et al.*, 1996); open diamond: primitive and evolved jotunite from Rogaland (Vander Auwera *et al.*, 1998b); open triangle: Tellnes dyke series (Wilmart *et al.*, 1989; Vander Auwera *et al.*, 1998b).

tion and separation of cumulate. In particular, the Nd model ages for the anorthosites, which have Sm/Nd ratios significantly lower than the Sm/Nd ratios in the parent materials, should be interpreted as minimum ages for this material. However, if this were the sole cause

for the scatter, the Nd isotopic compositions should converge at the age of intrusion, 1.5 Ga. This is not the case, as demonstrated in Fig. 6. If the scatter was caused by mixing of 1.5 Ga depleted mantle material with a crustal component having a homogeneous ε_{Nd} (1.5 Ga) composi-

tion of −5 to −6, this would depend on assumed concentrations requiring that the samples with the least negative ε_{Nd} value contain more than 50% crustal material. This is difficult to reconcile with the geochemical characteristics of the jotunite melts. We therefore suggest that the scatter is mainly caused by isotopic inhomogeneities in a crustal source material.

The magmatic sulphides and oxides from the Suwalki massif provided the first Re–Os evidence used to suggest a possible crustal origin for anorthosite (Stein *et al.*, 1998; Morgan *et al.*, 2000). The proposed crustal origin for anorthosite was subsequently corroborated in a Re–Os study of the Rogaland anorthosite complex in southern Norway (Schiellerup *et al.*, 2000). Subsequent modelling, however, leaves open the possibility that sulphide and silicate may be of different origin (Hannah and Stein, 2002). At Suwalki, high $^{187}Os/^{188}Os$ initial ratios (e.g. 1.16 ± 0.06) for several deposits require a crustal source for the Os in sulphides. Given the Proterozoic age for the Suwalki massif, these $^{187}Os/^{188}Os$ ratios clearly indicate involvement of crustal rocks in the genesis of the sulphide-bearing anorthosite massif. Morgan *et al.* (2000) suggest that a $^{187}Re/^{188}Os$ of ∼ 140 in the Suwalki source accommodates the ∼ 2.0 Ga, Nd model ages, noting that ratios of ∼ 140 may be found in a crust with a mafic component. These Re–Os data may be used to support a crustal origin for Suwalki.

Conclusions

The NE Polish anorthosites, which are part of the Mazury Complex, show typical petrological and geochemical characteristics of massif-type anorthosites, particularly occurrence of HAOM, of iridescent plagioclase megacrysts and of jotunite chilled rocks. Geochemical modelling points to a parental magma of jotunite composition, and phase diagrams constrain the gabbronoritic nature of the source rock in the lower crust. Negative ε_{Nd} values between −2 and −5 at the age of intrusion (1.5 Ga) and high $^{187}Os/^{188}Os$ initial ratios (1.16) are consistent with a ∼ 2.0 Ga sulphide-bearing mafic crustal source.
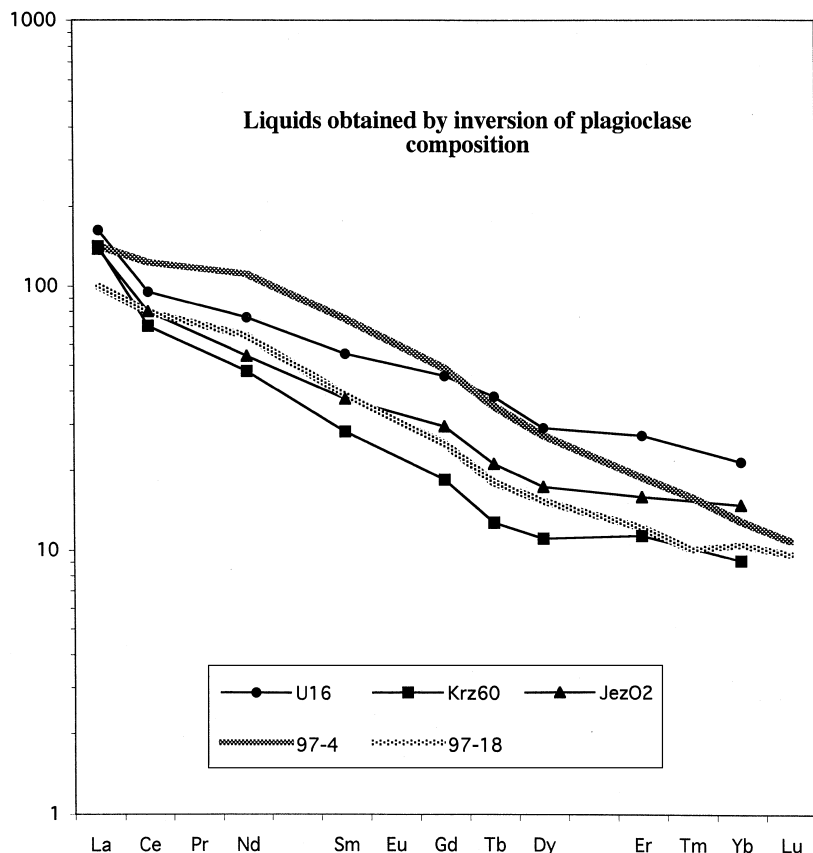
Fig. 5 REE distributions calculated by inversion of the plagioclase composition separated from norites and compared to jotunite compositions (JW97-4 and 97-18). Note that Eu is not considered because of the unknown control of the fO_2 on the distribution coefficient.

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Table 4 Nd isotope data from NE Polish anorthosites and related rocks

Sample	Rock type	Sm (p.p.m)	Nd (p.p.m)	^{147}Sm/^{144}Nd	^{143}Nd/^{144}Nd	±	T_{DM}* (Ga)	ε_{Nd}† (1.5 Ga)
4.2 Krzemianka IG 13	Anorthosite	0.39	2.50	0.0949	0.511506	0.000009	1.97	− 2.5
4.3 Krzemianka IG 26	Anorthosite	0.73	4.45	0.0992	0.511431	0.000008	2.15	− 4.8
4.4 Jeleniewo IG 2	Anorthosite	0.31	2.10	0.0884	0.511411	0.000008	1.99	− 3.1
4.9 Szlinokiemie IG 1	Anorthosite	0.36	2.41	0.0897	0.511448	0.000008	1.96	− 2.7
W15 Jez. Okr. IG 2	Anorthosite	0.82	4.91	0.1006	0.511458	0.000006	2.14	− 4.6
W19 Udryn IG7	Anorthosite	0.81	4.96	0.0987	0.511461	0.000016	2.06	− 3.5
W23 Udryn IG16	Anorthosite	0.33	2.26	0.0896	0.511443	0.000020	1.97	− 2.7
W26 Udryn IG18	Anorthosite	0.31	2.20	0.0862	0.511404	0.000017	1.96	− 2.8
4.8 Zaboryszki IG 1	Norite	22.9	126	0.1099	0.511512	0.000004	2.25	− 5.3
JW 97–4	Jotunite	14.0	69.4	0.1217	0.511672	0.000003	2.28	− 4.4
JW 97–4 repeat		14.1	69.9	0.1215	0.511675	0.000010	2.27	− 4.3
JW 97–18	Jotunite	7.50	39.9	0.1134	0.511628	0.000005	2.15	− 3.7
JW 97–18 repeat		7.50	40.2	0.1127	0.511646	0.000010	2.11	− 3.2
4.1 Udryn IG 11	Jotunite	14.6	79.7	0.1108	0.511537	0.000003	2.24	− 5.0
BCR-1		6.59	28.75	0.1385	0.512633	0.000005		
BCR-1 repeat		6.64	28.94	0.1386	0.512639	0.000010		

The samples were analysed on a Finnigan MAT 261 TIMS using total spiking with a mixed ^{149}Sm/^{150}Nd spike. The analyses were made over an extended period of time, during which both the chemical and the mass spectrochemical procedures were modified several times. The mass spectrometer was operated in both static and multidynamic mode. A conservative estimate of the reproducibility (external precision) for most samples, based on repeated runs of the La Jolla Nd standard, is ± 40 p.p.m. (0.000020), while the reproducibility for 'JW97-4 repeat' and 'JW 97-18 repeat' was 25 p.p.m. The internal precision was always better than, or similar to, the reproducibility and typically ± 10–20 p.p.m. (0.000005–0.000010). The Nd isotope ratios were normalized to ^{146}Nd/^{144}Nd = 0.7219. The repeat analyses of samples JW97-4 and JW 97-18 were made on separate aliquots of rock powder. The two analyses of the rock standard BCR-I, which are shown as examples from a larger number of BCR-1 runs, were made together with the analyses of JW 97-4 and JW 97-18.

*The depleted mantle model age after DePaolo (1983).

†ε_{Nd} values vere calculated assuming present day chondritic values ^{147}Sm/^{144}Nd = 0. 1967 and ^{143}Nd/^{144}Nd = 0.512638.
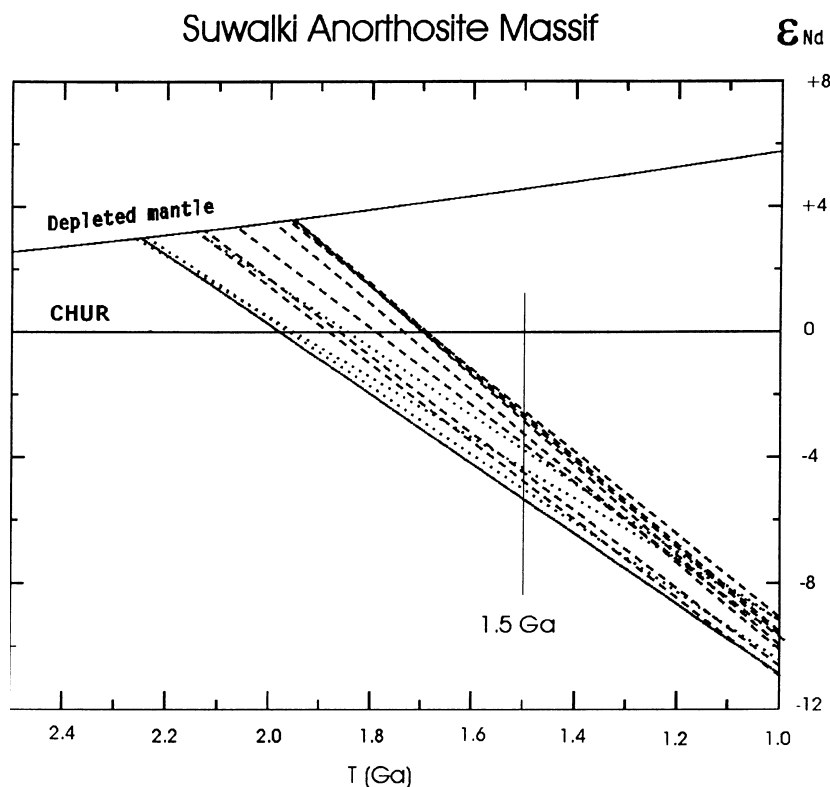
Fig. 6 ε_{Nd} vs. age for samples of NE Polish anorthosite massifs. The indicated age of 1.5 Ga approximates the intrusive age of the Suwalki massif. Dotted line: jotunite; dashed line: anorthosite; densely dotted line: norite (data in Table 4).

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