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Genesis and evolution of the Sudetic late Hercynian volcanic rocks inferred from the trace element modelling

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Abstract The late Hercynian volcanic complexes in the Sudetes originated due to decompressional melting of the subcontinental lithospheric source region. The volcanic activity started with the calc-alkaline andesite magma in an Early Permian, followed by the picritic relicts and the andesitic assemblage rocks both of tholeiitic affinity. The tholeiitic andesites originated by AFC processes involving mantle-and lower crust-derived material. The differentiation of the andesitic parental magma within high-level magma chamber(s) by AFC processes involving upper crust components yielded the acid volcanic varieties in the area. The geodynamic processes and geological relations correspond with those of continental rift zones.

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INTRODUCTION

Volcanic rock suites associated with Upper Palaeozoic molasse deposits are best exposed at the northern margin of the Bohemian Massif in the Sudetes Mts. These rock associations occur within separate depressions, within which relatively compact volcanic fields can be delineated (Fig. 1). The molasse deposits, mostly continental in character, represent Early Carboniferous to Autunian intervals in the Intra-Sudetic Depression (ISD) and Carboniferous/Permian to Autunian intervals in the North Sudetic Depression (NSD). They most likely rest upon the Palaeozoic, pre-Carboniferous greenschist facies metamorphic sequences outcropping at the eastern flank of these depressions.

Preliminary geochemical studies of mafic volcanic rocks from the Intra-Sudetic Volcanic Field (ISVF) revealed their cogenetic nature and origin due to decompressional melting of a lithospheric mantle source (Dziedzic, 1996). In view of that, geochemical investigations of volcanic rocks from the North Sudetic Volcanic Field (NSVF) were undertaken and a comparison with those of the Intra-Sudetic Volcanic Field was made to explain possible genetic relationships between these rock suites in the Sudetes.

In this paper, intermediate volcanic rocks of the North Sudetic Volcanic Field and intermediate and acid rocks of the Intra-Sudetic Volcanic Field are considered on the basis of their trace element data. Genetic aspects of intermediate volcanites in both the areas, relationships of acid volcanic rocks relative to intermediate ones, model developmental trends of acid volcanites, and the geotectonic setting of southwestern Poland during the Palaeozoic are discussed on the basis of these data.

GENERAL FEATURES OF THE VOLCANIC ROCKS

The analysed volcanic rocks comprise intermediate rocks from the North-Sudetic Volcanic Field and intermediate and acid rocks from the Intra-Sudetic Volcanic Field. All occur within the Lower Permian (Autunian) molasse deposits.

All the studied samples from the North Sudetic Volcanic Field (PA1–PA5) (Fig. 1a) and from the Intra-Sudetic Volcanic Field (RA) (Fig. 1b), are massive rocks uniformly dark grey-black in colour. They contain phenocrysts of plagioclase and clinopyroxene. Olivine or replacement pseudomorphs after it occur less commonly.

In general, the rocks display various degrees of spilitization as first discovered here by Dziedzicowa (1958). These metasomatic processes took the form of albitization

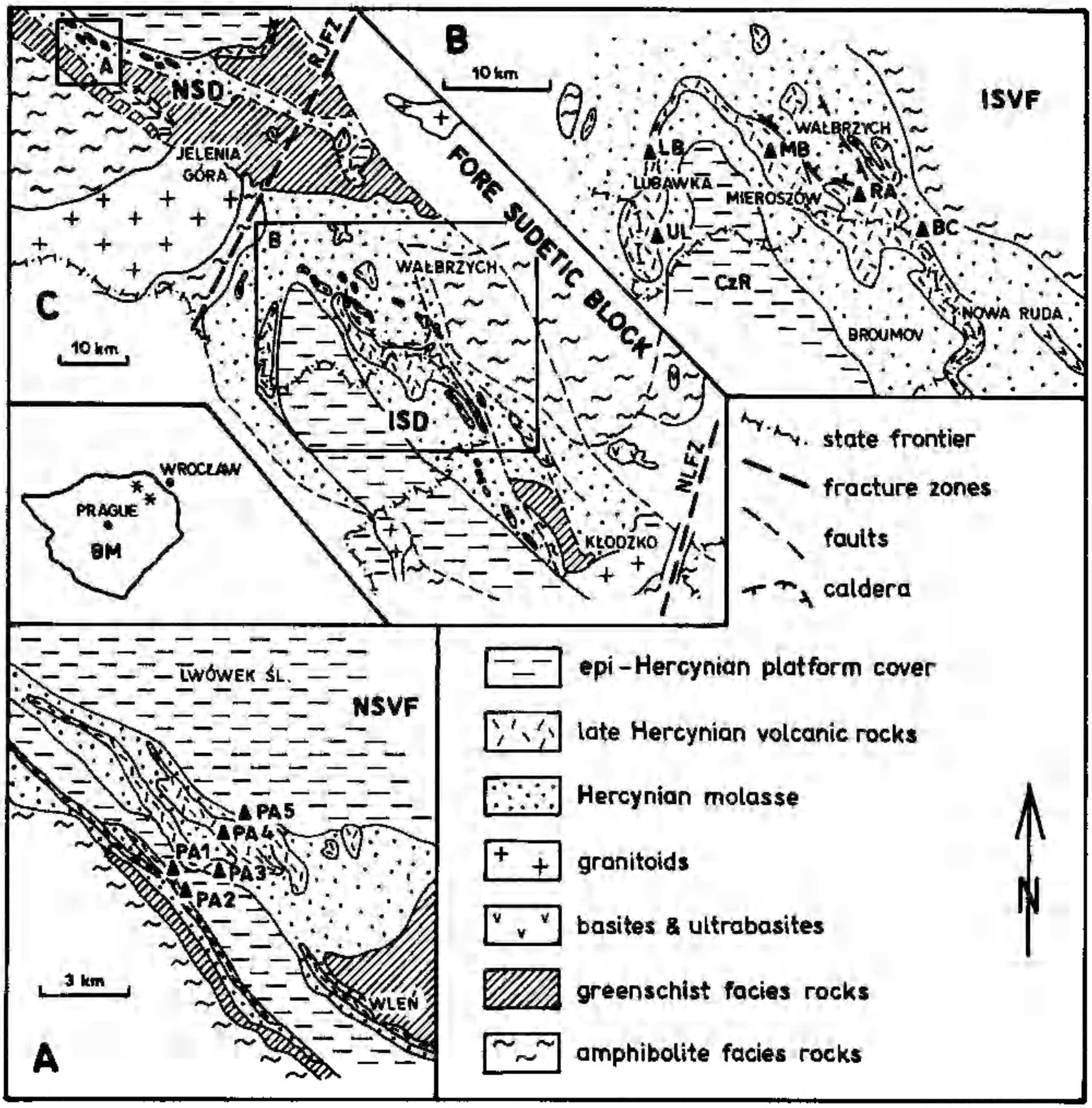

Fig. 1. Schematic geological maps and localization of the discussed volcanic rocks (filled triangles and letters).
a – North Sudetic Volcanic Field (NSVF) (after Milewicz, 1965, modified); b – Intra-Sudetic Volcanic Field (ISVF); c – distribution of Hercynian volcanic rocks in the North Sudetic Depression (NSD) and Intra-Sudetic Depression (ISD). Stars – position of the volcanic fields in the Bohemian Massif (BM). Abbreviations: RJFZ – Rudawy Janowickie Fracture Zone; NLFZ – Niemcza–Lądek Fracture Zone; CzR = Czech Republic.

of plagioclase, with interstitially crystallized K-feldspar and quartz, development of hornblende and chlorite after clinopyroxene, pseudomorphs (seladonite, serpentine) after olivine, and the appearance of hematite and calcite.

The Intra-Sudetic Volcanic Field volcanic complexes, which are mainly represented by intermediate and acid rocks, subordinately by basic varieties, form a volcanic chain subparallel to the axis of the depression (Fig. 1b). Along this volcanic chain these rocks were sampled in the vicinities of Lubawka (LB, UL) and Mieroszów (MR), and southeast of Walbrzych (RA, BC). Some samples (UL, MR and BC) were collected during drilling by Przedsiębiorstwo Geologiczne we Wroclawiu, most were taken from outcrops (Fig. 1b).

In the western part of the Intra-Sudetic Volcanic Field, the acid rocks occur at the uppermost part of the volcanic complex. An exception is a small oval contoured igneous body (sample UL) within the acid rocks near Lubawka, which megascopically resembles intermediate rather than acid rock. It possibly represents a portion of less differentiated magma injected into a more evolved one.

The acid rock varieties, usually reddish-brown in colour, contain phenocrysts of K-feldspar, quartz and plagioclase in various proportions. Biotite is rare and almost completely replaced by iron oxides

ANALYTICAL DATA

Major and trace element analyses were carried out in Activation Laboratories Ltd., Ancaster, Canada, and results are listed in Table 1. The analysed rocks appeared to be fresh but a few samples were intensively altered and have been excluded from discussion and the data set in Table 1.

Chemically, the intermediate rocks from both volcanic fields, considered without volatiles and with $Fe^{2+} = 0.85\Sigma Fe^{2+}$ do not differ significantly with contents of silica (55–57% SiO_2), aluminum (16–17% Al_2O_3) and total alkalies (5.59–6.62%), with K_2O/Na_2O ratios < 1. Slightly higher variations are noted in magnesium contents (1.68–

Table 1

Chemical analyses of the Hercynian volcanics of the Sudetes

	North Sudetic Volcanic Field (NSVF)					Intra-Sudetic Volcanic Field (ISVF)				
	PA5	PA4	PA3	PA2	PA1	RA	UL	LB	BC	MR
	A	A	A	A	A	A	QL	R	R	R
SiO_2	55.22	56.95	56.68	56.00	57.21	57.05	64.93	70.72	76.56	78.57
TiO_2	1.53	1.51	1.43	1.39	1.42	1.63	0.69	0.13	0.15	0.11
Al_2O_3	17.54	17.38	16.68	15.98	16.24	15.93	15.29	14.06	11.75	10.53
Fe_2O_3	1.39	1.38	1.06	1.51	1.44	1.63	0.89	0.82	0.40	0.42
FeO	7.13	7.06	5.36	7.65	7.40	8.35	4.56	4.19	2.03	2.21
MnO	0.06	0.07	0.17	0.08	0.11	0.13	0.03	0.03	0.01	0.01
MgO	3.27	1.68	5.57	4.53	3.35	2.58	2.35	0.52	0.19	0.34
CaO	7.38	7.03	7.11	6.71	6.41	5.43	1.75	0.57	0.14	0.13
Na_2O	3.41	3.58	3.15	3.02	3.12	4.02	2.45	3.42	0.67	0.81
K_2O	2.50	2.79	2.44	2.57	2.70	2.60	6.78	5.51	8.07	6.83
P_2O_5	0.57	0.57	0.59	0.57	0.59	0.65	0.29	0.02	0.03	0.04
LOI	4.56	3.07	3.91	2.87	2.37	1.60	3.20	0.22	0.81	0.77
Mg#	45	40	65	51	45	36	48	18	14	22
Trace elements in ppm										
Rb	39	71	60	86	89	66	147	135	156	136
Ba	537	518	580	567	588	1096	808	612	77	107
Th	6.4	6.5	8.9	8.7	9.7	8.7	16.8	21.9	17.9	19.6
Ta	DL	2	DL	2	2	DL	2	DL	2	2
Nb	29	29	33	32	33	33	33	32	32	21
Sr	432	278	308	300	286	285	51	36	19	21
Hf	6.2	6.2	7.6	7.6	8.6	9.4	7.6	7.3	4.6	4.2
Zr	300	300	371	370	379	473	336	261	133	156
Y	36	35	38	39	40	49	42	52	31	46
Ni	25	22	135	129	129	14	11	7	5	6
La	42.1	44.3	53.6	55.3	60.6	59.2	95.7	88.6	34.4	103
Ce	89	89	108	116	123	114	180	163	85	140
Nd	36	40	49	52	57	50	63	69	42	86
Sm	7.1	7.4	8.5	9	9.6	9	11	12	8	15.8
Eu	1.9	2	2.1	2.1	2.2	2.4	1.6	1.1	2	1.6
Tb	0.9	1.3	1.2	1.3	1.3	1.6	1.4	1.9	1.1	2
Yb	3.2	2.9	3.2	3.3	3.8	3.8	3.4	4.5	3.2	4
Lu	0.48	0.43	0.48	0.50	0.57	0.49	0.47	0.63	0.45	0.51
$(La/Yb)_N$	9.44	10.96	12.02	12.02	11.44	11.18	20.19	14.12	7.71	18.47
Eu/Eu*	0.84	0.79	0.76	0.71	0.71	0.77	0.46	0.27	0.77	0.08

Major elements in % normalized to 100%, volatile free with $Fe^{2+} = 0.85\Sigma Fe^{2+}$. $Mg\# = 100Mg/(Mg+Fe^{2+})$. Major elements, Ba, Sr, Zr, Y, Ni were determined by ICP; Rb, Th, Ta, Hf, REE by INAA and Nb by XRF in Activation Laboratories Ltd., Ancaster, Canada. DL – detection limit; A – andesite; QL – quartz latite; R – rhyolite

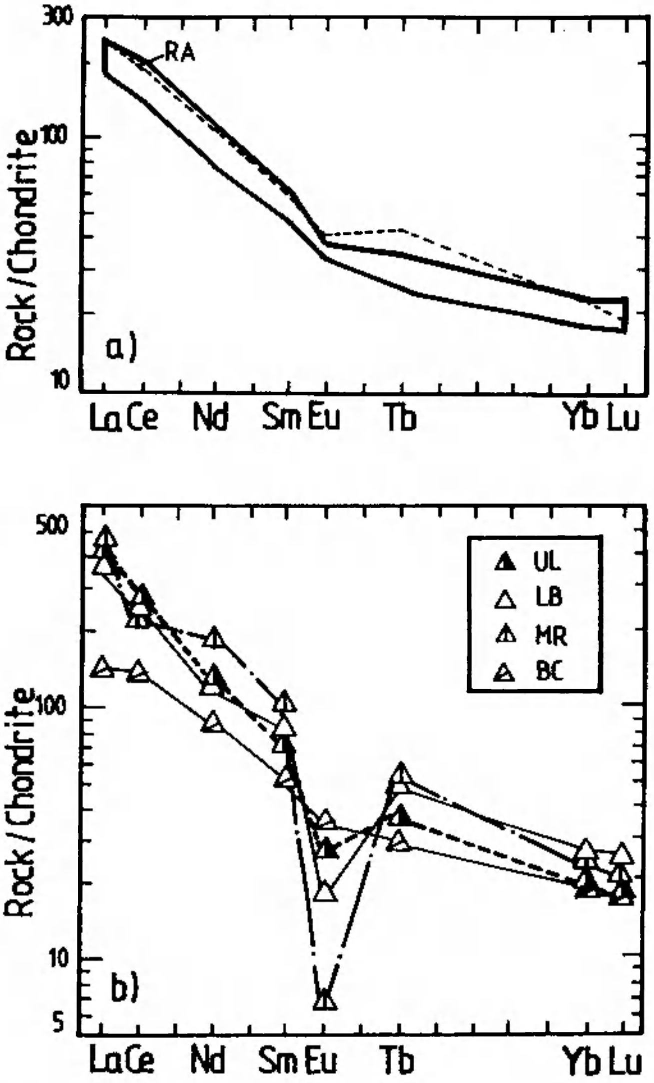

Fig. 2. Chondrite-normalized (Sun & McDonough, 1989) rare earth element patterns of the intermediate (a) and acid (b) rocks. For localization of acid rocks see Fig. 1b.
UL – Uniemyśl; LB – Lubawka; MR – Mieroszów; BC – Bartnica.

5.57% MgO) and the range of Mg# (36–65); Mg# = 100Mg/(Mg+Fe^{2+}) in atomic proportions with Fe^{2+} = 0.85 ΣFe^{2+}

The intermediate rocks are metaluminous with an alumina saturation index (ASI) corrected for apatite of < 1. They show moderately negative Eu anomalies (Eu/Eu* = 0.71– 0.84) calculated from Sm and Tb contents (Tab. 1). Chondrite-normalized (Sun & McDonough, 1989) rare earth elements (REE) partly overlap and display moderate slope (Fig. 2a) with $(La/Yb)_N$ ratios in a range of 9–12 (Table 1) and Yb values ranging from 17 to 22 x chondrites. These rocks are characterized by moderate concentrations of large ion lithophile elements (LILE) and light rare earth elements (LREE). However, the high field strength elements (HFSE) are almost constant and moderately high, with Nb ~ 32 ppm. Therefore the intermediate rocks are characterized by relatively low LILE/HFSE (Ba/Nb) and LREE/HFSE (La/Nb) ratios.

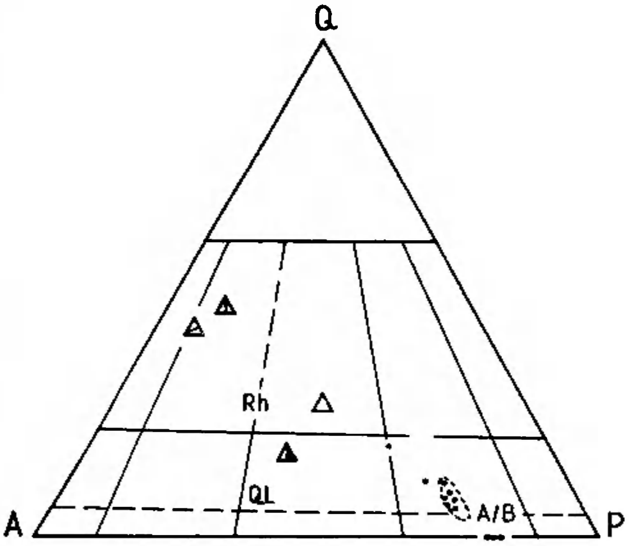

Fig. 3. QAP– classification diagram (Streckeisen, 1980) for the studied volcanic rocks.
Dashed contour encloses andesites discussed in this paper. Andesites from the ISVF (dots) (Dziedzic, 1996) are given for comparison. Fields: A/B – andesite/basalt; QL – quartz latite; Rh – rhyolite. Other symbols as in Fig. 2.

The acid rocks are peraluminous (1 < ASI < 1.2) and are potassium-enriched (5.51–8.07% K_2O) (Tab. 1), so that the K_2O/Na_2O ratios are high and in "ultrapotassic" varieties these ratios range from 8 to 12 at almost constant total alkali contents. An early post-emplacement ion exchange was very probably responsible for increasing K and reduction Na contents (Fischer & Schmincke, 1984). In case of intermediate rocks an increase of potassium contents upwards was due to the migration of this element along with volatile components at the place of magma deposition (Dziedzic, 1980a).The acid rocks exhibit well pronounced negative Eu anomalies (Fig. 2b).Their LILE/HFSE and LREE/HFSE ratios are variable but resemble those of intermediate rocks suggesting that they may be inherited.

On the QAP classification diagram (Streckeisen, 1980) the intermediate rocks cluster in the andesite field similarly to those from the Intra-Sudetic Volcanic Field. Members successively richer in silica occupy the quartz latite and rhyolite fields (Fig. 3). Most of these rocks are tholeiitic and only some of them (PA2, PA3 and UL) are transitional to calc-alkaline types according to the Jensen (1976) classification scheme. For the sake of clarity, all the andesites discussed in this paper and the sub-tholeiitic rocks from the Intra-Sudetic Volcanic Field (Dziedzic, 1996) are referred to in this paper as tholeiitic andesites.

MODELLING OF THE VOLCANIC ROCKS

ORIGIN OF THE INTERMEDIATE ROCKS

The geochemical similarity of intermediate rocks from the North Sudetic Volcanic Field and from the Intra-Sudetic Volcanic Field allows the assumption that the generation of magmas in both volcanic fields was controlled by similar mechanisms. On the La versus Ni projection, these volcanic rocks form an array oriented obliquely to the trend resulting from a closed-system fractional crystallization of mantle-derived melts after 5% fusion of the source material (Fig. 4). It indicates that in addition to fractional crystallization crustal contamination was also a likely critical process during the differentiation of these magmas.

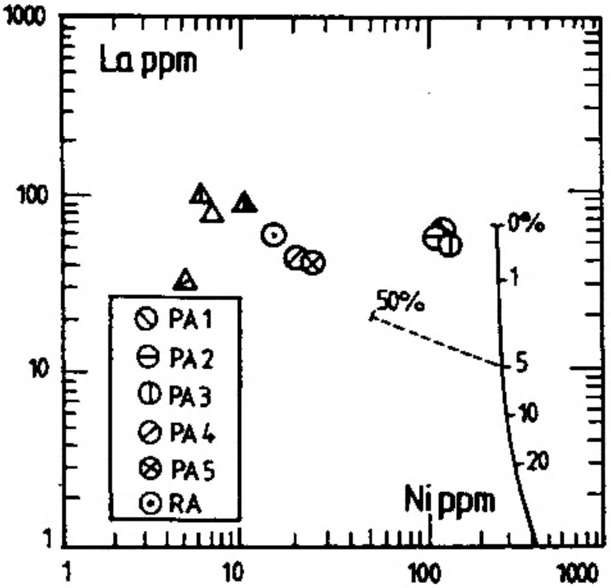

Fig. 4. Diagram of La vs. Ni of Late Hercynian volcanic rocks relative to the paths of batch melting mantle source (solid line) and fractional crystallization (dashed line). Percentages of melting and fractionation are indicated. For localisation of intermediate rocks see Figs. 1a, b. Samples: PA1–PA5 – Pławna 1 to Pławna 5; RA – Rybnica. Other symbols as in Fig. 2.

Combinations of Ba, Y, La, and Sm were utilized to test such a possibility. In a model of partial melting modal composition and melt proportions of a garnet peridotite (James & Henry, 1991) and trace element abundances of a primitive mantle (Hofmann, 1988) were chosen as a starting material. Ba, Y, La, and Sm concentrations were calculated (Shaw, 1970) after 10% batch partial melting. Assuming mantle-derived melts as a parental magma, contaminant represented by lower continental crust material (Taylor & McLennan, 1985), and using the algorithm of De Paolo (1981), the concentrations of trace elements in the final products were calculated.

The mineral/melt partition coefficients (k_d's) used in the calculations are the same as in the earlier paper (Dziedzic, 1996) and completed including the k_d's for fractionation of Ti (ol = 0.005, cpx = 1.07, pl = 0.05, ap = 0.0055; Villemant *et al.*, 1981; Honjo & Leeman, 1987), and of $Ni_{pl} = 0.05$; Brandon, 1989), which were mistakingly omitted there.

Trace element ratios in andesites relative to the assimilation/fractional crystallization paths (AFC) after 10% batch partial melting of mantle material are shown in Figure 5. In the presented model crystallized phases oli-

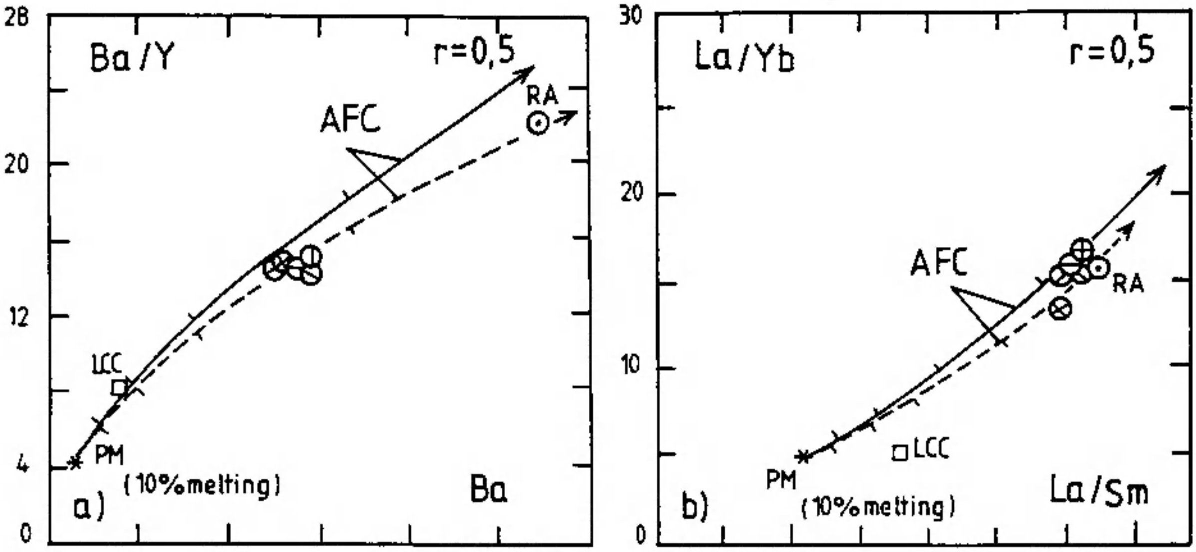

Fig. 5. Assimilation-fractional crystallization (AFC) models of mafic volcanic rocks a – Ba/Y vs. Ba; b – La/Yb vs. La/Sm. AFC models relate to garnet peridotite with trace element contents of primitive mantle (PM) (Hofmann, 1988) after 10% fusion and to lower continental crust (LCC) as an assimilant. Sample RA modelled separately (dashed curves) using the same parameters as in earlier paper (Dziedzic, 1996, Fig. 7, curves B). The values of *r* rate are shown at the right sides of the diagrams, and tick-marks along the curves represent 20% increments of fractionation.

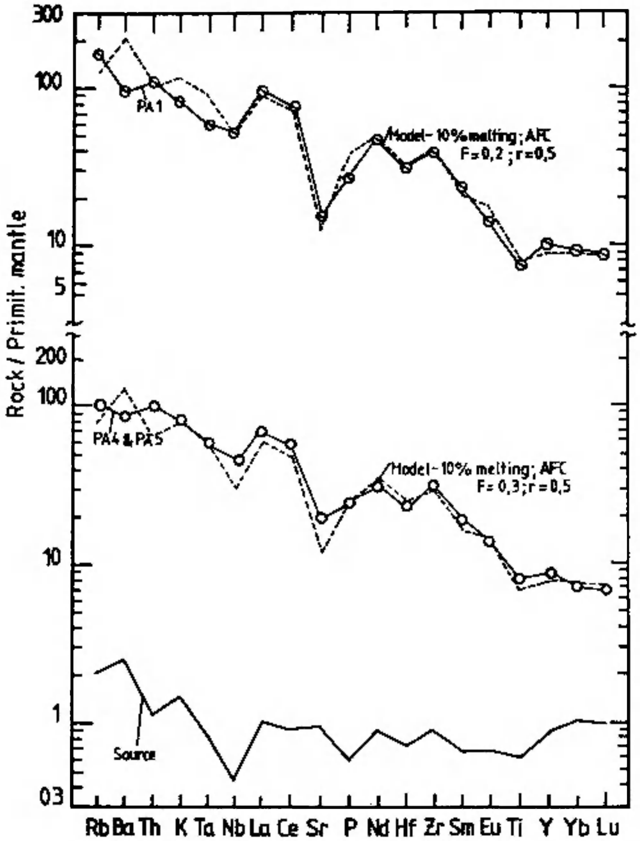

Fig. 6. The primitive mantle-normalized relationships between the calculated source composition and computed trace element concentrations in selected volcanic rocks of the NSVF. The degrees of partial melting of the model source (Dziedzic, 1996) and calculated element concentrations at respective *r* values as well as the fractions of magma remaining (F) are indicated.

vine:clinopyroxene:plagioclase:apatite were applied in the proportions of 21:49:29.91:0.09, which do not differ significantly from those obtained for this type of rocks in the ISVF (Dziedzic, 1996). A mass assimilated/mass fractionated ratio (*r*) of 0.5 was applied in the calculations. From the projections it follows that various groups of the applied trace elements (LILE, HFSE, REE) approximate the real ratios of these elements in the andesites of the North Sudetic Volcanic Field relatively well. The andesite from the Intra-Sudetic Volcanic Field (RA) can be modeled (Fig. 5, dashed curves) by applying the same parameters as in the earlier paper (Dziedzic, 1996; Fig. 7, curves B).

Subsequently, the same procedure was applied using the trace element abundances in a source computed from a picrite of the Intra-Sudetic Volcanic Field (Dziedzic, 1996). The small distance between the two volcanic fields does not prevent such an assumption because picritic magma in rift zones forms subcrustal intrusive complexes up to 200 km wide and 6 km thick (Borgia, 1994). It is more likely, however, that in the case of continental rifts picrites may form small, isolated diapirs. Hercynian picritic rocks associated with tholeiitic basalts have been also reported from several regions in Germany (Werner & Rösler, 1979).

For the modelling of the North Sudetic Volcanic Field rocks the external andesitic bodies were chosen (PA1 and PA4 & PA5; Fig. 1a), differing slightly in their trace element abundances (cf. Tab.1). As a result of modelling a satisfactory fit between the observed and calculated concentrations of the trace elements was obtained (Fig. 6). In both cases the differences are rather insignificant ($<20\%$) and greater deviations are noted only for some elements, particularly for LILEs, which may be related to the mobility of these elements.

The presented results support a thesis on the deep-level interaction of mantle-derived melts with heated crust ($r = 0.5$). Under such circumstances a significant lower crust material contamination should be expected. The estimates of assimilation rates resulting from the AFC models suggest that the tholeiitic andesites contain on average a 35–40% crustal component, which compares well with those inferred for this type of rock in the Intra-Sudetic Volcanic Field (cf. Dziedzic, 1996). Magmas produced due to the removal of the density barrier ascended and accumulated within intracrustal magma chambers where they underwent further evolution.

ORIGIN OF THE ACID ROCKS FROM AN INTERMEDIATE PARENTAL MAGMA

The relationships of the acid volcanites relative to the intermediate ones will be considered using the Intra-Sudetic Volcanic Field (Fig. 1b) as an example.The two types of volcanic rocks lie close together and exhibit inverse volumetric proportions, i.e: an increasing volume of one member usually corresponds to an appropriate volume reduction of the other member. Moreover, in drill cores, transitions from acid to more mafic rocks were observed with increasing depth suggestive a tapping of melts, at various stages of their differentiation, from a common magma chamber(s).

A small body of quartz latite (UL) near the village of Lubawka, croscutting a more extensive occurrence of rhyolite serves as an example of such a possibility. Taking into account experimental and model papers (Blake & Ivey, 1986; Bower & Woods, 1997), one may infer that within the magma chamber the boundary interface between the less dense, volatile-rich rhyolite layer at the top and the quartz latite layer below, initially occurred deeper and later shallower than the critical draw-up depth (Bower & Woods, 1997). Therefore the upper rhyolite layer was tapped earlier and the quartz latite later, injecting into the rhyolite.

It was showed that the studied rocks comprise quartz latite and rhyolites, high-silica variants inclusive.Their Mg# values are variable and, with the exception of the quartz latite, rather low (14–22), whereas negative Eu anomalies (Fig. 2b), are significant ($Eu/Eu^* = 0.77–0.08$). The Eu anomalies and low Ti and P contents and high Th/Nb ratios (0.51–0.93), more than twice as high as those in andesites, are typical of the upper continental crust and imply the contribution of such material in the production of the silicic magmas. This is well illustrated by the contrasting distribution, on a Yb/Th vs. Th plot (Fig. 7), of the intermediate rocks, contaminated with lower conti-

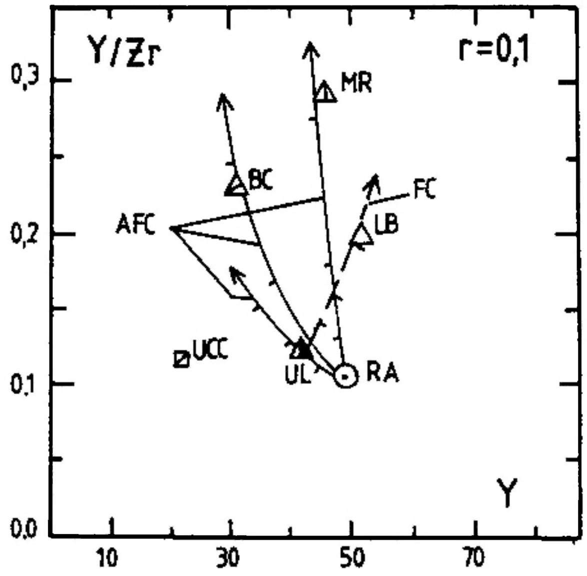

Fig. 7. Diagram of Yb/Th vs. Th of the Hercynian volcanics. Note contrasting position of the intermediate and acid rocks contaminated by lower and upper crust material, respectively.

nental crust material and acid rocks contaminated with upper continental crust material. The geochemical data and field relations, mentioned previously, indicate genetic connections between the parental intermediate magma and acid daughter magmas, gathered in the stratified zone within the upper part of a magma chamber(s) (Dziedzic, 1980b). The admixture of the upper crustal material and the acid volcanic rocks suggests that contamination within the chamber(s) took place at a high crustal level.

It follows from the trace element relationships that contamination did not result from a simple mixing but from a combination of assimilation and fractional crystallization (AFC). Theoretical and experimental papers (Huppert & Sparks, 1984; McBirney *et al.*, 1985) suggest that crystallization in systems comparable with those of the Intra-Sudetic Depression proceeds at the roof and margins of the chamber, and this type of solidification controls the chemical evolution of the magma, rather than phenocrysts preserved in the erupted products.

Table 2

Proportions of fractionated mineral phases used inmodelling of the acid volcanic rocks

Minerals	AFC andesite-quartz latite RA-UL	 andesite-rhyolite RA-MR	 andesite-rhyolite RA-BC	FC quartz latite-rhyolite UL-LB
olivine	–	2	2	2
clinopyroxene	13	5	11	–
plagioclase	55	65	55	37
K-feldspar	20	15.9	20.8	–
magnetite	8.5	8	7	7
ilmenite	2	2.4	2.2	2.2
apatite	1.5	1.7	2	1.7
zircon	0.02	0.026	0.03	0.017

AFC – assimilation-fractional crystallization, FC – fractional crystallization.

Considering possible evolutionary trends of the acid magmas, andesite (sample RA) was chosen as a parental magma and upper continental crust components as a contaminant (Taylor & McLennan, 1985), with a ratio of mass assimilated/mass fractionated (*r*) of 0.1 and the algorithm of De Paolo (1981). The choice of appropriate mineral/melt partition coefficients was not easy because they may vary greatly in acid rocks and generally tend to be higher in more silicic magmas (Mahood & Hildreth, 1983; Michael, 1988; Nash & Crecraft, 1985).

The partition coefficients for trace elements used in the calculations were taken from several sources: Rb, Th, Sr, Ti (Villement *et al.*, 1981); Ba, Nb, Hf, Zr, Tb, Y, Yb, Lu (Spell & Kyle, 1989), and the k_d of these elements for olivine from Honjo & Leeman (1987); La, Ce, Nd, Sm, Eu (Luais & Hawkesworth, 1994), in the case of apatite (Honjo & Leeman, 1987; Spath *et al.*,1996; Nabelek *et al.*,1988; Brandon & Lambert,1994), and in the case of olivine (Honjo & Leeman, 1987, except for Eu, from Villemant *et al.*, 1981); K, P, Ni (Honjo & Leeman, 1987).

The assemblages of the mineral phases removed, used for the modelling (Tab. 2), do not differ significantly from those approximated by the least squares method (Dziedzic, 1986).

The results of calculations, illustrated in the Y/Zr vs. Y plot (Fig. 8), approximate the real abundances and ratios of these trace elements in the acid volcanic rocks quite well. The quartz latite could have been produced from an andesitic parent by AFC processes with insignificant as-

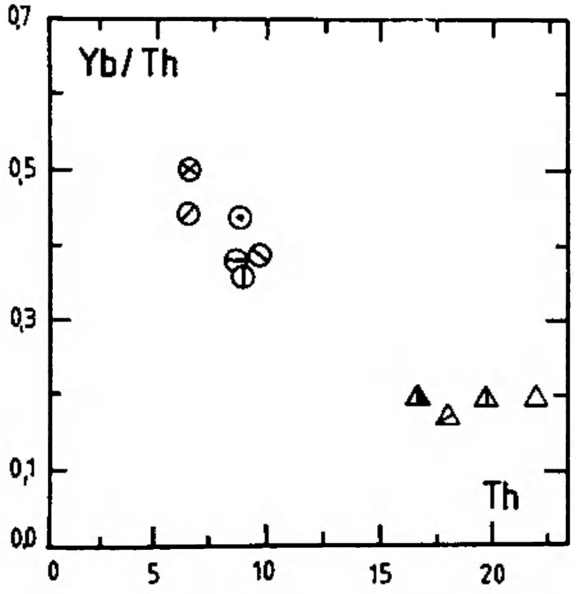

Fig. 8. Diagram of Y/Zr vs. Y illustrating the derivation of the acid volcanites from the intermediate parental magma (RA) and upper continental crust (UCC) as an assimilant. The value of *r* rate is shown at the upper right side of diagram and the tick-marks along the curves represent 20% increments of fractionation.

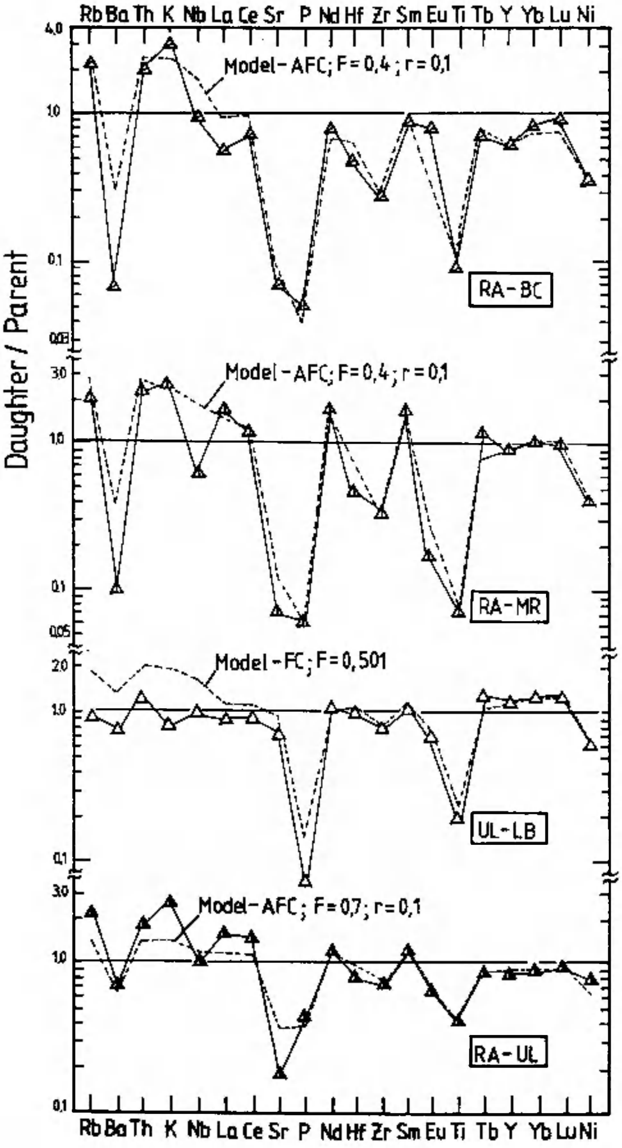

Fig. 9. Comparison of observed enrichments and depletions of trace element concentrations in the acid rocks relative to abundances in parental ones and, to those calculated in the models.Respective models, the fractions of magma remaining (F) and values of *r* rate are indicated.

similation of upper crustal material and 30% fractional crystallization. The quartz latite magma after ~ 50% removal of respective mineral phases (Tab. 2), provided the rhyolite (LB), poorer in silica, into which the former was locally injected. This agrees with the results obtained for these rocks by the least squares approximation method (Dziedzic, 1986). The high-silica rhyolites require a slightly higher assimilation of upper crust component and removal of 60% material dominated by feldspars (Tab. 2).

When the above data are applied to the other trace elements, the computed results normalized relative to their concentrations in a parental magma, indicate enriched or depleted abundances of the daughter products (Fig. 9). The calculated trace element concentrations approximate the observed abundances relatively well.The observed/calculated ratios for the andesite–quartz latite (RA–UL) and quartz latite–rhyolite (UL–LB) are in the range of 0.8–1.4 and 0.8–1.3 except for Rb, K, and Sr for the first group and Rb, Ba, Th, K, Sr and P for the second group. Likewise, trace element modelling of andesite – high-silica rhyolites (RA–MR and RA–BC) yielded observed/calculated ratios in the range of 0.8–1.2 and 0.8–1.3, except for Ba, Nb, Sr, Hf and Tb, and for Ba, Nb, La, Sr, and Eu, respectively. Greater deviations may be related to the mobility of some elements (Ba, Sr, K) and inadequately chosen distribution coefficients. For example, the model abundances of Nb mostly do not reproduce the observed ones, because Nb requires higher partition coefficients, suggesting some influence of accessory phases in the inclusions.

The AFC data show that the acid magmas were formed in a relatively cold environment ($r=0.1$) which efficiently limited (3–6%) the assimilation of crustal material.

In summary, it is clear that not all acid magmas were formed in the same way and not always by closed-system fractional crystallization as believed previously. In view of the missing isotopic data the usage of trace elements indicated the essential role of the upper crustal components in the production of the acid end-member. From the two end-members of the andesite–rhyolite suite the high-silica varieties and quartz latite were formed by AFC processes. The low-silica rhyolite represents the only product of fractionation via a quartz latite stage within a closed-system. Thus, the question of the formation of acid rocks cannot be generalized from a single estimation but must be considered with respect to individual magma bodies.

GEOTECTONIC SETTING

The geotectonic setting of late Hercynian volcanism in the Sudetes will be considered taking into account 1) the mantle source rocks, 2) the early stage magmatic products, 3) the probable geothermal regime, 4) the circumstances of the assimilation of continental crust material, and 5) the tectonic position of the volcanic complexes in southwestern Poland.

It was pointed out previously that the intermediate volcanic rocks may be best approximated by the model melting of mantle source rocks containing about 5% garnet. Melting of garnet peridotite under certain conditions, however, requires greater depths (>70 km) and silica-undersaturated alkaline igneous complexes would be expected, which are not observed. On the contrary, the late Hercynian (Autunian) volcanic rocks of the Sudetes are exclusively sub-alkaline, silica-saturated and mostly tholeiitic in nature, whereas the calc-alkaline varieties occur subordinately in the Intra-Sudetic Volcanic Field where relicts

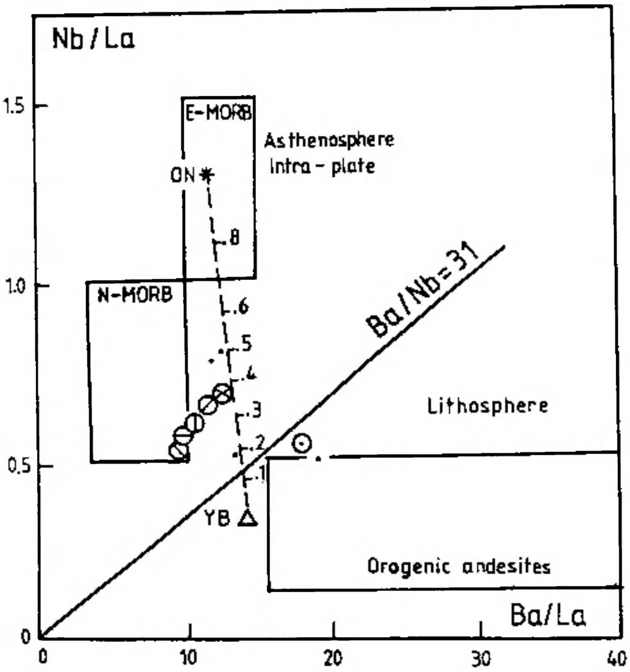

Fig. 10. Nb/La vs. Ba/La plot (projection after Menzies et al., 1991) of late Hercynian tholeiitic andesites from the Sudetes. The tholeiitic andesites from the ISVF (Dziedzic, 1996) are shown for comparison (dots). Note that the rocks are aligned near the mixing line (dashed) between the volcanic rocks presumed to be derived from the lithospheric (YB; Asmerom *et al.*, 1994) and asthenospheric (ON; Wedepohl *et al.*, 1994) mantle sources. The percentage mass fraction of an asthenospheric mantle in the mixture is indicated by numbered tick-marks The diagonal line (Ba/Nb = 31) marks the boundary between the volcanic rocks believed to be derived from the lithospheric or asthenospheric sources.

of picritic rocks are also preserved locally.

The volcanic activity started in the lowermost Permian (Early Autunian) when calc-alkaline andesites showing some geochemical signatures of island arc volcanic (IAV) rocks were formed locally under relatively hydrous conditions and a low degree of partial melting of the mantle rocks (Dziedzic, 1996). Isotopic age (K-Ar) determinations (Alibert *et al.*, 1987) on hornblende seperated from similar rocks nearby Bogatynia have yielded 285+5 Ma. Their place near the EM II (Enriched Mantle II) position on the isotopic plot ε_{Nd} - $^{87}Sr/^{86}Sr$ suggests an older material. A successive increase of partial melting yielded in turn picritic and andesitic magmas of tholeiitic affinity.

The picritic rocks. induced by hot asthenosphere and having signatures of mid-ocean ridge basalts (MORB) (Dziedzic, 1996), may be derivatives of picritic diapirs underplating the continental crust. Such diapirs are a common feature in rift systems (Deniel *et al.*, 1994).

Under the thermal anomaly and relatively anhydrous conditions more evolved magmas of tholeiitic andesite composition were formed through the interaction and mixing of mantle-derived melts with the products of the melting of crustal rocks. These volcanic rocks, originated through AFC processes at a lower crust component assimilation rate of 35–45%, display geochemical signatures of within-plate basalts (WPB). An arrangement of these volcanites near a two-component mixing trend on the Nb/La vs. Ba/La plot suggests that melts can be formed by mixing lithospheric and asthenospheric mantle components in proportions ranging from about 20 to 50% asthenospheric mantle (Fig. 10).

Subsequent differentiation of the tholeiitic andesite magma by AFC processes within high-level magma chamber(s) and limited (3–6%) assimilation of upper crust material provided the acid daughters which, like the parental magma, injected subvolcanically (Dziedzic, 1980b), during the Late Autunian (~270 Ma) volcanic paroxysm. A rift regime has been proposed (Dziedzic, 1980a, 1986) to explain the development of the molasse stage. This stage in the Sudetes began in the Late Devonian.

Moreover, it is worth stressing that in southwestern Poland (between Zielona Góra and Kraków) the lower Permian volcanic rocks and the molasse deposits, together with pre-molassic greenschist facies rocks comprising spilite- keratophyre volcanic suites, form a relatively narrow belt some 400 km long. All these originally volcanic–sedimentary piles, occupy en echelon arranged rift-like depressions (basins), such as the Lower Odra Depression, North Sude-tic Depression, Intra-Sudetic Depression, and Moravo- Silesian Depression (Fig. 11) considered to be rift-related structures (Dziedzic, 1986, 1989) which form the Silesian Rift.

The depression structures were formed under extensional conditions, with rifting and magmatic activity. These processes resulted in the formation of relatively thick sedimentary and igneous sequences which are characteristic of rift zones. The earlier stage of this sequence is represented by marine series while the latter one is mainly manifested by continental molasse deposits. Both stages were accompanied by volcanic rocks of tholeiitic affinity which might be derived from a common picritic type parental magma as was demonstrated by the least squares method (Dziedzic, 1984, 1986). Although the stratigraphic position of older volcanites is rather uncertain they probably belong to the same Devonian/Carboniferous and/or Late Devonian spilite/keratophyre associations of central Europe (Wong & Degens, 1983; Dziedzic, 1986, 1989; Pin & Paquette, 1997).

The depressions are separated by submeridionally oriented transversal fracture zones of transform-fault type spaced at an average distance of 60 km, which as zones of weakness were penetrated by the basaltoids of the Alpine cycle (Dziedzic, 1993). Such a crustal segmentation type is a rule rather than an exception within both young and old intracontinental rift zones, e.g., Rio Grande (Keller *et al.*, 1991), Gulf of California (Lonsdale & Lowver, 1980), Baikal (Sherman, 1978), Red Sea (Cohran & Martinez, 1988), and Oslo (Sundvoll *et al.*, 1990).

Having taken into account the geodynamic conditions, the preponderant tholeiitic nature of the Paleozoic volcanites developing at spreading rates estimated at 0.37 to 0.50 km/Ma (Dziedzic, 1986) and the regional field relationships, it is suggested that the rift setting may best explain the tectonic structure (Silesian Rift) of the area. If a Late Devonian age is assumed for the spilite-keratophyre extension-related volcanic associations, some of which belong to volcanic complexes of a continental rift (Werner & Rosler, 1979), then the ensialic Silesian Rift may correspond to the reactivated Early Devonian rift zone of central Europe delineated by Sawkins & Burke (1980).

DISCUSSION

The volcanic assemblages originated under dominantly extensional conditions, as indicated by their sub-alkaline chemistry, and elevated temperatures, as suggested by the presence of picritic relicts. An increasing melting of the source rocks might be a consequence of thermal progression accompanied by a rising geothermal gradient up to > 50°C/km or, alternatively, it may reflect successively higher levels of melt production under extension during crustal rifting of a ca. 60 km thick lithosphere (Dziedzic, 1989). The latter alternative seems to better explain the sub-alkalic character of the produced magmas and the decrease in lithospheric thickness.

A thinned lithosphere implies an upwelling of the asthenospheric mantle and preferred transmission of heat, responsible for an increase in melting and the appearance of picrite as well as for the assimilation of lower crust material and the contribution of asthenospheric components.

The inferred thickness of the lithosphere compares well with the results calculated for the North Sea rift system, where during the Mesozoic extension and rifting, the asthenosphere could rise up to c. 60 km in some segments. At normal potential temperature (1280–1300°C) and under rapid upsurge it can melt when a depth of < 80 km has been reached (Latin & Waters, 1992). Lithospheric thicknesses of 80 km or c. 60 km were also recently estimated in an area of Cenozoic volcanic activity and rifting in the Massif Central, France (Werling & Altherr, 1997).

The Late Hercynian volcanic assemblages in the Sudetes may be related to decompressional melting of a subcontinental lithospheric mantle source during regional extension. Their geochemical variations seem to reflect the progressive melting of this source with combined trace element signatures which are manifested in the volcanites. These trace elements either resided in the lithospheric

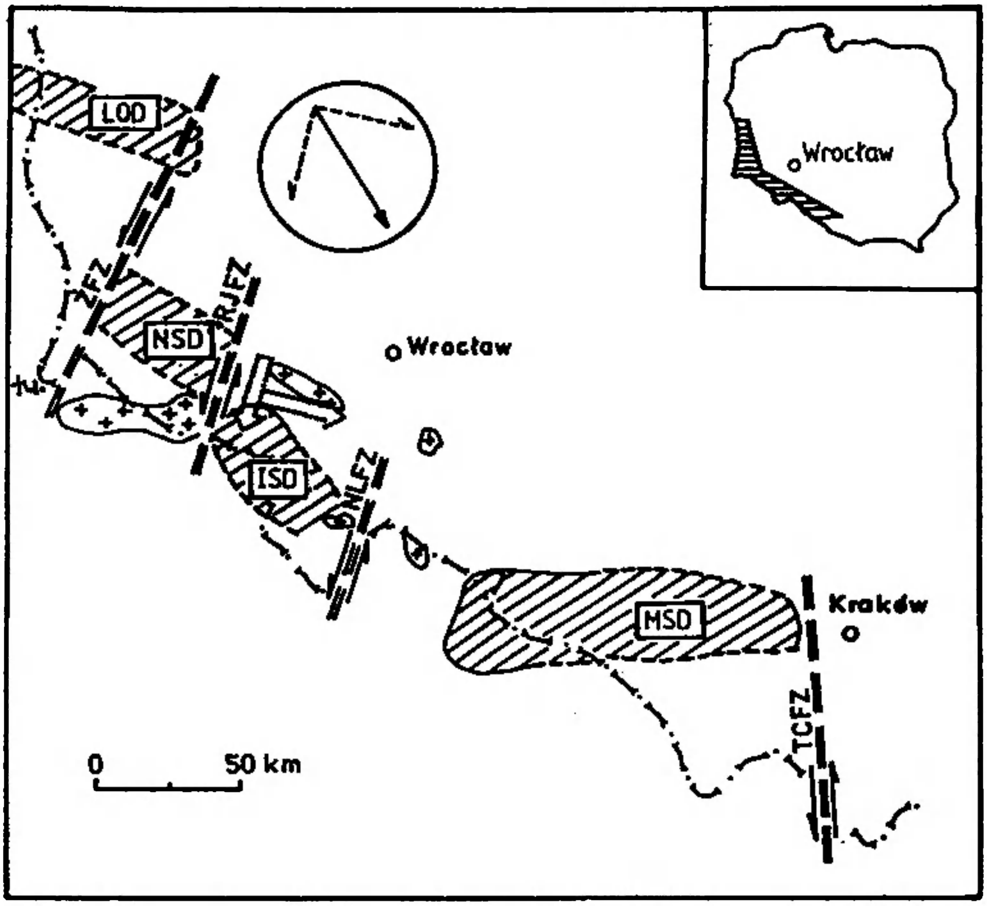

Fig. 11. En echelon arrangement of the rift depressions (after Dziedzic, 1989, modified). Note subparallelism of the Sudetic depressions and the resultant vector resulting from southward displacement and eastward migration of the crustal segments.
ISD – Intra-Sudetic Depression; LOD – Lower Odra Depression; MSD – Moravo-Silesian Depression; NLFZ – Niemcza–Lądek Fracture Zone; NSD – North Sudetic Depression; RJFZ – Rudawy Janowickie Fracture Zone; TCFZ – Trans-Carpathian Fracture Zone; ŻFZ – Żytawa Fracture Zone.

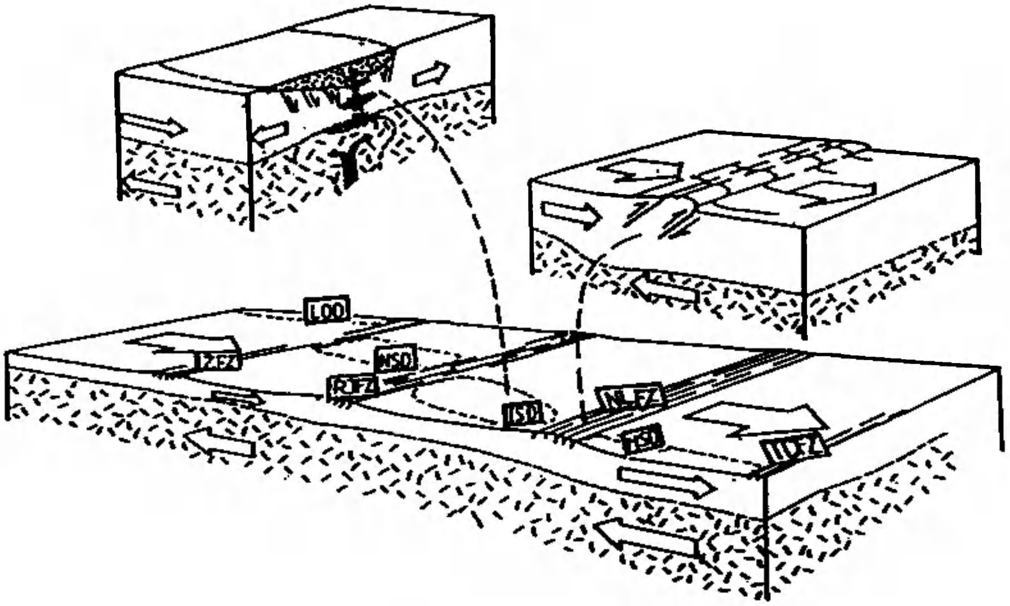

Fig. 12. Idealized geodynamic model of the Silesia region during the Hercynian episode (after Dziedzic,1989, modified). The two blockdiagrams at the top are representative, in outline, of the depressional and transversal structures. Designations as in Fig. 11.

mantle (MORB-type) or accreted to it or modified it during an earlier (EM II) subduction-related episode (IAV-type). The asthenospheric signature manifested with time (WPB-type) resulted from a mixing of lithosphere and asthenosphere derived melts, and was additionally modified by assimilation-fractional crystallization (AFC) processes (Dziedzic, 1996).

In the Silesian Rift the Permian volcanic rocks accompanying the molasse stage were confined to spatially limited depressions. A close spatial association of these volcanic rocks with Late Devonian volcanic rocks (referred to as pre-molasse stage volcanics) indicates a periodic operation of deep magmatic foci in these areas. Both groups of volcanites are associated with Palaeozoic sequences. These relationships suggest that in fact the pre-molasse and molasse volcanic suites represent some evolutionary stages of the persistently conservative geodynamic trend characterized by extensional stresses, sedimentary depressions, thermal anomalies and volcanism, altogether typical for rift zones. Such a trend demonstrated in the Silesian Rift during almost the whole Palaeozoic suggests the operation of a similar rifting mechanism at least during the Late Devonian–Autunian time interval. During this time span pre-existing rift-related transversal fracture zones were reactivated along which extensional crustal segments were displaced southward (Fig. 11) causing en echelon arrangement of the rift depressions (Dziedzic, 1989). It is likely that the curvatures of greater granitoid massifs in the vicinity of the Rudawy Janowickie Fracture Zone and Niemcza–Lądek Fracture Zone (Fig. 1c) are also related to these right-lateral displacements. Palaeozoic volcanic fields within the individual rift depressions and crustal segment imbrication have been regarded (Dziedzic, 1989) as the result of westward displacement or the subduction of continental lithospheric mantle.

The molasse stage volcanic fields are shifted relative to the pre-molasse ones. This manifests in both the Sudetic depressions in which the more abundant Autunian volcanites are concentrated at a distance of about 20–30 km west of the Devonian volcanic rocks. Similar situations, although the distances between the volcanites may vary a little, are noted in southern Krknoše (the Czech Republic) and also at the eastern periphery of the Moravo–Silesian Depression (MSD). On a regional scale it may reflect the eastward migration (~0.25 km/Ma in the Sudetes; Dziedzic, 1989) of the Silesian block over "fixed" subcrustal sources generating the melts.

The proposition that the distribution of Hercynian volcanic rocks is a result of crustal migration agrees with the well-known opinion about the eastward overthrust of a significant part of the Lower Silesian rifted block during the Hercynian episode. In this process diagonal crustal weaknesses (fracture zones) were preferentially deformed and transformed into transversal tectonic zones with dominating eastward (ESE) vergence implying a polarization of the westerly oriented displacement or subduction of the continental lithospheric mantle (Fig. 12) decoupling from the upper crustal layer. Thus, during the molasse stage two, diverse, roughly perpendicularly oriented geodynamic regimes acted concurrently resulting in NNE–SSW extension within the segments (transtensional-type structure with listric faults) and WNW–ESE compression in transversal zones (transpressional-type structure). This was reflected by a concentration of igneous rocks, both volcanic and plutonic, within the extensional structures (segments) and their restraint in the compressional ones (transversal zones). Both kinds of structures appear to be a consequence of co-operating mechanisms, that is diapirism and subduction-related crustal imbrication. The orientation of the Sudetic depressions approaching a NW-SE trend could correspond to the resultant vector (Fig. 11) of eastward migration and southward displacement of the crustal blocks (Dziedzic, 1989). If the rift model of the Palaeozoic sequences, as presented here, is correct, then the same mechanism could also be expected for the adjacent geological units.

The structure of the Silesian Rift has been considered (Dziedzic, 1986, 1989, and references therein), as part of the rift zone of central Europe. The proposed model is able to explain the conservative geodynamic trends and also the fold-block structure as a consequence of Hercynian diapirism, rifting and crustal segments displacement.

CONCLUSIONS

The volcanic suites were developed under extensional conditions and progressive melting of the subcontinental lithospheric mantle source region The calc-alkaline andesites came first followed by picritic rocks which were most probably derived from picritic diapirs underplating the continental crust.

An interaction of the mantle-derived melts with lower continental crustal material and AFC processes produced tholeiitic andesite magmas which were parental to acid magmas which evolved via AFC processes involving upper crust components.

The geochemical variations within the studied mafic rock assemblages seem to reflect compositional heterogeneities of the source region. The volcanic complexes originated due to decompressional melting of the subcontinental lithospheric mantle source during regional extension. The volcanic activity started in the Early Permian (~285 Ma) when the limited extent of melting allowed for local formation of calc-alkaline andesites corresponding to an EM II (Enriched Mantle II) source. Successive increases in melting yielded picritic relicts and subsequently a higher volume of tholeiitic andesites generated by AFC processes at an assimilation rate of 35–45% of lower crustal material. Differentiation of the parental tholeiitic andesite magma within shallow level magma chamber(s) with an assimilation of 3–6% of upper crustal components provided quartz latite and high-silica rhyolites. The low-silica rhyolite arose from closed-system fractional crystallization via a quartz latite stage.

The Palaeozoic basins of southwestern Poland, filled up by volcano-sedimentary piles, represent depressions of the Silesian Rift considered as part of the Early Devonian rift zone in central Europe.

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