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Genetic relationships between metabasalts and related gabbroic rocks: an example from the Fore-Sudetic Block, SW Poland

Kazimierz Dziedzic[1] **& Helena Dziedzic**[2]

[1] *Instytut Nauk Geologicznych, Uniwersytet Wrocławski, 50-205 Wrocław, Cybulskiego 30*
[2] *Polska Akademia Nauk, Instytut Nauk Geologicznych, Zakład Geologii Sudetów, 50-449 Wrocław, Podwale 75*

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Abstract The metabasalts occurring within the gneisses of the eastern part of the Góry Sowie gneiss block, have either enriched or depleted LREE patterns. However, these two types of metabasic rock are indistinguishable in terms of their major elements and many trace elements. Their intimate association indicates that their parent magmas erupted nearly contemporaneously. They originated through the decompressional, two-stage incremental melting of a mantle diapir source. The LRRE enriched variety of the basalts was formed from a spinel/garnet peridotite melt mixture which was followed by spinel peridotite melts. From the latter melts, cumulate gabbros crystallized and extracted portions of these melts provided the LRRE depleted variety of basalts. This strongly suggests that the metabasalts are compatible in age with the gabbros.

Both of the metabasalts varieties developed mainly by AFC processes involving mantle source melts and lower continental crust components. From a comparison of these metabasalts with those in adjacent areas, it is possible to draw the conclusion that volcanic activity in the whole region had the same time-span.

The metabasalts were metamorphosed to LP hornblende granulites, synchronously with the surrounding amphibolite facies gneisses of the margins of the Góry Sowie Block, which were metamorphosed to LP-HT cordierite gneisses. They originated due to the transition from high-grade amphibolite to granulite facies conditions associated with a near-isothermal decompression, during the time of the late Variscan (Carboniferous).

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INTRODUCTION

In the north-eastern part of the Bohemian Massif, the amphibolite and greenschist facies metamorphic complexes consist of abundant metabasalts and, less common plutonic basic and ultrabasic assemblages. A similar situation can be found in the Fore-Sudetic Block, where metabasalts, gabbros and ultrabasites occur in close vicinity (Fig. 1), concentrating in the submeridionally oriented fracture zone near the eastern margin of the Góry Sowie gneiss block and in the Niemcza Dislocation Zone. This may be a key area for the investigation of the genetic relationships between the metabasalt and gabbro suites.

This paper considers the metabasalts intercalated with the gneisses at the eastern margin of the Góry Sowie Block near Bielawa and the gabbros of the Braszowice Massif, both located in the south-eastern part of the Fore-Sudetic Block. The origin of the metabasalt diversity, the genetic and temporal relationships of the metabasalt and gabbroic suites, and the tectonic regime which existed at the time of the magmatic activity in this area were studied using microscopic and microprobe analyses and trace element modelling. Additionally, the probable linkage of these metabasalts with those in adjacent areas, and a possible geotectonic setting and timing for the mafic magmatism in the whole region are discussed.

GEOLOGICAL OUTLINE

The **Góry Sowie** complex of the Fore-Sudetic Block mainly contains the garnet-sillimanite paragneisses, which lie along the eastern margin of the complex, overprinted by LP-HT cordierite-sillimanite-andalusite assemblages.

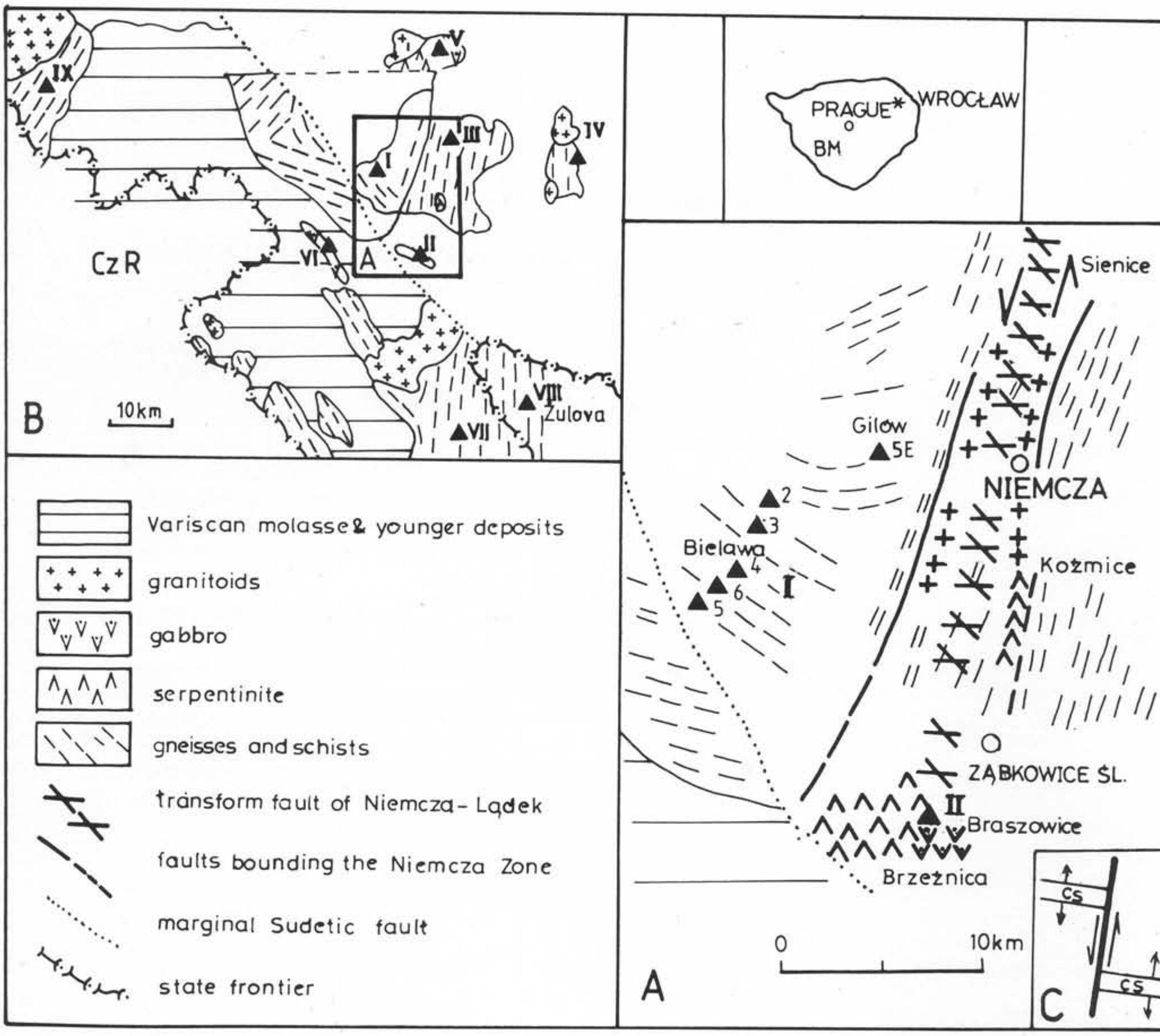

Fig. 1. Schematic geological maps and localization of the magmatic complexes (filled triangles and numbers). A – the metabasalt (I) and gabbro (II) rocks discussed in the text; their position in the Bohemian Massif (BM) is indicated by a star. B – the metabasalts used for comparison refer to: III – Niemcza Zone; IV – Strzelin crystalline complex; V-VI – Ślęża and Nowa Ruda Massifs; VII – Śnieżnik Dome; VIII – Stare Město Crystalline Unit (SMCU); IX – Rudawy Janowickie Range. C – a plan view of the right-lateral displacement of the crustal segments (cs), containing the Palaeozoic depressions, along the transform fault zone. Abbreviation: CzR = Czech Republic.

Mafic and felsic granulites, garnet- or garnet-free amphibolites, hornblende gneisses and minor marbles, gabbro-amphibolites, serpentinites, alaskites, and subordinate fine-grained dioritoid and tonalites, belonging to the Variscan granitoids suite of the adjacent Niemcza Zone, occur as intercalations in the paragneisses (Dziedzic, 1985, 1994; Żelaźniewicz, 1995).

The **Niemcza Dislocation Zone** strikes meridionally and is bordered by subvertical discontinuities on both its west and east sides (Fig. 1a). The Niemcza Zone rock complexes are represented by three main lithologies: a metasedimentary rock series, an ultrabasic/basic complex and granitoids.

At the base, the rock series consists of quartzites and graphitic metasiliceous rocks, overlain by fine-grained metapsammites and, in the NW part of the Zone, by discovered Carboniferous (uppermost Visean/lowermost Namurian) metasiltstones with small lenses of calc-silicate metamudstones (Dziedzic & Górecka, 1965; Dziedzic, 1979, 1987). In late Variscan times, the sedimentary rocks were metamorphosed under LP-HT cordierite-andalusite facies conditions (Dziedzic, 1963, 1980), together with the cordierite paragneisses of the eastern margin of the Góry Sowie Block (Dziedzic, 1985). The calc-silicate rocks weakly recrystallized with LP-LT assemblages (Dziedzic, 1987).

The ultrabasic/basic complex of the Niemcza Zone is represented by fault-bounded slivers of serpentinite, layered and unlayered gabbro, and amphibolites. These rocks form the serpentinite-gabbroic Braszowice-Brzeźnica

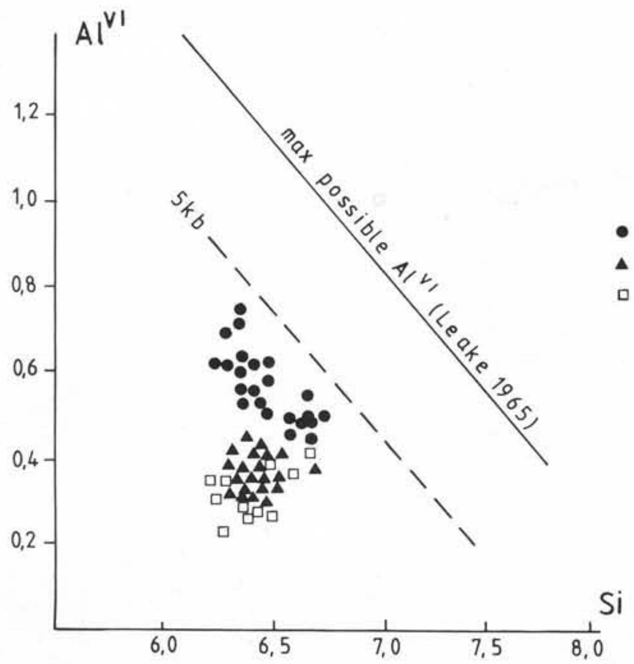

Fig. 2. Relation between the Al^{VI} and Si of the low pressure hornblendes (<5 kb) and middle pressure hornblendes (>5 kb) after Raase (1974). The diagonal solid line indicates the maximum possible Al^{VI} contents (after Leake, 1965). 1 – hbl in sample 4A, 2 – hbl in sample 3A and 3C, 3 – hbl from the Quairading district granulites of Western Australia after Davidson (1971).

massif in the south-west, and the Szklary serpentinite massif in the east (Fig. 1a). The latter continues northward into the amphibolite (metabasalt) bodies of Koźmice–Sienice (Dziedzic, 1979; Białek *et al.*, 1994). The gabbro of Braszowice disappears to the west beneath serpentinites along a NE–SW striking boundary. There is evidence from boreholes in the Braszowice area, that over a large area the gabbro is overlain by serpentinites, containing numerous gabbroic bodies. In the profile of one borehole, three intercalations of recrystallized amphibolites were found.

The trace element chemistry of the gabbro suggests normal MORB. The whole rock isochron Sm-Nd age was obtained for the gabbro of Braszowice and Ślęża (Fig. 1b) of 357±12 Ma (Pin *et al.*, 1988). Oliver *et al.* (1993) obtained a U-Pb zircon age of 420±20 Ma for the Ślęża gabbro. The U-Pb age of single zircons from rodingized plagiogranite within the serpentinite of the Ślęża complex is 400+4/-3 Ma (Żelaźniewicz *et al.*, 1998).

Late-tectonic hornblende-biotite granitoids occur in the Niemcza Zone and the adjacent gneisses. They also underline the eastern fault zone bordering the Niemcza Zone to the east (Fig.1a). The granitoids are relatively small sheet-like bodies, subparallel to the trends of surrounding rocks (Dziedzic, 1963). On the basis of the chemical composition of the hornblende, the crystallization depth of the granitoids was calculated. It varies from about 17 km to 10 km respectively for the granitoids of Kośmin and Koźmice. The U-Pb zircon age of the granitoids is about 338 Ma (Oliver *et al.*, 1993; Kroner & Hegner, 1998). The Rb-Sr data for individual bodies of the granitoids do not allow a well-constrained age to be calculated (Lorenc, 1998; Kennan *et al.*, 1999).

The discontinuities bordering the Niemcza Zone on both its west and east sides, dip steeply toward its centre. These shear zones are marked by controversially interpreted mylonitisation (Dziedzic 1979; Mazur & Puziewicz, 1995; Żelaźniewicz, 1995). Shearing is evident in the Braszowice gabbro, which also displays regionally developed foliations (S_1 and S_2) and a subhorizontal lineation (Dziedzic, 1988).

The occurrence of large ultrabasite bodies near the western and on the eastern boundaries of the Niemcza Zone demonstrates the geotectonic importance of this unit in the structural evolution of the Sudetes. It was probably already active as a right-lateral offset transform fault (Fig.1a, 1c) at the time of the generation of the ultrabasic/basic complex (Dziedzic, 1979, 1985).

THE PETROGENETIC CHARACTERISTICS OF THE METABASALTS AND GABBROS

The metabasalts of Bielawa. In the neighbourhood of Bielawa and Gilów (Fig.1a), in a few places within the sillimanite-biotite gneisses, there are sill-like bodies of LP hornblende granulites and garnet amphibolites (Dziedzic, H., 1996). All these rocks have tholeiitic affinity, and the hornblende granulites also show N-MORB and E-MORB geochemical signatures (Dziedzic, 1995a). The surrounding gneisses were metamorphosed at pressures of 3–5 kb and temperatures from 600 to 630°C (Żelaźniewicz 1995).

The metabasalts are isotropic and undeformed, fine-grained rocks with porphyroblasts of garnet. They can be divided into pyroxene-bearing hornblende granulites and pyroxene-free garnet amphibolites. The former (4A, 3A and 3C) are quartz-poor (~1%), whereas the latter (6A and B) contain about 5% quartz. In both, ilmenite and apatite are present as accessory phases. Orthopyroxene appears as a minor phase only in granulites 4A and 3C. Amphibole is mainly represented by pargasite hornblende in the hornblende granulites and mainly by tschermakite hornblende in the garnet amphibolites (Dziedzic, H., 1996, Fig. 5).

The content of the metamorphic minerals of the hornblende granulites (Table 1) is compatible with the whole rock chemistry (Table 2), in particular as regards the Mg : Fe ratio of the garnet, orthopyroxene, clinopyroxene and also of the hornblende. The low Al contents in both the pyroxenes and amphiboles point to their crystallization at pressures less than 5 kb, presumably 3–4 kb (Fig. 2).

The occurrence of orthopyroxene+plagioclase symplectites or clinopyroxene+orthopyroxene+plagioclase coronas, which are characteristic for the hornblende

Table 1

Microprobe analysis of garnets, orthopyroxenes, clinopyroxenes and hornblendes in the metabasalts of Bielawa

	garnet			orthopyroxene			clinopyroxene		hornblende		
	4A	3C	6A	4A	3C		4A	3C	4A	3C	6A
SiO_2	39.793	38.234	37.337	53.160	50.593	SiO_2	53.013	51.092	43.158	42.557	42.601
Al_2O_3	22.220	20.848	21.066	0.754	0.345	Al_2O_3	0.712	1.476	14.037	11.760	10.601
TiO_2	0.068	0.182	0.047	0.034	0.067	TiO_2	0.121	0.253	2.224	2.040	2.212
Cr_2O_3	0.015	0.090	0.000	0.025	0.039	Cr_2O_3	0.050	0.000	0.298	0.087	0.057
FeO	18.539	25.389	29.904	21.819	31.974	FeO	7.230	11.615	10.818	16.970	19.372
MnO	0.491	0.798	1.095	0.494	0.264	MnO	0.216	0.155	0.140	0.065	0.205
MgO	11.538	4.473	4.439	23.057	15.705	MgO	14.518	11.952	12.884	10.063	8.857
CaO	6.988	9.775	6.443	0.569	0.766	CaO	24.086	22.272	12.377	11.728	11.760
Na_2O	0.084	0.226	0.030	0.000	0.000	Na_2O	0.212	0.266	1.942	1.792	1.211
K_2O	0.000	0.029	0.007	0.000	0.000	K_2O	0.034	0.009	0.848	0.511	1.026
Total	99.738	99.943	100.169	99.913	99.754	Total	100.192	99.091	98.728	97.573	98.019
crystallochemical formulae based on											
	12 oxygens			6 oxygens			6 oxygens		23 oxygens		
Si	2.990	3.000	2.961	1.976	1.982	Si	1.969	1.953	6.256	6.347	6.411
Al^{IV}	0.010	0.000	0.039	0.024	0.016	Al^{IV}	0.031	0.047	1.744	1.653	1.589
	Z3.000	3.000	3.000	**T**2.000	1.998		**T**2.000	2.000	**T**8.000	8.000	8.000
Al^{VI}	1.958	1.928	1.930	0.009	0.000	Al^{VI}	0.012	0.020	0.654	0.412	0.312
Ti	0.004	0.011	0.003	0.001	0.002	Ti	0.004	0.007	0.242	0.229	0.250
Cr	0.001	0.006	0.000	0.001	0.002	Fe^{3+}	0.021	0.033	0.040	0.410	0.408
Fe^{3+}	0.043	0.044	0.103	0.012	0.011	Fe^{2+}	0.162	0.338	1.271	1.707	2.030
	Y2.006	1.989	2.036			Mn	0.003	0.005	0.017	0.008	0.025
						Mg	0.813	0.681	2.783	2.237	1.896
Fe^{2+}	1.122	1.616	1.867	0.666	1.037				**C**5.007	5.003	5.012
Mn	0.031	0.053	0.073	0.016	0.009						
Mg	1.293	0.523	0.525	1.278	0.917	Ca	0.978	0.912	1.922	1.874	1.896
Ca	0.563	0.822	0.525	0.023	0.032	Na	0.014	0.020	0.078	0.123	0.104
Na	0.012	0.034	0.004	0.000	0.000				**B**2.000	2.000	2.000
K	0.000	0.003	0.000	0.000	0.000						
	X3.021	3.051	3.008	**M**2.006	2.009	Na			0.468	0.395	0.249
						K	0.002	0.000	0.157	0.097	0.198
							M2.008	2.016	**A**0.625	0.492	0.459

Fe^{3+} calculated by considering charge balance constraints.

granulites of Bielawa, indicates the (M_2) near-isothermal decompression stage (Dziedzic, H., 1996). The hornblende+ plagioclase symplectites possibly formed during the M_3 stage, as a result of isobaric cooling and retrograde metamorphism (Zhao *et al.*, 2000). During the older MP high amphibolite facies metamorphism, about 380 Ma (Van Breemen *et al.*, 1988; Bröcker *et al.*, 1998; Timmermann *et al.*, 2000), the pressure could have been higher, perhaps 8–9 kb, as was recently evaluated for typical high amphi- bolite-facies assemblages of Bielawa rocks by Winchester *et al.* (1998).

The temperatures of the hornblende granulites calculated using the grt-cpx-hbl geothermometer (Nasir & Al-jarayesh, 1992) and using grt-opx (Harley, 1984) are highest for sample 4A: 829 to 737°C, and 848–686°C , respectively, whereas sample 3C gave 730–715°C and 776–694°C. The temperatures of the garnet amphibolites calculated using grt-hbl (Graham & Powell, 1984) are from 704 to 644°C in sample 6A and from 670 to 569°C in sample B.

The above calculated temperatures and low pressure stages indicate the transition from older high-grade amphibolite to late Variscan hornblende granulite facies, associated with near-isothermal decompression. A similar conclusion can be drawn from the large distribution of Ti in hornblende (Fig. 3).

Evidence for a Variscan amphibolite to LP granulite facies transition for MORB-derived amphibolites was also found in SW Spain (Castro *et al.*, 1996).

The gabbro of Braszowice. The plutonic member of the Braszowice–Brzeźnica ultrabasic/basic massif is coarse-grained, or, as in the boreholes, medium and fine-

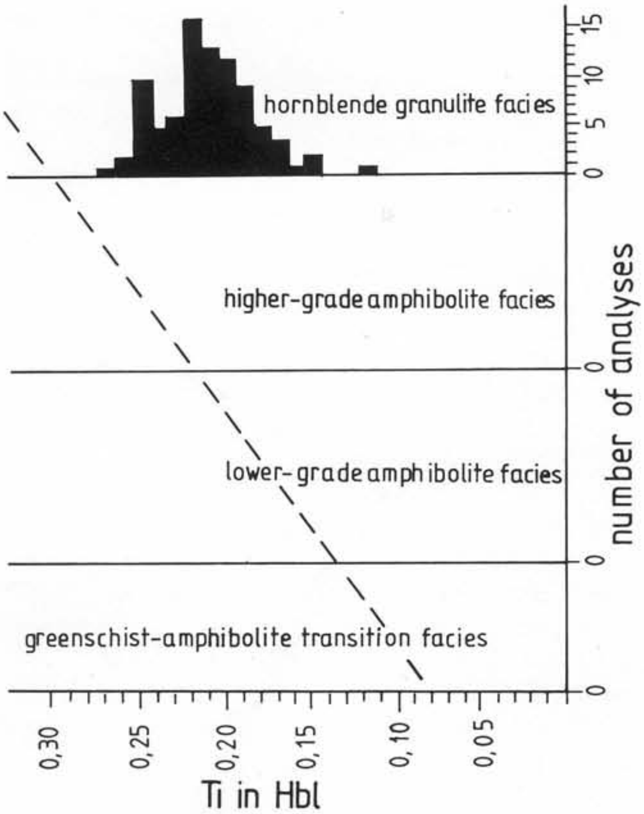

Fig. 3. Histogram of the Ti contents of the hornblende (on basis of 23 oxygens) after Raase (1974) from samples 4A, 3A, 3C, 6A and B. The distances between the metamorphic facies on the abscissa are aranged so that the maximum Ti contents give a straight line, drawn as a broken line in the diagram.

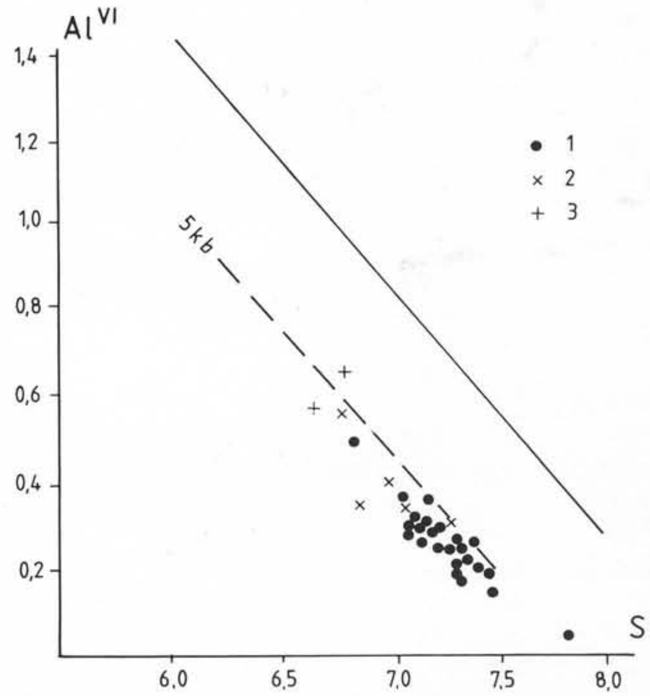

Fig. 4. The relationship between the Al^{VI} and Si of the low pressure hornblendes (< 5 kb) and middle pressure hornblendes (> 5 kb) after Raase (1974). The diagonal solid line indicates the maximum possible Al^{VI} contents (after Leake, 1965). 1 – green hornblende, 2 – green hornblende in fissures of pyroxene, 3 – brown hornblende.

grained. These rocks have an olivine-free hornblende-rich gabbroic assemblage. On the Pl-Hbl-Px diagram their representative points lie in the pyroxene-hornblende and hornblende gabbro fields (Dziedzic, H., 1989). Microprobe analyses of the plagioclase, clinopyroxene and hornblende of the gabbro are listed in Dziedzic (1995b, Tab. 1–3).

On the basis of the method of Lindsley and Anderson (Lindsley 1983; Lindsley & Anderson, 1983) and on the basis of the method of Kretz (1982) temperatures between 1120 and 500°C, and 1097 and 515°C, respectively, have been estimated as the crystallization temperatures of the Braszowice gabbro clinopyroxenes. There is a positive dependence of temperature on pressure: the Al_2O_3 content diminished synchronically with the temperature reduction; therefore there was also a reduction in pressure (Dziedzic, 1995b).

The hornblende is rich in magnesium. The Al^{IV} versus $(Na+K)_A$ diagram (Leake, 1978) shows a trend controlled by crystallization temperatures from edenite hornblende to common and actinolite hornblende (Dziedzic, 1995b). Their crystallization began under pressures of about 5 kb (Fig. 4). The main crystallization stage took place at 5–4 kb, limiting the lower pressure boundary for the Braszowice gabbro at around 4 kb.

The crystallization of the hornblende depended on an increase in H_2O activity which simultaneously condi-

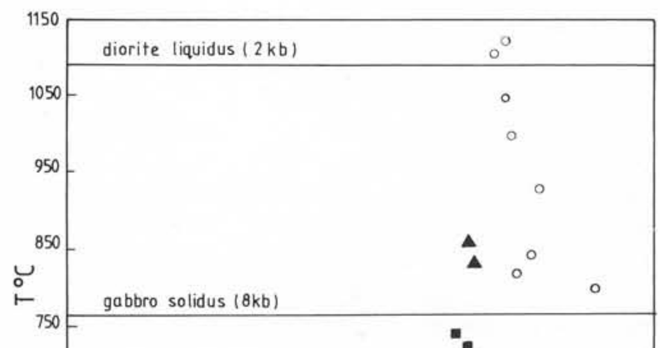

Fig. 5. The temperature variations of pyroxenes and pairs hornblende-plagioclase versus the Mg# $(Mg/Mg+Fe^{2+})$ of pyroxenes and hornblendes (after Blundy & Holland, 1990, modified). Symbols are: *circles* – clinopyroxenes, *filled triangles* – pairs: brown hornblende-plagioclase, *filled squares* – pairs: green hornblende-plagioclase.

tioned the crystallization of more basic plagioclases, and the reverse zoning of the plagioclases (Dziedzic, 1995b). The equilibrium temperatures for the pair of hornblende-plagioclase (Blundy & Holland, 1990) were calculated as 858–526°C. Hornblende crystallization began at significantly lower temperatures than that of the pyroxenes (Fig. 5).

The chemical compositions of the higher-temperature pyroxenes and the edenite hornblende indicate that the minerals crystallized under pressures of 10–5 kb. The

Table 2

Major, trace, and rare earth element analysis of metabasalts and gabbro

	Metabasalts									Gabbro					
	4A	4B	3A	3C	3F	2	5	5E	6A	14	9	12	6	18	15
SiO_2	50.36	48.94	49.64	49.91	49.65	50.26	51.52	46.94	45.99	50.27	50.74	51.07	51.31	51.89	52.12
TiO_2	1.12	0.92	1.70	1.76	2.52	2.21	1.45	2.11	4.32	0.62	0.54	0.56	0.55	0.57	0.34
Al_2O_3	17.52	17.41	14.66	14.01	14.69	14.68	15.64	14.61	14.49	11.85	17.40	15.18	15.33	17.19	20.51
Fe_2O_3	1.06	1.01	1.65	1.51	2.11	1.83	1.51	1.80	2.29	1.22	0.85	0.91	0.87	0.75	0.69
FeO	7.04	6.74	10.98	11.21	14.04	12.17	10.01	11.99	15.25	8.08	5.69	6.07	5.77	4.98	4.63
MnO	0.15	0.14	0.19	0.20	0.25	0.23	0.21	0.22	0.27	0.18	0.12	0.13	0.13	0.11	0.09
MgO	9.06	9.57	6.35	6.81	6.94	6.64	7.59	7.26	5.53	13.11	9.62	10.70	10.03	8.26	6.11
CaO	11.42	12.69	12.37	12.26	8.83	9.89	10.54	11.45	9.07	13.02	12.34	12.96	14.21	13.54	12.38
Na_2O	1.92	2.13	2.29	2.21	0.65	1.27	1.15	2.97	2.01	1.45	2.49	2.22	1.62	2.51	2.93
K_2O	0.31	0.40	0.13	0.05	0.20	0.66	0.31	0.51	0.41	0.20	0.20	0.20	0.17	0.20	0.20
P_2O_5	0.05	0.05	0.06	0.06	0.12	0.17	0.08	0.12	0.36	0.0	0.0	0.0	0.0	0.0	0.0
Mg#	70	72	51	52	47	49	57	52	39	74	75	76	76	75	70
	trace elements in ppm														
Cr		560	170	154	92	310	258	275		180	210	180	250	540	230
Hf		2.1	4.1	3.7	4.3	5.6	3.1	3.0		0.47	0.48	0.50	0.71	0.56	0.45
Nb		3	2	2	6	7	4	7		1.1	0.3	0.3	1.2	1.0	1.0
Ni		97	62	62	52	73	42	115		145	146	114	139	153	112
Rb		9	<5	<5	5	18	10	5		na	na	na	na	na	na
Sr		247	208	147	55	199	95	250		82	169	147	355	149	137
Ta		0.4	<0.3	<0.3	0.6	0.6	0.4	0.5		0.03	0.03	0.03	0.11	0.17	0.26
Th		0.7	<0.1	0.3	0.3	1.6	1.6	0.4		0.13	0.21	0.10	1.0	0.10	0.16
V		na	na	na	na	na	na	na		231	195	210	170	96	222
Y		12.7	34.6	32.5	45.5	39.8	33.4	26.1		10.2	10.1	9.6	8.6	5.1	11.0
Zr		55	101	105	147	170	99	109		6	5.6	5.6	7.3	5.0	6.6
La		4.6	2.1	2.9	3.2	14.3	7.6	16.0		1.11	1.31	1.18	3.58	1.16	1.14
Ce		11.8	9.4	12.0	12.8	33.4	20.4	19.5		9.6	9.5	10.0	12.9	9.2	9.7
Pr		1.4	1.4	1.5	1.4	4.4	2.4	3.1		0.34	0.35	0.32	0.72	0.27	0.36
Nd		7.6	10	10	8.0	21	11	10.5		2.48	2.46	2.35	3.64	1.66	2.70
Sm		2.5	4.2	4.6	3.1	6.0	3.5	3.3		1.12	0.96	1.07	1.31	1.04	1.14
Eu		0.8	1.4	1.5	1.2	1.7	1.1	1.4		0.55	0.54	0.58	0.66	0.40	0.57
Gd		2.4	4.2	4.5	4.8	5.9	4.4	3.9		1.3	1.2	1.1	1.1	0.7	1.3
Tb		0.4	0.9	1.0	1.1	1.1	0.7	0.7		0.20	0.17	0.19	0.18	0.10	0.20
Dy		2.8	5.0	5.1	6.7	6.6	4.8	4.3		2.0	1.7	1.8	1.5	0.9	2.0
Ho		0.6	1.3	1.4	1.8	1.5	1.1	0.9		0.42	0.34	0.38	0.31	0.17	0.40
Er		1.5	3.3	3.2	5.0	4.0	3.2	2.2		1.3	12	1.2	1.1	0.6	1.3
Tm		0.2	0.6	0.6	0.8	0.7	0.5	0.5		0.21	0.18	0.20	0.17	0.09	0.20
Yb		1.5	3.2	3.2	4.6	3.9	3.1	2.8		1.12	0.95	1.0	0.88	0.46	1.13
Lu		0.2	0.56	0.56	0.74	0.68	0.45	0.4		0.19	0.15	0.16	0.13	0.07	0.16
Eu/Eu*		0.98	1.00	1.00	0.95	0.86	0.86	1.19		1.39	1.54	1.62	1.63	1.35	1.43
$(La/Yb)_N$		2.20	0.47	0.65	0.50	2.63	1.76	4.10		0.71	0.99	0.85	2.92	1.80	0.72
$(Sm/Yb)_N$		1.85	1.46	1.60	0.75	1.71	1.25	1.31		1.11	1.12	1.19	1.65	2.51	1.12

Major elements in wt%, volatile free, Mg# = 100 Mg/Mg+Fe^{2+}; Fe_2O_3/FeO = 0.15; na = not analysed

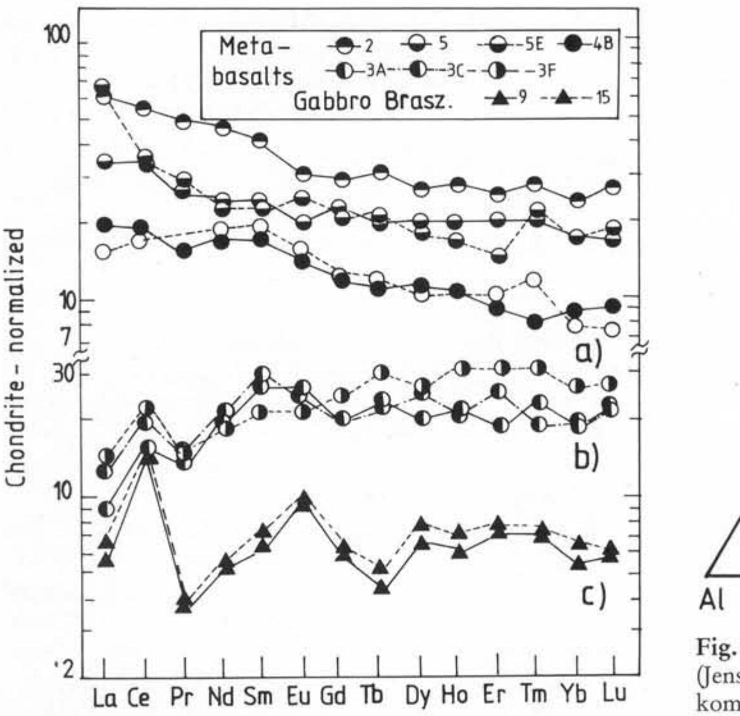

Fig. 6. Chondrite-normalized (Sun & Mc Donough, 1989) REE content plots for the enriched (a) and depleted (b) metabasalts as well as for gabbros (c). In (a) calculated parental melt concentrations (empty circles) are shown relative to those of sample 4B.

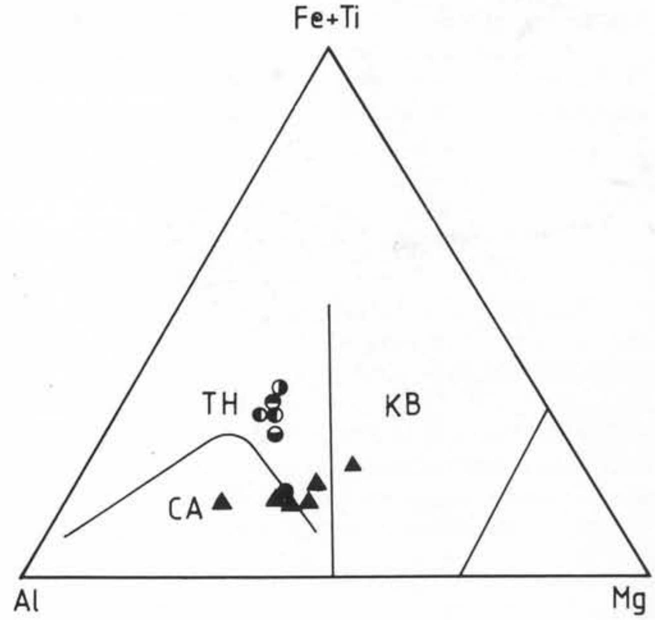

Fig. 7. Metabasalts and gabbros on an (Fe + Ti)-Al-Mg plot (Jensen, 1976). Fields: TH - tholeiitic, CA - calc-alkaline, KB - komatiitic basalt. Symbols as in Fig. 6.

crystallization pressure in the gabbro was 10–4 kb, with two maxima: at 8–7 kb and 5 kb (Dziedzic, H., 1989). Supposing the pressure under which the studied gabbros crystallized was 8 kb, their solidus temperature should have been 765°C, because such a solidus temperature was experimentally determined for basic tholeiitic magmas at a pressure of 8 kb (Holloway & Burnham, 1972). Therefore the temperatures estimated using the clinopyroxene and hornblende-plagioclase pair geothermometer of the Braszowice gabbro below 765°C are the subsolidus temperatures at which the re-equilibration of the chemical compositions of the minerals occurred under subsolidus conditions, pressure decreased to 5–4 kb. This is confirmed by the Al^{VI}/Si ratio (Fig. 4) of the lower-temperature hornblendes. The re-equilibration ended, in lower amphibolite facies conditions, and the pressure did not fall below 4 kb, which suggests a depth of about 13 km. However, the beginning of the gabbro crystallization could have taken place at a depth of about 30 km.

During the uplift to shallow crustal levels of the crystallizing Braszowice gabbros, acompanied by mantle upwelling, narrow shear zones, marked by dynamic recrystallization of plagioclase, diallage and hornblende formed in the gabbros (Dziedzic, 1988). These zones probably started forming under conditions close to the gabbro solidus. Dynamic deformation is also manifested by planar structures (S_1 and S_2) and subhorizontal lineation, due to strike-slip movements along a transform fault. The deformation continued until the rocks reached amphibolite facies conditions, when the gabbro locally became amphibolitized (up to 1m thick strips) along the margins of the shear planes. This ductile deformation must have taken place at depths below 10 km.

ANALYTICAL DATA

The investigated rock complexes consist of metabasalts occurring within the Góry Sowie gneisses and the gabbros of the Braszowice massif, both situated at eastern margin of Góry Sowie gneiss block. Chemical analyses of the metabasalts and the REE, Th, Hf, Nb, Ta and Y of the metabasalts and gabbros (except Y) were carried out in Activation Laboratories, Ltd., Ancaster, Canada. The trace contents of Rb, Sr, Zr, Cr and Ni in the metabasalts were determined by XRF in the Inst. Geowiss. u. Lithosphärenforschung, Justus-Liebig Univ. Giessen, Germany.

The major element data for the gabbros were determined by wet method in the Chemical Laboratory of PG, Katowice, Poland, whereas Sr, P, Zr, Ti, Y, V, Cr and Ni assays were carried out by ICP in the Chemical Laboratory of GFZ, Potsdam, Germany. Selected geochemical data are listed in Table 2.

The metabasalt samples were collected from outcrops in the Góry Sowie Block near Bielawa and Gilów, whereas the gabbroic ones were taken from a borehole at

Braszowice.

Selected samples of "unaltered" rocks were chosen for further analysis.These were rocks unaffected by simple secondary processes and, with the exception of the gabbros, not cumulate varieties. Although this procedure reduced the material to be analyzed (> 35% was eliminated), it gives more accurate results.

WHOLE ROCK GEOCHEMISTRY AND MAGMATIC AFFINITY

The chemical compositions of the metabasalts, considered volatile-free and with Fe_2O_3/FeO = 0.15, do not differ significantly in terms of silica (46–51.5% SiO_2), aluminum (14–17% Al_2O_3) and magnesium contents (5.53–9.57% MgO). Their Mg# values (Mg# = 100 Mg/(Mg + FeO^{2+}) in atomic proportions with Fe_2O_3/FeO = 0.15), range from 39 to 57 if the highest Mg# (= 72) of sample 4B, most likely related to magnetite fractionation, is omitted.

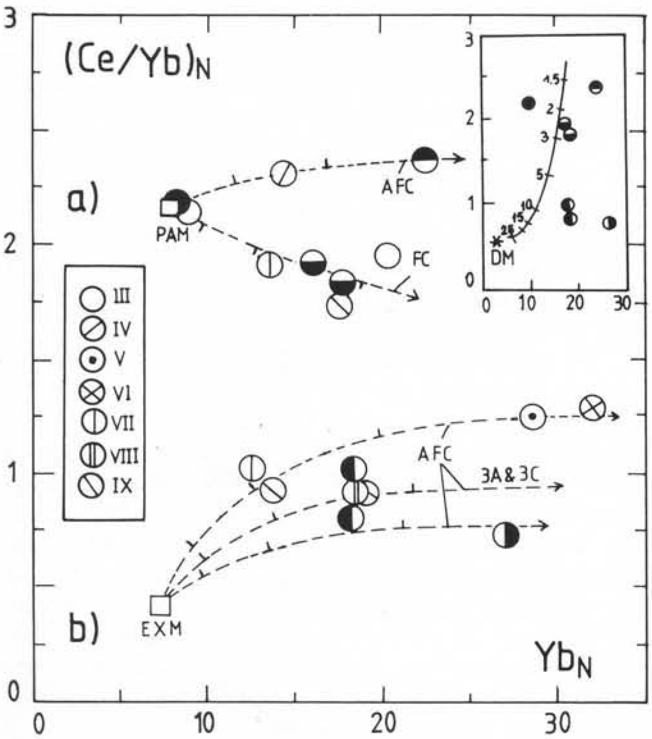

Fig. 8. Chondrite-normalized Ce/Yb versus Yb plots of the enriched (a) and depleted (b) metabasalts from the Bielawa vicinity. Assimilation-fractional crystallization (AFC) and fractional crystallization (FC) models relate to parental melt (PAM) or extracted melt (EXM) (see text). Tick-marks along the curves represent 20% increments of fractionation. Inset: distribution of the metabasalts relative to the spinel peridotite melting path (solid line with percentages of melting), and the trace element content of depleted mantle (DM). The metabasalts used for comparison were taken from an adjacent areas numbered as in Fig. 1b: III – Niemcza Zone (Koźmice, Wilków Wlk. – Białek *et al.*,1994); IV – Strzelin crystalline complex (GD-2 – Szczepański & Oberc-Dziedzic, 1998); V-VI – Ślęża (IV21) and Nową Ruda (NR13) massifs, respectively, – Pin *et al.*, 1988); VII – Śnieżnik Dome (S-20, S-67 – Floyd *et al.*, 1996); VIII – Stare Město Crystalline Unit (33 – Poubova & Sokol, 1992), and IX – Rudawy Janowickie Range (FR30, FR34 – Kryza *et al.*, 1995; OK2 – Winchester *et al.*, 1995).

The metabasalts display moderate range in their compatible trace element contents, with Ni and Cr concentrations ranging from 42 to 115 ppm and 92 to 560 ppm, respectively, although the latter value of Cr may be cumulate. Their rare earth element (REE) concentrations range from 8 to 60 x chondritic values.

The chondrite-normalized (Sun & Mc Donough, 1989) REE patterns of the metabasalts are shown in Fig.6. To allow for better data presentation, these samples were plotted in two groups, according to their element concentrations. The metabasalts of the first group (Fig. 6a) are light rare earth element (LREE) and middle rare earth element (MREE) enriched [$(La/Yb)_N$ >1.8; $(Sm/Yb)_N$ > 1.25] (subscript N denotes chondrite normalized), and may or may not show small negative Eu anomalies. The ratios of high field strength elements (HFSE) relative to MREE oscillate around unity [$(Zr/Sm)_N \sim 1.1$]. The metabasalts of the second group (Fig. 6b) are LREE(MREE)/HREE depleted (<0. 65 and >0.75, respectively) and have no Eu anomalies.Their HFSE/MREE ratios (~1.2) are similar to those of the first group of metabasalts.

Compared to the metabasalts, the gabbros have generally higher MgO (6.11–13.11) and Mg# (70–76) contents, while their FeO and TiO_2 contents are, in general, much lower (cf. Table 2). Their Ni and Cr contents are in the range 112–153 ppm and 180–250 ppm, respectively, whereas their REE concentrations range from 3 to 21 x chondritic values. Most of the gabbro samples are characterized by LREE depletion, moderate positive Eu anomalies (Eu/Eu* =1.35–1.63) and flattening in the heavy rare earth elements (HREE) on the chondrite-normalized plot (Fig. 6c). Their relatively low ratios of $(La/Sm)_N$ (av. 0.89), Nb/Zr (av. 0.10) and their positive correlation between V and Zr are typical of tholeiitic suites. Most the gabbro samples plot in the tholeiitic basalt field (Fig. 7) and shift into the calc-alkaline or komatiitic basalt fields, which may be due to plagioclase accumulation or an admixture of picritic material, respectively. In general, the lower position on the diagram of the gabbro relative to the tholeiitic metabasalts may be due to an early fractionation of magnetite as a result of high oxygen fugacity.

MODELLING OF TRACE ELEMENT CONTENTS

From the presented data, it follows that, in general, the metabasalts belong to two groups. One group of the metabasalts (4B, 2, 5, 5E) has higher LREE(MREE)/

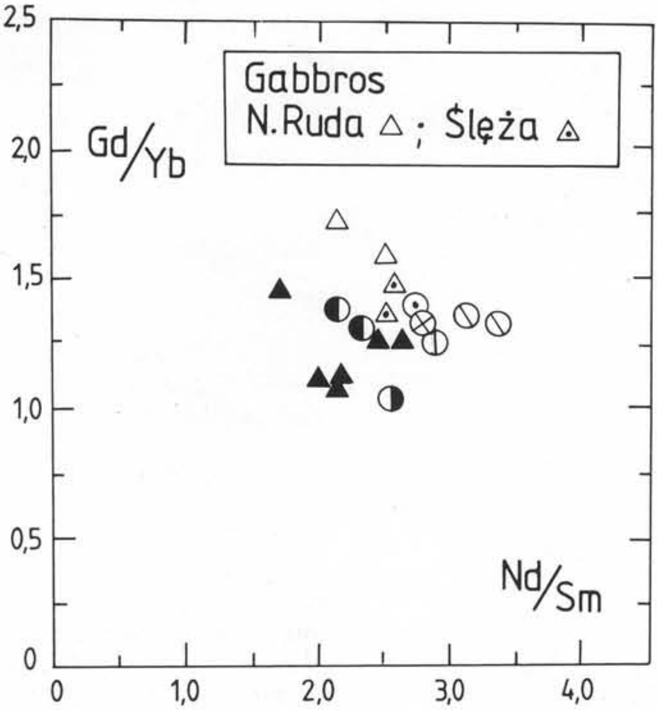

Fig. 9. Gd/Yb versus Nd/Sm ratios for the gabbros and depleted metabasalts. Gabbros of Nowa Ruda (NR04, NR08) and Ślęża (7G, 18G - Pin *et al.*, 1988).

HREE ratios than the other group (3A, 3C, 3F). Accordingly, the first group of metabasalts has been compared to enriched mid ocean ridge basalts (E-MORB) and the second one to normal MORB (N-MORB) rocks (Dziedzic, 1995a).

The differences between these metabasalts are clearly demonstrated on the $(Ce/Yb)_N$ versus Yb_N plot (Fig. 8, Inset). The calculated non modal batch melting path of the spinel peridotite was also included on this plot.

The distribution of the enriched varieties relative to the melting path may suggest the influence of garnet in the source. For example, the most primitive rock of this group, plotted to the left of the melting path, and also related to the more evolved samples, could not have been produced by a fusion of the spinel peridotite alone. On the other hand, the depleted metabasalts, which occur in proximity to the enriched forms and witchin the same structural setting, could have originated from the melting of spinel-bearing peridotite alone.

As derivation of the lavas from two separate and unrelated sources is rather inprobable, it is likely that both of the groups of metabasalts could have been derived from a source which changed slightly during the melting processes. Given that fractional melting is a more likely process than batch melting (Johnson *et al.*,1990; Sobolev & Shimizu, 1993), then this magma diversity can be explained by two-stage melting. The first stage started with a melt derived from a spinel/garnet peridotite source and provided the enriched variety of basalts, whereas the second stage, after the transformation of the garnet remaining in the residue (Johnson *et al.*, 1990), ended with spinel

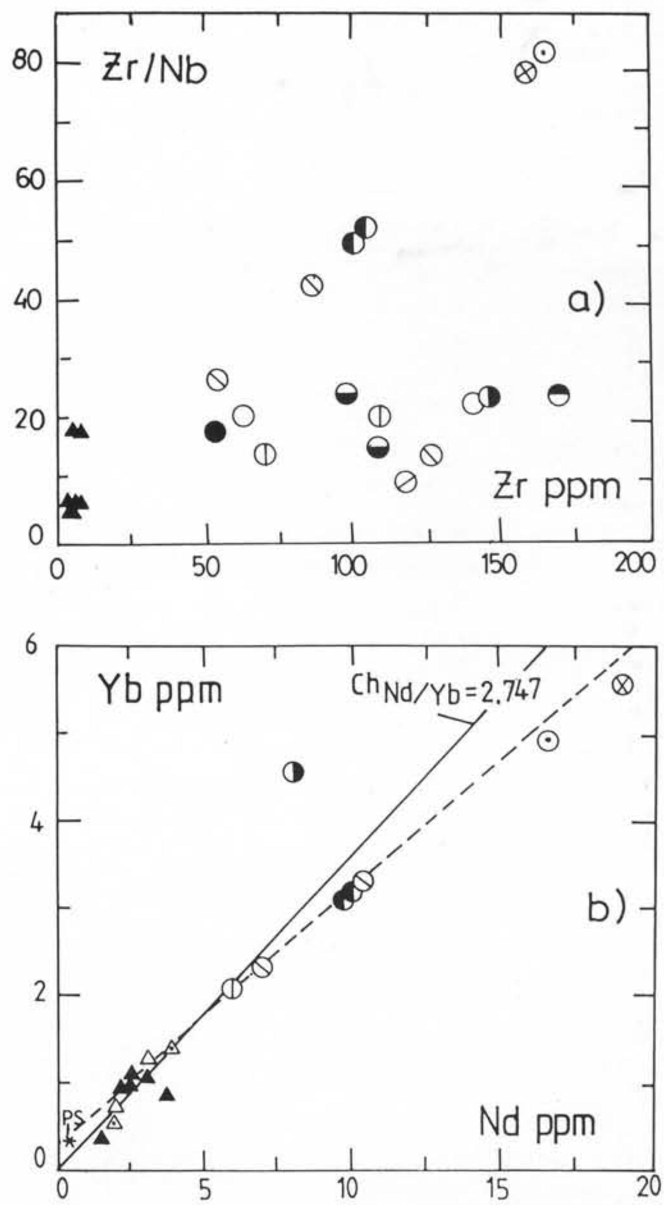

Fig. 10. a) Distribution of the basic rock assemblages in the Zr/Nb vs. Zr plot. Note the linear trend for the depleted metabasalts and gabbro. b) Nd versus Yb abundances in the gabbros and depleted-type metabasalts. A regression line (dashed, r = 0.95) through both the rock series is nearly collinear with the origin suggesting that their Nd/Yb ratios are the same as that of their parental source (PS). The chondritic Nd/Yb ratio is also shown (solid line).

peridotite derived melts, which, in combination with the plutonic member, produced the depleted basalts.

To test that possibility, several models were tested using non-modal near-fractional melting (Shaw,1970) of spinel and garnet peridotite (Johnson *et al.*,1990) with a depleted mantle source (McKenzie & O'Nions, 1991, 1995; Wood *et al.*, 1979 - of La only) as a starting material. The percentages of spinel and garnet peridotite undergoing melting were calculated, and the mineral/melt partition coefficients were taken from the literature as follows: La, Ce, Nd, Sm, Eu, Gd, Dy, Er and Yb (Kelemen *et al.*,

Table 3

Parameters used in modelling of the metabasalts and gabbros

	Metabasalts						Gabbros		
	FC		AFC				FC		
	4B	5	2	3A& 3C	3F	NR13	9, 15	SL	NR
olivine	-	19	7	15	15	6	-	-	-
clino-pyro-xene	-	-	19	7	-	27	4	4	5
plagio-clase	-	25	12	5	20	4	-	-	-
magne-tite	4	6	3	-	1	-	3	3	4
apatite	-	0.5	0.3	0.3	0.15	-	-	-	-
r	-	-	0.4	0.2	0.15	0.2	-	-	-
F	0.96	0.495	0.48	0.45	0.31	0.23	0.93	0.93	0.91
Ma/ Mo	-	-	0.35	0.14	0.12	0.19	-	-	-
palgio-clase cumu-late	-	-	-	-	-	-	20	17	15

FC – fractional crystallization, AFC – assimilation-fractional crystallization, F – fraction of magma remaining; SL=7G & 18G; NR=NR04 & NR08.

1993), except Nd for the clinopyroxene of spinel peridotite from Blundy *et al.*, (1998); Nb, Sr, P, Hf, Ti, Tm, Lu, and Y (McKenzie & O'Nions, 1991, 1995), except the last two elements for clinopyroxene from Hart & Dunn (1993); Zr and Tb (Beard & Johnson, 1997); Cr and Ni (Wedepohl, 1985).

The results of the modelling indicate that the parental melt (PAM) of the enriched group of the metabasalts can be generated by the liquid produced after the melting of 3% (in 1% increments, with one percent of the liquid remaining in the residue at each step of melting) of the spinel and garnet peridotite in the proportions of 70% and 30%, respectively. This parental melt approximates the rare earth element compositions of the least differentiated sample (4B) relatively well (Fig. 6a).This metabasalt can be modelled by removing about 4% of the magnetite from the parental melt. The more evolved varieties require a higher extent of fractionation and can be satisfactorily modelled (Fig. 8a), either by fractional crystallization (FC, sample 5), or by concomitant assimilation-fractional crystallization (AFC, sample 2) involving the parental melt and lower continental crust components (Taylor & McLennan, 1985), and using the algorithm of De Paolo (1981).

The fractionated mineral assemblages and ratios of Ma/Mo (the mass assimilated to the original mass of magma) (Farmer & De Paolo,1983), along with the other parameters used in the computations are listed in Table 3. All the fractional crystallization calculations were performed assuming the partition coefficient values of the

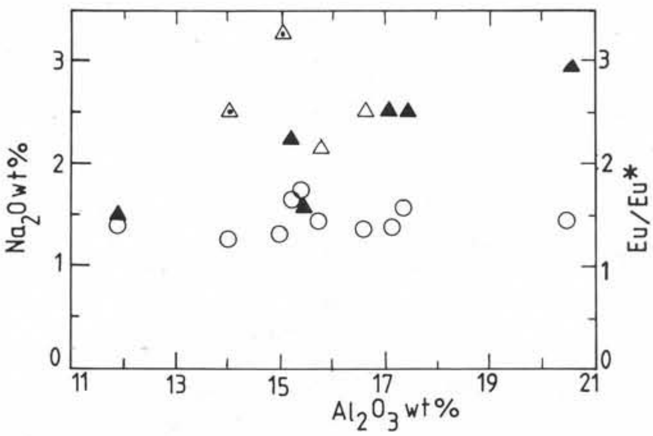

Fig. 11. A variation diagram of Al_2O_3 vs. Na_2O and related Eu/Eu* values in the gabbros. Note the lack of correlation between the Eu/Eu* values (empty circles) and the positive trends of the oxides.

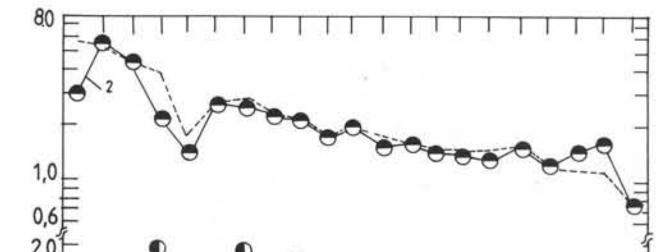

Fig. 12. A comparison of modelled (dashed lines) trace element contents with observed ones for selected metabasalts (circles) and gabbros (triangles). The symbols are the same as in Figs. 6 and 9.

REE (except Tm), and Sr as in Luais & Hawkesworth, 1994; Nb, P, Hf, Zr, Y, Tm and Ni as in Honjo & Leeman, 1987; Ti and Cr as in Villemant *et al.*, 1981.

The depleted metabasalts have low $(La/Yb)_N$ ratios and nearly flat HREE patterns. It is noteworthy that the gabbros in this area also display similar features and ratios of selected REE (cf. Figs. 6 and 9); these gabbros have been compared with normal mid ocean ridge basalts (Pin *et al.*, 1988: Dziedzic & Kramer, 1994). Although there are some differences in the trace element concentrations of both the suites, this did not prevent the treatment of these suites together when considering the diverse processes which accompanied their development. For example, the higher values of incompatible trace elements and lower values of compatible ones in the depleted metabasalts indicate that they crystallized from a more evolved melts than the gabbros.

It appears probable that the depleted metabasalts could be the potential volcanic counterparts of the plutonic members which display a similar distinct linear arrangement on the Zr/Nb vs. Zr plot (Fig. 10a). If that is true then a common source magma would be expected. That possibility appears to be supported by the Nd/Yb plot (Fig. 10b). Compared to the gabbro samples the depleted metabasalt samples have a wide range in Nd and Yb contents but both the groups of rocks plot about a line which is nearly collinear with the origin. This suggests that Nd and Yb were roughly equally incompatible (no garnet in the residue or fractionated assemblages) and that their Nd/Yb ratios are the same as that of their parental source (Hanson, 1989). The conclusion is that the depleted metabasalt and gabbro series were derived from a common parental source. In this respect, it might resemble the relationships at the Oman ophiolitic suites, where cumulate gabbros were formed by partial crystallization of a melt from which a large liquid fraction had been removed (Kelemen *et al.*,1997; Korenaga & Kelemen, 1997).

Batch melting clearly fails to produce a common parental melt. On the contrary, the subsequent melt, produced in two successive steps of incremental melting of the spinel peridotite mantle source, can have been common for both volcanic and plutonic series, if ~51% (based on the average Ni content in the gabbro of Braszowice), of the mass of the parental liquid was precipitated as a cumulate gabbro during the crystal fractionation of the cooling magma and the remaining 49% was removed. These results are comparable with those estimated or suggested for a selected section of the Oman cumulate gabbros (Browning,1984; Kelemen *et al.*,1997).

The extracted melt (EXM) uprose as porous flows (Korenaga & Kelemen, 1997) and differentiated by AFC processes providing the depleted variety of basalts (Fig. 8b). The composition of these volcanic rocks can be approximated by the fractional crystallization of low pressure mineral assemblages ranging from 27 to 37%, the limited (0.12–0.19) ratios of Ma/Mo, and mostly at moderate (~0.20) *r* values (rate of mass assimilated to mass fractionated) (Tab.3).

The concept that lower crust material was assimilated appears to be supported by the high enough initial (at 415 Ma, see further in the text) value of εNd (~ +7.0) obtained using the isotopic data (Pin *et al.*, 1988) of the indicated metabasalts and those of mafic granulite (PHN 1919 - Rogers & Hawkesworth, 1982), assumed to represent lower crustal material, and using a suitable algorithm (De Paolo,1981). The calculated (r = 0.05) εNd value is a little lower (~1.3 epsilon unit) than the average for the compared metabasalts.

The gabbros have limited ranges of compatible element contents consistent with the removal of magnetite from the liquid by crystallization during an early stage of differentiation. Their positive Eu anomalies may be indicative of plagioclase accumulation, as this mineral is known as the main carrier of europium. However, comparing the structural constituents of the plagioclase such as Al_2O_3 and Na_2O which correlate positively, so that feature is not reflected in the Eu/Eu* ratios (Fig. 11). This relationship suggests that a positive Eu anomaly is characteristic for the melt from which these rocks crystallized and is not due to mineral accumulation. On the other hand, under oxidized conditions, as supported by the crystallization of magnetite, the partition coefficients of Eu in the plagioclase are likely to have decreased (McKenzie & O'Nions, 1991), resulting in respective Eu/Eu* values that only weakly correspond with the oxides mentioned.

In addition, an enrichment in cerium (Fig. 6c) may have been due to either selective melt metasomatism or it reflects an oxidized conditions or both. Similar enrichments in Ce were also noted in the gabbros of Ślęża (Fig. 12), and Nowa Ruda (not shown) imply that these processes might be common to all these rocks.

The modelling suggests, as indicated by the averaged value of two samples (9 & 15), that the trace element contents in the gabbro may be approximated by the removal (< 10% total) of magnetite and clinopyroxene in roughly equal proportions, followed by the crystallization (~20%) of plagioclase (Table 3). These mineral proportions do not differ significantly from those used to model the gabbros of Nowa Ruda and Ślęża. In all cases the differences between observed and calculated values for a number of the trace elements are rather moderate (<30%, and even less), although much higher (> 50%) differences are also displayed for La, Ce, P, Hf, Zr and Tb (Fig. 12). Various factors, such as metasomatic and hydrothermal processes, inadequately selected distribution coefficients and the mobility of some elements, may account for these deviations.

DISCUSSION

COMPARISON WITH THE METABASALTS OF ADJACENT AREAS

The geochemical and geological characteristics of the discussed magmatic complexes correspond with those of the oceanic rifts where mafic and ultramafic suites occur togheter particularly along major fracture zones (Bonatti, 1973, 1978; Fox *et al.*, 1976). Also, in the discussed area such complexes in the form of isolated massifs or various scale magmatic bodies, accompanied in places by metabasalts, are concentrated along a submeridionally oriented zone. However, in most cases the plutonic and volcanic assemblages occur separately, the latter mostly in form of sill-like bodies. Accordingly, all the features were considered as delineating an ensialic rift-related right-lateral offset transform fault (the Niemcza–Lądek F. Z.) (Dziedzic, 1979, 1985, 1986).

The metabasalt assemblages, similar to those considered previously, occur in several geological units including the Sudetes. In order to compare the discussed metabasalts with those of adjacent areas, in which the enriched varieties are more abundant (see the references in the description to Fig. 8), selected samples of the metabasalts from the crystalline complexes, numbered as III–IX (Fig. 1b) were examined; (references in the description to Fig. 8). They contain rocks for which there are suitable geochemical data, of (III) - the Niemcza Zone, (IV) - the Strzelin area, (V–VI) - the Ślęża and Nowa Ruda Massifs, (VII) - the Śnieżnik Dome, (VIII) - the Stare Město Crystalline Unit (SMCU, Czech Rep.), and (IX) - the Rudawy Janowickie Range. The latter is believed to be a rift-related fracture zone of the transform fault-type (Dziedzic, 1986).

All the rocks used for this comparison represent "unaltered" and non cumulate varieties. Some of them have relatively high chondrite-normalized LREE (MREE)/HREE ratios of ~ 1.9 and ~ 1.5, respectively, and HFSE/MREE ratios (~ 1. 3). On the other hand, there are also rock samples that have low LREE (MREE)/HREE ratios of ~ 0.8 and ~ 1.1, respectively, and HFSE/MREE ratios (~ 1). These data and the Mg # values compare well with those of the enriched and depleted metabasalts of Bielawa. In fact, all the compared rocks plot within their respective domains (Fig. 8), either enriched (Strzelin area, Niemcza Zone) or depleted (Ślęża, Nowa Ruda, SMCU) or both (Śnieżnik Dome, Rudawy Janowickie Range), and the similarities mentioned suggest that the magma of all the volcanic suites underwent comparable differentiation processes. The trace element contents of these rocks can be relatively well approximated by the presented model. The differences between the observed and calculated values are generally rather insignificant (< 20%), although, for some trace elements (Nb, Sr, and rarely for Cr and Ni), higher differences (> 40%) are also noted. The latter may be due to the various proportions of the fractionated mineral assemblages in specific areas, compared to those used in the modelling of the rocks from Bielawa.

GEOLOGICAL IMPLICATIONS

From the comparison outlined above it follows that the metabasalt suites in the region exhibit of a number similar features. The occurrence of two varieties of volcanic rocks that can be separately related to common parental melts, in an area of maximal distance of about 80 km, seems to be not accidental, but may be a natural consequence of deep magmatic processes. For the same reasons, the assignment of any specific environments or episodes of volcanic activity to the particular geological units, may be unnecessary if it is assumed that the repetition of similar processes and rock assemblages may be questionable. Under the extensional conditions manifesting during volcanism, the melts generated by the decompressional melting of a diapir might explain the duality and spatial distribution of the magmas during a volcanic episode.This possibility may be supported by the transition from spinel/garnet peridotite mixed melts, reflected in the enriched rocks, to those produced within the spinel peridotite mantle source, reflected in the depleted rock assemblages. Also, the spatial distribution of the volcanic rocks can be acceptable if the diameters of the seismically estimated melting zones of the mantle diapirs are taken into account. These range between 50 and 100 km (Hasegawa *et al.*, 1991; Tryggvason *et al.*, 1983).

The "garnet signatures" of enriched MORBs have been explained either as the influence of mantle plume- or hot spot-related melts (Robillard *et al.*, 1992; Cousens *et al.*, 1995; Haase *et al.*, 1997), or by the partial melting of garnet pyroxenite, and eclogitic lithologies (Hirschmann & Stolper, 1996; Niu *et al.*, 1999). Additionally, in many areas the enriched and depleted rock assemblages, which display intimate associations, are related to fracture zones and usually exhibit limited ranges of isotopic and temporal relationships (Wright & Wyld, 1994; Guivel *et al.*, 1999).

The genetic correlation between the mafic plutonic and volcanic series in the Fore-Sudetic Block, and the corresponding volcanic complexes in adjacent areas imply a comparable geotectonic regime, and also a similar timing of the magmatic activity. That rift setting (the Silesian Rift - Dziedzic, 1998), accompanied by the decompressional melting of a mantle diapir could have been principal factors controlling the magmatism in the region. An ensialic setting for the metabasalts is indicated by the assimilation of lower continental crust material. That setting was also suggested for the metabasalts from the Śnieżnik area (Floyd *et al.*, 1996).

If it is assumed that the time-span of diapir activity may have been as long as 100 Ma (Duncan, 1981) and the corresponding thinning of the continental lithosphere at the base of the crust required about 50–75 Ma (Crough, 1983), then the granitoid plutonism and regional metamorphism studied in the Fore-Sudetic area could have been related to a diapir as a potential source of heat. This suggests that in a frame of mantle diapirism both of these phenomena may be causally related. The fact that the amphibolite facies metamorphic rocks are usually accompanied by the Variscan granitoids seems to support this view. The age differences between the suites could be a consequence of thermal progression resulting in metamorphic and anatectic products preceding the plutonic assemblages in time. It is possible that this mechanism also operated in the adjacent areas.

The thermal anomaly related to this diapir was associated with thinning of the continental lithosphere to about 60 km, high heat flow up above 50°C/km during the culmination of the granitoid plutonism, LP-HT granulite facies metamorphism (Dziedzic, H., 1985, 1996), and finally, with the participation of asthenospheric mantle material in the late Variscan volcanic rocks (Dziedzic, K., 1989, 1996). It is also possible to demonstrate that particular episodes of magmatic activity in the region almost coincide with an interval between two magmatic episodes, of about 30 Ma resulting from the rifting above the diapir (White & McKenzie, 1989). All these features appear to support an idea about the role of mantle diapirism in the geological evolution of the Silesian region.

The cogenetic relationships of the gabbro and metabasalt series support their origin during one magmatic event, in geological time scale terms. A U-Pb zircon age of 420 Ma for the Ślęża gabbro (Oliver *et al.*, 1993) suggests the earliest Devonian, i.e., ~ 415 Ma, according to Tucker *et al.*, (1998), as a plausible period for the emplacement of the magmatic complexes. As some gabbro is locally transgressively overlain by the Upper Devonian deposits, then the emplacement and subsequent uplift of the gabbro corresponds well with the Acadian event, as well as with the Early Devonian rift zone of central Europe delineated by Sawkins & Burke (1980).

CONCLUSIONS

The trace element geochemistry of the metabasalts from the Bielawa vicinity requires two parental melts to explain the diversities among these rock series. This is not obvious, however, if these series occur in intimate association and within the same geologic/structural setting. The enriched and depleted basaltic varieties were derived from a changing mantle source rather than from two separate sources.

The modelling suggests that both the metabasalt varieties resulted from two successive stages of the incremental melting of the peridotitic mantle source. The enriched variety was derived from a spinel/garnet peridotite melt mixture followed by melts of spinel peridotite. The cumulate gabbros crystallized from the latter melts, and an extracted large fraction of these melts provided the LREE-depleted variety of basalt. Thus, the timing of at least these basalts, and by implication, of the LREE-enriched ones should be compatible with the age for the gabbro.

Both the metabasalt varieties developed by mostly AFC processes involving the parental melts and lower continental crust material. The results of the comparison of these metabasalts and those of adjacent areas imply the same time-span of volcanic activity in the region.

In the late Variscan episode, the metabasalts of the eastern margin of the Góry Sowie Block were metamorphosed to LP hornblende granulites, synchronously with the gneisses of the border zone, which were metamorphosed to LP-HT cordierite gneisses. This process occurred along with the transition from high-grade amphibolite to granulite facies, associated with the near-isothermal decompression of the rocks. It suggests a rapid emplacement to a shallow crustal level, and a short duration of thermal progression to high temperatures (Dziedzic, H., 1996). A mantle diapir upwelling and magma underplating might have been responsible for this thermal progression.

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