SYMPOSIUM II: VARISCAN BASEMENT AND VOLCANISM IN THE ALPINE OROGENIC BELT

ALLANITE SOLUBILITY AND THE ROLE OF ACCESSORY MINERAL PARAGENESIS IN THE CARPATHIAN GRANITE PETROLOGY

I. BROSKA[1], A. S. CHEKMIR[2] and J. HATÁR[3]

[1]Geological Institute of the Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; geolbros@savba.savba.sk
[2]Institute of the experimental mineralogy of RAS, 142432 Chernogolovka, Moscow, Russia
[3]Geological Survey of Slovak Republic, Mlynská dolina 1, 817 04 Bratislava, Slovak Republic

Key words: granites, allanite stability, monazite-allanite dichotomy.

Introduction

Despite the low abundance of accessory minerals which usually do not exceed 1 % of rock volume in granites they may contain a substantial fraction of the whole-rock content of trace elements and geochemically important isotopes. These characteristics have resulted in the widespread use of accessory phases for different geochronometry methods inclusive fission track method, geothermometry and mass balance calculations of crustal magmatic processes. The purpose of this paper is to outline how the study of accessory mineral paragenesis has contributed to granite petrology in the Western Carpathians and to introduce part of the results on allanite solubility in felsic melts.

Methods

Accessory minerals were studied from crushed rocks. The experiments with allanite solubility in obsidian from locality Hraň (Eastern Slovakia) were carried at pressure 1, 2 and 3 kbars and temperature 700, 750, 800 and 850 °C; products of experiments were analysed on a Jeol JCXA-733 microanalyser using 15 kV accelerating voltage, 30 nA beam current and 3–5 µm spots.

Results

Studies of accessory minerals from granites began in the middle fifties by M. Mišík when he published a paper concerning accessory minerals from the eluvium of the Malé Karpaty granites. In the sixties and seventies classique separation techniques were developed by arranging laboratories for concentration of heavy minerals what enabled an extensive study of accessory phases from in situ samples. This resulted in the characterization of basic parageneses of accessory minerals in main granite types of the Western Carpathians (e.g. Hovorka & Hvožďara 1965; Hovorka 1968; Veselský 1972; Hvožďara & Határ 1978; Chovan & Határ 1978; Gbelský 1982) as well as the description of the dichotomy monazite/allanite, and magnetite/ilmenite.

Recently the dichotomy of monazite and allanite is explained by different solubilities of these minerals (Fig. 1) when the solubility of allanite is higher than monazite in peraluminous granites. On the other hand, the solubility of allanite in a lower temperature melt (e.g. below 670 °C) is close to monazite and in some conditions could be even lower (Fig. 1). This, in nature, may be the case

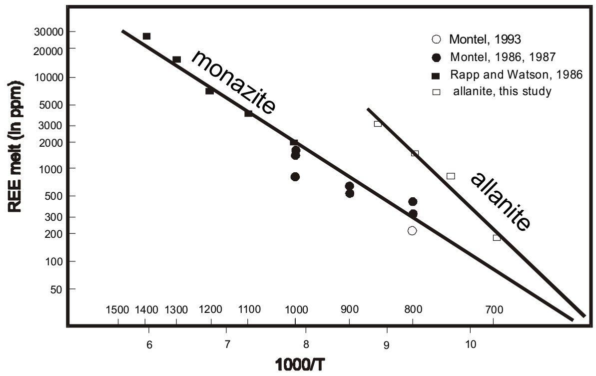

Fig. 1. Allanite vs. monazite in melt with D=1.08 and 1.03–1.17 respectively under water-rich conditions. D= [(Na+K+Li+2Ca)/Al] * 1/(Al+Si).

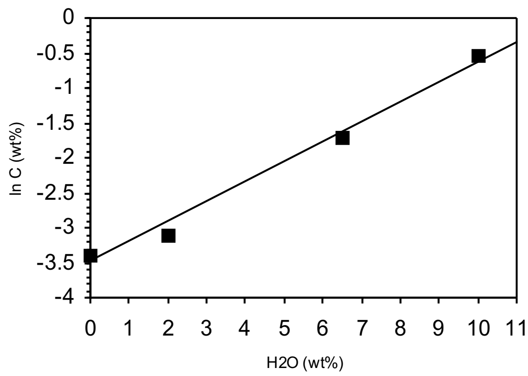

Fig. 2. Solubility of allanite in obsidian vs. melt water content. C is concentration of Ce_2O_3.

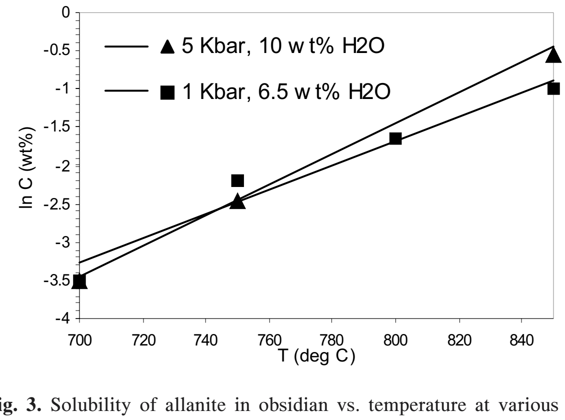

Fig. 3. Solubility of allanite in obsidian vs. temperature at various pressures. C is concentration of Ce_2O_3.

of pegmatite environment, or dykes where sometimes allanite prevails over monazite.

The solubility of allanite in felsic melts similarly as that of monazite increases with water content (Fig. 2) slightly depending on pressure when allanite at relatively high pressure conditions has higher solubility (Fig. 3). The dichotomy of magnetite and ilmenite is caused by level of the oxygen fugacity in melt: the higher fO_2 is connected with magnetite precipitation (Petrík & Broska 1992).

The distribution and chemistry of accessory minerals inclusive the fact of allanite/monazite and magnetite/ilmenite dichotomy along with the Pupin zircon typology became a basis for the subdivision of Western Carpathians Paleozoic granites into three groups: orogenic S and I-type granites (Broska & Uher 1991; Petrík et al. 1994; Petrík & Kohút 1998) and anorogenic A-type (Uher & Gregor 1992); the collisional S-type granites are of middle Carboniferous age, the extensional I-types are Late Carboniferous, the A-type granites are Permian.

Discussion and conclusions

Experiment runs show that the allanite solubility is higher than monazite in felsic melts. Although experiments on solubility of allanite in metaluminous melts are still missing, from point of view of monazite/allanite dichotomy we can to predict an oposite solubility effect for metaluminous melts compared to peraluminous — a higher solubility of monazite and a lower solubility of allanite.

A further study of accessory mineral parageneses would bring a possibility to recognize better the relict minerals in granite accessory mineral assemblages and so contribute more significantly to the understanding of granite protoliths and melting processes. Also, the future of accessory mineral studies is in the wider usage of their thermodynamic parameters, chemical dating and, in modern single grain isotopic analyses.

References

Broska I. & Uher P., 1991: Regional typology of zircon and their relationship to allanite-monazite antagonism (on example of Hercynian granitoids of the Western Carpathians). *Geol. Carpathica*, 42, 271–277.

Gbelský J., 1982: Characteristics of accessory minerals of the selected samples of the West Carpathian granitoids. *Geol. Zbor. Geol. Carpath*., 33, 5, 651–662.

Hovorka D. & Hvožďara P., 1965: Accessory minerals in the veporic granite rocks. *Acta Geol. Geogr. Univ. Comen., Geol.,* 9, 145–179 (in Slovak).

Hovorka D., 1968: Accessory minerals of some granite types from Malá Magura, Malá Fatra and Tríbeč Mountains. *Acta geol. geogr. Univ. Comen.,* 13, 165–189.

Hvožďara P. & Határ J., 1978: Accesory minerals of some magmatic and metamorphic rocks of veporides. *Acta Geol. Geogr. Univ. Comen.,* 33, 113–128 (in Slovak).

Chovan M. & Határ J., 1978: Accessory minerals of some types of veporic crystalline rocks. *Miner. slovaca,* 10, 9, 24–32 (in Slovak).

Montel J.M., 1993: A model for monazite/melt equilibrium and application to the generation of granitic magmas. *Chem. Geol.,* 110, 127–146.

Petrík I. & Broska I., 1994: Petrology of two granite types from the Tribeč Mountains, Western Carpathians: an example of allanite (+magnetite) versus monazite dichotomy. *Geol. J.,* 29, 59–78.

Petrík I., Broska I. & Uher P., 1994: Evolution of the Western Carpathian granite magmatism: source rock, geotectonic setting and relation to the Variscan structure. *Geol. Carpathica*, 45, 283–291.

Petrík I. & Kohút M., 1997: The evolution of granitoid magmatism during the Hercynian orogen in the Western Carpathians. In: Grecula P. et al. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca, Monogr.,* 235–252.

Rapp R.P. & Watson E.B., 1986: Monazite solubility and disolution kinetics: implications for the thorium and light rare earth chemistry of felsic magmas. *Contr. Mineral. Petrology*, 94, 304–316.

Uher P. & Gregor T., 1992: The turčok granite: product of post-orogenic A-type magmatism? *Miner. slovaca,* 24, 301–304 (in Slovak).

Veselský J., 1972: Accessory minerals of the granitoids of the Malé Karpaty Mts. *Geol. Zbor. Geol. Carpath.,* 23, 1, 115–131 (in German).

THE EARLY MIOCENE ERUPTIVE COMPLEX OF BORAC (CENTRAL SERBIA): VOLCANIC FACIES AND EVOLUTION OVER TIME

V. CVETKOVIC[1] and Z. PÉCSKAY[2]

[1] Faculty of Mining and Geology, Djusina 7, 11000 Belgrade, Yugoslavia; cvladica@eunet.yu

[2]ATOMKI, Bem ter. 18/C, 4001 Debrecen, Hungary; pecskay@moon.atomki.hu

Key words: Early Miocene, Lava flows, Pyroclastic deposits, Volcanic facies, Borac Eruptive Complex, Vardar Zone, Serbia.

The Borac Eruptive Complex (BEC) is situated in southwestern Sumadija (central Serbia). It is a part of the Vardar Zone composite terrane (Karamata et al. 1994), that is a suture zone along which the Vardar branch of the Tethys Ocean closed. Recent data (Cvetkovic 1997; Cvetkovic et al. 1998) revealed the presence of various volcanic rocks originated from high- to medium-K calc-alkaline and high-K alkaline (shoshonitic) magmas. These magmas formed within the post-collisional geotectonic setting by melting of the lower crust and subcontinental lithosphere, the latter being previously modified by the subduction component (high LILE/HFSE). Parental calc-alkaline and alkaline melts were later subjected to various differentiation processes dominated by fractionation, thermogravitational diffusion, contamination and probably magma mixing.

The BEC is built of different volcanic facies of dacites, lamprophyric rocks, quartzlatites and andesites to basaltic andesites. Relics of two caldera structures dominate this volcanic terrain: the Ostrica situated at south and the Borac caldera, which comprises the central and eastern area of the BEC. The major volcanic structures and volcanic facies are presented on Fig. 1 while the petrographical characteristics as well as radiometric ages of the Borac volcanics are summarized in Table 1.

The first eruptive events in the BEC occurred at the very beginning of the Early Miocene, about 23 Ma ago and volcanic activity lasted until about 20 Ma ago corresponding to the Egerian-Eggenburgian boundary. The repose periods between the volcanic phases were propably very short, most probably shorter than the analytical error of the age determinations (0.77–1.24 Ma). The stratigraphical evidence inferred two main eruptive periods during which the Ostrica and Borac caldera originated. These periods comprise several phases of volcanic activity which gave rise to different volcanic facies. The final character of these products were strongly influenced by various factors for example changes in the composition of the erupting magma (including gas and crystal content), eruption discharge rate, the surface properties etc.

The caldera of Ostrica originated during the first, dacitic volcanic phase. At the very beginning of the evolution of the BEC widespread dacitic lava flows, followed by analogous lava domes and extrusions originated. Due to a rather high eruption rate, this effusive activity emptied the upper parts of the magma chamber the roof of which, lacking physical support, collapsed. During the reactivation of the volcanism the wall of the northwestern side of the caldera was tectonically displaced and new portions of magma penetrated its bottom.

The second period of quartzlatite-rhyodacite explosive volcanic activity followed by quartzlatite, andesite and basaltic andesite effusions produced the collapse of the large Borac caldera. Apart from the rather simple evolution of the Ostrica structure, the Borac caldera is characterized by a diversity of eruption styles and volcanic products. After the effusive dacitic phase, a large zoned magmatic chamber originated below the central part of the recent Borac

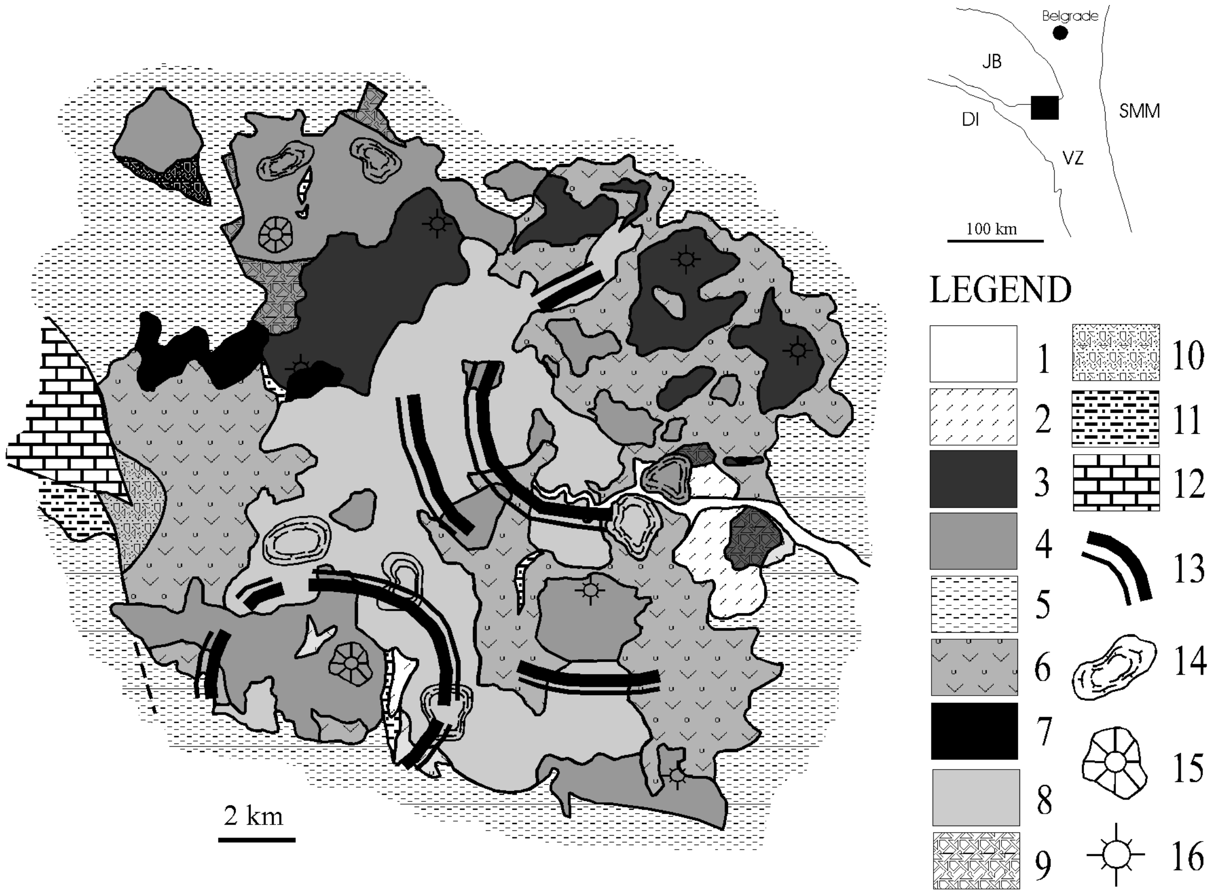

Fig. 1. Volcanological map of the Borac Eruptive Complex (the inset shows a simplified geotectonic sketch, after Karamata et al. 1994); EXPLANATION: 1 — Alluvium, 2 — Delluvium, 3 — Andesites and basaltic andesites, 4 — Quartzlatites, 5 — Middle Miocene sediments, 6 — Quartzlatite/rhyodacite primary and resedimented pyroclastics, 7 — Lamprophyres, 8 — Dacites, 9 — Autoclastic deposits in situ, 10 — Resedimented autoclastic deposits, 11 — Early Miocene sediments, 12 — Cretaceous sediments, 13 — Caldera relics, 14 — Lava dome, 15 — Volcanic neck, 16 — Supposed volcanic center; ABBREVIATIONS: JB — the Jadar Block terrane, VZ — the Vardar Zone composite terrane, DI — the Drina-Ivanjica terrane, SMM — the Serbian-Macedonian mass terrane.

Table 1: Lithological and petrographical characteristics and age of the Borac volcanic rocks.

Dacites	**Lamprophyric rocks**	**Quartzlatite-rhyodacite pyroclastics**	**Quartzlatites**	**Andesites and basaltic andesites**
Lava flows, lava domes, primary and redeposited autobreccia	Hypabyssal bodies, autobrecciated lava flows, different volcaniclastic breccia	Lapilli fall deposits, unwelded ignimbrites – primary and reworked	Lava flows, lava domes and necks, hyaloclastitic and autobrecciated facies	Lava flows and hyaloclastic deposits
Pl (An_{30-50}), ±Q, Ho (Mg-, Ed-, Act-), Bi (MgO~13-16%), Opx ($Fs_{\sim 50}$), Cpx ($Wo_{\sim 40}$), Aln, Sph	Cpx ($Wo_{\sim 40}$), Phl, San ($Or_{\sim 55}$), ±Lc, ±Pl, ±alter. Ol, Ap, Opq, *Xen (Q, Pl)*	Juvenile: Pumices, dense volcanic fragments, phenocryst clasts - Q, San ($Or_{\sim 60}$), Pl ($An_{\sim 35}$), Bi (MgO ~ 13–14%), ±Ho, ±Cpx, ±Opx; Accidental and accessory lithics: Different volcanic and nonvolcanic fragments	Q, San (Or_{62-78}), Pl (An_{25-45}), Bi (MgO~10-14%), Ho (Mg-, Ed-), Cpx ($Wo_{\sim 45}$), Ap, Sph, Opq, *xenocr. (Ol, Phl?), lamprophyric xenoliths*	Pl (An_{45-77}), Ho (Parg), Opx (En_{60-80}), Cpx ($Wo_{\sim 45}$) with Pig margin (Wo_{2-7}), Sf, Ap, *xenocr. (Phl, Cpx?), numerous xenoliths*
wr: ~20–23 Ma Pl: ~ 21 Ma	wr: ~ 22 Ma Phl = ~ 22	Pu(Bi): ~ 18 Ma (?)	San: ~ 23 Ma	wr: ~ 21–23 Ma

Abbreviations: Pl — Plagioclase, Q — Quartz, Ho — Hornblende (Magnesio /Mg/, Edenitic /Ed/, Actinolitic /Act/, Pargasitic /Parg/), Bi — Biotite, Opx — Ortopyroxene, Cpx — Clinopyroxene, Aln — Allanite, Sph — Sphene, Phl — Phlogopite, Ol — Olivine, San — Sanidine, Lc — Leucite, Ap — Apatite, Opq — Opaque minerals, Xen — Xenocrysts, Pu — Pumice.

complex. The uppermost parts of the chamber consisted of a volatile oversaturated, glassy quartzlatitic-rhyodacitic magma. Due to relaxation numerous volcanic channels were formed; they represented centers of maximal decompression and places of rigorous explosive activity. Hence, the Borac pyroclastic deposits originated after a Plinian explosive phase, which included formation of pyroclastic fall and flow deposits. The strong paroxysm started after an indulgence of the roof of the magma chamber which caused decompression and increased vesiculation of magma. The injection of hot and volatile oversaturated lamprophyric magma as a kind of starting bullet could effectively initiate a strong explosive phase. Some petrographical evidence of magma mingling and mixing processes, for example xenoliths and xenocrysts of lamprophyric magma within the host calc-alkaline rocks and viceversa, largely support this hypothesis. The formation of the pyroclastic fall deposits was succeeded by eruption column collapse(s) thereby causing pyroclastic flows and formation of unwelded ignimbrites. The large masses of pyroclastic material emptied the roof parts of the magma chamber and induced the caldera collapse. However, the subsidence of the Borac caldera was rather irregular because the volcanic activity continued at different places with different intensity and character. Less vesiculated and crystal rich magma gave rise to short lava flows with autoclastic marginal parts as well as lava dome and neck facies. Final andesite to basaltic andesite effusions post-date the formation of the Borac caldera. They overlay earlier products forming thin lava flows predominately represented by hyaloclastic breccia, rarely as coherent rock masses.

References

Cvetkovic V., 1997: Petrostructural and volcanological characteristics of the Borac Eruptive Complex (in Serbian, Abstract in English). *Ph.D. Thesis. Faculty of Mining and Geology*, Beograd, 241.

Cvetkovic V., Pécskay Z. & Prelevic D., 1998: Petrology of lamprophyric rocks of the Miocene Borac Eruptive Complex (Central Serbia, Yugoslavia). *Abstr. Carpathian-Balkan Geol. Assoc. XVI Congress*, 30.08.–2.09.1998. Vienna, 111.

Karamata S., Krstic B., Dimitrijevic M.D., Knezevic V., Dimitrijevic M.N. & Filipovic I., 1994: Terranes between the Adriatic and Carpatho-Balkanian arc. Bull. CVIII de l'Acad. Serbe des Sci. et des Arts, Classe des Sci math. et natur. *Sci. et nature,* 34, 47–66.

Ar/Ar DATING OF HORNBLENDE SAMPLES FROM CARPATHIAN IGNEOUS ROCKS

K. DUDEK[1], K. BUKOWSKI[1], A. WÓJTOWICZ[2], T. DURAKIEWICZ[3] and ST. HAŁAS[3]

[1]University of Mining and Metallurgy, Faculty of Geology, Geophysics and Environmental Protection, 30-059 Kraków, Poland
[2]State Geological Institute, Upper Silesian Branch, SPGRL, 20-418 Lublin, Poland
[3]Maria Curie-Skłodowska University, Institute of Physics, 20-031 Lublin, Poland

Key words: Ar/Ar dating, Carpathians, hornblende, andesite, Ditrau.

Introduction

The Ar/Ar method of radiometric age determination was applied to hornblende samples separated from three different kinds of igneous and pyroclastic rocks occurring in various parts of the Carpathians. These are:

1) Amphibole tuff from the salt mine in Wieliczka near Kraków. The tuff layer occurs in marly clays of the Skawina series, several metres below Badenian evaporites of the Carpathian Foredeep in S Poland. The sample W-10 was taken from gallery 2-323, in the WT-1 level of the salt mine.

2) Amphibole-augite andesite from Mt Wżar in Polish part of the Pieniny Klippen Belt. The sample An-1, representing the younger, second-generation andesite dykes, was collected in the abandoned quarry "Lisi lom", on the S slope of Mt Wżar.

3) Hornblendite from the Ditrau Massif in the Romanian Internal Carpathians. The sample J-2, representing a hornblendite body enclosed in gabbro-dioritic rocks, was collected in the area of prospection for Mo and REE ores in the Jolotca (Orotva) Valley, NW part of the Massif.

All these rocks were earlier dated with K/Ar and others radiometric methods (fide Birkenmajer et al. 1987; Bukowski et al. 1996; Kräutner & Bindea 1995, and references therein). The presented analyses were carried out with the aim of controlling the accuracy of earlier datings as well as gaining additional information on thermal/geological events, registered in changes of Ar isotopes content. It should be noted that the Ar/Ar method does not require determination of potassium content in the analysed material. In the case of hornblende samples, usually containing not more than 1–2 % of K_2O, this enables us to avoid errors related to inexact determination of potassium content.

Method

The ages of the investigated samples have been obtained by the $^{40}Ar/^{39}Ar$ method in the Mass Spectrometry Laboratory at Maria Curie-Skłodowska University. The samples were prepared as follows. Aliquots of mineral species of about 60–80 mg each were wrapped in Al-foil and sealed in two evacuated silica-glass ampoules. The ampoules were irradiated in the reactor Maria in Świerk (Poland). Samples An-1, J-2 and one aliquot of spike (MMhb-1 hornblende), were irradiated for 55 hours, whereas sample W-10 and the second aliquot of MMhb-1 were irradiated for 50 hours. Both sets of samples were irradiated in fast neutron flux of about 10^{13} $cm^{-2}s^{-1}$. After irradiation the ampoules were stored in a lead container for a few weeks in order to reduce their activity. Then the samples were loaded into the extraction-purification line, where the argon was released in steps ranging from 600 to 1300 ^oC. The peaks ^{36}Ar, ^{37}Ar, ^{39}Ar and ^{40}Ar were measured by use of the static-vacuum mass spectrometry for each fraction of argon extracted. The measurements were performed with a mass spectrometer MS10 in which the original magnet was replaced by a stronger one (with B = 0.44 T, produced by AMAG, Kraków) and self-designed control of the electron emission (Durakiewicz 1996).

Results

The ages were calculated according to the formula (McDougall & Harrison 1988):

$$t = \frac{1}{\lambda}\left(J \frac{^{40}Ar^*}{^{39}Ar} + 1 \right)$$

where $\lambda = 5.543 \cdot 10^{-10}\ a^{-1}$ is total decay constant of ^{40}K, J is the neutron fluence parameter, $^{40}Ar^*$ is the quantity of radiogenic argon released and ^{39}Ar is the quantity of argon obtained from the nuclear reaction $^{39}K(n, p)^{39}Ar$. The isotope ratio used in formula (1) was calculated from mass spectrum peak ratios corrected for interfering nuclear reactions:

$$\frac{^{40}Ar^*}{^{39}Ar} = \frac{\left(\frac{^{40}Ar}{^{39}Ar}\right)_m - 295.5\cdot\left(\frac{^{36}Ar}{^{39}Ar}\right)_m + 295.5\cdot\left(\frac{^{37}Ar}{^{39}Ar}\right)_m \cdot a - c}{1-\left(\frac{^{37}Ar}{^{39}Ar}\right)_m \cdot b}$$

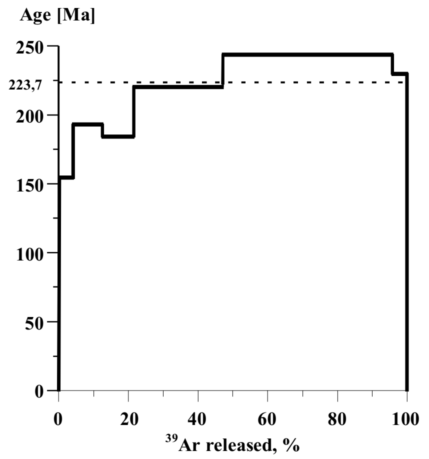

Fig. 1. Age spectrum of sample J-2. Average age is marked by dashed line.

Table 1: Ages of samples W-10 and An-1.

Sample	Fluence	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$ *)	$^{36}Ar/^{39}Ar$	% $^{40}Ar_{rad}$	Age [Ma]
W-10	9.199e-3	1.685	5.490	3.533e-3	38.0	18.0±1.0
An-1	10.230e-3	2.130	4.505	5.972e-3	17.2	13.5±1.0

*) — ratio of ion currents corrected for decay of ^{37}Ar after irradiation ($T_{1/2}$ = 35.5 d).

Table 2: Age of sample J-2. Fluence = 10.230e-3.

Temperature [°C]	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$ *)	$^{36}Ar/^{39}Ar$	% $^{40}Ar_{rad}$	% of total ^{39}Ar released	Age [Ma]
600	130.57	1.664	412.773e-3	6.6	4.11	154.4±3.1
800	16.83	0.358	19.694e-3	65.4	8.21	193.0±2.2
930	13.04	0.599	8.875e-3	79.9	9.18	183.4±2.3
1060	13.51	2.968	3.635e-3	92.1	25.60	220.3±1.7
1200	14.82	2.884	3.193e-3	93.6	48.71	243.6±1.4
1250	17.04	3.525	13.770e-3	76.1	4.19	229.8±2.7

*) — ratio of ion currents corrected for decay of ^{37}Ar after irradiation ($T_{1/2}$ = 35.5 d).

where the measured peak of ^{37}Ar was corrected to radioactive decay, and a, b and c are small corrections for the interfering reactions. The values of a, b and c were taken from Dalrymple & Lanphere (1971). The fluence parameter J was estimated for both ampoules from equation (1) on the basis of known date for hornblende MMhb-1 standard = 520.4 ± 1.7 Ma [a4]. For the first ampoule (55 hours of irradiation) we obtained the fluence parameter J = 10.230e–3 ± 0.10e–3, whereas for the second (50 hours of irradiation) — J = 9.199e–3±0.09e–3.

For samples W-10 and An-1, large amounts of argon were obtained only in one step at temperature about 1200 °C. Large errors occur in the calculated dates for lower temperature steps, therefore they are not presented here.

The age spectrum of the sample J-2 exhibits partial escape of radiogenic argon during its geological history. The results are shown in Fig. 1 and Table 2.

Discussion

The measured age of the Mt Wżar andesite (13.5 ± 1.0 Ma) is in agreement with the isochron age (12.6 Ma) of the youngest andesite dykes, yielded by the K/Ar method (Birkenmajer et al. 1987). Earlier K/Ar datings of the Wieliczka tuff gave results between 11.4 and 28.3 Ma (Bukowski et al. 1996). The obtained Ar/Ar age (18.0 ± 1.0 Ma) coincides with one of those K/Ar values, which can support hypothesis about two or three generations of amphiboles in the pyroclastic material.

The obtained maximal value (243.6 ± 1.4 Ma) of the age of the hornblendite from the Ditrau Massif confirms the opinions that these rocks represent the oldest parts of the whole intrusive complex. It should be noted that the K/Ar ages of hornblendites quoted by Kräutner & Bindea (1995) are somewhat lower (226–237 Ma), and these values are close to the weighted average calculated from our analysis (223.7 Ma). This, together with the whole age spectrum yielded by the analysis, demonstrates high usefulness of the Ar/Ar method of age determination for samples representing rocks of complicated, multiphase history.

Acknowledgements: The study was sponsored by the AGH grants nos. 10.140.112 and 10.140.153.

References

Birkenmajer K., Delitala M.C., Nicoletti M. & Petrucciani C., 1987: K/Ar dating of andesite intrusions (Miocene), Pieniny Klippen Belt, Carpathians. *Bull. Pol. Acad. Sci.,* 35, 1, 11–20.

Bukowski K., Wójtowicz A. & Durakiewicz T., 1996: Preliminary results of K/Ar dating of hornblende from the tuff level in Wieliczka salt mine. 3rd Scientific Session: Dating of Minerals and Rocks. *Book of Abstracts.* UMCS Lublin 24–25 X 1996, 13–18 (in Polish).

Dalrymple G.B. & Lanphere M.A., 1971: $^{40}Ar/^{39}Ar$ technique of K-Ar dating: A comparison with the conventional technique. *Earth Planet. Sci. Lett.,* 12, 300–308.

Durakiewicz T., 1996: Electron emission controller with pulsed heating of filament. *Int. J. Mass Spectr. Ion Proc.*, 156, 31–40.

Kräutner H.G. & Bindea G., 1995: The Ditrau Alkaline Intrusive Complex and its Geological Environment. *Rom. J. Miner.,* 77, 3, 44.

McDougall I. & Harrison T.M., 1988: Geochronology and Thermochronology by $^{40}Ar/^{39}Ar$ Method. *Oxford Univ. Press*, New York.

TATRIC BASEMENT FRAGMENTATION IN THE MALÉ KARPATY MTS. (REACTION EXTENT, P-T AND GARNET CSD DATA)

M. DYDA

Department of Mineralogy and Petrology, Comenius University, Mlynská dolina, 842 15 Bratislava, Slovak Republic; dyda@fns.uniba.sk

Key words: Mineral reactions, dehydration, geothermobarometry, crystal size distribution, garnet, tectonics.

Introduction

The Tatricum basement rocks of the Malé Karpaty Mts. are represented mainly by medium to high-grade paragneisses and other metamorphic rock types which were later intruded by granitoidic rocks of Variscan age.

Earlier tectonic concepts (Cambel 1954) did not favour any Alpidic orogenic activity in the crystalline basement complexes, and faulting and retrograde processes were considered as the only Alpidic developments. However, recent research indicates a complex tectonic and structural development of this "core mountain" (Dyda 1980, 1994; Cambel et al. 1981; Korikovskij et al. 1984; Miklóš 1987), including several superimposed nappe units consisting of pre-Alpine crystalline basement and its Mesozoic cover (Putiš 1991; Plašienka et al. 1991). Due to the tectonic disturbances of the Variscan orogen, it is still unclear whether or not the crystalline complexes in some "core mountains" are in autochthonous positions. The metamorphic process took place during the Variscan era. The biotite pairs Rb/Sr isochron gave an age of 344 ± 3 My (Cambel et al. 1990). This age is considered to date the Bratislava granitoid massif intrusion.

Methods

The mineral assemblages, modal and chemical composition of rocks in periplutonic zones have been used for a detailed study of metamorphic processes in the chosen profiles. The rock protolith mineralogy has been computerised and approximated, giving thus the comparative basis for the volume changes and extent of the dehydration the reactions. The P-T data were obtained on the basis of chemical composition of minerals coexisting in the equilibrium assemblages and correlated with published calibrated mineral equilibria. Crystal size distribution (CSD) has been determined microscopically and the relationships between population density and particular crystal size have enabled us to obtain kinetic characteristics of garnet's is progressive nucleation and growth.

Results

Particular periplutonic zones (Fig. 1) do not offer the metamorphic reaction to argument about a continual dehydration caused by the granitoid rock body. This petrological observation, together with the index mineral appearance confirms disturbances in metamorphic zonality with respect to the metamorphic reaction sequence and the dehydration extent. Should the granitoid body intrusion be complex and/or comagmatic, the development of metamorphic zonation would be more complicated. However, the mineralogical differences in crystal size distribution and morphology of accessory zircon (Dyda 1999a) clearly indicate the polyphase nature of this granitoid intrusion.

Calculated approximative P-T trajectories in the range of 570–650 ^{o}C and 3.5–6.1 Kbar (Dyda 1994) express the first order tectonic motion and represent specific uplift conditions of the particular tectonic blocks (see location in Fig. 1). Some of the samples express uplift trajectories determined dominantly by decompression during cooling while the others may represent more isothermal, probably rapid decompression during the uplift period. The thermodynamic data are in accordance with the index mineral appearances, mineral equilibrium domains, apparent garnet morphology and garnet zonation development. The occurrence of retrograde mineral domains, the microscopic appearance of garnets and their crystal size distribution all confirm the individuality of these tectonic blocks.

Crystal size distribution (CSD) of a garnet population determines garnet nucleation and growth rates, garnet producing reaction overstepping and average residence time of a garnet population. High nucleation rates occur during the contact metamorphic conditions, whereas high-grade regional metamorphism is represented by slow rate of nucleation. Thus the calculated garnet nucleation rates $2.9 * 10^{-8}/cm^3/s$ to $1.0 * 10^{-7}/cm^3/s$ indicate regional metamorphic regime for the analysed paragneissic rock samples. The calculated estimations of garnet growth residence times based on garnet CSD may be bracketed within the time span of 1000–2000 years. Time estimates based on CSD of garnets match well with the metamorphic reaction conditions of $\Delta T = 0.15$ ^{o}C and $\Delta S_{reaction}$= 25 cal/mol/deg. The reaction temperature overstep of about 0.15 ^{o}C is consistent with the regional thermal event and the modelled heating rate trends obtained ($4*10^{-5}$ ^{o}C/year) are in a good agreement with the thermal regional metamorphic regime. The regional recrystallization products are usually subjected to prolonged cooling after thermal culmination and most CSD histograms are significantly modified. Small unstable garnet crystals dissolve and the material is precipitated on larger crystals. The calculated annealing estimates indicate that in the rock samples approximately 0.10–0.53 garnet mass fraction was transferred during the annealing process that lasted after the peak metamorphic conditions were completed. Thus, the apparently quenched mineral aseemblage of a particular tectonic block differs from other periplutonic assemblages which exhibit significantly higher garnet mass transfer estimates. Such numerical values testify more about the regional metamorphic thermal history than the contact recrystallization conditions for samples which now have periplutonic tectonic position.

Discussion

The reconstruction of the Hercynian orogenic cycle in the Western Carpathians is complicated because of the presence of Alpine tectonic overprinting. Fragments of Hercynian structure were incorporated into new Paleoalpine units (Bezák 1993). This development includes the nappe formation, the motions in the area (Plašienka 1989), the structural distinction of two Hercynian and four Alpine deformation stages (Putiš 1987, 1991) and post-Oligocene tectonic motions represented by tectonic block units rotation in the neighbouring areas (Marko et al. 1991).

The crystalline basement experienced two dominant distinct progressive metamorphic events. The earlier, predating the Variscan granitoid magma intrusions, is regionally represented by medium to high pressure and medium to high temperature mineral assemblages, which passed the kyanite stability field, as the unreacted rare kyanite rests are present in some samples (Dyda 1999b).

Orogenic block transport and tectonic structuring occurred during the Variscan era and later during the Alpine movements which destroyed and displaced the polymetamorphic Tatric crystalline basement into new structural positions. The calculated approximate trajectories thus verify different uplift conditions of individ-

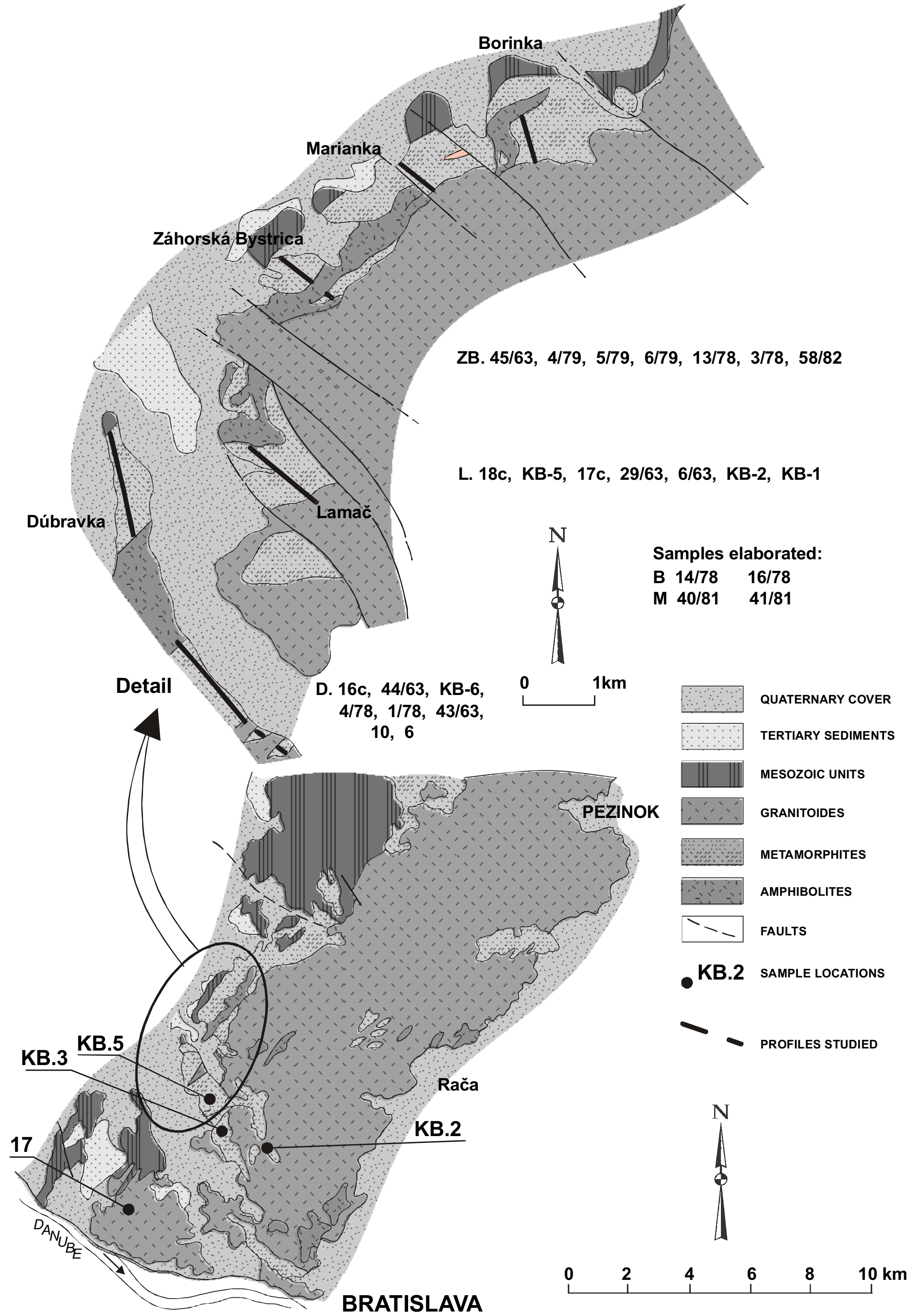

Fig. 1. Geological sketch map of the western part of the Bratislava granitoid massif compiled by Putiš (1987, 1991) with the sample profiles of the author. The profiles chosen for dehydration and volume changes reaction studies were as follows: Dúbravka — (D.), Lamač — (L.), Záhorská Bystrica — (Z.B.), Marianka — (M.), Borinka — (B.). In some profiles rock data of Janák (1980) and Dyda & Mikloš (1993) were used. The solid dots locate the paragneisic rock samples for which P-T and garnet CSD characteristics were determined. Legend for rock types designation: 1 — Quarternary, 2 — Neogene and Paleogene, 3 — Mesozoic, 4 — Granitoides, 5 — Metamorphites, 6 — Amphibolites, 7 — Faults, 8 — Sample location.

ual tectonic blocks during the final stages of the Variscan orogen events. The data obtained represent the arguments in favour of the Tatricum crystalline basement disturbance and its allochthonity in the Malé Karpaty Mts.

References

Bezák V., 1993: Hercynian and Alpine tectogenesis of the western part of Slovenské rudohorie: principal stages of development. In: Rakús M. & Vozár J. (Eds.): *Geodynamický model a hlbinná stavba Západných Karpát.* GÚDŠ Bratislava 1993, 243–247 (in Slovak).

Cambel B., 1954: The problem of crystalline schists between Cajla and Horné Orešany in Malé Karpaty Mts. *Geol. Práce, Spr.*, 1, 16–20 (in Slovak).

Cambel B., Dyda M. & Spišiak J., 1981: Thermodynamic measurement of origin of minerals in metamorphites in the area of crystalline of Malé Karpaty Mts. *Geol. Zbor. Geol. Carpath.*, 32, 6, 745–768.

Cambel B., Kráľ J. & Burchart J., 1990: *Isotope Geochronology of the Western Carpathians, Veda, Vyd. SAV,* Bratislava, 183 (in Slovak).

Dyda M., 1980: Physical properties and temperatures of crystallization of coexisting garnets and biotites from paragneisses of the Little Carpathians. *Geol. Zbor. Geol. Carpath.,* 31, 2, 201–213.

Dyda M., 1994: Geothermobarometric characteristics of some Tatric crystalline basement units (Western Carpathians). *Mitt. Österr. Geol. Gesell.,* 86, 45–59.

Dyda M., 1999a: Magmatic melts and zircon crystallization in the Malé Karpaty Mts. granitoidic rocks. *Miner. slovaca*, 31, 1, 49–60 (in Slovak).

Dyda M., 1999b: Fragmentation of the Bratislava massive metamorphic cover indicated by garnet properties. *Miner. slovaca*, 31, 1, 39–48 (in Slovak).

Dyda M. & Miklóš J., 1993: Periplutonic zones of Bratislava granitoidic massif (Malé Karpaty Mts.): recrystallization, dehydration and volume changes. *Miner. slovaca*, 25, 2, 93–103 (in Slovak).

Janák M., 1980: Study of metapelites in the area of Bratislava granitoidic massive. *Manuscript*, UK Bratislava, 80.

Korikovskij S.P., Cambel, B., Miklóš J. & Janák M., 1984: Metamorphism of Malé Karpaty Mts. crystalline complex. Zonality and relation to granitoidic rocks. *Geol. Zbor. Geol. Carpath.*, 35, 4, 437–462 (in Russian).

Marko F., Fodor L. & Kováč M., 1991: Miocene strike-slip faulting and block rotation in Brezové Karpaty Mts. (Western Carpathians). *Miner. slovaca*, 23, 3, 189–200.

Miklóš J., 1987: Contribution to study of Malé Karpaty Mts. crystalline metamorphic rock protoliths. In: *Sborník príspevku z II. Celostátní konference mineralogu a petrologu.* Brno-Blansko, 93–95 (in Slovak).

Plašienka D., 1989: Structural development of Tatricum in Malé Karpaty Mts. *Miner. slovaca*, 21, 2, 538 (in Slovak).

Plašienka D., Michalík J., Kováč M., Gross P. & Putiš M., 1991 : Paleotectonic evolution of the Malé Karpaty Mts. An overview. *Geol. Zborn. Geol. Carpath.*, 42, 2, 195–208.

Putiš M., 1987: Geology and tectonics of the southwest and northern part of Malé Karpaty Mts. crystalline complex. *Miner. slovaca*, 19, 2, 135–157 (in Slovak).

Putiš M., 1991: Tectonic styles and Late Variscan-Alpine evolution of the Tatric-Veporic crystalline basement in the Western Carpathians. *Zbl. Geol. Paläont.*, l, 1, 181–204.

NATURE AND GEOTECTONIC INTERPRETATION OF THE MELIATA BLUESCHISTS

S.W. FARYAD

Department of Geology and Mineralogy, Technical University Košice, Slovak Republic; faryad@tuke.sk

Key words: blueschists, protolith, metamorphism, Meliata Unit, Western Carpathians.

Introduction

The Meliata Unit (Fig. 1) consisted of very low-grade oceanic-related rocks with slices and tectonic blocks of serpentinites and blueschists, occurs along the southwards dipping and EW-striking zone — the Rožňava tectonic zone in the southern margin of the Western Carpathians. It is wedged between the early and late Paleozoic of the Gemericum to the bottom and unmetamorphosed or very low-grade metamorphosed Triassic platform carbonates of the Silica nappe and Permian evaporite-limestone rocks of the Turna nappe to the tope (Mello et al. 1996). The oceanic rocks are represented by very low-grade metamorphosed deep sea pelagic sediments, radiolarian limestones, turbidites and dismember serpentinite bodies (Mock 1978; Kozur & Mock 1997; Harangi et. al. 1996; Dostály & Józsa 1992; Hovorka et al. 1980; Reichwalder 1973; Arkai & Kovacs 1986; Faryad 1995; Ivan & Kronome 1996). In addition to isolated slices of high-pressure rocks, exposed on surface, the blueschist facies rocks form small blocks within low-grade clastic sediments, even within carbonate-evaporite sequences of the hanging wall Turna nappe. Regarding protoliths, the blueschists formed from rock of different geotectonic positions, however most of them represent former continental crust material. This contribution is aimed to petrological and geochemical composition of blueschist facies rocks that originated by subduction of continental margin and possibly continental rift volcanic material.

Petrology of the blueschists

Besides the most common marbles and phyllites, three varieties of blueschist facies metabasites can be distinguished in the Meliata Unit: 1. Metabasites forming layers in marbles have composition between MORB-arc basalts (Faryad 1995) or E-MORB (Ivan & Kronome 1996). From typical ocean ridge basalts in the literature they differ by major oxide contents (relatively high TiO_2 and P_2O_5, low Na/K and CaO) and enriched LREE. 2. Tectonic blocks of metagabbro, occurring in evaporite melange have composition, comparable with alkaline rocks. These rocks were recently found in the Hungarian territory (Horváth 1997) and investigated also from Slovakia near Bohuňovo. 3. Metabasites associated with micaschists, having relic early Paleozoic mica, are comparable with within plate basalts. The blueschist facies phyllites were formed pelites, psammites and conglomerates. Amphibolite facies basement rocks, overprinted by blueschist facies metamorphism, together with conglomerates are clear evidence for subduction continental crust.

Large number of metamorphic minerals were analysed from the blueschist facies rocks. Beside garnet, hornblende and muscovite in the basement rocks, the pre-blueschist facies relic phases are igneous diopside and richterite. The blueschist facies amphibole group minerals are glaucophane, riebeckite, magnesioriebeckite and winchite. Metamorphic pyroxene are represented by aegirine, omphacite and jadeite. Maximum jadeite content in pyroxene is 70 mol. %. Other blueschist facies minerals are albite, phengite, paragonite, titanite, epidote, zoisite, chloritoid, garnet, locally also actinolite. Some pseudomorphs of phengite, probably after lawsonite, were also found.

Two contrasting mineral assemblages found in metabasites adjacent to marbles are zoisite-muscovite-paragonite (1) and glaucophane-Na-pyroxene-phengite-chlorite-epidote (2). Muscovite, paragonite and zoisite form inclusions in glaucophane. Zoisite is almost rimmed by epidote. Na-pyroxene has strong compositional variation (jadeite-omphacite-aegirine) and seems to be a result of local equilibrium and replacement of an older phase, or may represent jadeite-rich crystals formed at relatively low temperature. Maximum P-T conditions of 1.2 GPa and 450 ^{o}C were estimated for metabasites with omphacite and some metasediments with jadeite and garnet. Mineral zonation and textural relations indicate a prograde clockwise P-T path trajectory from greenschist to blueschist facies, followed by nearly isothermal decompression and exhumation back to greenschist facies conditions. In addition to chlorite, albite, the retrograde phases are actinolite, biotite, mixed-layer philloslicate and white mica.

The metagabbro from evaporite melange has composition of alkaline rocks with normative olivine and nepheline. As relic igne-

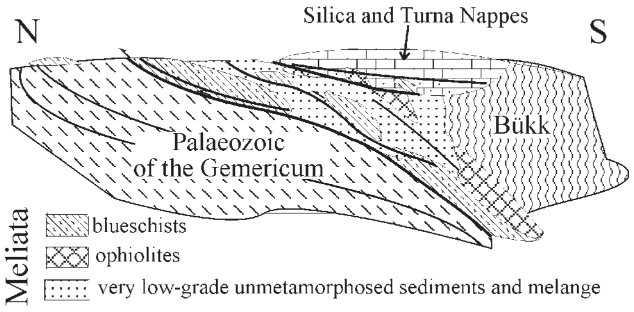

Fig. 1. Schematized cross-section of the Meliata Unit.

ous phase they contain richterite that is usually known from anorogenic environments and interpreted to be formed in mantle conditions. Metamorphic amphibole formed by replacements of richterite are winchite, riebeckite, actinolite and rarely also edenite and magnesiohornblende. Beside riebeckite the blueschist phase is Si-rich phengite in these rocks.

^{40}Ar-^{39}Ar age spectra from metabasites and some phyllites indicate Middle Jurassic high-pressure metamorphism (Maluski et al. 1993; Faryad & Henjes-Kunst 1997; Dallmeyer et al. 1996). Some phengitic white mica from phyllites, however gave Lower Jurassic age of medium to high-pressure metamorphism (Faryad & Henjes-Kunst 1997). Furthermore, a very late overprint probably related to low-temperature deformation processes with maximum ages of ca. 80–90 Ma is supposed.

The exhumation path is characterized by appearance of actinolite and biotite at the expense of glaucophane and garnet probably the result of pressure decrease and introduction of a hydrous fluid. Deformation at greenschist facies conditions is characterized by the presence of quartz, white mica and chlorite in strain shadows of glaucophane porphyroblasts. Structural relations of chloritoid in glaucophane-free phyllites indicate that syntectonic chloritoid porphyroblasts and rosettes that overgrown the foliation S_1 are rotated and surrounded by white mica and quartz. This suggests that S_1 system was used during exhumation of the blueschists. In contrast to some black phyllites where SW-NE striking penetrative cleavage of greenschist facies conditions completely transposed the S_1 foliation, most phyllites are characterized by tighter or discrete cleavage. Beside cleavage the late stage deformation is represented by boudinage which can be well observed in metaconglomerates and some albite-rich carbonate rocks.

Discussion

Based on the presence of slices and tectonic blocks of blueschists and ultramafic rocks within very low-grade matrix or tectonically emplaced into the hanging wall of Permian sequence as well as overthrusted on the Paleozoic of the Gemericum, the Meliata Unit can be interpreted as accretionary complex. It was formed by subduction of rocks different geotectonic positions and reveals a multiple tectonometamorphic evolution. Because of melange character of this unit, it is difficult to classify the rocks into different lithostratigraphic units. The presence of blueschist that formed from basement rocks and clastic sediments is a strong evidence that the Meliata suture was operated adjacent to continental crust. Nature of continental crust material is evidenced also from metabasites associated with micaschists. Although some geochemical correlation of metabasites, forming layers in marbles, with E-MORB basalt can be found, but the small amount of these metabasites (only 5 % from the whole marbles) is not typical feature for MORB basalts. Furthermore the presence of alkaline metagabbro with relic richterite is comparable with ocean island basalts or with rift-related igneous rocks. All these rocks, including serpentinites form slice and blocks within very low-grade matrix formed by deep-see and continental margin sediments that suffered only accretionary metamorphism.

References

Árkai P. & Kovács S., 1986: Diagenesis and Regional metamorphism of Aggtelek-Rudabánya Mountains (Northern Hungary). *Acta Geol. Hung.,* 29, 349–373.

Dallmeyer D., Neubauer F., Handler R., Fritz, H., Muller, W., Pana, D. & Putiš M., 1996: Tectonothermal evolution of the internal Alps and Carpathians: Evidence from ^{40}Ar/^{39}Ar mineral and whole-rock data. *Eclogae Geol. Helv*., 89, 203-227.

Dosztály L. & Józsa S., 1992: Geochronological evolution of Mesozoic formations of Darnó Hill at Recsk on the basis of radiolarians and K-Ar data. *Acta Geol. Hung.,* 35, 371–394.

Faryad S.W., 1995: Phase petrology of mafic blueschists of the Meliata Unit (West Carpathians)-Slovakia. *J. metamorphic Geol.*, 13, 432–448.

Faryad S.W. & Henjes-Kunst F., 1997: K-Ar and Ar-Ar age constraints of the Meliata blueschist facies rocks, the Western Carpathians Slovakia). *Tectonophysics,* 280, 141-156.

Harangi Sz., Szabó Cs., Józsa S., Szoldán Zs., Árva-Sós E., Balla M. & Kubovics I., 1996: Mesozoic igneous suites in Hungary: Implication for genesis and tectonic setting in the northwestern part of Tethys. *Int. Geol. Rev.,* 38, 336–360.

Horváth P., 1997: High-pressure metamorphism and P-T path of the metabasic rocks in the borehole Komjáti-11, Bódva Valley area, NE Hungary. *Acta Mineral. Petrogr.,* Szeged, 38, 151–163.

Hovorka D., Ivan P., Jaroš J., Kratochvíl M., Reichwalder P., Rojkovič I., Spišiak J. & Turanová L., 1985: Ultramafic rocks of the Western Carpathians, Czechoslovakia. *Geol. Inst. Dionýz Štúra*, Bratislava, 258.

Ivan I. & Kronome B., 1996: Predmetamorfný charakter a geodynamické prostredie vzniku vysokotlakovo metamorfovaných bázitov meliatskej jednotky na lokalitách Radzim, Bôrka, Hačava a Rudník. *Miner. slovaca*, 28, 26–37.

Kozur H. & Mock R., 1997: New paleogeographic and tectonic interpretations in the Slovakian Carpathians and their implications for correlations with the Eastern Alps and other parts of the Western Tethys. Part II: Inner Western Carpathians. *Miner. slovaca*, 29, 164–209.

Maluski H., Rajlich P. & Matte Ph., 1993: ^{40}Ar/^{39}Ar dating of the Inner Carpathian Variscan Basement and Alpine mylonitic overprinting, *Tectonophysics,* 223, 313-337.

Mello J., Elečko J., Pristaš J., Reichwalder P., Snopko L., Vass D. & Vozárová A., 1996: Vysvetlivky k geologickej mapy slovenského krasu, mierka 1: 50,000.

Mock R., 1978: Niektoré nové poznatky o južnej časti Západných Karpát. in: Vozár J. (Ed.): Paleogeografický vývoj Západných Karpát. *Geol. Ústav Dionýza Štúra*, Bratislava, 322–341.

Reichwalder P.1973: Geologische Verhältnisse des jüngeren Paläozoikums im Südteil des Zips-Gemerer Erzgebirges. *Západné Karpaty, Geol. Inst. D. Štúr, Bratislava,* 18, 99-139.

MINERAL CHEMISTRY OF METAMORPHIC TARAMITE FROM LOW-GRADE METABASITES IN THE GEMERICUM, WESTERN CARPATHIANS

S.W. FARYAD[1], R. ABART[2], K. ETTINGER[2], F. WALTER[2] and K. JAKABSKÁ[1]

[1]Department of Geology and Mineralogy, Technical University Košice, Slovak Republik; faryad@tuke.sk
[2]Institute of Mineralogy-Crystallography and Petrology, University of Graz, Austria

Key words: taramite, chemistry, metabasites, Rakovec Unit, Western Carpathians.

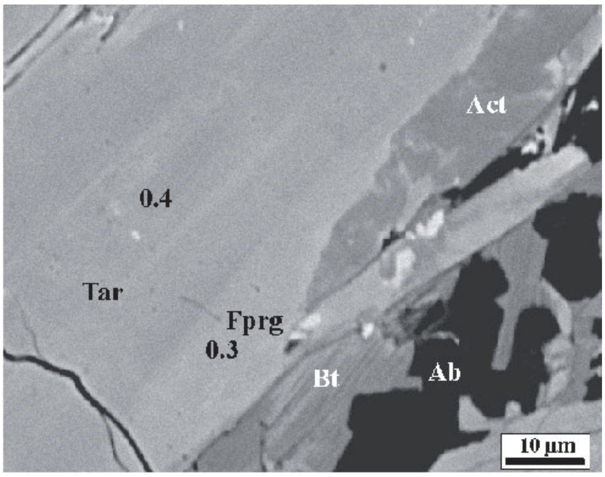

Fig. 1. Back scattered electron image of zoned taramite with rim of ferro-pargasite and actinolite. Number indicate XMg contents in amphibole.

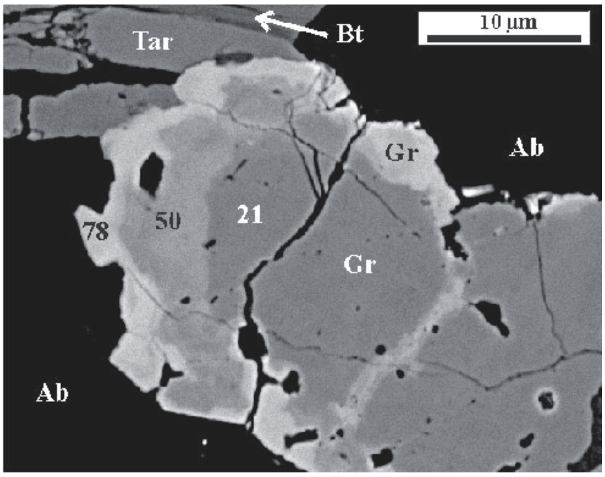

Fig. 3. Grossular-rich garnet rimmed by andradite in psuedomorphs after plagioclase. Numbers show andradite content in garnet.

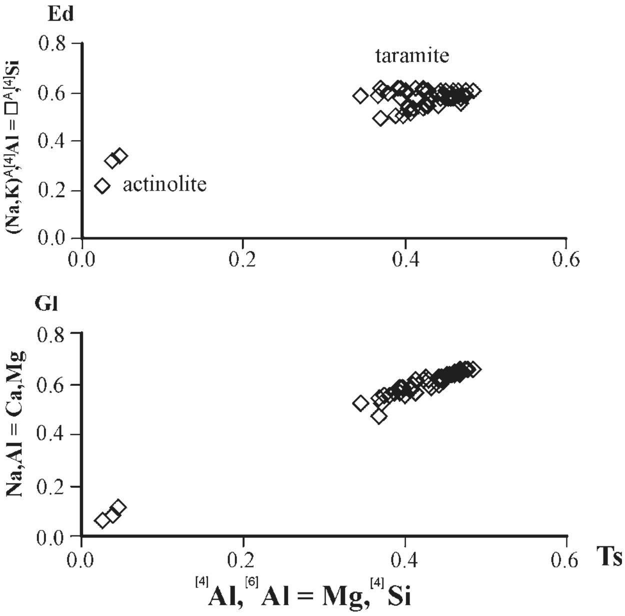

Fig. 2. Plot of edenite and glaucophane substitution vectors against tschermakite for taramite and actinolite rim in Fig. 1.

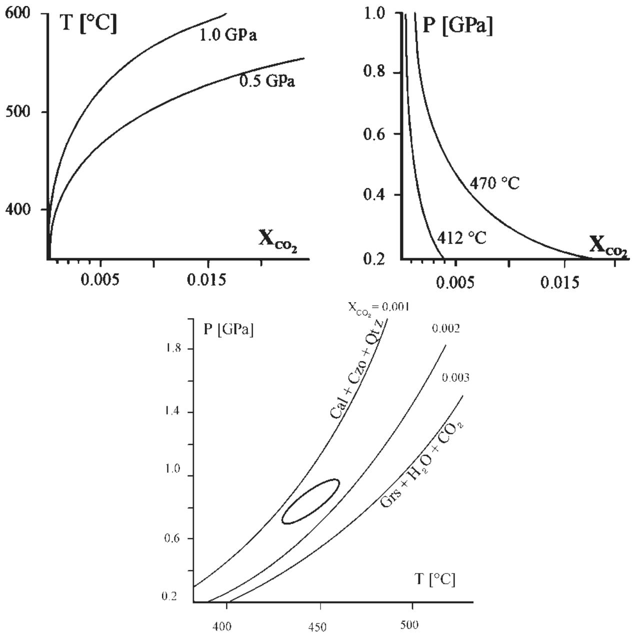

Fig. 4. T-X_{CO_2}, P-X_{CO_2} and P-T diagrams, calculated using TWEEQ program for the equilibrium curves of reaction calcite + clinozoisite + quartz = grossular + H_2O + CO_2. Dashed field shows temperature range of 412–470 °C obtained from isotope thermometry. Ellipse indicates P-T conditions calculated using GeO-Calc program (distributed by Berman 1992) for amphibole, chlorite, epidote, and garnet (Faryad & Bernhardt 1996),

Introduction

A Na-Ca amphibole of taramite composition from the Early Paleozoic greenschist metabasites (the Rakovec Unit) in the Gemericum was petrographically investigated by Hovorka et al. (1988) and Faryad & Bernhardt (1996). It occurs in metabasalts with pillow lava and porphyric structures. The presence of taramite is restricted to a 2 km long area, where metabasalts form lenses in metabasites with greenschist facies assemblage. The most common minerals in the metabasites are albite, epidote, actinolite. Some quartz phyllites intercalated with metabasites contain white mica, albite and chlorite. The metabasalts contain relic igneous pyroxene which is partly replaced by greenish blue taramite, chlorite and biotite, and pseudomorphs of albite after phenocrysts of plagioclase. Taramite following foliation in some metabasalts indicates its syntectonic formation. Small amygdales in the rock are filled by chlorite, epidote, calcite and rarely also by albite. Since taramite is usually know from retrogressed eclogites, we present some textural relations and crystal-chemical data from a prograde taramite in low-grade metabasalt. In order to improve P-T estimate of taramite formation, oxygen isotope data from associated epidote, albite and magnetite are also give in this paper.

Textural relations and mineral chemistry

The taramite is associated with albite, epidote, titanite, magnetite and rarely also with Ca-rich garnet, calcite, biotite, chlorite, quartz and actinolite. It intergrows with biotite and forms inclusions in, or cross-cuts albite. Chlorite is mostly a secondary phase, but some chlorite is in textural equilibrium with biotite and taramite. The accessory actinolite rims taramite (Fig. 1). Some calcite aggregates are cut by taramite. Garnet occurs in pseudomorphs of albite plagioclase and in the matrix with epidote and taramite.

Table 1: Selected microprobe analyses from core (c) and rim (r) of zoned taramite crystal (sample Fg-16/95).

	c	r		c	r
SiO_2	40.81	40.82	Si	6.0759	6.2533
TiO_2	0.36	0.24	Al^{IV}	1.924	1.747
Al_2O_3	16.69	13.50	Al^{VI}	1.005	0.691
Fe_2O_3	4.70	5.10	Ti	0.0409	0.0273
FeO	15.68	18.44	Fe^{3+}	0.5267	0.5885
MnO	0.07	0.23	Fe^{2+}	1.9530	2.3620
MgO	7.08	6.48	Mn	0.0094	0.0297
CaO	7.55	7.51	Mg	1.5713	1.4805
Na_2O	5.53	4.88	Ca	1.2041	1.2333
K_2O	0.48	0.70	Na^{M4}	0.689	0.588
			Na^{A}	0.908	0.863
Total	98.95	97.9	K	0.0921	0.1374

Table 2: X-ray powder diffraction data of taramite from Rakovec (Western Carapathians) and from Mbozi Complex-Tanganyika (W Tanzania).

	Rakovec			Mbozi Complex	
hkl	d obs.	d calc.	I/Io	d	I/Io
110	8.430	8.440	100	8.530	70
040	4.530	4.520	8	4.530	8
220	4.220	4.220	5	-	-
041	3.400	3.400	10	3.420	20
240	3.284	3.283	14	3.290	25
310	3.133	3.134	100	3.150	100
221	2.954	2.954	11	2.970	10
330	2.814	2.814	10	-	-
151	2.160	2.716	26	2.732	50
061	2.601	2.601	10	2.605	30
-351	2.345	2.344	9	2.347	30
351	2.024	2.024	7	2.033	25
461	1.656	1.656	5	1.663	25

Table 3: Unit cell parameters of taramite.

Cel parameter	Mbozi Complex	Rakovec	Δ	Nybö (Norway)
a [Å]	9.952	9.869 (11)	0.083	9.766
b [Å]	18.101	18.104 (41)	-0.003	17.835
c [Å]	5.322	5.314 (1)	0.008	5.312
ß [°]	105.4	104.70 (3)	0.7	104.79
V [$Å^3$]	924.3	918.0	6.1	894.5

The Na-Ca amphibole from metabasalts corresponds to alumino-taramite with Si = 6.0–6.4, X_{Mg} = 0.30–0.45 and X_{Fe}^{3+} = 0.2–0.5 a.f.u. (cations calculated on the basis of 23 oxygens and 15 cations + K). Large amphibole crystals are weakly zoned (Fig. 1) with Fe, Si and Ca increasing, Al and Na decreasing towards rims of grains (Table 1). Some rim analyses have compositions of ferropargasite. Departure from ideal end-member taramite is due mainly to tschermakite substitution, leading to Si totals of 6.0–6.4, Al^{VI} = 0.55–1.1 and Mg 1.6–1.1 a.f.u. and to glaucophane substitution, resulting in an increase of Na from 0.5–0.9 a.f.u. on the M4-site (Fig. 2).

Garnet has average composition of $Grs_{76.6}And_{19.2}Sps_2Alm_2Py_{0.2}$ and it is mostly rimmed by andradite-rich garnet (Fig. 3) of $Grs_{22}And_{78}$. Epidote has average Al_2Fe = $[Fe_{tot}/(-2+Al_{tot}+Fe_{tot})]$ ratio of 0.66, but some grains are rimmed by nearly pure epidote with Al_2Fe = 0.95. The X_{Mg} content in biotite and chlorite are 0.47 and 0.46, respectively.

Mineral structure

Powder diffraction data of studied taramite and from Mbozi Complex-Tanganyika (W Tanzania) are listed in Table 2 and 3. Compared to taramite (mboziite) from syenite (Brock et al. 1964) with V = 924 $Å^3$ and Fe^{2+}+Fe^{3+} = 4.1 a.f.u., the studied taramite from the Rakovec Unit has lower volume V = 919 $Å^3$ that is result from low total Fe = 2.76 a.f.u. content. Taramite from Nybö (Norway) amphibolized eclogite (Ungaretti et al. 1981) is rich in Mg (Fe_t = 0.9 a.f.u) and has lower volume.

Oxygen isotope data and P-T conditions

Because of very small grains of taramite that are mostly intergrown with other phases, oxygen isotope data could only be obtained for magnetite, epidote and albite which are common phases in the metabasalts. Oxygen-isotope thermometry yields metamorphic temperatures for the taramite-bearing assemblage of 412 ± 7 °C for magnetite-albite (using calibration of Chiba et al. 1989) and 470 ± 15 °C for magnetite-epidote (using calibration of Matheus 1994). Such temperature (400–480 °C at 0.7–0.9 GPa) is consistent with that calculated for end-member composition of taramite and other associated phase by using thermodynamic data of Bermann 1988 (Faryad & Bernhardt 1996). Equilibrium curves of garnet-calcite and epidote-bearing reaction (Fig. 4) show maximum pressure ranging from 0.6–1.1 GPa at 450 °C for X_{CO_2} = 0.001–0.003. With increas of X_{CO} in the reaction fluid, decrease pressure conditions.

References

Brock P.W.G., Gellatly D.C. & von Knorring O., 1964: Mboziite, a new sodic amphibole end-member. *Mineral. Mag.,* 33, 1057–1065.

Chiba H., Chacko T., Clayton R.N. & Goldsmith J.R., 1989: Oxygen isotope fractionations involving diopside, forsterite, magnetite and calcite: application to geothermometry. *Geochim. Cosmochim. Acta,* 53, 2985–2995.

Faryad S.W. & Bernhardt H.J., 1996: Taramite-bearing metabasites from Rakovec (Gemeric Unit, the Western Carpathians). *Geol. Carpathica*, 47, 349–357.

Hovorka D., Ivan P., Jilemnická L. & Spišiak J., 1988: Petrology and geochemistry of metabasalts from Rakovec (Paleozoic of Gemeric Group, Inner Western Carpathians), *Geol. Zbor. Geol. Carpath.*, 39, 395–425.

Matthews A., 1994: Oxygen isotope geothermometers for metamorphic rocks. *J. Metamorphic Geol.*, 12, 211–219.

Ungaretti L., Lombardo B., Domeneghetti C. & Rossi G., 1983: Crystal-chemical evolution of amphiboles from eclogitised rocks of the Sasia-Lanzo Zone, Italian Western Alps. *Bull. Miner.*, 106, 645–672.

ORIGIN AND TECTONIC SIGNIFICANCE OF I-TYPE GRANITOIDS IN THE VARISCAN FOLD BELT

F. FINGER

Institut für Mineralogie der Universität Salzburg, Hellbrunnerstraße 34, A-5020 Salzburg, Austria

Key words: I-type granites, Variscan fold belt, magma sources, mantle melts, crustal melts, subduction.

The term "I-type granite" has been introduced by Chappell & White (1974) for a group of metaluminous to weakly peraluminous, relatively sodium-rich granitoids in the Australian Lachlan Fold Belt which were considered to be derived from the melting of (meta)igneous crustal sources. These I-type granites were compared with another group of more peraluminous, sodium-poor "S-type" granites which were considerd to be magmas formed through the melting of metasedimentary crust.

The I-type/S-type classification is now regarded to be one of the most fundamental subdivisions in modern granite petrology. However, in recent literature, the term I-type granite is used in many conflicting ways, which are not always in agreement with the original genetic definition. For example, I-type granitoids are often considered to be genetically related to subduction processes, and to basaltic/andesitic parental melts, which formed in the mantle and evolved to felsic melts through FC (fractional crystallisation) or AFC (assimilation and fractional crystallization). Some schools of thought considered these to be the major I-type granite forming processes with intracrustal igneous sources playing little part (e.g. Pearce et al. 1984).

The central European Variscan fold belt provides a good opportunity to study granitic rocks of variable origin, which all broadly fit the chemical definition of I-type granitoids:

One group of normal-K, I-type granodiorites and tonalites probably contains a mantle component and may have formed via AFC processes. Examples of such magmas are the isotopically primitive Sázava suite in central Bohemia (Janoušek et al. 1995) and the Cetic granitoids in Austria (Finger et al. 1997). In terms of existing geological models, both may be related to pre- or early-collisional subduction process with the closure of Early Variscan oceans.

Other large, essentially late- to post-collisional (Visean) I-type plutons in the central Moldanubian zone of the Variscan fold belt are of the high-K subtype and are crustal-derived. They formed though anatexis of intermediate crust (biotite-plagioclase-gneisses), either by fluid-absent lower crustal melting, following reactions of the type "Bt + Pl + Qtz $\rightarrow$ Melt + Opx $\pm$ Kfs", or by fluid-present mid-crustal melting in the contact aureoles of the intruding lower crustal granites. Major examples of such infra-crustal I-type plutons are the Weinsberg granite and the Schlierengranite in the South Bohemian Batholith (Finger & Clemens 1995).

A third group of Variscan I-type plutons are apparently hybrids that formed through mixing of crustal melts and enriched mantle melts. Such a mixing process has been documented in the Rastenberg pluton in Lower Austria (Gerdes 1997) and in the so-called Durbachite plutons in the Czech Republic (Holub 1977, 1997). These hybrids are also of Visean age. Furthermore, large masses of S-type granite plutons intruded in the Visean period. There is however, still much debate about the tectonic processes, which caused such large scale crustal melting at that time. Models involving magmatic underplating compete with those involving burial and enhanced radioactive heat flow (Gerdes 1997).

An interesting feature of the Variscan orogeny is the occurrence of many late-stage, epizonal, medium- to high-K, I-type granodiorite/tonalite plutons with ages between ca. 310 and 270 Ma (Finger et al. 1997). Chemical and isotope data provide no clear evidence as to whether these plutons contain a mantle component. They may be purely intracrustal I-type melts, derived from a metaigneous lower crustal source. These late I-type plutons are concentrated in the southern (intra-Alpine) domains of the Variscan orogen. This fact has been used as an argument in favour of a late-Variscan northward dipping subduction zone at the southern fold belt flank (Finger & Steyrer 1990). The viability of such a late-Variscan subduction model is a matter of contoversy (Neubauer 1991; Finger & Steyrer 1991; Schaltegger & Corfu 1995; Stampfli 1996). A relationship between the ca. 270 Ma old, I-type granitoids of the Alps and a large-scale mid-Permian rifting event is becoming more and more evident (Neubauer et al. 1999).

In conclusion, it is still unclear, as to what extent the Variscan I-type plutons are actually subduction related. In some cases it appears reasonable to assume a relationship to subduction processes. On the other hand, there are examples where there is no evidence for such a genetic context. It may be the case that all of the major Variscan I-type plutons formed fully independent of any subduction event. Respective melts may have been simply generated due to enhanced temperature conditions in distinct sections and evolution stages of the Variscan orogen.

References

Chappell & White, 1974: Two contrasting granite types. *Pacif. Geol.*, 8, 173–174.

Finger F. & Steyrer H.P., 1990: I-type granitoids as indicators of a late Paleozoic convergent ocean/continent margin along the southern flank of the Central European Variscan Orogen. *Geology,* 18, 1207–1210.

Finger F. & Steyrer H.P., 1991: Replies on „I-type granitoids as indicators of a late Paleozoic convergent ocean-continent margin along the southern flank of the central European Variscan fold orogen". *Geology,* 1245–1248.

Finger F. & Clemens J.D., 1995: Migmatization and „secondary" granitic magmas: effects of emplacement and crystallization of „primary" granitoids in Southern Bohemia, Austria. *Contr. Mineral. Petrology*, 120, 311–326.

Finger F., Roberts M.P., Haunschmid B., Schermaier A. & Steyrer H.P., 1997: Variscan granites of central Europe: their typology, potential sources and tectonothermal relations. *Mineral. and Petrol.,* 61, 67–96.

Gerdes, 1997: Geochemische und thermische Modelle zur Frage der spätorogenen Granitgenese am Beispiel des Südböhmischen Batholiths: Basaltisches Underplating oder Krustenstapelung. *Thesis Univ. Göttingen*, 113.

Holub, 1977: Petrology of inclusions as a key to petrogenesis of the durbachitic rocks from Czechoslovakia. *Tschermaks Mineral. Petrogr. Mitt. TMPM,* 24, 133–150.

Holub, 1997: Ultrapotassic plutonic rocks of the durbachite series in the Bohemian Massif: Petrology, geochemistry and petrogenetic interpretation. *Sbor. Geol. Věd.,* 31, 5–26.

Janousek et al., 1995: Sr-Nd isotopic constraints on the petrogenesis of the Central Bohemian Pluton, Czech Republic. *Geol. Rdsch.,* 84, 520–534.

Neubauer, 1991: Comment on "I-type granitoids as indicators of a late-Palaeozoic convergent ocean-continent margin along the southern flank of the Central European Variscan orogen". *Geology,* 19, 1246.

Neubauer et al., 1999: Pre-Alpine metamorphism of the Eastern Alps. *SMPM*, 79/1 (in print).

Pearce et al., 1984: Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. *J. Petrology*, 25, 956–983.

Schaltegger & Corfu, 1995: Late Variscan "Basin and Range" magmatism and tectonics in the Central Alps: evidence from U-Pb geochronology. *Geodynam. Acta,* 8, 82–89.

Stampfli, 1996: The Intra-Alpine terrain: a Palaeotethyan remnant in the Alpine Variscides. *Eclogae Geol. Helv.,* 89/1, 13–42.

PETROGRAPHY, GEOCHEMISTRY AND AGE OF GRANITIC PEBBLES FROM THE MORAVIAN PART OF THE CARPATHIAN FLYSH

P. HANŽL[1], F. SCHITTER[2], F. FINGER[2], O. KREJČÍ[1], K. BURIÁNKOVÁ[1] and Z. STRÁNÍK[1]

[1]Czech Geological Survey, Leitnerova 22, 658 69 Brno, Czech Republic; hanzl@cgu.cz
[2]University of Salzburg, Hellbrunnstrasse 34, A-5020, Salzburg, Austria

Key words: Carpathian flysh conglomerates, granitic pebbles, geochemistry, monazite age.

The Cretaceous to Oligocene flysh sediments of the Magura and Ždánice Units contain conglomerate layers with relatively abundant granitic pebbles. Nearly 80 pebbles with a size between 5 and 150 cm were sampled from conglomerates of Paleogene age in the area of the Hostýn Hills, Chřiby Hills and Ždánický les Hills (Moravian part of the Carpathian flysh belt). The aim was to characterize the rocks in terms of petrography and geochemistry and to reveal their possible source area. In previous studies on the granitic "exotics" from the Magura flysh (Němcová-Hlobilová 1964; Němcová 1967; Štelcl 1989, 1993a,b) a source area on the consolidated eastern margin (Brunovistulicum) of the Bohemian Massif was supposed. However, this interpretation is in conflict with the results of Soták (1992), who proposed a relationship with the Pieniny Klippen Belt or eastern Dacides, based on a study of sedimentary pebbles and clasts in conglomerates.

The pebbles were analysed in the laboratories of the Czech Geological Survey in Prague using wet chemical analysis for the major oxides. The trace elements were analysed by XRF and INAA. REEs were determined by ICP-OES in 20 selected samples. The age of some of pebbles was roughly constrained through chemical U-Th-Pb monazite dating with a Jeol JX 8600 electron microprobe at Salzburg University (Finger & Helmy 1998).

The pebbles cover a wide compositional spectrum and include diorites, tonalites, granodiorites and granites. However, medium grained granodiorites and granites are clearly predominant (Fig. 1). Biotite is the prevailing mafic mineral. Frequently it is accompanied by muscovite. Hornblende bearing granitoids are rare. Titanite, zircon, apatite, rare allanite, monazite and garnet have been found as accessory minerals. A brittle to brittle-ductile deformation and a low-T alteration as sericitization, epidotization and chloritization are common features of the rocks. Fine fissures are often filled with carbonate, which filled them after the deposition of sediments.

The geochemistry of the granitic pebbles points to calc-alkaline to high-K calc-alkaline granitoid terrain as the source area. Most samples (Fig. 2) have trace element signatures of "volcanic-arc-granites" in the sense of Pearce et al. (1984). Element distributions in Harker´s variation diagrams (e.g. Ba, Sr, Zr) suggest that the pebbles belong to at least two different magmatic suites.

A wide variation of Na_2O/K_2O ratios and high A/CNK index (> 1.5) in some samples could be related to alteration. Relatively low Rb, high Ba and Sr abundances indicate rather I-type than S-type sources, although most samples are quite peraluminous. The K/Rb (100–500) and Rb/Sr (0.1–1) ratios (Figs. 3, 4) are generally in the range of the mature continental crust. The granitoids show a variable enrichment of LREE relative to HREE (La_N/Yb_N=3–28) and a usually pronounced negative Eu anomaly (Eu/Eu*= 0.7–0.3).

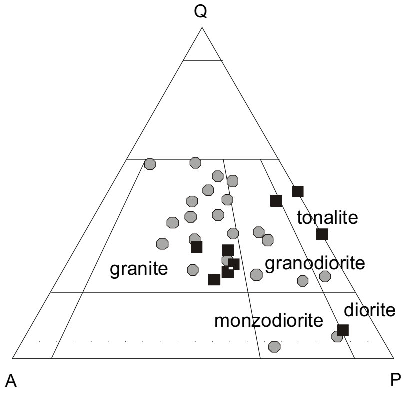

Fig. 1. Classification of granitic pebbles from the Magura (circles) and Ždánice units (squares) in QAP diagram.

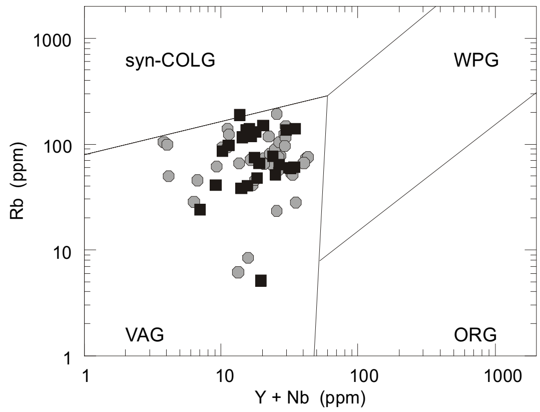

Fig. 2. Rb vs. Y+Nb discrimination diagram according to Pearce et al. (1984). Circles — granitic pebbles from the Magura Unit, squares — granitic pebbles from the Ždánice Unit.

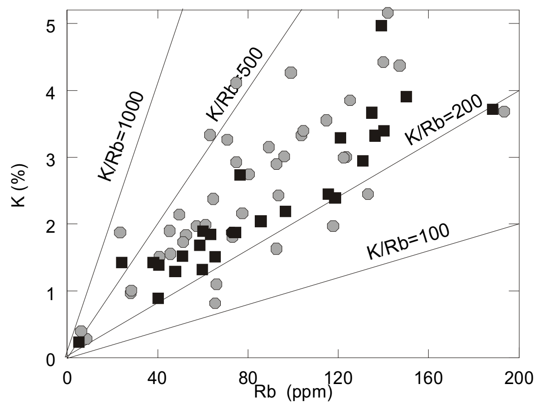

Fig. 3. K vs. Rb diagram for granitic pebbles from the Magura (circles) and Ždánice units (squares).

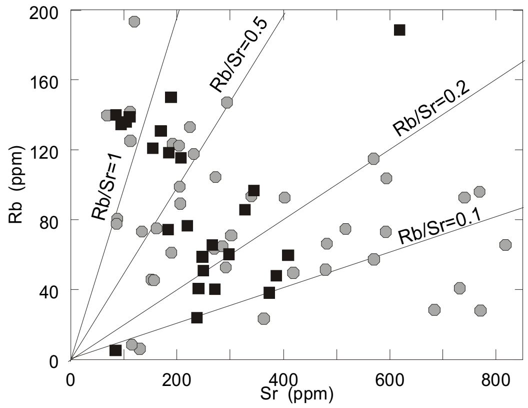

Fig. 4. Rb vs. Sr diagram for granitic pebbles from the Magura (circles) and Ždánice units (squares).

The age of six samples of granitic composition was roughly constrained using the method of chemical U-Th-Pb dating of monazite. Model ages of ca. 330–340 Ma were obtained in four samples from the Magura flysh (Hanžl et al. 1998), two samples from the Ždánice Unit gave somewhat younger model age of 323 ± 21 and 295 ± 18 Ma, respectively. This shows that the granitoid pebbles in the Moravian parts of the Carpathian flysh are derived from a Variscan batholitic terrain. Low grade alterations and low- to medium grade deformation of a substantial part of the granitoids indicate that this granite terrain underwent a tectonothermal overprint in the upper part of crust. No significant petrographic differences were found between granitic pebbles from the Magura and Ždánice units.

Considering the geochemistry and age, a correlation of the studied pebbles with the granites of the Western Carpathians appears to be the most straightforward interpretation.

References

Finger F. & Helmy H.M., 1998: Composition and total-Pb model ages of monazites from high-grade paragneisses in the Abu Swayel area, southern Eastern Desert, Egypt. *Mineral. and Petrol.,* 62, 262–289.

Hanžl P., Finger F., Krejčí O., Schitter F., Buriánková K. & Stráník Z., 1998: Typology and age of granitoid pebbles in Carpathian flysh conglomerates. *Acta Univ. Carolinae*, *Geol.*, 42, 2, 26.

Němcová-Hlobilová J., 1964: Príspevek k exotickým horninám karpatské soustavy. *Acta Univ. Palackianae Olomouc, Geogr. Geol.*, 17.

Němcová J., 1967: Príspevek k exotickým horninám Chřibu. *Acta Univ. Palackianae Olomouc*, *Geogr. Geol.,* 26, 61–81.

Pearce J.A., Harris N.B.W. & Tindle A.G., 1984: Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. *J. Petrology,* 25, 956–983.

Soták J., 1992: Evolution of the Western Carpathian suture zone — principal geotectonic events. *Geol. Carpathica*, 43, 6, 355–362.

Štelcl J., 1989: Petrology and geochemistry of granitoid exotitcs of the Rača Unit of the Magura Flysh. *Geol. Zbor. Geol. Carpath*., 40, 6, 735–748.

Štelcl J., 1993a: Hlavní výsledky petrologického výzkumu krystalinických exotik racanské jednotky magurského flyše. *Zpr. Geol. Výzk. v r. 1991,* 137–138.

Štelcl J., 1993b: Výsledky petrochemického studia krystalinických exotik račanské jednotky magurského flyše. *Zpr. Geol. Výzk*., 138–140.

SIHLA – IPEĽ GRANITE STAGE IN THE FRAME OF THE WESTERN CARPATHIAN HERCYNIAN MAGMATISM

Ľ. HRAŠKO[1], V. BEZÁK[1] and I. BROSKA[2]

[1] Geological Survey of the Slovak Republic, Mlynská dolina 1, 815 00 Bratislava, Slovak Republic
[2] Geological Institute of the Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; geolbros@savba.savba.sk

Key words: granite, Veporic unit, magmatism.

Introduction

The role of Neohercynian orogenic stage (340–260 Ma — Bezák et al. 1997) is really significant in Hercynian granite magmatism of the Western Carpathians. After main collisional events and formation of the principal Hercynian lithotectonic units (Bezák et al. 1997) the tectonic movements dominated on deep faults systems in transtensional and extensional regimes. This resulted in the generation of deep seated melts in lower crust and their interaction with a mantle material. The Neohercynian stage followed the Mesohercynian collisional orogenic events producing S-type granites in the Western Carpathians (Petrík & Kohút 1997). The most complete granite products of Hercynian orogenesis are observable in the basement of the Veporic tectonic unit.

The main products of the Hercynian magmatism in Veporic unit formerly described in literature as Sihla (Zoubek 1936; Broska & Petrík 1993a) and Ipeľ granite types (Krist 1979) are characterized in this contribution from point of view their positions, relationships, mineralogy as well as their geochemistry. Both mentioned granite types belong, on the basis of recent mapping, and geochronology, evidently to the Upper Carboniferous magmatic stage with U/Pb zircon ages 303 ± 2 Ma for the Sihla type (Bibikova et al. 1990) and 305 ± 5 Ma for the Ipeľ type (Michalko et al. 1998). Recent mapping and studies of these granitoids in this area shed new light on their complicated intrusive and differentiated history (Fig. 1).

Characterization of main Upper Carboniferous magmatic bodies in the Veporic unit

1. Sihla type of granites are outcropped in the middle part of the Kráľova Hoľa zone. They form larger or smaller, mostly high angle bodies prolongated in direction NE-SW and less E-W. They are plagioclase bearing granitoids with more then 50 % of plagioclase of oligoclase-andesine composition, sometimes in porphyritic development and with rather low biotite Fe/(Fe+Mg) ratio: 0.48–0.52. Magnetite, allanite and titanite along with zircons of S_{12}, S_{13}, S_{16} subtypes after Pupin classsification (Pupin 1980) are their typical accessory assemblage. The Sihla type s.l. comprise also further petrographic types: tonalite–granodiorite with tiny idiomorphic K-feldspar and pink porphyritic K-feldspar. Sporadically, the Sihla granite is cut by leucocratic granite with high amount of plagioclases (above 50 %). Their contacts against the older granites (hybridic or Vepor porphyritic granite type — Fig. 1) or metamorphic rocks in the wallrocks have a clear intrusive character. Locally mafic microgranular magmatic enclaves derived from basic magmas are abundant in the Sihla granitoids. A typical loaf shape of enclave resulted from their com-mingling with the Sihla felsic melt. The mafic enclaves may represent remnants of a basic magmatic body which may triggered melting of the lower crust (Broska & Petrík 1993b).

2. *Porphyritic Ipeľ type* of granitoids with characteristic pink K-feldspar are widespread in the SE part of the Kráľová Hoľa zone in contact with high metamorphic mantle. They are represented by granodiorites, granites, less frequently by tonalites, with similar accessory mineral associations as in the Sihla type granite (zircons S_3, S_4, S_{12}, S_{13}). The Fe/(Fe+Mg) biotite ratio is around 0.52 and Rb/Sr ratio around 0.2.

3. The Sihla type of granites are intruded through steeply inclined structures by fine- and medium-grained *leucocratic granites*, partly also porphyritic tonalites and granodiorites (locally granites). This type has SiO_2 contents varying between 69 to 73 % and common xenoliths of crustal origin (gneisses and migmatites captured in granitoids forming several dm ellipsoid xenoliths in size) and quartz xenocrysts up to 3–5 cm in size. The presence of included biotite-bearing rocks (up to 5 cm) are considered as restite. The chaotic distribution of xenoliths indicate high melt mobility. More acid types of granites contain monazite instead typical allanite, titanite usually is missing; the amount of xenotime is very low. Zircon morphology S_1, S_2, S_6, S_7 is characteristic of a lower temperature S-type granitoids. The Fe/(Fe+Mg) biotite ratio varied from 0.51–0.52 but their diferenciated porphyritic varietes which are on the rims of intrusive bodies have this ratio — 0.55 up to 0.57.

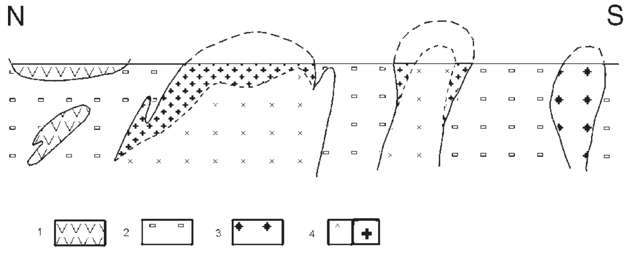

Fig. 1. Idealized profile in the central part of the Veporicum (Sihla-Látky). Sihla s.l. and Ipeľ s.l. type granitoide bodies — spatial relationship. Explanations: 1 — older Vepor porphyric type, 2 — Sihla s.l. type, 3 — Ipeľ type l. stage; coarse-grained, 4 — Ipeľ type (a — non-porphyric, b — porphyric) 2. stage.

Genetic aspects of Upper Carboniferous magmatism

The genesis of Sihla-Ipeľ magmatic stage of the Upper Carboniferous Hercynian orogenesis is proposed after following time succesion with decreasing age:

1. The oldest Sihla granite type s.l. was probably, a lower crustal subduction-related melting, product triggered by underplated mafic melts with contribution of mantle.

2. Younger intrusions of coarse-grained porphyric K-feldspar granites (Ipeľ type) is derived from lower crustal melting of a K-rich immature protholith of "arcose nature".

3. The youngest fine(medium)grained leucocratic granodiorite (2. stage of Ipeľ type) originated by lower temperature melting of quartz-plagioclases (± biotite ± muscovite) lower crustal protholith. The middle crustal contamination of gneisses and migmatites are suggested.

References

Bezák V., Jacko S., Janák M., Ledru P., Petrík I. & Vozárová A., 1997: Main Hercynian lithotectonic units of the Western Carpathians. In: Grecula P. et al. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca, Monogr.,* 261–268.

Bezák V. & Hraško Ľ., 1998: Geological map 1: 25 000 and explanation to the geological map, sheet 36 414 — (Hriňová). *Manuscript, GSSR* (in Slovak).

Bibikova E.V., Korikovsky S.P., Putiš M., Broska I., Goltzman Y.V. & Arakeliants M.M., 1990: U-Pb, Rb-Sr a K-Ar dating of Sihla tonalites of the Vepor pluton (Western Carpathians Mts.). *Geol. Zbor. Geol. Carpathica,* 41, 4, 427–436.

Broska I. & Petrík I., 1993a: Tonalite of the Sihla type s.l.: a Variscan plagioclase-biotite I-type magmatite in the Western Carpathians. *Miner. slovaca*, 25, 1, 23–28 (in Slovak with English resume).

Broska I. & Petrík I., 1993b: Magmatic enclaves in the Variscan granites of the Western Carpathians. *Miner. slovaca*, 25, 2, 104–108 (in Slovak with English resume).

Hraško Ľ., Bezák V., Broska I. & Dublan L., 1998: Geological map and explanation to the geological map, 1: 25,000, sheet 36-412 (Drábsko). *Manuscript, GSSR* (in Slovak).

Krist E., 1979: Granitoid rocks of the south-western part of the Veporide crystalline complex. *Geol. Zbor. Geol. Carpath.,* 30, 2, 157–179.

Michalko J., Bezák V., Kráľ J., Huhma H., Mäntäri I., Vaasjoki M., Broska I., Hraško Ľ. & Határ J., 1998: U/Pb zircon data from the Veporic granitoids (West Carpathians). *Krystallinikum*, Praha, 24, 91–104.

Petrík I. & Kohút M., 1997: The evolution of granitoid magmatism during the Hercynian orogen in the Western Carpathians. In: Grecula P. et al. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca, Monogr.,* 235–252.

Pupin J.P., 1980: Zircon and Granite Petrology. *Contr. Mineral. Petrology*, 73, 207–220.

Zoubek V., 1936: Remarks to the crystalline rocks of the Western Carpathians. *Věst. Stát. geol. Úst. Čs. Rep.*, 12, 1–2, 207–239 (in Czech).

NITROGEN-BEARING FLUIDS IN VARISCAN BASEMENT OF THE TATRA MOUNTAINS, WESTERN CARPATHIANS

V. HURAI[1] and M. JANÁK[2]

[1]Geological Survey of the Slovak Republic, Mlynská dolina 1, 817 04 Bratislava, Slovak Republic; hurai@gssr.sk
[2]Geological Institute of the Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; geolmjan@savba.savba.sk

Key words: nitrogen-bearing fluids, high-pressure metamorphism, Western Carpathians, Variscan orogeny.

Introduction

Main objective of this study are fluid inclusions in Qtz-Plg segregations formed during partial melting of metapelites and metabasites of the Tatra Mts. More detailed fluid inclusion study was initiated by the previous findings of compositionally contrasting nitrogen-bearing carbonic fluids associated with the Grt-Cpx metabasites (retrograded eclogite relics) and metapelite migmatites (Janák et al. 1996, 1999).

Geological setting and metamorphic P-T paths

The Tatra Mts. expose highly condensed crustal section of the pre-Alpine basement with inverted metamorphic zonation resulting from top-to-the-south-east overthrusting of high grade unit onto a lower-grade one (Fritz et al. 1992; Janák 1994).

Eclogite facies metamorphism in the basement rocks is indicated by diopside-plagioclase symplectites in the garnet-clinopyroxene metabasites exhumed along Variscan ductile shear zone and preserved as boudins in banded amphibolites at the base of the upper unit. The symplectites have formed by breakdown of primary Jd_{36} omphacite crystallized at pressures in excess of 15 kbar. Hence, the garnet-clinopyroxene metabasites represent retrogressed eclogites, which have undergone extensive re-equilibration at ~650 °C, 8–10 kbar during their exhumation (Janák et al. 1996).

Metapelitic migmatites contain kyanite and staurolite relics, recording initial prograde *PT* conditions at ~600 °C and 9–10 kbar. During decompression, the metapelites have undergone incipient melting by breakdown of muscovite and biotite at >700–750 °C, 6–12 kbar, producing a garnet-bearing granitic (tonalite-trondhjemite) leucosome (Janák et al. 1999). It is inferred that these rocks represent strongly retrogressed granulites transformed during exhumation along a clockwise *P-T* path.

The garnet-bearing trondhjemitic leucosomes occur also in banded amphibolites, containing the eclogite relics. Origin and formation *P-T* conditions of the leucosomes in the metabasites are not fully elucidated (dehydration melting of amphibole ?).

Fluid inclusions

Primary, negative crystal-shaped gaseous inclusions, 2–15 μm in diameter, monophase at room *T,* occur in weakly deformed or undeformed quartz grains of the migmatite leucosome. The gaseous inclusions closely associate with inclusions of graphite, halite-containing brines (32–36 wt. % NaCl) and carbonate spherules. The latter two phases prevail in pseudo-secondary systems.

On cooling, the gaseous inclusions exhibit H-type (homogenization to liquid as the last phase transition) and S-type phase transformations (melting in the liquid as the last phase transition), indicating the presence of high-density (liquid-like) gaseous mixture.

While the S-type inclusions are present merely in the garnet-containing leucosome of banded amphibolite, the H-type inclusions typically occur in the metapelite migmatites. Both types of gaseous inclusions have been encountered in the garnet-absent leucosome, intersecting the banded amphibolite. Raman microprobe and microthermometry data indicate N_2-CO_2 composition (X_{N2} = 0.8–1.0) for the S-type inclusions hosted by banded amphibolites and the CO_2-N_2 composition (X_{N2} = 0.05–0.5) for metapelite migmatites (Fig. 1).

Conclusions

Retrogression of high-grade metapelites and metabasites in the Tatra Mts. was accompanied by CO_2- and N_2-rich fluids similar in terms of composition and density with those commonly attributed

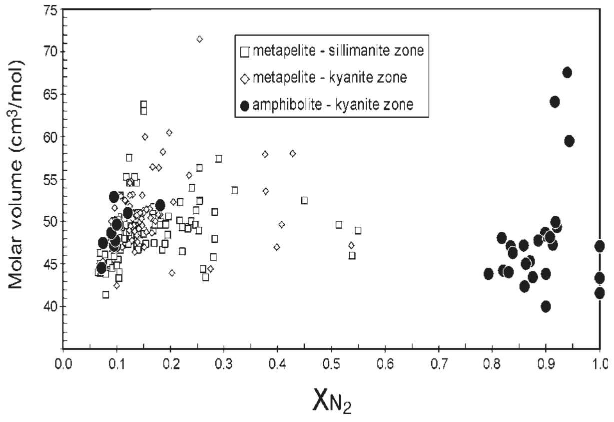

Fig. 1. Compositions and molar volumes of the gaseous inclusions in leucosome of the metapelite and metabasite migmatites.

to granulite and eclogite facies metamorphism. The high density gases and associated brines have been probably exsolved during solidification of the tonalite-trondhjemite melts under amphibolite facies conditions. Composition of the gaseous phase was controlled by lithology (nitrogen-dominated in metabasites and CO_2-dominated in metapelites), thus indicating a local fluid source. An immiscible carbonate liquid corresponding chemically to Fe-Mg carbonate in ferroan metapelites or calcite in metabasites, has coexisted with the N_2-bearing carbonic fluids and brines in the temperature range of 750–550 °C.

References

Fritz H., Neubauer F., Janák M. & Putiš M., 1992: Variscan mid-crustal thrusting in the Carpathians II: Kinematics and fabric evolution of the Western Tatra basement. *Terra Abstr., Suppl.* 2, 4, 24.

Janák M., 1994: Variscan uplift of the crystalline basent, Tatra Mts., Central Western Carpathians: evidence from $^{40}Ar/^{39}Ar$ laser probe dating of biotite and P-T-t paths. *Geol. Carpathica*, 45, 293–300.

Janák M., O´Brien P.J., Hurai V. & Reutel C., 1996: Metamorphic evolution and fluid composition of garnet-clinopyroxene metabasites from the Tatra Mountains, Western Carpathians. *Lithos,* 39, 57–79.

Janák M., Hurai V., Ludhová L., O´Brien P.J. & Horn E.E., 1999: Dehydration melting and devolatilization during exhumation of high-grade metapelites: the Tatra Mountains, Western Carpathians. *J. Metamorphic Geol.,* 17 (in press).

DECIPHERING ALPINE AND PRE-ALPINE METAMORPHISM IN THE WESTERN CARPATHIANS: AN OVERVIEW

M. JANÁK and D. PLAŠIENKA

Geological Institute, Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; geolmjan@savba.savba.sk; geol.plas@savba.savba.sk

Key words: Western Carpathians, Variscan and Alpine orogeny, metamorphism, tectonics.

Introduction

The Western Carpathians are a classical area of the Alpine orogeny in the northern branch of European Tethys zone, which evolved as a complex subduction-collisional orogenic belt during Meso-Cenozoic times. The Central Western Carpathians, located between the Meliatic and Penninic-Vahic oceanic sutures, originated by shortening and stacking of a continental domain which was related to Europe during the Late Paleozoic and Triassic and to Apulia during the Cretaceous and Tertiary. The crustal-scale basement/cover sheets (Tatric, Veporic and Gemeric units) and detached cover nappes (Fatric, Hronic and Silicic systems) build up the Slovakocarpathian tectonic system that is well correlable with the Austroalpine units of the Alps. All these units originated during the Cretaceous collisional shortening and progradational stacking of the lower plate following the closure of the Meliatic ocean by the Late Jurassic (Plašienka et al. 1997). Pre-Mesozoic rock complexes belong to the Variscan basement that developed generally in the inner and southern part of the outer Variscan zones of the Central Europe (Neubauer & von Raumer 1993).

In spite of long-lasting geological research in the Western Carpathians, many aspects of the Alpine vs. pre-Alpine metamorphic processes remain problematic, especially in the polymetamorphosed basement complexes (e.g. Krist et al. 1992).

Pre-Alpine metamorphism

Pre-Alpine metamorphic evolution of the Western Carpathians is best recorded in the Tatric unit because of only very low Alpine metamorphic overprint. The crystalline basement of the Tatric unit is exposed in several core mountains, where the pre-Mesozoic metamorphic rocks and granitoids are overlain by unmetamorphosed or only anchizonal Mesozoic and Cenozoic sedimentary cover and nappes.

In the Tatra Mts., the basement consists of two superimposed tectonic units — lower and upper, differing in lithology and metamorphic grade (Janák 1994). The lower unit is composed of the micaschists. Kyanite-, staurolite- and fibrolitic sillimanite-bearing metapelites alternate with quartz-rich metapsammites, indicating former flysch sediments. The upper unit is composed of migmatized orthogneisses, paragneisses and amphibolites, intruded by a sheet-like granitoid pluton. Variscan deformation D_1 is demonstrated by top-to-the-south thrusting of the upper unit onto the lower one, whereas deformation D_2 with dextral, or top-to-the-east shearing indicates Variscan orogen-parallel extension.

Pre-Alpine, high-pressure eclogite facies metamorphism is indicated by diopside-plagioclase symplectites in the garnet-clinopyroxene metabasite, sporadically preserved as boudins in layered amphibolites (so called leptyno-amphibolite complex sensu Hovorka et al. 1994, 1997) in several core mountains of the Tatric unit (Janák et al. 1997). The symplectites have formed by breakdown of primary omphacite crystallized at pressures in excess of 15 kbar. Hence, the garnet-clinopyroxene metabasites represent retrogressed eclogites, which have undergone extensive re-equilibration in the amphibolite facies during their exhumation (Janák et al. 1996). The fluid inclusions contain nitrogen-dominated, water-absent fluid similar to that reported from typical high-pressure eclogites. Trondhjemitic to tonalitic leucosomes in layered amphibolites developed by dehydration melting of amphibole.

Pre-Alpine high-grade metamorphism is also well documented in the migmatized metapelites. As example, in the Tatra Mts. migmatites, the high-grade metamorphism resulted in dehydration-melting of muscovite and biotite. Decompression from *HP/HT* conditions (>750–800 °C; 8–10 kbar) produced the cordierite at *LP/HT* stage (ca. 750 °C; 4 kbar). The CO_2-rich fluid was generated during the interaction of melt-derived water with metapelite graphite. Consequently, peak metamorphism granulite facies stage has been strongly obliterated by retrogression at subsolidus conditions (Janák et al. 1999a). It is inferred that many migmatites in the Western Carpathian´s basement (e.g. Malá Fatra Mts., Low Tatra Mts., Strážovské vrchy Mts.) represent strongly retrogressed granulites transformed during exhumation along a clockwise *P-T* path.

In the southern Central Carpathian zones, the *P-T* conditions of Variscan metamorphism decrease down to the greenschist facies in Early Paleozoic volcanosedimentary complexes of the Gemericum (e.g. Faryad 1997). Metaophiolites of the Rakovec group have been affected partly by amphibolite facies overprint (e.g. Hovorka & Spišiak 1997). Early Paleozoic complexes of the Inner Carparthians (Szendrö, Upponyi and Bükk Mts.) show no Variscan and only Alpine metamorphism in the anchi and epizonal conditions (Árkai 1983).

Geochronological data indicate a polystage metamorphic evolution of the Western Carpathian basement during Paleozoic time (e.g. Cambel et al. 1990), and no pre-Cambrian metamorphic events. However, based on Sm-Nd, Rb-Sr and Pb-Pb isotopic data (Poller et al. 1999a) old, mostly Proterozoic source material of mixed crustal/mantle origin was involved in the Variscan orogenic processes. Single zircon grain age determinations from the Tatra Mts. (Poller et al. 1999b) indicate that granitic orthogneiss precursor crystallised at ca. 405 Ma. A multistage granitoid magmatism from Late Devonian to Late Carboniferous (360–315 Ma) is recorded by single grain zircon dating in the Tatra Mts. (Poller et al. 1999b). Cooling ages of micas from granitoids and migmatites range from 330 to 300 Ma, according to ^{40}Ar-^{39}Ar data (e.g. Maluski et al. 1993; Janák 1994).

It is inferred that pre-Alpine high-pressure metamorphism resulted from northwards subduction in the Eo-Variscan (Upper Silurian–Lower Devonian) time. Following collision and lithospheric thickening in the Upper Devonian–Lower Carboniferous time, delamination, or detachment of lithospheric mantle with concomitant asthenospheric upwelling provided the heat for generation of granulite facies migmatites and granitoid melts by dehydration melting of micas and amphibole. This might trigger rapid decompression and uplift of the thermally weakened lower crust. Overturn of a hot deep crustal root with synkinematic granitoid intrusions could produce an extra advected heat sorce at mid crustal levels, resulting in low-pressure metamorphism. Exhumation was accomplished by south-wergent thrusting and nappe stacking, balanced by concomitant west-east (orogen-parallel) extension.

Alpine metamorphism

In the Central Carpathian zones, mainly in the Veporic unit, widespread sedimentary, metamorphic, magmatic and structural records point to a primary importance of the early Alpine (generally Cretaceous) orogenic processes in the generation of the nappe edifice. Both the pre-Alpine basement and Late Paleozoic-Mesozoic cover have been affected by regional metamorphism, closely related to Early Cretaceous, neo-Cimmerian collision after suturing of the Meliata ocean.

The high-pressure/low-temperature metamorphism of the Meliatic unit was reported by Faryad (1995), its Late Jurassic age is constrained by the $^{40}Ar/^{39}Ar$ thermochronology (Maluski et al. 1993; Dallmeyer et al. 1996; Faryad & Henjes-Kunst 1997).

The Veporic unit exhibits a core complex structure, composed of the Variscan basement and Permian-Mesozoic cover rocks. The penetrative Alpine deformation records the post-collisional extension coeval with the Late Cretaceous top-to-the-E exhumation. The exhumation was orogen-parallel and roughly perpendicular to the earlier, top-to-the-N, NW thrusting. Present data from the Veporic unit (e.g. Vrána 1964, 1966; Vozárová 1990; Méres & Hovorka 1991; Plašienka et al. 1999) led to conclusions on significant extent and intensity of Alpine metamorphism. The Permo-Mesozoic cover has been metamorphosed in the anchi- and epizone, up to the greenschist facies conditions (Plašienka et al. 1989; Korikovsky et al. 1997a,b; Lupták et al. 1999). In the basement, Alpine mineral assemblages of the pelitic schists, metabasites and mylonitic Variscan granites show synkinematic growth with respect to penetrative Alpine deformation. Alpine metamorphism in the SE part shows increasing grade from chloritoid to staurolite + chlorite and staurolite + biotite zone (Janák et al. 1999b). Thermobarometric data yield up to ~600 °C and 10 kbar conditions of Alpine metamorphism (Plašienka et al. 1999). Metamorphic *P-T* paths are generally clockwise, consistent with collisional crustal thickening and burial, followed by relatively fast, extension-related exhumation. Locally, the intrusion of the Cretaceous Rochovce granite emplaced into the shear zone along the Veporic-Gemeric boundary, caused low-pressure metamorphism with formation of cordierite and andalusite (e.g. Korikovsky et al. 1986; Vozárová 1990).

Timing of Alpine metamorphism in Veporicum is constrained by available Ar-Ar data (Maluski et al. 1993; Dallmeyer et al. 1996; Kováčik et al. 1996), recording Cretaceous cooling ages of around 110 Ma (amphiboles) and 90–80 Ma (micas). Alpine granite magmatism is documented by intrusion of Rochovce granite at 81 Ma (U-Pb on zircons, Hraško et al. 1995).

Alpine metamorphism in the Gemeric unit is hardly to distinguish. Greenschist facies metamorphism with occurrence of chloritoid in the metasediments has been suggested (Varga 1973). Chloritoid occurs in the Upper Carboniferous graphitic schists of the Ochtiná Formation in the north-Gemeric zones (Korikovsky et al. 1997a). Alpine metamorphic overprint is indicated by presence of grossular-rich garnet in the Gemeric granites (Faryad & Dianiška 1989). Ar-Ar dating of white mica from the Late Paleozoic metasediments gave 86–101 Ma age (Dallmeyer et al. 1996).

Alpine metamorphic signatures in the Tatric and Fatric units were only preliminarily studied in some regions, mostly by means of illite-crystalinity (Plašienka et al. 1989 for the Krížna nappe; Plašienka et al. 1993 for the Tatric cover in the Malé Karpaty Mts.). These studies indicate anchizonal conditions of the Alpine recrystallization, structural control and scarce K-Ar isotopic ages point to its Late Cretaceous age.

Present data suggest that Alpine metamorphism in the southern part of the Central Western Carpathians is related to: (1) deep burial due to continental collision thickening during the latest Jurassic–Early Cretaceous, after the closure of the Meliata ocean, (2) Early Cretaceous thermal equilibration at lower to mid-crustal levels, (3) rapid Late Cretaceous exhumation due to orogen-parallel extension. In the northern Tatric zones, the very low-grade metamorphism is connected with the thin-skinned nappe stacking. These Eoalpine events, however, were in other Alpine orogenic segments, e.g. in the Alps, mostly overprinted and obliterated by a later, Tertiary collision.

Tertiary Alpine metamorphism is well documented in the east Slovak Neogene Basin, floored by the Penninic-like series of the Iňačovce-Krichevo unit (Soták et al. 1994). Metamorphism in the Iňačovce-Krichevo unit reached the temperature up to 350–400 °C. In metabasalts and metapelites, earlier higher pressure (7–8 kbar) greenschist to blueschist metamorphism has been overprinted by decompression in greenschist facies conditions. The Middle Eocene black phyllites and metasandstones reflect the higher anchizone to lower epizone metamorphism (Biroň et al. 1993).

References

Árkai P., 1983: Very low-grade Alpine regional metamorphism of the Paleozoic and Mesozoic Formations of the Bükkicum, NE Hungary. *Acta Geol. Hung.,* 26, 83–101.

Biroň A., Soták J., Spišiak J., Bebej J. & Magyar J., 1993: Bündnerschiefer metasediments of the Pozdišovce-Iňačovce unit: metamorphic petrology data. *Geol. Carpathica*, 44, 65–79.

Cambel B., Kráľ J. & Burchart J., 1990: Isotopic geochronology of the Western Carpathian crystalline complex with catalogue of data. *Veda,* Bratislava, 1–183.

Dallmeyer R.D., Neubauer F., Handler R., Fritz H., Müller W., Pana D. & Putiš M., 1996: Tectonothermal evolution of the internal Alps and Carpathians: Evidence from 40Ar/39Ar mineral and whole-rock data. *Eclogae geol. Helv.,* 89, 203–227.

Faryad S.W., 1995: Phase petrology and P-T conditions of mafic blueschists from the Meliata unit, Western Carpathians, Slovakia. *J. Metamorphic Geol.,* 13, 701–714.

Faryad S.W. & Dianiška I., 1989: Garnet from granitoids of the Spišsko-gemerské rudohorie Mts., *Geol. Zbor. Geol. Carpath.,* 40, 715–734.

Faryad S.W. & Henjes-Kunst F., 1997: K-Ar and Ar-Ar age constraints of the Meliata blueschist facies rocks, the Western Carpathians (Slovakia). *Tectonophysics,* 280, 141–156.

Faryad S.W., 1997: Metamorphic petrology of the Early Paleozoic low-grade rocks in the Gemericum. In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca-Monogr.*, 309–314.

Hovorka D. & Spišiak J., 1997: Medium-grade metamorphics of the Gemeric unit (central Western Carpathians). In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca-Monogr.*, 315–332.

Hovorka D., Méres Š. & Ivan P., 1994: Pre-Alpine WesternCarpathians basement complexes: lithology and geodynamic setting. *Mitt. Österr. Geol. Gesell.,* 86, 33–44.

Hovorka D., Ivan P. & Méres Š., 1997: Leptyno-amphibolite complex of the Western Carpathians: its definition, extent and genetical problems. In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca-Monogr.,* 269–280.

Hraško Ľ., Michalko J., Határ J., Hók J., Vaasjoki M. & Kotov A.B., 1995: Upper Cretaceous granite in Western Carpathian region, *Terra Abstr., 7,* 1, 271.

Janák M., 1994: Variscan uplift of the crystalline basement, Tatra Mts., Central Western Carpathians: evidence from ^{40}Ar/^{39}Ar laser probe dating of biotite and P-T-t paths. *Geol. Carpathica*, 45, 293–300.

Janák M., O'Brien P.J., Hurai V. & Reutel Ch., 1996. Metamorphic evolution and fluid composition of garnet-clinopyroxene amphibolites from the Tatra Mts., Western Carpathians. *Lithos,* 39, 57–79.

Janák M., Hovorka D., Hurai V., Lupták B., Méres Š., Pitoňák P. & Spišiak J., 1997: High-pressure relics in the metabasites of the Western Carpathians pre-Alpine basement. In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca-Monogr.*, 301–308.

Janák M., Hurai V., Ludhová L., O´Brien P.J. & Horn E.E., 1999a: Dehydration melting and devolatilization of high-grade metapelites: the Tatra Mountains, Western Carpathians. *J. Metamorphic Geol.* (in print).

Janák M., Plašienka D., Lupták B. & Frey M., 1999b: Alpine metamorphism in the Central Western carpathians: relationships to Cretaceous continental collision. *J. Conf. Abstr.*, 4, 1, *EUG 10 Abstract,* 74.

Korikovky S.P., Janák M. & Boronikhin V.A., 1986: Geothermometry and phase relations during recrystallization of the garnet-micaschists to cordierite-hornfelses in the aureole of the Rochovce granite (Slovak ore Mountains, area Rochovce–Chyžné), *Geol. Zbor. Geol. Carpath.,* 37, 5, 607–633.

Korikovsky S.P., Putiš M. & Plašienka D., 1997a: Cretaceous low-grade metamorphism of the veporic and North-Gemeric Zones: a result of collisional tectonics in the central Western Carpathians. In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca-Monogr.*, 107–130.

Korikovsky S.P., Putiš M., Plašienka D., Jacko S. & Ďurovič V., 1997b: Cretaceous very low-grade metamorphism of the Infratatric and Supratatric domains: an indicator of thin-skinned tectonics in the central Western Carpathians. In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological evolution of the Western Carpathians. *Miner. slovaca-Monogr.*, 89–106.

Kováčik M., Kráľ J. & Maluski H., 1996: Metamorphic rocks in the Southern Veporicum basement: their Alpine metamorphism and thermochronologic evolution, *Miner. slovaca,* 28, 185–202.

Krist E., Korikovsky S.P., Putiš M., Janák M. & Faryad S.W., 1992: Geology and Petrology of Metamorphic rocks of the Western Carpathian Crystalline Complexes. *Comenius Univ. Press*, Bratislava, 324.

Lupták B., Frey M., Janák M., Schmidt S.Th. & Plašienka D., 1999: Alpine metamorphism of the Mesozoic cover rocks from the Veporic unit, Western Carpathians: X-ray and EMP study of white mica and chlorite (this volume).

Maluski H., Rajlich P. & Matte P., 1993: ^{40}Ar-^{39}Ar dating of the Inner Carpathians Variscan basement and Alpine mylonitic overprint. *Tectonophysics,* 223, 313–337.

Méres Š. & Hovorka D., 1991: Geochemistry and metamorphic evolution of the Kohút crystalline complex mica schists (Western Carpathians), *Acta Geol. Geogr. Univ. Comen., Geol.*, 47, 15–66.

Neubauer F. & von Raumer J.F., 1993: The alpine basement-linkage between variscides and East-Mediterranean mountain belts. In: von Raumer J.F. & Neubauer F. (Eds.): Pre-Mesozoic geology of the Alps. Springer Berlin – Heidelberg – New York, 641–663.

Petrík I. & Kohút M., 1997. The evolution of granitoid magmatism during the Hercynian orogen in the Western carpathians. In: Grecula P., Hovorka D. & Putiš M. (eds.): Geological Evolution of the Western Carpathians. *Miner. slovaca-Monogr.*, 235–260.

Plašienka D., Janák M., Hacura A. & Vrbatovič P., 1989: Prvé údaje o kryštalinite illitu z alpínsky metamorfovaných hornín veporika, Centrálne Západné Karpaty. *Miner. slovaca*, 21, 43–51.

Plašienka D., Korikovsky S.P. & Hacura A., 1993: Anchizonal Alpine metamorphism of Tatric cover sediments in the malé Karpaty Mts., Western Carpathians. *Geol. Carpathica*, 44, 365–371.

Plašienka D., Grecula P., Putiš M., Hovorka D. & Kováč M., 1997: Evolution and structure of the Western Carpathians: an overview. In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological Evolution of the Western Carpathians. *Miner. slovaca-Monogr.,* 1–24.

Plašienka D., Janák M., Lupták B., Milovský R. & Frey M., 1999: Kinematics and metamorphism of a Cretaceous core complex: the Veporic unit of the Western Carpathians. *Phys. Chem. Earth* (in print).

Poller U., Todt W., Janák M. & Kohút M., 1999a: The relationships between the Variscides and the Western Carpathians basement: new Sr, Nd and Pb-Pb isotope data from the Tatra Mountains (this volume).

Poller U., Todt W., Janák M. & Kohút M., 1999b: The geodynamic evolution of the Tatra Mountains constrained by new U-Pb single zircon data on orthogneisses, migmatites and granitoids (this volume).

Soták J., Spišiak J. & Biroň A., 1994: Metamorphic Sequences with "Bündnerschiefer" lithology in the Pre-Neogene Basement of the East Slovakian Basin. *Mitt. Österr. Geol. Gesell.,* 86, 111–120.

Varga I., 1973: Mineral assemblages of regional metamorphism and its zonation in the Spišsko-gemerské rudohorie Mts. *Miner. slovaca,* 5, 115–134.

Vozárová A., 1990: Development of metamorphism in the Gemeric/Veporic contact zone (Western Carpathians). *Geol. Zbor. Geol. Carpath.*, 41, 475–502.

Vrána S., 1964: Chloritoid and kyanite zone of Alpine metamorphism on the boundary of the Veporides and the Gemerides, *Krystalinikum*, 2, 125–143.

Vrána S., 1966: Alpidische metamorphose der Granitoide und der Foederata serie im Mittelteil der Veporiden, *Zbor. Geol. Vied, Západ. Karpaty*, 6, 29–84.

Na-RICH AND HIGH-Al GRANITOID MAGMA IN THE TATRA MTS. (WESTERN CARPATHIANS, SLOVAKIA) — MELTING OF THE AMPHIBOLITIC LOWER CRUST?

M. KOHÚT[1], U. POLLER[2], W. TODT[2],
P. NABELEK[3] and M. JANÁK[4]

[1]Geological Survey of Slovakia, Mlynská dolina 1, 817 04 Bratislava, Slovak Republic
[2]Max-Planck Institute für Chemie, Postfach 3060, D-55020 Mainz, Germany; poller@mpch-mainz.mpg.de
[3]Department of Geological Sciences, University of Missouri, Columbia, MO 65211, USA
[4] Geological Institute, Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; geolmjan@savba.savba.sk

Key words: Western Carpathians, Tatra Mountains, granitoid rocks, geochemistry, isotopes.

Introduction

Granitoid rocks form the major part of the pre-Alpine basement in the Western Carpathians and are generally considered to be products of the Hercynian (Variscan) orogeny. Abundant granitoid plutons — so called "core mountains" are exposed in the Tatric Unit of the Central Western Carpathians (CWC), the Tatra Mts. being one of the best examples. The crystalline basement of the Tatra Mts. is composed of pre-Mesozoic metamorphic and granitoid rocks. This crystalline massif is overlain by the sedimentary Mesozoic cover sequence as well as the Krížna and the Choč nappes. The Tatra pluton is composed of several granite types, ranging from two-mica granite to biotite tonalite, locally with dioritic enclaves (MME). The sheet-like intrusion is accommodated in the upper tectonic unit of an inverted metamorphic sequence (Janák 1994). Structural studies (Fritz et al. 1992) suggest generally SE kinematic of the Variscan thrusting, involving granitoids in the hanging wall. The granitoids show a generally dextral E-W shear sense in ductile conditions. In contrast to the Variscan, the Alpine deformation is mostly in brittle or brittle-ductile conditions.

Results

The following granitoid types have been distinguished (Kohút & Janák 1994, 1996) in the Tatra Mts. pluton:

a) *biotite-amphibole quartz diorite* — form only small bodies and lenses of several meters to tens of meters within the muscovite-biotite granodiorite of the common Tatra type and/or small MME in the High Tatra type. Mineral composition: plagioclase, hornblende, biotite, quartz ± K-feldspar; accessories are: apatite, sphene, zircon, magnetite, pyrite, ilmenite ± allanite.

b) *biotite tonalite and muscovite-biotite granodiorite "High Tatra type s.s."* (HTT) — appears only in the central High Tatra Mts. part. The wall-rock xenoliths are common in this type, but in the Velická Valley one can observe xenoliths and mafic enclaves (MME) at the same location. Mineral composition: plagioclase, quartz ± K-feldspar, biotite ± muscovite, accessory phases: apatite, allanite, zircon, magnetite, monazite, ilmenite ± magnetite.

c) *biotite and muscovite-biotite granodiorite to granite, slightly porphyric (the common Tatra type* — CTGD) — this type is dominant in both parts of the massif — Western and High Tatra Mts. Except structure and colour, the common Tatra type is very similar to the previous type, with respect of a different mineral composition.

d) *porphyric granites to granodiorites with pinkish K-feldspar "Goryczkowa type"* (GG) — appears mainly in the northernmost part of the Tatra pluton. Porphyric fabric with phenocrysts of up to 2–3 cm in size, make this type well distinguishable.

Petrographically, all Tatra granites represent common crustal anatectic rocks with magmatic muscovite. Biotite is the dominant Fe-Mg mafic mineral, whereas hornblende occurs only rarely in the dioritic enclaves. Prevalence of plagioclases (An_{20-45}) over K-feldspar is a typical attribute. The mafic enclaves (MME) and metamorphic xenoliths can be observed together, enclosed in rather homogenous granite rocks. Silica contents of the Tatra Mts. granitic rocks vary from 64 to 74 wt. %. The dominance of Na_2O over K_2O is a common feature for HTT and CTGD (K_2O/Na_2O = 0.3–0.9), while GG have this ratio rather equal (K_2O/Na_2O = 0.85–1.05). All granite types are Na_2O-rich (average Na_2O > 4 wt. %). Majority of granitic rocks have high Al_2O_3 content (15–18.5 wt. %). Although there is some scatter, Al_2O_3, TiO_2, MgO, FeO^t and CaO show good negative correlation with respects to increasing SiO_2, i.e. from more basic HTT to more fractionated GG. When plotted on in Ab-An-Or normative diagram of Barker (1979), they show the trend from tonalite through tight group of throndhjemite field, to the granite field (GG). These granites have high Sr content > 300 p.p.m., Sr/Y ratio is generally elevated (25–80), Y is low, usually less than 15 p.p.m., Rb/Sr ratio is low 0.05–0.35, and content of Sc is as low as 1.5–4 p.p.m. Chondrite-normalized patterns of the REE exhibit absence or slight negative Eu anomalies and uniform fractionated distribution trends for various granite types (La_N/Yb_N = 42–15), Yb_N = 2.5–4. Granitic rocks of the Tatra Mts. on a K_2O vs. SiO_2 plot represent medium to high-potassium calk-alkaline (throndhjemite & monzonite) series of magmatic rocks. Metaluminous to peraluminous (subaluminous) character is reflected by ASI = 0.8–1.2 and/or Peacock's index ALI = 62–63, correspond rather to the Ca-series, which is also supported by the high content of CaO reaching up to 3.5 wt. % in some samples. Many of these characteristics match those of volcanic rocks considered to be the product of slab melting (Peacock et al. 1994 and Drummond et al. 1996). Compositions are also similar, particularly in trace element and REE content, to Archaean high-Al throndhjemites, which have been also attributed to oceanic slab melting (Martin 1986). There are two environments in which this kind of magma might originated: 1) melting of subducted oceanic lithosphere; 2) melting of lower continental crust — thicker than 40 km.

The Rb/Sr ratio range from 0.05 to 0.35 indicates an inhomogeneous weakly differentiated lower crustal source, possibly influenced by the mantle input, albeit initial strontium ratios I_{sr} = 0.705–0.706 call for lower crust dominance. The $\varepsilon Nd_{(340)}$ values for common tonalites–granites varying from –1.6 to –3.4, are similar to other European crustal granites (Liew & Hofmann 1988; Janoušek et al. 1995). Dioritic enclaves having positive $\varepsilon Nd_{(340)}$ = 0.7 to 1.9 values indicate possible mantle origin. Neodymium crustal index NCI = 0.53 to 0.72 for HTT–CTGD is similar to other CWC granites, whereas dioritic enclaves having lower NCI = 0.22–0.40 suggest strong lower crustal/mantle influence. In the $\varepsilon Nd_{(i)}$ vs. $\varepsilon Sr_{(i)}$ diagram, the Tatra Mts. granite samples plot within the quadrant IV. Noteworthy is their slight Sm enrichment and/or affinity to quadrant I, which is not typical for ordinary igneous rocks but common for old mafic lower crust, amphibolitic in composition (Keay et al. 1997; Poller et al. 1998, 1999b). In contrast, dioritic enclaves plot in quadrant II, just in the "mantle array". The apparent crustal residence ages, indicated by Nd model ages t_{DM} = 1.20 ~ 1.49 Ga are similar to the other segments of the Hercynian Europe, however dioritic enclaves with t_{DM} = 975 Ma have only analogue in Limousin, Massif Central (Shaw et al. 1993). Particularly low $\delta^{18}O$ values 8.5–9.6 ‰ SMOW are consistent with infracrustal metaluminous whole-rock chemistry. The whole rock Pb-Pb isotopic compositions show only small spread in $^{206}Pb/^{204}Pb$ = 18.30–18.85 and $^{207}Pb/^{204}Pb$ = 15.58–15.66 ratios (Poller et al. 1999b), indicating rather lower continental crust and/or EM II. There is a number of models which explain the origin of enriched mantle. In general term, enrichment is likely related to subduction, when crustal material (pelagic sediments) is injected into the mantle, or it reflects altered oceanic crust and/or old subcontinental lithosphere. Keeping in mind all these facts, amphibolitic lower crust could be the best source candidate for the Tatra Mts. granite rocks, although parental magma was subsequently mixed with some amount of "purely crustal melt", and/or modified by an AFC mechanism. New single grain zircon U/Pb data (Poller et al. 1999a,c) suggest emplacement of the Western Tatra Mts. granites between 369 ± 19 Ma and 347 ± 14 Ma, or 335 ± 4 Ma to 314 ± 4 Ma in the High Tatra Mts.

Discussion and conclusions

There is commonly accepted opinion that Variscan granitoid rocks were produced as the consequence of intra-continental subduction and collisional processes (Kohút & Janák 1994; Petrík et al. 1994; Petrík & Kohút 1997) in the Western Carpathians. All geochemical features suggest that the Tatra Mts. granitoids are analogous to VAG (CAG) granites, related to subduction processes. However, metamorphic, sedimentary and structural data rule out this scenario and suggest continental collision processes. These collisional processes are documented by large-scale deep-crustal thrusting, involving granite melt. Field evidence together with *P-T-t* paths (generally clockwise) suggest tectonic inversion of metamorphism in the Western Tatra Mountains by thrusting of hot, highly metamorphosed slab (migmatites, gneisses, amphibolites) over a cooler parautochthon dominated by mica schists (Janák 1994; Janák et al. 1999). The continental crust had to be sufficiently thick (more than 40 km) and hot enough for melting of amphibolitic source. Alternatively, an additional heat flux from underplated, mantle-derived magma would be required. Nevertheless, there is no evidence (geologic or isotopic) for a direct input of juvenile basaltic magma from mantle, which could possibly trigger anatectic and/or differentiation processes during the Variscan orogen in the Western Carpathians. Dehydration (fluid absent) melting is considered to explain the generation of the Tatra Mts. anatectic melts both in metapelites (Janák et al. 1999) as well as metabasites (Janák et al. 1995, 1996). Melt fraction of 25–40 vol. % (critical melt fraction percentage CMFP) is required to enable melt segregation and transportation via fractures to the middle/upper crustal levels. It is assumed that pristine amphibolite-derived melt was mixed with metapelite-derived one being partly assimilated by host rocks. The heterogeneity of the above distinguished granitoid types seems to be more controlled by melting of a heterogeneous — vertically-zoned lower crustal source, than by the differentiation of magma itself in the magma chamber. Granitoid rocks of the Tatra Mts. represent several batches of magma, which ascended and collected, forming the pluton in the upper part of the middle crust, without isotope and substance homogenisation. Final emplacement of granitoids in the Tatra Mts. was probably syn-kinematic with respect to the Variscan uplift of the upper tectonic unit, as shown by the presence of granitoids solely within the overthrust unit.

References

Barker F., 1979: Trondhjemite: definition, environment and hypotheses of origin. In: Barker F. (Ed.): Trondhjemites, dacites and related rocks. *Elsevier,* Amsterdam, 1–2.

Drummond M.S., Defant W.J. & Kapezhinskas P.K., 1996: Petrogenesis of slab-derived throndhjemite-tonalite-dacite/adakite magmas. *Trans. Roy. Soc. Edinb., Earth Sci.*, 87, 205–215.

Fritz H., Neubauer F., Janák M. & Putiš M.,1992: Variscan mid-crustal thrusting in the Carpathians. Part II: Kinematics and fabric evolution of the Western Tatra basement. ALCAPA Graz, *Terra abstr.,* 4, 24.

Janák M., 1994: Variscan Uplift of the Crystalline Basement Tatra Mts., Central Western Carpathians: Evidence from $^{40}Ar/^{39}Ar$ Laser Probe Dating of Biotite and P-T-t Paths. *Geol. Carpathica*, 45, 5, 293–300.

Janák M., Pitoňák P., Spišiak J., Petrík I. & O'Brien P.J., 1995: Tonalitic-trondhjemitic melts in the Western Carpathian basement: implications for partial melting of amphibolite and the differentiation of the lower crust. *Abstr., suppl.* 1, 7, 375.

Janák M., O'Brien P.J., Hurai V. & Reutel Ch., 1996: Metamorphic evolution and fluid composition of garnet-clinopyroxene amphibolites from the Tatra Mts., Western Carpathians. *Lithos,* 39, 57–79.

Janák M., Hurai V., Ludhová L., O´Brien P.J. & Horn E.E., 1999: Dehydration melting and devolatilization of high-grade metapelites: the Tatra Mountains, Western Carpathians. *J. Metamorphic Geol.* (in print).

Janoušek V., Rogers G. & Bowes D.R., 1995: Sr-Nd isotopic constraints on the petrogenesis of the Central Bohemian pluton, Czech republic. *Geol. Rdsch.,* 84, 520–534.

Keay S., Colins W.J. & McCulloch M.T., 1997: A three-component Sr-Nd isotopic mixing model for granitoid genesis, Lachlan fold belt, eastern Australia. *Geology,* 25, 4, 307–310.

Kohút M. & Janák M., 1994: Granitoids of the Tatra Mts., Western Carpathians: Field Relations and Petrogenetic Implications. *Geol. Carpathica*, 45, 5, 301–311.

Kohút M. & Janák M., 1996: Granitoids of the Tatra Mts., Western Carpathians (Slovakia) — Geochemistry and Petrology. *Pol. Miner. Soc., Spec. Pap.,* Krakow, 7, 39–41.

Liew T.C. & Hofmann A.W., 1988: Precambrian crustal components, plutonic associations, plate environment of the Hercynian fold belt of Central Europe: Indications from a Nd and Sr isotopic study. *Contr. Mineral. Petrology*, 98, 129–138.

Martin H., 1986: Effect of steeper Archean geothermal gradient on geochemistry of subduction zone magmas. *Geology,* 14, 753–756.

Peacock S.M., Rushmer T. & Thompson A.B., 1994: Partial melting of subducting oceanic crust. *Earth Planet. Sci. Lett.,* 121, 227–244.

Petrík I., Broska I. & Uher P., 1994: Evolution of the Western Carpathian granite magmatism: age, source rock, geotectonic setting and relation to the variscan structure. *Geol. Carpathica*, 45, 5, 283–291.

Petrík I. & Kohút M., 1997: The Evolution of Granitoid Magmatism During the Hercynian Orogen in the Western Carpathians. In: Grecula P. et al. (Eds.) 1997: Geological evolution of the Western Carpathians). *Miner. slovaca*, 235–252.

Poller U., Todt W., Janák M. & Kohút M., 1999a: U-Pb single zircon TIMS data in combination with CL and EMP studies on granitoids of the High Tatra Mts. (Slovakia). *J. Conf. Abstr.,* 4, 669.

Poller U., Todt W., Janák M. & Kohút M., 1999b: The relationship between the Variscides and the Western Carpathians: New Sr, Nd, and Pb-Pb isotope data of the Tatra Mountains. *Geol. Carpathica* (this volume).

Poller U., Todt W., Kohút M. & Janák M., 1999c: The geodynamic evolution of the Tatra Mountains considered by new U-Pb single zircon data on orthogneisses, migmatites and granitoids. *Geol. Carpathica* (this volume).

Shaw A., Downes H. & Thirlwall M.F., 1993: The quartz diorites of Limousin: elemental and isotopic evidence of Devono-Carboniferous subduction in the Hercynian belt of the French Massif Central. *Chem. Geol.,* 107, 1–18.

PALEOVOLCANIC RECONSTRUCTION OF NEOGENE VOLCANOES IN THE CENTRAL SLOVAKIA VOLCANIC FIELD

V. KONEČNÝ and J. LEXA

Geological Survey of the Slovak Republic, Mlynská dolina 1,
817 04 Bratislava, Slovak Republic; lexa@gssr.sk

Key words: volcanoes, volcanic rocks, Neogene, central Slovakia, Western Carpathians.

Introduction

The paleovolcanic reconstruction is the final step in the geological analysis of ancient volcanic terrains. It is an actualistic interpretation of the observed lithological phenomena using active volcanoes and observations of ongoing volcanic eruptions and their products as the analogue. Methods of systematic paleovolcanic reconstruction have been introduced by the Russian school (e.g. Maleyev 1959, 1963) and were subsequently also applied to analysis of the Neogene volcanics in Slovakia (e.g. Konečný 1971; Konečný in Vass et al. 1979; Lexa 1971; Bezák & Lexa 1979). Further advance in methodology has been reached by careful studies of recent eruptions and their products (e.g. Fisher & Schmincke 1984) and the use of sedimentological approaches (e.g. Cas & Wright 1991).

A successful paleovolcanic reconstruction depends on the availability of relevant lithological information. The first attempt to carry out systematic paleovolcanic reconstruction of the Central Slovakia Volcanic Field accompanied compilation of the geological map on the scale 1:100,000 (Konečný & Lexa 1984), which was the first map based on lithostratigraphic principles, using the results of facies analysis of volcanoclastic rocks. Since that time a systematic detailed geological mapping including volcanological and lithological studies has created a much better basis for the paleovolcanic reconstruction in the form of geological maps on the scale 1:25,000 (unpublished maps and Open file reports in the Geological survey archive) and published geological maps on the scale 1:50,000 (Dublan et al. 1997; Konečný et al. 1978, 1998a, 1998b; Lexa et al. 1998; Šimon et al. 1997). Essential results of paleovolcanic reconstruction have been discussed by Konečný et al. (1995) and Konečný & Lexa (1995).

General setting

The Central Slovakia Volcanic Field (CSVF), of the Badenian to Pannonian age (16.5–8.5 Ma), is a part of the Carpathian volcanic arc. Volcanic activity was contemporaneous with the active subduction in front of the Carpathian arc and backarc extension processes giving rise to the contemporaneous horst and graben structure. The composition of lavas varies from basalts to rhyolites, however, andesites dominate. Variable viscosity and explosivity of magmas gave rise to an almost complete set of volcanic forms and products. Further modifications arise from the changing terrestrial to shallow marine environment.

The essential results of the CSVF paleovolcanic reconstruction are demonstrated schematically in Fig. 1. The following volcanoes and volcanic structures have been distinguished: (1) an early complex of scattered andesite extrusive domes and related reworked

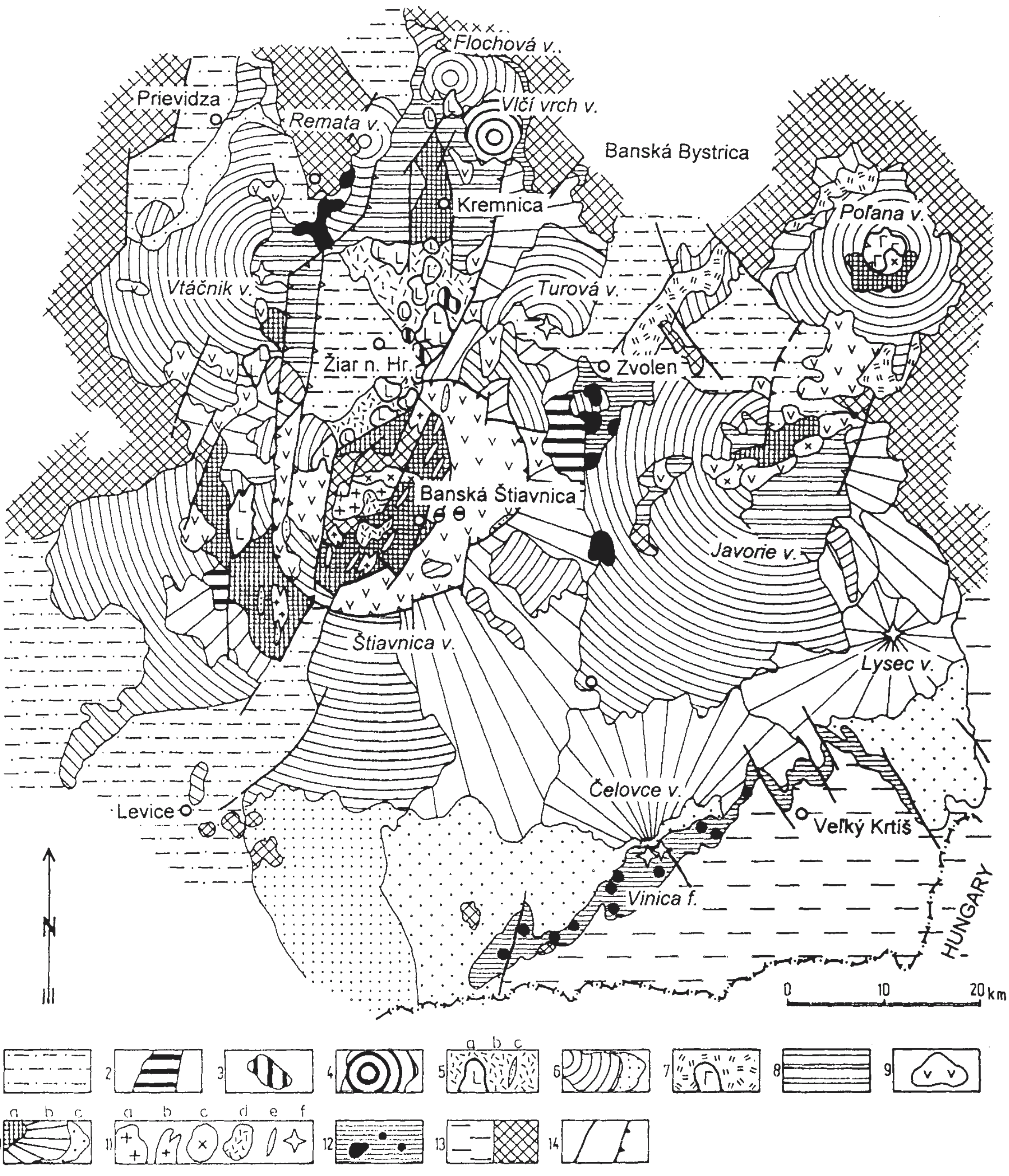

Fig. 1. Structural scheme of the Central Slovakia Volcanic Field (according to Konečný et al. 1995). 1 — sediments of intravolcanic depressions, 2 — products of the alkali basalt volcanism (Late Pannonian–Quaternary), 3 — lava flows and sills of aphanitic calc-alkali basalts/basaltic andesites (Early Pannonian), 4 — stratovolcano of porphyritic calc-alkali basalts/basaltic andesites (Early Pannonian), 5 — rhyolite domes/dome flows (a), dykes (b) and pyroclastic and epiclastic rocks (c) of the Jastrabá formation (Late Sarmatian), 6 — Sarmatian andesite stratovolcanoes and reworked marine facies, 7 — rhyodacite domes/dome flows and related pumice tuffs and reworked tuffs of the Strelníky formation (Early Sarmatian), 8 — effusive complexes of basic to intermediate andesites filling grabens (Late Badenian), 9 — domes/dome flows of intermediate to acid andesites filling grabens and caldera (Late Badenian), 10 — Early to Middle Badenian andesite stratovolcanoes: a — propylitized complex of the central zone, b — stratovolcanic complex of the proximal zone, c — reworked marine or fluvial facies, 11 — intrusions: a — granodiorite, b — granodiorite porphyry, c — diorite and diorite porphyry, d — quartz-diorite porpfyry sills, e — quartz-diorite porpfyry dykes, f — necks, 12 — extrusive domes and reworked breccias of garnet-bearing andesites (Early Badenian), 13 — pre-volcanic basement: a — Early Miocene sediments, b — older rocks, 14 — faults: a — normal, b — limiting grabens and calderas.

breccias and epiclastic rocks, (2) the Čelovce and Lysec andesite pyroclastic volcanoes, (3) the relatively large multiple-stage Javorie andesite stratovolcano, (4) the multiple-stage Poľana andesite stratovolcano, (5) the very large multiple-stage Štiavnica stratovolcano, (6) the multiple-stage volcanic complex of the Vtáčnik mountain range, (7) the multiple-stage volcanic complex of the Kremnické vrchy mountain range, (8) the rhyolite dome/flow complex and related volcanoclastic rocks of the Jastrabá formation, (9) a late stage calc-alkali basalt/basaltic andesite volcano and scattered shallow intrusions and lava flows, (10) rare alkali basalt necks, lava flows and cinder cones. The volcanoes evolved in terrestrial environments with the exception of the southern part of the CSVF, where a shallow marine environment persisted.

Early complex of scattered andesite extrusive domes

The initial Early Badenian hornblende-pyroxene ± biotite ± garnet andesite volcanic activity (16.5–16.3 Ma) is represented by scattered extrusive domes, with aprons of coarse breccias in the proximal zone and fluvial conglomerates and sandstones in the distal zone. In the shallow marine environment extrusive domes are extensively brecciated, and coarse hyaloclastite breccias accumulate around domes in a large volume. Gravity driven sliding created extensive submarine breccia flow/slump breccia deposits. In the distal zone these deposits alternate with shallow marine epiclastic volcanic conglomerates and sandstones and marine sediments.

Čelovce and Lysec pyroclastic volcanoes

The Early Badenian pyroxene andesite volcanoes of *Čelovce* and *Lysec* are situated at the SE part of the CSVF (Fig. 1). Remnants of the volcanoes are represented in the central zone by necks (Čelovce) or late stage tholoids (Lysec) and dominantly coarse pyroclastic breccias with periclinal dips 15–20^{o}. In the proximal zone agglomerates and block and ash pyroclastic flow deposits alternate with epiclastic breccias. Epiclastic volcanic conglomerates and sandstones with rare pyroclastic flow and/or mudflow deposits with intercalations of siltstones and reworked tuffs are characteristic of the distal zone. The marine environment is responsible for more intense reworking into conglomerates and sandstones.

Large compound stratovolcanoes

The large stratovolcanoes and stratovolcanic complexes of *Javorie, Poľana, Štiavnica and Kremnica-Vtáčnik* (Fig. 1) represent a result of multistage volcanic activity during the Badenian and Sarmatian, involving volcanotectonic depressions (grabens and calderas), differentiated rocks and subvolcanic intrusions. The Early stratovolcanoes of pyroxene and hornblende-pyroxene andesite composition are dominantly effusive with periods of pyroclastic flow activity. Their central zones are eroded to subvolcanic levels — late stage andesite porphyry sills and laccoliths are exposed. Lava flows (± extrusive domes), hyaloclastite breccias or block and ash pyroclastic flow deposits, and coarse mudflow and debris flow deposits dominate in their wide proximal zone. Epiclastic volcanic conglomerates and sandstones laid down in the ephemeral stream, fluvial and/or marine environment dominate in the distal zone, grading outward into volcanosedimentary formations.

During Middle to Late Badenian times (15.8–14.5 Ma) evolution of these stratovolcanoes continued with the subsidence of grabens and calderas, associated with activity of both — mafic undifferentiated rocks and differentiated intrusive and volcanic rocks. At the *Javorie stratovolcano* the graben fill is represented by basalts to pyroxene andesites, forming a complex of lava flows and hyaloclastite breccias, and by the pyroxene-hornblende andesite dome-flow complex including coarse blocky breccias in the proximal zone. Diorite stocks intruding the graben fill are accompanied by extensive hydrothermal alterations. Rhyodacite extrusive domes and related coarse breccias fill up a small caldera of the *Poľana stratovolcano*, intruded by diorite porphyry stocks. Pumice flow deposits and reworked pumice tuffs make up a corresponding horizon in the proximal and distal zones of the volcano. At the *Štiavnica stratovolcano,* the caldera is filled by a complex of biotite-hornblende andesite to dacite extrusive domes and dome-flows, including coarse blocky breccias, block and ash pyroclastic flow deposits and pumice flow deposits. Corresponding paleovalley fill on slopes of the stratovolcano is represented by lava flows and epiclastic volcanic breccias. An extensive subvolcanic intrusive complex of diorite, granodiorite, and their porphyries was emplaced by underground cauldron subsidence. The *Kremnica graben* is filled by basalts to px andesites, forming a complex of lava flows, hyaloclastite breccias, and phreatomagmatic pyroclastics (including reworked facies), covered by a thick effusive complex of hornblende-pyroxene andesites.

At the *Javorie* and *Štiavnica stratovolcanoes* renewed activity of less differentiated andesites during the Sarmatian period (14.5–12.5 Ma) formed discontinuous complexes on the slopes of older stratovolcanoes, often laid down in radial paleovalleys — effusive complexes ± hyaloclastite breccias and complexes dominated by explosive activity with ignimbrites, pumice flows, fall tuffs, and reworked tuffs, are present to variable extents, and grade southward into a marine volcanosedimentary formation. In the northern part of the CSVF it formed new central type stratovolcano (*Vtáčnik, Remata, Flochová, Sielnica, Turová, Poľana*) with centers situated mostly on marginal faults of grabens. *Poľana, Vtáčnik, Remata and Flochová volcanoes* are formed of periclinally dipping lava flows, pyroclastic breccias and agglomerates in the central zone, while alternating lava flows, pyroclastic flow deposits and epiclastic volcanic breccias grading outward into conglomerates and sandstones, form the proximal and distal zones. Diorite stocks and dykes are present in the eroded central zone of the *Poľana volcano*, while necks and dykes form the *Vtáčnik volcano* center. Rocks are pyroxene andesites with rare late stage hornblende-pyroxene andesites. The *Sielnica volcano* involves biotite-hornblende-hypersthene andesite extrusive dome and lava flows in the central zone, while hornblende-pyroxene andesite block and ash pyroclastic flow deposits with coarse epiclastic breccias dominate in the proximal zone — grading outward into distal zone conglomerates and sandstones. The younger *Turová volcano* of pyroxene andesite composition involves central explosive necks, pyroclastic breccias and agglomerates in the central zone, and pumice tuffs, reworked pyroclastic breccias, epiclastic volcanic breccias and late stage lava flows in the proximal zone. Reworked tuffs, epiclastic volcanic sandstones and conglomerates occur in the distal zone, grading into a volcanosedimentary formation in the Zvolen depression.

Rhyolites of the Jastrabá formation

An extensive Middle to Late Sarmatian rhyolite volcanic activity (13–10.7 Ma) gave rise to a dome/flow complex and related volcanoclastic rocks in the western part of the CSVF, along the N-S to NE-SW trending fault system. In most volcanic centers early phreato-magmatic and/or Plinian eruptions, forming tuff rings and cones build of pyroclastic fall, surge and flow deposits, were succeeded by the growth of extrusive domes and dome-flows (Bezák & Lexa 1979). Reworked pumiceous tuffs and epiclastic volcanic breccias, or conglomerates and sandstones make up the outer proximal and distal zones.

Late stage basalts and basaltic andesites

Early Pannonian high-Al basalts and basaltic andesites are the youngest products of calc-alkali volcanism in the CSVF. The small scale *Vlčí vrch volcano* is situated in the northern part of Kremnické vrchy on the marginal fault of Kremnica graben (Lexa 1971). The central olivine-bearing diorite porphyry neck is surrounded by remnants of volcanic cone formed of agglomerates, agglutinates and thin aa-type lava flows. The volcanic cone is surrounded by an effusive complex.

Scattered lava flows, necks, dykes, sills (in rhyolite tuffs of the Jastrabá formation) and a phreatic tuff-cone of fine-grained high-Al basalts and basaltic andesites occur in the southern part of the Kremnické vrchy (*Šibeničný vrch complex*), the northern part of the Štiavnické vrchy and the central part of Vtáčnik.

Alkali basalts

Scarce nepheline basanite/trachybasalt volcanic activity took place during the Pliocene, represented by two lava necks near Banská Štiavnica and by a small effusive plateau south of Zvolen. The cinder/spatter cone with related lava flows near Nová Baňa is of the Quaternary age (Šimon & Halouzka 1996).

References

Bezák V. & Lexa J., 1979: Genetic types of rhyolite volcaniclastic rocks in surroundings of Žiar nad Hronom. *Geol. Práce, Spr.,* 79, 83–112.

Cas R.A F. & Wright J.V., 1991: Volcanic successions: Modern and ancient. *Allen and Unwin*, London, 528.

Dublan L., Bezák V., Bujnovský A., Halouzka R., Hraško Ľ., Vozárová A. & Vozár J., 1997: Geological map of Poľana Mts. *Geol. služba SR,* Bratislava.

Fischer R.V. & Schmincke H.U., 1984: Pyroclastic rocks. *Springer Verlag*, Berlin, 472.

Konečný V., 1971: Evolutionary stages of the Banská Štiavnica caldera and its post-volcanic structures. *Bull. Volcanol.*, 35, 95–116.

Konečný V., Pristaš J. & Vass D., 1978: Geologická mapa Ipeľskej kotliny a južnej časti Krupinskej planiny. *GúDŠ*, Bratislava.

Konečný V. & Lexa J., 1984: Geological map of the Central Slovakia Neogene Volcanic Field. *GúDŠ*, Bratislava.

Konečný V. & Lexa J., 1995: Structure of the Neogene stratovolcanoes in central Slovakia. Proceedings, XVth CBGA congress, Athens. *Geol. Soc. Greece, Spec. Publ., 4,* 539–544.

Konečný V., Lexa J. & Hojstričová V., 1995: The Central Slovakia Neogene Volcanic Field: a review. *Acta Vulcanol.*, Pisa, 7, 63–78.

Konečný V., Lexa J., Halouzka R., Dublan L., Šimon L., Stolár M., Nagy A., Polák M., Vozár J., Havrila M. & Pristaš J., 1998a: Geological map of Štiavnické vrchy Mts. and Pohronský Inovec Mts. (Štiavnica stratovolcano). *Geol. služba SR*, Bratislava.

Konečný V., Bezák V., Halouzka R., Stolár M. & Dublan L., 1998b: Geological map of the Javorie Mts. *Geol. služba SR*, Bratislava.

Lexa J., 1971: Stratovulkán bazaltoidného andezitu v severnej časti Kremnického pohoria (Basaltic andesite stratovolcano in the northern part of Kremnické vrchy mountain range). *Geol. Práce, Spr.,* Bratislava, 55, 139–146.

Lexa J., Halouzka R. & Havrila M., 1998: Geological map of the Kremnické vrchy Mts. *Geol. služba SR*, Bratislava.

Maleyev E.F., 1959: Volcanoclastic rocks (in Russian). *Nauka,* Moscow, 163.

Maleyev E.F., 1963: Neogene volcanism of Zakarpatye (in Russian). *Nauka,* Moscow, 250.

Šimon L. & Halouzka R., 1996: Pútikov vŕšok volcano — the youngest volcano in the Western Carpathians. Slovak Geological Magazine 2/96, *Geol. služba SR*, Bratislava, 103-123.

Šimon L., Elečko M., Lexa J., Pristaš J., Halouzka R., Konečný V., Gross P., Kohút M., Mello J., Polák M., Havrila M. & Vozár J., 1997: Geological map of the Vtáčnik Mts. and Hornonitrianska kotlina depression. *Geol. služba SR*, Bratislava.

Vass D., Konečný V. & Šefara J., 1979: Geologická stavba Ipeľskej kotliny a Krupinskej planiny (Geology of the Ipeľ Basin and Krupina Plateau). *GÚDŠ*, Bratislava, 277.

NEOGENE - QUATERNARY ALKALI BASALT VOLCANISM OF SLOVAKIA: REVIEW OF VOLCANIC FORMS AND EVOLUTION

V. KONEČNÝ[1], J. LEXA[1] and K. BALOGH[2]

[1]Geological Survey of the Slovak Republic, Mlynská dolina 1, 817 04 Bratislava, Slovak Republic; lexa@gssr.sk

[2]Institute of Nuclear Research, Hungarian Academy of Science, Bem tér 18/c, H-4026 Debrecen, Hungary

Key words: alkali basalt, volcanism, maars, diatremes, K/Ar dating, Slovakia.

Introduction

Alkali basalt volcanism in central and southern Slovakia was active from the Late Miocene to Quaternary time, and following the extensive Middle Miocene calc-alkali, it was dominantly andesitic volcanism. Following pioneer geological studies by Jugovics (1944) and Kuthan et al. (1963) and petrographic studies by Šímová (1965) and Mihaliková (1966) volcanic forms were analyzed during detailed geological mapping by Konečný and Lexa and summarized in the geological map of the Lučenec Basin and Cerová vrchovina highland (Vass et al. 1992). Various aspects of volcanic forms were also discussed by Konečný et al. (1995a, in press), Šimon & Halouzka (1996) and Vass et al. (1997, 1998). The problem of the timing of volcanic activity was repeatedly addressed by Kantor & Wiegerová (1981), Balogh et al. (1981), Vass & Kraus (1985), Vass et al. (1992), Konečný et al. (1995a,b, in press) and Šimon & Halouzka (1996), using gradually increasing amounts of data, especially the results of K/Ar dating on whole rock samples, to a lesser extent also on mineral separates (details of methodology are given in the original papers). The aim of this contribution is to review the present state in analysis of volcanic forms and timing of alkali basalt volcanic activity.

Central Slovakia

Only scarce eruptive centers are known in the area of the Central Slovak Neogene Volcanic Field dominated by andesites and rhyolites (Fig. 1). The initial manifestation of alkali basalt volcanic activity follows closely the last manifestations of calc-alkali volcanism, which was terminated by activity of high alumina basalts and/or basaltic andesites with K/Ar ages 12–8.2 Ma. A relic of alkali basalt lava flow at the locality Devičie (1 km south of Krupina) filling an E-W oriented paleovalley has been dated to 8.0 ± 0.54 Ma. An extensive lava plateau Ostrá Lúka (southwest of Zvolen) consists of several lava flows, which have moved primarily northward over an undulating surface at the bottom of a wide valley. The relative elevations of the base and top of lava flows point to a source at the southern edge of the lava plateau, however, the corresponding cinder/spatter cone has been eroded. The lava flow at the locality of Dobrá Niva has been dated to 6.59 ± 0.29 Ma.

The lava neck of Kalvária near Banská Štiavnica forms a conspicuous hill. Remnants of phreatomagmatic breccias at the base and the outward dipping columnar jointing indicate, that the neck represents a remnant of a lava lake filling a crater (maar) in the center of the tuff-cone, which was created during the initial phreatomagmatic stage of eruption. A deeper eroded lava neck at the locality Kysyhýbel (east of Banská Štiavnica) also shows an early phreatomagmatic (tuffisite) breccia followed by emplacement of lava. The necks have been dated to 7.29 ± 0.41 or 6.77 ± 0.48 Ma (Kantor & Wiegerová 1981).

Fig. 1. Relics of alkali basalt volcanism in the Central Slovak Neogene Volcanic Field (according to Konečný et al. in press). 1 — lava flows and lava complexes of alkali basalts, 2 — a) cinder cone; b) supposed cinder cone/removed by erosion); 3 — lava necks; 4 — fluvial limnic sediments of depressions dammed by lava flows a) Riss/Würm, b) Late Miocene/Early Pliocene; 5 — margins of Central Slovak Neogene Volcanic Field; 6 — state boundary.

While the above mentioned alkali basalt and basanite lava flows and necks represent the oldest manifestations of alkali basalt volcanism in Slovakia, the Pútikov vŕšok volcano near Nová Baňa is the youngest one. Related lava flows have been dated to 0.53 ± 0.16 Ma, however, their age must be younger, as the lava flows rest over the Riss terrace of the Hron river with the assumed age 0.22–0.13 Ma (Šimon & Halouzka 1996). Thanks to its young age the volcano is well preserved. Its products fill up a former SE-NW oriented valley opened northwestward towards the Hron river. Hawaiian to Strombolian eruptions gave rise to a sizable cinder/spatter cone, feeding contemporaneous lava flows outflowing towards the Northwest. After reaching the alluvial flat of the Hron river individual aa-type lava flows spread into a small plateau with recognizable lava tongues (Šimon & Halouzka 1996).

Southern Slovakia

Most of the alkali basalt to basanite volcanic activity took place in the area of southern Slovakia (Lučenec basin, Cerová vrchovina highland), extending over the state boundary into northern Hungary (Fig. 2). Numerous cinder cones, lava flows, necks, diatremes and maars have been observed (Vass et al. 1992).

The results of K/Ar dating and variable relationship of volcanic products towards sedimentary deposits and morphology indicate, that volcanic activity of alkali basalts and basanites in southern Slovakia took place in five volcanic phases. The oldest are volcanic products of the *Podrečany basalt formation* (several lava flows and two maars) extending in the western part of the Lučenec Basin, dated to 7.2–6.4 Ma. This age is confirmed by the Pontian age of the Poltár formation sediments, which alternate with volcanic products and/or cover them (Vass & Kraus 1985; Planderová 1986). Lava flows, divided by erosion into several isolated remnants, rest on a relatively flat surface. Due to extensive erosion their sources (related cinder/spatter cones) are not known. Phreatomagmatic activity stimulated by contact between ascending lava and water in sediments (or lakes) resulted in formation of maars and tuff-rings, which were subsequently filled by diatomitic clay and/or alginite (Vass et al. 1997, 1998).

The alkali basalts and basanites of the *Cerová basalt formation* extending over the Cerová vrchovina highland are younger, of the Pliocene to Quaternary age. Owing to synvolcanic updoming of the Cerová vrchovina highland (Vass et al. 1986), the older volcanic products occupy relatively higher positions on ridges, while the younger volcanic products occupy relatively lower elevations in valleys. The youngest maars are at the level of recent alluvial flats. Relative elevation, degree of volcanic form preservation, K/Ar dating (Balogh et al. 1981; Kantor & Wiegerová 1981; Konečný et al. 1995a,b) and paleomagnetic data (Konečný et al. 1995b) allow us to distinguish five volcanic phases:

— 5.43–3.74 Ma dominantly inside and subordinately at the margins of the updomed area. Lava necks, cinder/spatter cones and related lava flows reveal an advanced degree of destruction.

— 2.92–2.60 Ma dominantly at the margins of the updomed area. Diatremes, cinder/spatter cones and related lava flows reveal a moderate degree of destruction.

— 2.25–1.6 Ma mostly at the margins of the updomed area. Remnants of maars, cinder/spatter cones and related lava flows reveal a weak to moderate degree of destruction.

— 1.51–1.16 Ma dominantly concentrated NE of Fiľakovo within the Lučenec Basin. Volcanic forms — maars, tuff-cones, cinder/spatter cones and related lava flows reaching the lowest levels of paleovalleys are relatively well preserved.

— 1–0.4? Ma represented by remnants of maars and tuff-cones at the level of alluvial flats near Fiľakovo and Hodejov. The relationship to river terraces indicates an age around the Günz-Mindel boundary.

Volcanic activity created a number of cinder/spatter cones accompanied by lava flows. Phreatomagmatic activity stimulated by interaction of ascending magma with water in deep aquifers at the base of Tertiary sedimentary the formations resulted in the formation of maars, tuff rings and tuff cones, which pass downward into diatremes, exposed nowadays due to erosion in uplifted parts of the region.

Cinder and spatter cones built up of agglomerates, agglutinates alternating with lapilli tuffs and basaltic bombs are the source of lava flows. Eroded cinder/spatter cones usually reveal an early phreatomagmatic phase represented by palagonite tuffs and/or pyroclastic breccias.

Lava flows with variable thickness (10–50 m) form either elongated bodies following the course of paleovalleys, reaching the length of several kilometers, or small lava plateaus spreading over flat surfaces. Successions of lava tongues have been observed at some localities. Massive lava shows variably platy, blocky or columnar jointing. Vesiculated lava at the top of lava flows passes into aa-type breccias. Lava breccias are almost absent at the base of flows.

Necks and dykes cutting underlying Early Miocene sediments in eroded parts of the volcanic area represent feeding channels to cinder/spatter cones and related lava flows. Lava necks and dykes often penetrate through early breccia fillings of diatremes. Columnar jointing patterns in the uppermost parts of several necks indicate a transition to lava lake bodies filling a maar, tuff-ring or tuff-cone at the top of the diatreme.

Maars are generally situated in less eroded parts of the region. Early phreatic and phreatomagmatic explosions reflecting an interaction of uprising magma with water-saturated sediments at the base of the Tertiary formations (at depths of around 800–1000 m) created tuff-rings and maar depressions, subsequently infilled by lakes. Following phreatomagmatic activity of the Surtseyan — wet surge type, reflecting interaction of magma with water in the maar lake, created palagonite tuff cones. Gradual exhaustion of water in

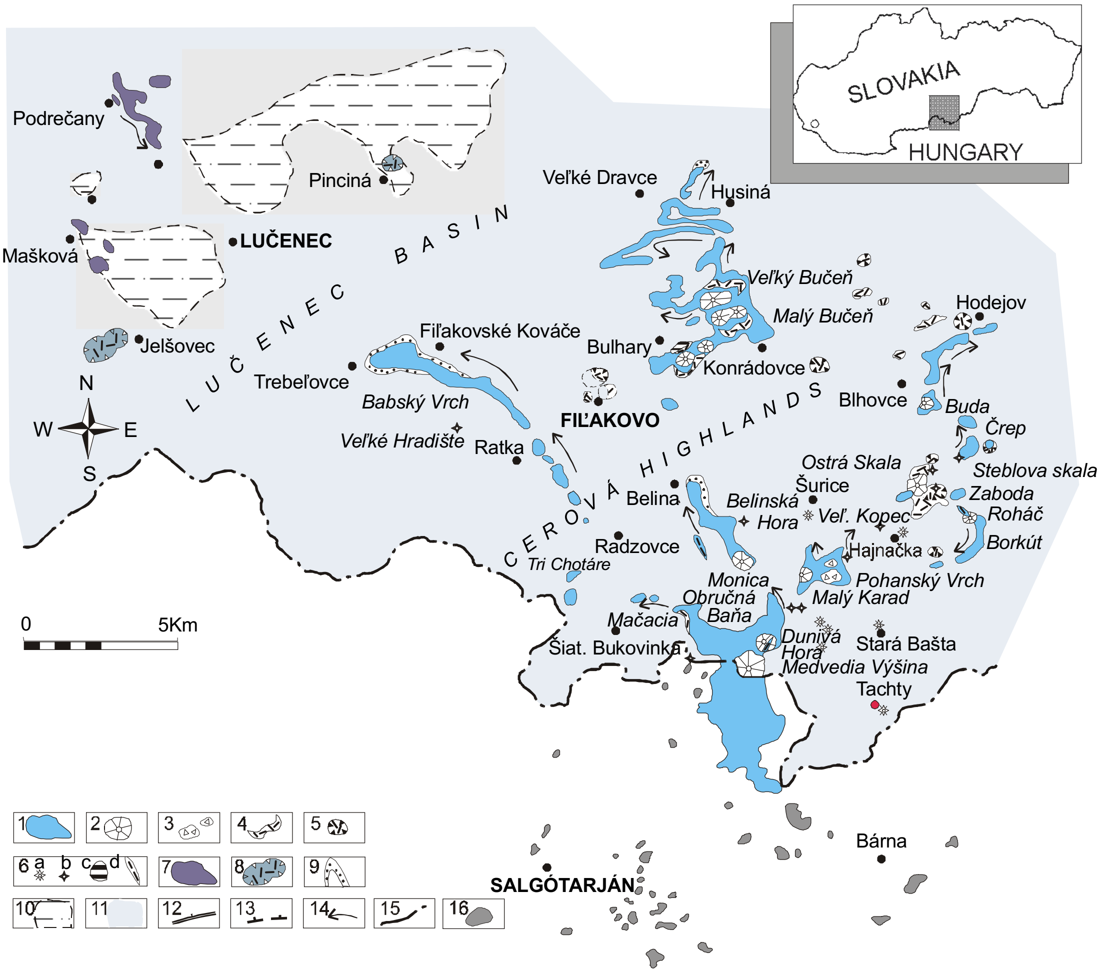

Fig. 2. Scheme of Late miocene to early Quaternary alkaline volcanics in southern Slovakia (according to Konečný et al. 1995). *Cerová Vrchovina formation (Middle Pliocene-Pleistocene)*: 1 — lava flow, 2 — scoria cone, 3 — agglomerates, 4 — lapilli tuffs, 5 — maar, 6 — eruptive centres: 6a) diatreme, 6b) neck, 6c) extrusion, 6d) dyke, *Podrečany formation (Early Pliocene):* 7 — lava fow, 8 — maar, Belina beds (Romanian?), 9 — gravels, clays, sands, Poltár formation (Pliocene), 10 — clays, sands, gravels, rare lignite lenses, 11 — Early Miocene sediments, other signs, 12 — updomed area, 13 — local scale elevation, 14 — direction of lava flows, 15 — state boundary, 16 — undivided basaltic rocks.

the maar lake eventually resulted in the transition towards the mixed Surtseyan-Strombolian eruptions (palagonite tuff of proximal base surge deposits are mixed with basaltic bombs) and in the final transition to Hawaiian type eruptions, which gave rise to the youngest spatter cones and/or lava lakes.

Diatremes are exposed conduits of maars, exposed by erosion. Their filling reflects evolutionary stages of maar structures. The early stage is represented by phreatic megabreccia of Early Miocene sediments cemented by sandy matrix with a tuffaceous admixture. Later products of phreatomagmatic activity are represented by palagonite tuffs with sandy admixture and pieces of vesiculated basalts and sediments. The transition towards Strombolian eruptions is indicated by younger scoraceous basalt breccia cutting palagonite tuffs. Feeders of the youngest Hawaiian type eruptions occur in the form of basaltic dykes cutting breccias and tuffs of earlier stages.

Conclusions

According to the results of K/Ar dating, alkali basalt volcanism in Slovakia was active from the Late Pannonian to Quaternary periods, following immediately the latest products of calc-alkali volcanism in the region of central Slovakia. The relatively low volume of individual eruptions and rather random timing in seven phases 8.0–6.4, 5.5–3.7, 3.0–2.6, 2.3–1.6, 1.5–1.1, 1.0?–0.5? and around 0.2–0.1? Ma indicate a generally low magma production rate characteristic of this type of intraplate volcanism. As the intervals between the previous volcanic phases are longer then the time which has elapsed since the last eruption, volcanic activity cannot be considered finished and we can still expect sporadic eruptions to take place in the future.

Hawaiian type eruptions giving rise to cinder/spatter cones and related lava flows are characteristic for less viscous alkali basalt magmas with relatively small content of volatiles represented predominantly by CO_2. However, if uprising magmas on their way to the surface meet water saturated sediments, the water/magma interaction results in phreatic and phreatomagmatic eruptions giving rise to maars and palagonite tuff rings. The Surtseyan type eruptions due to interaction of magma with water in maar lakes give rise to palagonite tuff cones. The decreasing water/magma ratio with time is the reason for the final transition towards the Hawaiian type eruptions.

References

Balogh K., Mihaliková A. & Vass D., 1981: Radiometric dating of basalts in Southern and Central Slovakia. *Západ. Karpaty, Ser. Geol.,* 7, 113–126.

Jugovics L., 1944: Somoskö-Fülek-Ajnacskö települo basalt-elöfordulá-

sok. Különlenyomat a M. Kir. *Földt. Intez. 1939-40, Évi jelent.,* Budapest, 1-50.
Kantor J. & Wiegerová V., 1981: Radiometric ages of some basalts of Slovakia by K/Ar method. *Geol. Zbor. Geol. Carpath.,* 32, 1, 29-34.
Konečný V., Mihaliková A., Halouzka R., Škvarka L., Stolár M., Planderová E., Panáček A. & Šefara J., 1986: Vysvetlivky k listu geologickej mapy 1:25,000 Krupina (36-341). Open file report, *Geol. Survey Slovak Rep.,* Bratislava.
Konečný V., Lexa J., Balogh K. & Konečný P., 1995a: Alkali basalt volcanism in Southern Slovakia: volcanic forms and time evolution. *Acta Vulcanol.,* 7, 167-171.
Konečný V., Balogh K., Orlický O., Lexa J. & Vass D., 1995b: Evolution of the Neogene-Quaternary alkali basalt volcanism in Central and Southern Slovakia (West Carpathians). *Proceedings XI. Congres of the Carpatho-Balkan Geol. Assoc. Athens*, 533-538.
Konečný V., Lexa J. & Balogh K., in press: Neogene-Quaternary alkali basalt volcanism in Central and Southern Slovakia (Western Carpathians). *Geoline,* Praha.
Kuthan M., Biely A., Böhm V., Čechovič V., Fusán O., Hovorka D., Mazúr E. & Regásek F., 1963: Vysvetlivky k prehľadnej geologickej mape 1:200,000 M-34-22 Zvolen. *GÚDŠ,* Bratislava, 1-131.
Mihaliková A., 1966: Petrografická a petrochemická charakteristika bazaltov juhovýchodného Slovenska *GÚDŠ,* Bratislava,. *Západ. Karpaty*, 5, 151-199.
Planderová E., 1986: Biostratigrafické zhodnotenie sedimentov poltárskeho súvrstvia. *GúDŠ,* Bratislava, *Geol. práce*, *Spr.,* 84, 113-118.
Šimon L. & Halouzka R., 1996: Pútikov vŕšok volcano — the youngest volcano in the Western Carpathians. *GSSR*, Bratislava, *Slovak Geol. Mag.,* 2/96, 103-123.
Šímová M., 1965: Petrografia a petrochémia produktov finálneho vulkanizmu Slovenského Stredohoria. *Acta Geol. Geogr. Univ. Comen.,* Bratislava, *Geologica*, 9, 9-89.
Vass D. & Kraus I., 1985: Double age of basalts in the Southern Slovakia and Their relation to Poltár beds. *Miner. slovaca,* 17/5, 435-440.
Vass D., Bezák V., Elečko M., Konečný V., Lexa J., Pristaš J., Straka P. & Vozár J., 1992: Geological map of the Lučenská kotlina depression and Cerová vrchovina highland. *GÚDŠ*, Bratislava.
Vass D., Elečko M. & Pristaš J., 1986: Updoming of Cerová vrchovina - young structure in the Southern Slovakia. *Geol. práce*, *Spr.,* 84, 135-140.
Vass D., Konečný V., Elečko M., Milička J., Snopková P., Šucha V., Kozáč J. & Škrabana R., 1997: Alginit — nový zdroj slovenského nerudného potenciálu (ložisko Pinciná). *Miner. slovaca*, 29, 1-39.
Vass D., Konečný V., Elečko M., Kozáč J., Molnár A. & Zakovič M., 1998: Ložisko diatomitu v bazaltovom maare pri Jelšovci a možnosti jeho využitia. *Miner. slovaca*, 30, 333-356.

PRESSURE EFFECT ON THE STABILITY AND ASSEMBLAGES OF ACID PLAGIOCLASE IN MEDIUM-TEMPERATURE METABASITES, ECLOGITES, AND ASSOCIATED GNEISSES

S.P. KORIKOVSKY

Institute of the Geology of Ore Deposits, Petrography, Mineralogy and Geochemistry, Russian Academy of Sciences, Pyzhevsky per. 7, 109017 Moscow, Russia

Key words: eclogites, plagioclase stability, P-T conditions, phase equilibria.

It is well known that as the depth of metamorphism increases, and the transformation of metabasics into eclogite-amphibolites and eclogites caused by it increases too, the stability and number of assemblages of acid plagioclase are restricted, whereas the maximum Jd content in Ca-Na clinopyroxene of the augite-omphacite-jadeite series rises. Typical crustal eclogites are usually devoid of albite-oligoclase, although this mineral may remain stable, at the same P-T parameters, in acid gneisses intercalating with the eclogites. Since medium-temperature eclogites are merely high-pressure analogues of epidote amphibolites and Fe-Ca-Mg metagraywackes, a comparative study of all types of metabasics and associated gneisses at the pressure interval of 5-8 to 15-16 kbar, makes it possible to trace, step by step, the narrowing stability of acid plagioclase with increasing pressure up to the complete disappearance of this mineral, and its replacement by jadeite at P ~ 14-18 kbar in accordance with the reaction Ab = Jd+Qtz (Holland 1980). The effect of pressure and bulk-rock composition on the stability of albite-oligoclase is best illustrated by the evolution of its assemblages with Ca-Na clinopyroxene, garnet, kyanite, and quartz at approximately 600 ^{o}C; these relations also make it possible to distinguish between the depth subfacies of crustal eclogites.

The model of Cpx-Gr-Ab(Olg)-Ky-Qtz phase relations in amphibolites, eclogites, rocks of intermediate composition, and gneisses is discussed using a series of (Mg,Fe)-Al-Na plots with excess Qtz, Czo, and Rt (Fig. 1), which are based on data from medium-temperature complexes of various depths in different areas. Points 1-6 in the diagrams correspond to the bulk compositions of the rocks with different (Mg,Fe) content, starting from mafic eclogites and clinopyroxene-garnet amphibolites, which are characterized by low total (Na + Al)/(Mg,Fe) ratios (points 1-2), to leucocratic gneisses in which this ratio is high (points 5-6).

The evolution of Ab(Olg)-bearing associations with increasing of pressure at T ~ 600 ^{o}C is as follows.

In ***garnet-epidote-clinopyroxene amphibolites*** from usual complexes metamorphosed under P = 5-8 kbar (Fig. 1-1), the clinopyroxene contains 1-8 % Jd in association with acid plagioclase, which is stable in rocks from the mafic (points 1-2) to leucocratic (p. 5-6) composition. As the pressure increases from 8 to 12 kbar, due to prograde reactions such as $Cpx_{Ca} + Ab = Cpx_{Ca\text{-}Na}$ (Omp) +Qtz (Holland 1980), the clinopyroxene becomes progressively enriched in jadeite and is transformed initially to Na-Aug and, then to omphacite with up to 35-40 % Jd. The acid plagioclase entirely disappears at P = 10-12 kbar in mafic rocks (Fig. 1-2, pp. 1-2), in which the previous assemblage Cpx_{Ca} + Grt + Ab-Olg is replaced by the higher pressure, plagioclase-free eclogite assemblage $Cpx_{Ca\text{-}Na}$ (10-35 % Jd) + Grt. But in the eclogites with elevated (Na + Al)/(Mg,Fe) ratios (p. 3 in Fig. 1-2), albite-oligoclase remains stable in the assemblage Cpx($\geq$ 35 % Jd) + Grt + Ab-Olg (Heinrich 1986; Korikovsky et al. 1997); in gneisses (pp. 4-6) acid plagioclase is also fully stable. Hence, over the pressure interval of 8-12 kbar, at whose the upper limit of the maximum Jd content in Cpx attains 35-42 %, an association of sodic augite or omphacite with primary albite-oligoclase is possible. The presence of Ab-Olg in eclogites and eclogite-amphibolites is controlled under these conditions only by the (Na + Al)/(Mg,Fe) ratio of the rock. This interval may be designated as ***the plagioclase eclogite depth subfacies*** (Fig. 1-2). This subfacies occurs most widely in many Phanerozoic complexes, and is characterized by the coexistence and intercalation of plagioclase-free eclogites with Pl-bearing eclogite-amphibolites and gneisses.

A further pressure growth ($\geq$ 12 kbar) results in the univariant reaction $Grt_{Fe\text{-}Mg\text{-}Ca}$ + Ab(Olg) $\rightarrow$ Cpx($\geq$ 42 % Jd) + Ky + Qtz, that leads to the first appearance of the omphacite plus kyanite assemblage (Fig. 1-3), and albite-oligoclase completely disappears from **assemblages with garnet**. The pressure interval of 12-15 kbar (Fig. 1-3, -4) can be termed ***the depth subfacies of kyanite eclogites.*** In its field albite-oligoclase cannot crystallize in eclogites with any (Na + Al)/(Mg,Fe) ratio (Fig. 1-3, pp. 1-3), but acid plagioclase remains fully stable in garnet-free omphacite-bearing and kyanite gneisses (Fig. 1-3, pp. 4-6). The normal intercalating plagioclase-free kyanite eclogites and Pl-bearing garnet-free gneisses, typical for this subfacies, give an amusing, but false impression of a tectonic collage of rocks of various depth types. However, as it follows from Fig. 1-3, -4, the equilibrium coexistence of such rocks is permitted for the kyanite-eclogite depth facies.

At pressures of 14-15 kbar, i.e., in the high-pressure area of the kyanite eclogite facies, the maximum Jd solubility in Cpx attains

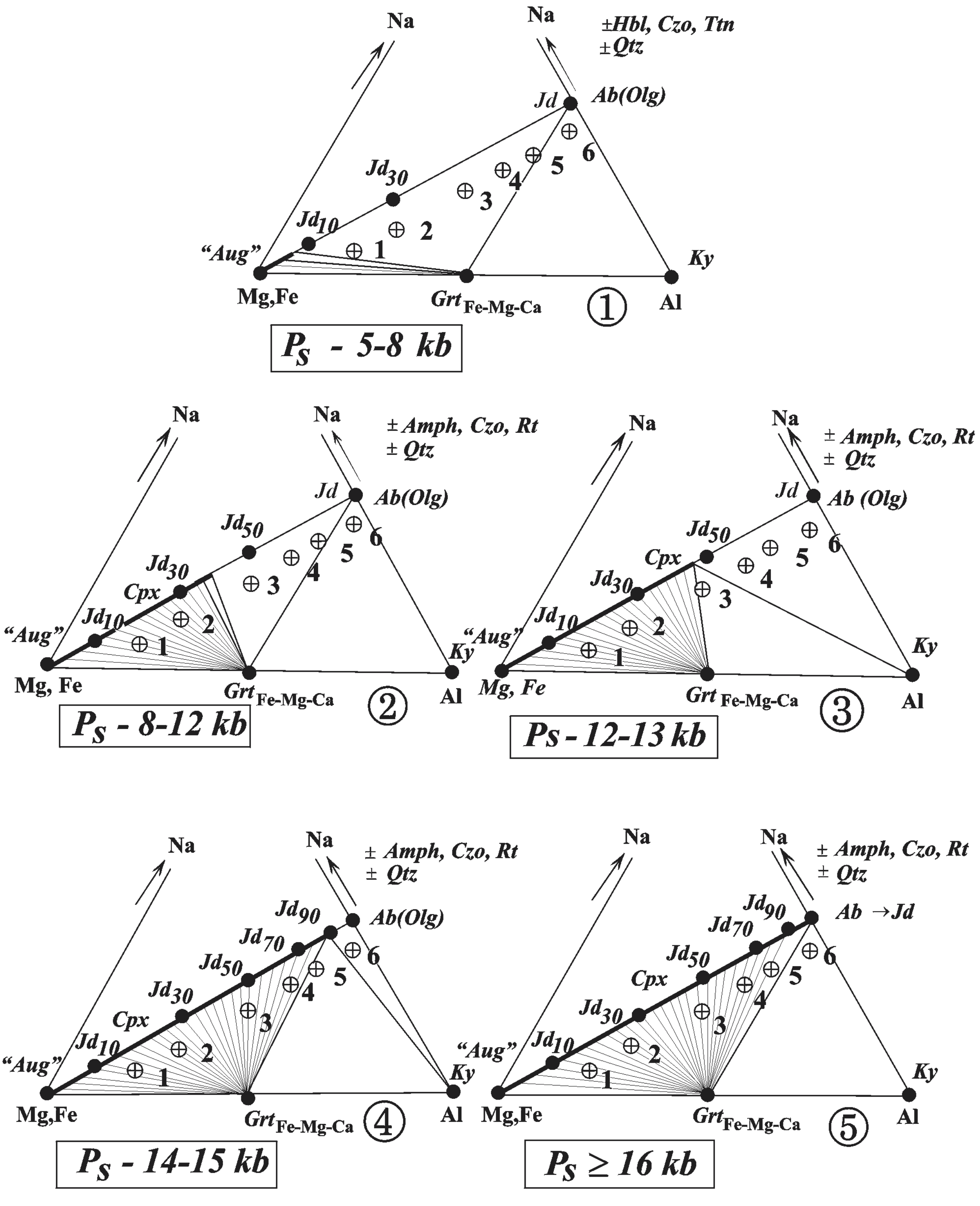

Fig. 1. (Mg,Fe)-Al-Na plots illustrating changes in phase relations of Ca-Na clinopyroxene, albite (oligoclase), garnet, kyanite, and quartz in eclogites, metabasites, and gneisses at T ~ 600 °C with increasing pressure: 1 — Grt-Cpx-Ep-Pl-Qtz amphibolites from moderate-pressure metamorphic complexes; 2–5 — eclogite-bearing complexes: 2 — the depth plagioclase eclogite subfacies, 3–4 — kyanite eclogite subfacies, 5 — boundary between kyanite and jadeite eclogite subfacies.

80–90 % (Fig. 1-4). Under these conditions, all crustal eclogites are omphacite-garnet or omphacite-kyanite-garnet in composition and bear no plagioclase, which remains preserved only in the most acid rocks with the maximum (Na + Al)/(Mg, Fe) ratio (Fig. 1-4, p. 6), for example, in jadeitized metagranites (Gil Ibarguchi 1995).

Finally, at P ≥ 16 kbar, albite completely disappears from all types of metabasics and gneisses due to the reaction Ab = Jd + Qtz, and oligoclase is replaced by jadeite with zoisite inclusions (Compagnoni & Maffeo 1973). The mixing in the Cpx_{Ca}(Aug)-Jd solid solution becomes full (Fig. 1-5), and rocks occur in ***the depth subfacies of jadeite eclogites***. In gneisses, acid plagioclase also disappears, but the jadeite-K-feldspar associations crystallize (Okay 1997).

Thus, the diagrams of Fig. 1 enable one to trail the succession of and reactions, which result in the decrease in acid plagioclase stability with depth in medium-temperature rocks: initially, Pl disappears from mafic eclogites and metabasic rocks, and a further pressure increase leads to its disappearance from mesocratic eclogites and other rocks of intermediate composition, and, finally, at the highest pressures Ab-Olg disappears from acid gneisses.

References

Compagnoni R. & Maffeo B., 1973: Jadeite-bearing metagranites, and related rocks in the Mount Mucrone area (Sesia-Lanzo zone, Western Italian Alps). *Schweiz. Mineral. Petrogr. Mitt.,* 53, 3, 355–378.

Gil Ibarguchi J.I., 1995: Petrology of jadeite metagranite and associated orthogneiss from the Malpica-Tuy allochton (Nortwest Spain). *Europ. J. Miner.,* 7, 403–415.

Heinrich C.A., 1986: Eclogite facies regional metamorphism of hydrous mafic rocks in the Central Alpine Adula Nappe. *J. Petrology,* 1986, 123–154.

Holland T.J.B., 1980: The reaction albite = jadeite+quartz determined experimentally in range 600-1200 °C. *Amer. Mineralogist*, 65, 129–134.

Korikovsky S.P., Mircovski V. & Zakariadze G.S., 1997: Metamorphic evolution and the composition of the protolith of plagioclase-bearing eclogite-amphibolites of the Buchim block of the Serbo-Macedonian Massif, Macedonia. *Petrology*, 5, 534–549.

Okay A.I., 1997: Jadeite-K-feldspar rocks and jadeitites from northwest Turkey. *Mineral. Mag*., 61, 835–843.

OPAQUE Fe-Ti MINERALS IN THE HIGH TATRA GRANITOIDS

M. KUSIAK[1], M. MICHALIK[2] and B. WOLDAŃSKA[2]

[1]Institute of Geological Sciences, Polish Academy of Sciences, Kraków Research Centre, ul. Senacka 1, 31-002 Kraków, Poland; ndkusiak@cyf-kr.edu.pl

[2]Institute of Geological Sciences, Jagiellonian University, ul. Oleandry 2a, 30-063 Kraków, Poland; michalik@ing.uj.edu.pl

Key words: accessory minerals, Fe-Ti-oxides, the High Tatra granitoids, Carpathians.

Introduction

Three genetic groups of granitoids which are subdivided in the Western Carpathians show S-, I-, and A-type characteristics with decreasing age. The S-type rocks are peraluminous two-mica granites and granodiorites, which contain monazite and ilmenite in the accessory minerals assemblage. The younger I-type metaluminous granodiorites and tonalites (rarely granites) are characterised by the presence of allanite and magnetite among accessory minerals. The A-type rock is represented by biotite leucogranite, rich in pink K-feldspar phenocrysts (Petrík et al. 1994; Petrík & Kohút 1997). The accessory mineral assemblages suggest that S-type granite magma was relatively water-poor and crystallisation took place in reducing conditions. The I-type magma was relatively water-rich and the mineral assemblage suggests oxidising conditions (Petrík & Kohút 1997). Broska & Uher (1991), Broska & Gregor (1992) recognised pairs of accessory minerals which correlate one with the other and the presence of these pairs is mutually exclusive. Allanite (typical of I-type) correlates with magnetite, and monazite correlates with ilmenite (typical of S-type).

Geological setting

Granodiorites and tonalites prevail in the studied, northern part of the High Tatra (Turnau-Morawska 1948). Granites are scarce and occur mostly on some ridges related to so-called pegmatite-aplite border zone (Michalik 1951). Most of tonalites and granodiorites are peraluminous; metaluminous rocks occur only subordinately (Degenhardt et al. 1996). The composition of biotite, a dominant mafic aluminosilicate, varies in a broad range. In biotites the Fe/(Fe+Mg) values are between 0.45–0.7. Apatite, monazite, xenotime and zircon are common accessory minerals; titanite occurs rarely (Woldańska et al. 1998). Allanite and epidote are (at least in dominant part) the secondary minerals. Hematite, magnetite, ilmenite, hemoilmenite, ilmenohematite, martite, and pyrite were identified among opaque minerals in reflected light (Woldańska 1995; Woldańska et al. 1998). The accessory minerals assemblage determined in the High Tatra granitoid rocks by Broska et al. (1993) is similar but relative frequencies are different — allanite predominates over monazite, and magnetite over ilmenite. Kohút & Janák (1994) described the following accessory minerals in the High Tatra type granitoid rock: apatite, zircon, monazite, ilmenite ± magnetite. The High Tatra granitoids exhibit numerous alterations — feldspar albitization and sericitization, the growth of secondary muscovite, chloritization of biotite, growth of epidote (and/or allanite), carbonatization and prehnitization (Michalik et al. 1998). The breakdown of monazite and development of irregular coronas of apatite in reaction zone was also noted (Michalik & Skublicki 1998).

Methods of study

For all sample transmitted and reflected light optical microscopy was applied. The variability of chemical composition and detailed structure of Fe-Ti oxides was determined using electron microprobe analysis (SEM-EDS). All analyses were performed with 20 kV voltage.

Results

The occurrence of opaque minerals. Opaque minerals occur very often in clots with biotite, apatite, zircon, and monazite (Fig. 1). Seldom, Fe-Ti oxides occur in feldspars and/or in quartz. Fe-Ti oxides (rutile, ilmenite, or Fe-oxide) are present also as by-products of biotite chloritization.

The structure and chemical composition of opaque minerals. During examinations in reflected light microscopy, hematite (two groups) was recognised as the most common opaque mineral. Hematite of the first group exhibits the presence of rutile lamellae. Hematite pseudomorphoses after magnetite (martite) form the second group. Previously mentioned magnetite (Woldańska 1995) is completely or almost completely hematitised. Products of alteration of hematite to magnetite (mushkietovite) are scarce.

The SEM-EDS analysis shows that almost all opaque minerals are inhomogeneous (Fig. 2). The presence of complicated system (systems) of fine intergrowths in host mineral is the reason of difficulties in determination of chemical composition of the host and exolutions. Beside one or two systems of lamellae which are 1–4 μm wide and up to 15 μm long, and of lensoid forms, small trails (<< 1 μm width and 1–2 μm long) are also visible in electron microscope but determination of their composition is impossible. The host is usually a member of hematite-ilmenite solid solution series. The proportion of end-members varies in wide range but the variability is probably related to analytical difficulties. In some grains the exsolved lamellae are represented by rutile. Rutile forms also rims around opaque grains. Rutile often contains Fe (up to 0.07 Fe atoms *pfu*) and V, Zn, Nb. Ilmenite lamellae or lenses can be present in hematite-ilmenite$_{ss}$ host. The presence of Mn is typical of most of ilmenite lamellae. In some opaque minerals lamellae rich in Al and Zn (beside Fe) are present (a member hercynite-gahnite$_{ss}$?). Because of small dimensions, the precise determination of their chemical composition is not possible.

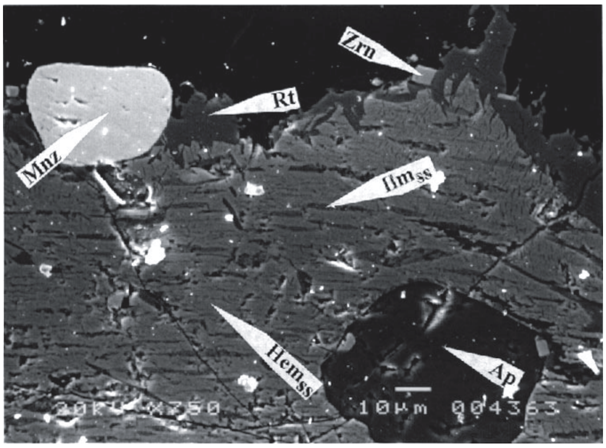

Fig. 1. Hematite (Hem_{ss}) with intergrowths of ilmenite (Ilm_{ss}) rimmed by rutile (Rt). Apatite (Ap) inside hematite and monazite (Mnz) and zircon (Zrn) in outer zone. Electron microscope, SE image.

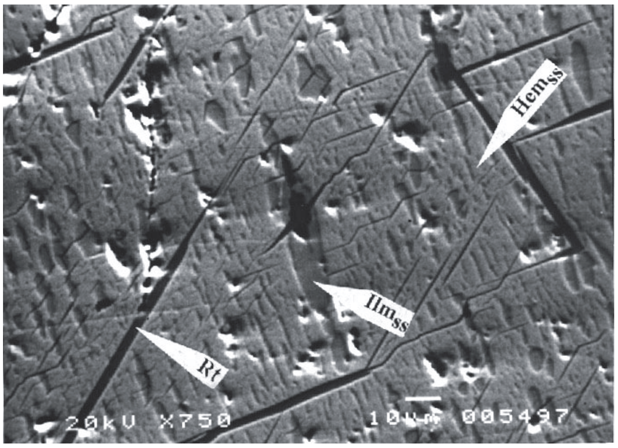

Fig. 2. Hematite (Hem_{ss}) with irregular lenses of ilmenite (Ilm_{ss}) and thin rutile (Rt) lamellae. Electron microscope, BSE image. Examples of EDS analysis results: Hematite — 6.1_{Hem} — 1.9_{Ilm}, ilmenite — $Ti_{0.81}Mn_{0.03}V_{0.20}Fe_{0.85}O_3$, rutile — $Ti_{0.83}V_{0.09}Fe_{0.07}O_2$.

Discussion and conclusions

Most of opaque minerals are members of ilmenite-hematite$_{ss}$. Based on the EDS determination and applying the terminology of Buddington et al. (1963) these minerals can be named as ilmeno-hematite (i.e. two-phase grains consisting of ferrianilmenite lamellae enclosed in titanohematite host). Both, in ilmeno-hematite and in titanohematite (or ferrianilmenite) exsolved lamellae of ferrian rutile can be present. Assuming that bulk composition of host and all exsolved phases represents the composition of primary mineral before exsolution, it is possible to state that ilmenite-rich member of ilmenite-hematite$_{ss}$ was probably the most common primary phase. The precise recalculation of studied minerals is not possible because of presence of very fine intergrowths (the determination of individual mineral constituent is beyond resolving capabilities of the electron microscope). The high content of Ti suggests that the possiblility of titanohematite-rutile assemblage development by oxidation of titanomagnetite, is less probable.

The complexity of systems of different Fe-Ti oxides phases intergrowths indicates, most probably, multi-stage processes of exsolution and, perhaps, oxidation. It is possible that oxidation of ilmenite started during crystallisation (Czamanske & Mihalik 1972; Czamanske & Wones 1973) and continued during deuteric stage (Duchesne 1972). Optical microscope study suggests, that magnetite is formed from hematite indicating that processes of transformation of Fe-Ti oxides could be very complex.

A further, detailed study of opaque minerals and mafic aluminosilicates is needed for better understanding the transformations of ilmenite (and possibly of magnetite) during the crystallisation of magma and alterations processes.

The presence of ilmenite (or ilmenite-rich member of ilmenite-hematite$_{ss}$), now in form of complex intergrowths of hematite-ilmenite-rutile, phases and monazite indicates reducing conditions during initial stages of crystallization. Absence or scarcity of primary allanite supports this suggestion. The presence of ilmenite and monazite in accessory mineral assemblage is considered to be typical of S-type granites in the Western Carpathians (Broska & Uher 1991; Broska & Gregor 1992; Petrík & Kohút 1997).

References

Broska I. & Gregor T., 1992: Allanite-magnetite and monazite-ilmenite granitoid series in the Tribec Mts. *Spec. Vol. of IGCP Project No. 276, GÚDŠ,* Bratislava, 25–36.

Broska I., Kohút M. & Uher P., 1993: Variscan granitoid rocks of the Tatry Mts. And their relation to other granitoids of the Western Carpathians. *Pre-Alpine events in the Western Carpathians' realm*, Stará Lesná, 252–253.

Broska I. & Uher P., 1991: Regional typology of zircon and its relationship to allanite/monazite antagonism (on an example of Hercynian granitoids in Western Carpathians). *Geol. Carpatica,* 42, 271–277.

Buddington A.F., Fahey J. & Vlisidis A., 1963: Degree of oxidation of Adirondack iron oxide and iron-titanium oxide minerals in relation to petrogeny. *J. Petrology*, 4, 138–169.

Czamanske G.K. & Mihalik P., 1972: Oxidation during magmatic differentiation, Finnmarka Complex, Oslo area, Norway: Part 1, The opaque oxides. *J. Petrology*, 13, 493–509.

Czamanske G.K. & Wones D.R., 1973: Oxidation during magmatic differentiation, Finnmarka Complex, Oslo area, Norway: Part 2, The mafic silicates. *J. Petrology*, 349–380.

Degenhardt J., Michalik M., Skublicki Ł. & Woldańska B., 1996: Petrology and geochemistry of granitoids from Polish High Tatra Mts. *Pol. Tow. Mineral., Prace Spec.*, 7, 44–47.

Duchesne J.C., 1972: Iron-titanium oxide minerals in the Bjerkrem-Sogndal Massif, South-western Norway. *J. Petrology,* 13, 57–81.

Kohút M. & Janák M., 1994: Granitoids of the Tatra Mts., Western Carpathians: field relations and petrogenetic implications. *Geol. Carpatica*, 45, 301–311.

Michalik A., 1951: Marginal zone of crystalline core in the Tatra Mts in the Kosista area. *Biul. PIG*, 61, 3–38 (in Polish).

Michalik M. & Skublicki Ł., 1998: Monazite breakdown during alterations of granitoids from The High Tatra Mts. *Pol. Tow. Mineral., Prace Spec.,* 11, 145–147 (in Polish).

Michalik M., Skublicki Ł. & Woldańska B., 1998: Postmagmatic alterations of the High Tatra granitoids (Tatra Mts., Poland). In: *Carpathian-Balcan Geological Association XVI Congress, Abstr., Vienna,* 381.

Petrík I., Broska I. & Uher P., 1994: Evolution of the Western Carpathian granite magmatism: age, source rock, geotectonic setting and relation to the Variscan structure. *Geol. Carpathica*, 45, 283–291.

Petrík I. & Kohút M., 1997: The evolution of granitoid magmatism during the Hercynian orogen in the Western Carpathians. In: Grecula P., Hovorka D. & Putiš M. (Eds.): *Geological evolution of the Western Carpathians, Miner. slovaca - Monogr.*, 235–252.

Turnau-Morawska M., 1948: Microgeological research in the central part of the crystalline Tatra. *Kosmos, Ser. A., Rozpr.*, 65, 59–100 (in Polish).

Woldańska B., 1995: Selected accessory minerals in granitoids from NE part of the High Tatra Mts. *Pol. Tow. Mineral., Prace Spec.*, 6, 54–56 (in Polish).

Woldańska B., Skublicki Ł., Michalik M. & Degenhardt J., 1998: Primary and secondary accessory minerals in the High Tatra granitoids (Poland). In: *Carpathian-Balcan Geological Association XVI Congress, Abstr., Vienna,* 641.

CRETACEOUS EVOLUTION OF THE VARISCAN BASEMENT OF THE SE WESTERN CARPATHIANS: COMBINATION OF CONTINENTAL UNDERTHRUSTING AND INDENTATION TECTONICS

O. LEXA and K. SCHULMANN

Institute of Petrology and Structural Geology, Charles University, 128 43 Prague, Czech Republic

Key words: Western Carpathians, Cretaceous collision, cleavage development, transpression.

Modern interpretations of the Mesozoic tectonic evolution of the Alpine chain underline lithospheric scale subduction of oceanic and European lithosphere below the African indenter (Allemand & Lardeaux 1997). A prominent feature of Mesozoic evolution of the Alpine chain is Jurassic to Cretaceous blueschist to ultrahigh pressure metamorphism (Chopin 1984). The exhumation of high-pressure rocks continued up to the Eocene, for instance, schist lustré and Mon Viso eclogites in French and Italian Alps (Spalla et al. 1996). The mechanism of exhumation of high-pressure rocks and associated structures is interpreted as imbrication and rapid exhumation of subducted oceanic and continental crust (Davy & Gillet 1986) or as a corner flow model of Platt (1993).

The Mesozoic evolution of Western Carpathians belt represents an integral part of the closing of the Tethyan ocean, and is characterized by Jurassic subduction and Late Jurassic exhumation of high pressure rocks (Faryad 1995; Mock & Reichwalder 1992). The Cretaceous and Tertiary evolution of the Western Carpathians does not exhibit the subduction-like metamorphism and associated tectonic evolution known from the Alpine belt. It seems to be evident that there is a lack of common features between Mesozoic tectonic and thermal evolution of the Alpine and Western Carpathian belts. The Western Carpathians structure is traditionally interpreted as crustal shortening of the Variscan basement resulting in ”décollement” of Mesozoic cover sequences and their transportation toward the north in the form of two large scale nappes; the lower Krížna nappe and the upper Choč nappe (Andrusov 1936; Andrusov 1958; Plašienka 1991). Therefore, the Mesozoic tectonic evolution of the Western Carpathians was mostly studied in sedimentary rocks with well-known stratigraphy (Biely 1989; Plašienka 1995).

The present structure of the southern part of the Western Carpathians consists of three main tectonometamorphic units. From north to south and from the bottom to the top they are: 1) Variscan crystalline basement with Late Paleozoic and Mesozoic cover subdivided into the Tatra unit to the north and the Vepor unit to the south, 2) Early to late Paleozoic basinal mostly turbiditic metasedimentary unit (Gemer unit) and 3) Mesozoic accretionary wedge containing blueschist facies relics (Bôrka, Meliata and Turňa units) overlain by the flat unmetamorphosed Silica nappe.

The crustal rocks of the Vepor basement are composed of two contrasting Variscan metamorphic domains exhibiting pre-Alpine thrust tectonics (Bezák 1992; Klinec & Miko 1987). The structurally lower domain, globally dipping to the north, is composed of Barrovian Ky-St micaschist. The structurally higher crystalline unit is represented by a domain of heterogeneous para- and ortho-derived migmatites intruded by peraluminous granites.

The Gemer unit consists of two contrasting series (Rakovec and Gelnica group) which differ in lithology and metamorphic grade. The pre-Alpine fabric of the Rakovec group dips to the north and structurally overlies the Gelnica group from the viewpoint of the Variscan southward structural pattern. The Rakovec group includes greenschist facies metabasites and phyllites but also contains slices of amphibolite facies rocks. The Variscan zonation is reversed with decreasing intensity from north to south (Faryad & Bernhardt 1996; Vozárová 1993).

The Gelnica group makes up the major part of the Gemer unit and consist of a thick flysch sequence and subordinate products of rhyolite-dacite volcanism. On the basis of fossil remnants, the age of the Gelnica unit has been established as Cambro-Ordovician–Devonian (Snopková & Ivanička 1981; Snopková & Snopko 1979; Vozárová et al. 1999). Late Carboniferous conglomerates and Permian clastics discordantly cover Early Paleozoic rocks of the Gemer unit and overlie Variscan tectonic contact between amphibolite and greenschist facies rocks of the Rakovec group.

The south verging Variscan fabric of the Gemer unit has been reworked by Alpine cleavage developed under low-grade metamorphic conditions. The cleavage exhibits asymmetrical fan structure with a narrower southern flank dipping to the north. Here, the Mesozoic sequences show early thrust-related structures overprinted by cleavage fabric similar to that developed in the southernmost extremity of Paleozoic sediments of the Gemer unit. The cleavage changes its inclination northwards, so that in the central part of the Gemer unit a steep vertical fabric predominates, and further to the north the cleavage becomes flat and dipping to the south. The intensity of cleavage also decreases to the north and becomes heterogeneous and weak in amphibolite and greenschist facies rocks of the Rakovec group. This change of fabric intensity is partly due to the disappearing Alpine deformation northwards, and partly also due to the mechanical properties of the rheologically stronger metabasites. The main cleavage in axial zone of the cleavage fan is affected by the second set of subhorizontal or gently inclined planar fabric characterized by development of kink bands perpendicular or oblique to strongly developed early Alpine anisotropy. This later cleavage forms concave regional structure in axial part of the cleavage fan.

The structure of the Gemer unit is different in the SW in contact with crystalline basement of the Vepor unit. Here, the whole succession is exceptionally narrow (several km) and exhibits extremely strongly developed cleavage fans. The geometry of planar structures, subhorizontal stretching and numerous sense of shear criteria suggests sinistral transpression.

The Alpine reworking of adjacent part of the Vepor unit was found to have also been very intense. Here, the Vepor gneisses show strong greenschist facies transpressive reworking, the intensity of which gradually decreases to the north. The internal part of the Vepor basement has been affected by two major NE-SW trending steep transpressive shear zones several kilometres wide. These zones are associated with intense refolding of Variscan anisotropy (often complete transposition of the early fabric) and dominant NE-SW trending subhorizontal stretching lineation. Exhumation of structurally lower micaschists emplaced into supracrustal levels is a prominent feature of intra-Vepor transpressive shear zones. The high-grade gneisses that occurred primarily above micaschists were affected by intense greenschist transtensional shearing with flat planar fabrics and E-W trending stretching.

The Cretaceous structural pattern of the SE Western Carpathian Variscan basement and its Mesozoic cover can be interpreted as a result of oblique convergence. The first event is responsible for thrusting of Mesozoic sequences over Variscan basement rocks. The formation of a Mesozoic accretionary wedge started during the Late Jurassic and culminated during the Cretaceous. The ongoing Cretaceous compression led to significant shortening of weak Early Paleozoic flysch sediments of the Gemer unit. The deformation of this unit is primarily controlled by the shape of the north-east trending body of the Vepor crystalline basement which acted as a rigid indenter relative to the weak Gemer metasediments. The northwards oriented stress was continuously transmitted farther to the north leading to underthrusting of the Vepor unit by the stable European lithosphere represented by the Tatra unit, producing E-W flat extension of the central part of the Vepor gneisses. Continuous shortening of Gemer metasediments resulted in development of transpression tectonics along the SE edge of the Vepor basement. Strongly shortened cleavage fan suffered lateral extrusion to the NE, and the adjacent Vepor gneisses were intensively sheared. The interaction of these two regimes resulted in the development of complex transpressional–transtensional tectonics within the Vepor basement described above.

References

Allemand P. & Lardeaux J.M., 1997: Strain partitioning and metamorphism in a deformable orogenic wedge; application to the Alpine belt. In: Schulmann K (Ed.): Thermal and mechanical interactions in deep-seated rocks. Elsevier, Amsterdam, Netherlands. *Tectonophysics,* 280, 157–169.

Andrusov D., 1936: Subtatranské príkrovy v Západných Karpatoch. In: Carpathica 1, Praha, 3–50.

Andrusov D., 1958: Geológia Československých Karpát Zv. 1. *Vydavateľstvo SAV*, Bratislava.

Bezák V., 1992. The relationship of granitoids and metamorphic rocks in the crystalline complexes of the Tatricum and Veporicum (Western Carpathians); a combination of the effects of Hercynian and Alpine tectonics. In: Vozár J. (Ed.): Special volume to the problems of the Paleozoic geodynamic domains; Western Carpathians, Eastern Alps, Dinarides. *Dionýz Štúr Inst. Geol*., Bratislava, Czechoslovakia, 21–24.

Biely A., 1989: The geological structure of the West Carpathians. In: Nairn A.E.M. (Ed.): Evolution of the northern margin of Tethys; the results of IGCP Project 198; Volume II. Occasional Publications ESRI, New Series 4. Earth Sciences and Resources Institute, *University of South Carolina*, Columbia, SC, United States, 51–57.

Chopin C., 1984: Coesite and pure pyrope in high-grade blueschists of the Western Alps; a first record and some consequences. *Contr. Mineral. Petrology,* 86, 2, 107–118.

Davy P. & Gillet P., 1986: The stacking of thrust slices in collision zones and its thermal consequences. *Tectonics,* 5, 6, 913–929.

Faryad S.W., 1995: Phase petrology and P-T conditions of mafic blueschists from the Meliata unit, West Carpathians, Slovakia. *J. Metamorphic Geol.,* 13, 6, 701–714.

Faryad S.W. & Bernhardt H.J., 1996: Taramite-bearing metabasites from Rakovec (Gemeric Unit, the Western Carpathians). *Geol. Carpathica,* 47, 6, 349–357.

Klinec A. & Miko O., 1987: Tectonic correlation of the Vepor crystalline units in the KV-1 borehole, Pohronská, Polhora. *Geol. Průzk.,* 29, 4, 103–105 (in Slovak).

Mock R. & Reichwalder P., 1992: The blueschist belt of the Slovak karst, Slovakia. In: ALCAPA; Geological evolution of the internal Eastern Alps, Carpathians and of the Pannonian Basin. (Edited by Anonymous). *Terra Abstr.*, 4. *Blackwell Sci. Publ.,* Oxford, International, 46.

Plašienka D., 1991: Mesozoic tectonic evolution of the epi-Variscan continental crust of the central Western Carpathians; a tentative model. *Miner. slovaca*, 23, 5–6, 447–457.

Plašienka D., 1995: Cleavages and folds in changing tectonic regimes; the Vel'ký Bok Mesozoic cover unit of the Veporicum (Nízke Tatry Mts., central Western Carpathians). *Slovak Geol. Mag*., 2–95, 97–113.

Platt J.P., 1993: Exhumation of high-pressure rocks; a review of concepts and processes. *Terra Nova,* 5, 2, 119–133.

Snopková P. & Ivanička J., 1981: Silurian age of phyllites of Stara Voda-SV-1 Borehole, Spiš-Gemer Mountains. *Geol. Práce, Zpr.,* 75, 19–26 (in Slovak).

Snopková P. & Snopko L., 1979: Biostratigraphy of the Gelnica Group in the Spiš-Gemer, based on palynological data (in Slovak).

Spalla M.I., Lardeaux J.M., Dal P.G.V., Gosso G. & Messiga B., 1996: Tectonic significance of Alpine eclogites. *J. Geodynamics*, 21, 3, 257–285.

Vozárová A., 1993: Variscan metamorphism and crustal evolution in the Gemericum *Západ. Karpaty, Sér. Miner., Petrol., Geochém. Metalogen*., 16, 55–117 (in Slovak).

Vozárová A., Soták J. & Ivanička J., 1999: Cambro-Ordovician fossils (conodontes, foraminifers, chitinous shields) from the methamorphic series of the Gemericum (Western Carpathians). In: Tenth meeting of European Union of Geosciences (Edited by Anonymous). *Abstr.*, 4. *Cambridge Publ.*, 266.

DECOMPRESSION AND EXHUMATION OF VARISCAN OROGENIC ROOT IN THE TATRA MOUNTAINS, WESTERN CARPATHIANS: EVIDENCE FROM HIGH-GRADE METAPELITES

L. LUDHOVÁ [1] and M. JANÁK [2]

[1]Department of Mineralogy & Petrology, Comenius University, Mlynská dolina G, 84216 Bratislava, Slovak Republic; ludhova@fns.uniba.sk
[2] Geological Institute, Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; geolmjan@savba.savba.sk

Key words: Western Carpathians, Tatra Mountains, Variscan orogeny, decompression, high-grade metapelites.

Introduction

The aim of this paper is to document decompression and exhumation of Variscan basement in the Western Carpathians based on study of high-grade metapelites in the Tatra Mountains. As documented in detail by previous studies (Janák et al. 1999; Ludhová & Janák 1999), these rocks followed a clockwise *P-T* path involving muscovite and biotite dehydration-melting, and cordierite formation during decompression. Synkinematic granite intrusion provided some extra advected heat source at mid crustal levels. Decompression and exhumation was accomplished by deep-crustal thrusting and orogen-parallel extension during Carboniferous time (ca. 355–335 Ma).

Geological setting

The Tatra Mountains are the most representative of the core mountains within the Tatric unit, a principal crustal-scale superunit of the Western Carpathians. In general, the crystalline basement of the Tatra Mountains is composed of pre-Mesozoic metamorphic rocks and granitoids, overlain by Mesozoic and Cenozoic sediments. The metamorphic rocks are most abundant in the western part (Western Tatra Mts.), whereas in the eastern part (High Tatra Mts.) the granitoids are more common. The basement belongs to two tectonic units, differing in metamorphic grade and lithology that are separated by major Variscan thrust fault (Janák 1994). The *lower unit* is exposed only in the Western Tatra Mts. It is composed of micaschists, exhibiting a medium grade metamorphism. Two metamorphic zones, the kyanite-staurolite zone and the kyanite-sillimanite zone have been distinguished (Janák 1994). The *upper unit* shows high-grade metamorphism and migmatization. The base of the unit is formed by orthogneisses, kyanite bearing paragneisses and banded amphibolites with eclogitic relicts (Janák et al. 1996). They belong to the kyanite zone of high-pressure and high-temperature metamorphism. Higher levels are occupied by sillimanite, K-feldspar and cordierite bearing migmatites. They belong to the sillimanite zone, characterised by medium to low pressure and high temperature metamorphism (Janák et al. 1999). The granitoids form a sheet like pluton, ranging from leucogranites to biotite tonalites and amphibole diorites (Kohút & Janák 1994).

A polyphase, Variscan and Alpine deformation under distinct P-T conditions was recognized in the Tatra Mountains (Kahan 1969; Fritz et al. 1992). The earliest, Variscan deformation D_1 is related to top to the S-E thrusting of the upper unit onto the lower unit. Deformation D_2 is manifested by re-orientation and transposition of the lineation in metamorphites into the west-east direction, the sense of shearing is dextral, or top-to-the east. The same W-E ori-

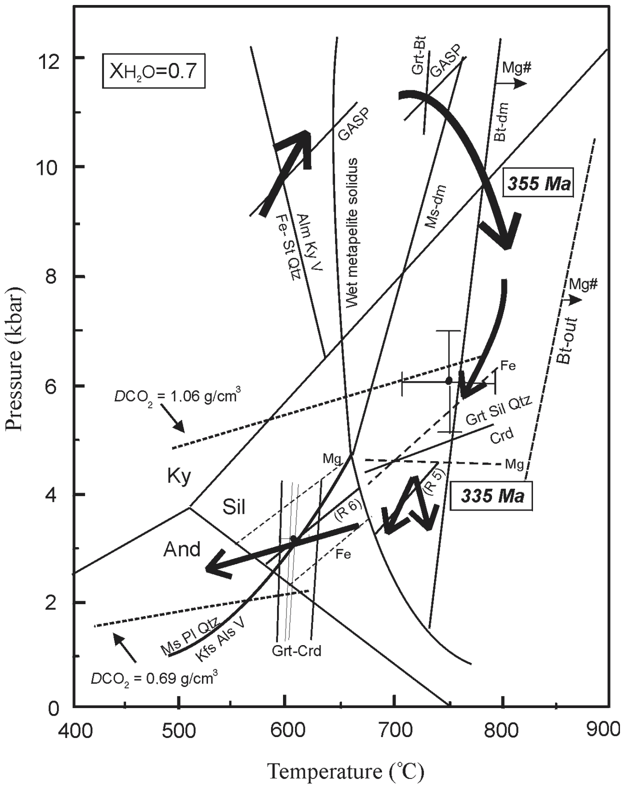

Fig. 1. Quantitative petrogenetic grid and *P-T* path of Tatra Mts. upper unit metapelites. Wet metapelite solidus as well as dehydration-melting curves of muscovite and biotite are according Thompson (1990).

entation of fabrics is in the marginal zones of the granitoid pluton that are affected by solid-state deformation. Deformation D_2 is attributed to Variscan, orogen-parallel extension (Janák et al. 1999). Alpine deformation D_3 at brittle conditions is manifested by top to the N-W shear and the last major deformation D_4 is related to updoming and normal faulting in a N-S to NW-SE direction during Tertiary extension.

The oldest tectono-metamorphic events in the Tatra Mountains seem to be Early Devonian (ca. 405 Ma), according to Pb-Pb single zircon data from orthogneisses (Poller et al. 1997, 1999b). A multistage granitoid magmatism from Late Devonian to Late Carboniferous (360–315 Ma) is recorded by zircon data (Poller et al. 1999a,b). Cooling ages of micas from granitoids and migmatites range from 330 to 300 Ma (^{40}Ar-^{39}Ar method, Maluski et al. 1993; Janák 1994).

Metamorphic P-T-t path

Metapelitic migmatites from the upper tectonic unit contain the following mineral assemblages:

a) in the kyanite zone: garnet + kyanite ± fibrolitic sillimanite + biotite + plagioclase ± K-feldspar ± muscovite with staurolite relics

b) in the sillimanite zone: garnet + prismatic and fibrolitic sillimanite + biotite + quartz + plagioclase ± K-feldspar ± muscovite;

cordierite + biotite + prismatic and fibrolitic sillimanite ± garnet + quartz + plagioclase ± K-feldspar ± muscovite

Based on reaction textures and petrogenetic grid for KFMASH system (Vielzeuf & Holloway 1988), metamorphic *P-T* path was reconstructed (Janák et al. 1999; Ludhová & Janák 1999). Pressure and temperature conditions were calculated by TWEEQU method (Berman 1991) with thermodynamic data of Berman (1988, updated in March 1997). Thermobarometric and phase equilibria calculations are summarized in Fig. 1.

Staurolite and kyanite relics document an early stage of the prograde metamorphism at ~ 600 °C and 9–10 kbar. An increase in temperature to >730 °C at 11–12 kbar resulted in partial melting and incipient migmatisation in the stability field of kyanite. Further heating at decreasing pressure during the earliest stage of exhumation led to the dehydration-melting of muscovite (Mu + Qtz ± Pl = Kfs + Als + L) and biotite (Bt + Als ± Pl + Qtz = Grt + Kfs + L) at >750–800 °C and 6–10 kbar, producing garnet-bearing granite as leucosomes in migmatite. However, it is assumed that thermometric results record the post-peak rather than maximum temperatures due to some retrograde Fe-Mg exchange between garnet and biotite during cooling. Subsequent cooling is documented by garnet resorption by biotite and sillimanite (a reversal of the prograde biotite dehydration-melting reaction). Growth of cordierite from garnet documents a decompression below 5 kbar. Further cordierite and melt formation from biotite and sillimanite decomposition (Bt + Als + Qtz ± Pl = Cor + Kfs + L) is suggested by the reaction (R 5). Crossing this equilibrium by the rock trajectories suggests either nearly isothermal decompression or some heating through heat advection from an external source, such as granite intrusion.

Thermal effect of synkinematic granite intrusion at the contact with cordierite bearing migmatite was tested by thermal modelling (Ludhová & Janák 1999), suggesting that intrusion caused only limited (~50 °C) increase in temperature at the contact with surroundings rocks (arrows crossing R 5, Fig. 1). This implies that advection of heat from synkinematic intrusion might maintain, at least close to the contact, the migmatites sufficiently hot during decompression, preventing rapid cooling and crystallization of melt.

Nearly isobaric cooling at a pressure of ~3 kbar led to pinitization of cordierite (R 6) at a temperature below 612 °C ± 20 °C.

Tectonic implications

A clockwise *P-T* path of high-grade metapelites from the upper, allochthonous unit of the Tatra Mountains suggests substantial crustal thickening followed by exhumation during the Variscan orogeny. The timing of high-pressure stage is believed to be related to crustal thickening at ca. 405 Ma, as indicated by available geochronological data (Poller et al. 1997, 1999b). High-pressure event in the Tatra Mountains is documented by eclogitic relics (Janák et al. 1996), occurring near the base of the upper unit, within the kyanite zone metapelites. It is inferred that the high-pressure metamorphism resulted from northwards oceanic subduction in the Eo-Variscan time. Following continental collision and lithospheric thickening, delamination, or detachment of the lithospheric mantle with concomitant asthenospheric upwelling, could provide sufficient heat to produce the granulite facies migmatites by dehydration melting. Moreover, this might trigger rapid decompression and uplift of the thermally weakened lower crust. Overturn of a hot crustal root with synkinematic granitoid magma intrusion could produce an extra advected heat source at mid crustal levels to produce cordierite-bearing migmatites. In fact, decompression of high-grade, partially melted migmatites in the Tatra Mountains (Fig. 1) overlapped intrusion of the granitoids between 355–335 Ma, as recorded by zircon ages (Poller et al. 1999a,b). Decompression and exhumation was accomplished by top-to-the-south-southeast thrusting as well as presumably concomitant west-east (orogen-parallel) extension. The thrust décollement was probably initiated by the presence of melt which localised a decoupling between the migmatites in the hangingwall and the micaschists in the footwall. Finaly, the rocks cooled between 300–330 Ma, as recorded by Ar-Ar data (Maluski et al. 1993; Janák 1994).

References

Berman R.G., 1988: Internally-consistent thermodynamic data for minerals in the system: Na_2O-K_2O-CaO-MgO- FeO-Fe_2O_3-Al_2O_3-SiO_2-TiO_2-H_2O-CO_2. *J. Petrology*, 29, 45–522.

Berman R.G., 1991: Thermobarometry using multi-equilibrium calculations: a new technique, with petrological applications, *Canad. Mineralogist*, 29, 833–855.

Fritz H., Neubauer F., Janák M. & Putiš M., 1992: Variscan mid-crustal thrusting in the Carpathians II: Kinematics and fabric evolution of the Western Tatra basement. *Terra Abstr.*, *Suppl.* 2, 4, 24.

Janák M., 1994: Variscan uplift of the crystalline basement, Tatra Mts., Central Western Carpathians, evidence from 40Ar/39Ar laser probe dating of biotite and P-T-t paths. *Geol. Carpathica*, 45, 293–300.

Janák M., O´Brien P.J., Hurai V. & Reutel C., 1996: Metamorphic evolution and fluid composition of garnet-clinopyroxene amphibolites from the Tatra Mountains, Western Carpathians. *Lithos,* 39, 57–79.

Janák M., Hurai V., Ludhová L., O´Brien P.J. & Horn E.E., 1999: Dehydration melting and devolatilization of high-grade metapelites: the Tatra Mountains, Western Carpathians. *J. Metamorphic Geol.* (in print).

Kahan Š., 1969: Eine neue Ansicht uber den geologischen Aufbau des Kristallinikums der West Tatra. *Acta Geol. Geogr., Univ. Comen.,* 12, 115–122.

Kohút M. & Janák M., 1994: Granitoids of the Tatra Mts., Western Carpathians: field relationships and petrogenetic implications. *Geol. Carpathica*, 45, 301–311.

Ludhová L. & Janák M., 1999: Phase relations and *P-T* path of cordierite-bearing migmatites, Western Tatra Mountains, Western Carpathians. *Geol. Carpathica,* 50*,* 283–293.

Maluski H., Rajlich P. & Matte P., 1993: ^{40}Ar-^{39}Ar dating of the Inner Carpathians Variscan basement and Alpine mylonitic overprint. *Tectonophysics,* 223, 313–337.

Poller U., Todt W., Janák M. & Kohút M., 1997: Pre-Variscan orthogneisses of the Western Tatra Mountains: U-Pb single zircon data by cathodoluminescence controlled dating (CLC-dating). *Abstr. suppl.* 1*,* 9, 488.

Poller U., Todt W., Janák M. & Kohút M., 1999a: U-Pb single zircon TIMS data in combination with CL and EMP Studies on granitoids of the High Tatra Mountains (Slovakia). *J. Conf. Abstr*., 4, 669.

Poller U., Todt W., Janák M. & Kohút M., 1999b: The geodynamic evolution of the Tatra Mountains constrained by new U-Pb single zircon data on orthogneisses, migmatites and granitoids. *Geol. Carpathica* (this volume).

Thompson A.B., 1990: Heat, fluids, and melting in the granulite facies. In: Vielzeuf D. & Vidal Ph. (Eds.): Granulites and Crustal Evolution. *Kluwer,* Dordrecht, 37–57.

Vielzeuf D. & Holloway J.R., 1988: Experimental determination of the fluid-absent melting relations in the pelitic system. Consequences for crustal differentiation. *Contr. Mineral. Petrology*, 98, 257– 276.

ALPINE METAMORPHISM OF THE MESOZOIC COVER ROCKS FROM THE VEPORIC UNIT, WESTERN CARPATHIANS: X-RAY AND EMP STUDY OF WHITE MICA AND CHLORITE

B. LUPTÁK[1], M. FREY[2], M. JANÁK[3], S. SCHMIDT[2] and D. PLAŠIENKA[3]

[1]Department of Mineralogy and Petrology, Comenius University, Mlynská dolina G, 842 15 Bratislava, Slovak Republic; bluptak@fns.uniba.sk
[2]Mineralogisch-Petrographisches Institut, Universität Basel, Bernoullistrasse 30, CH 4056 Basel, Switzerland
[3]Geological Institute, Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; geolmjan@savba.savba.sk

Key words: Western Carpathians, Veporic unit, Alpine orogeny, low-grade metamorphism.

Introduction

The Veporic superunit is the middle of three thick-skinned basement thrust sheets of the Central Western Carpathians (CWC). The imbricated crustal structure of the CWC originated during the Cretaceous continental collision after the Late Jurassic closure of the Meliata ocean. As a consequence of collisional thickening, the Veporic basement and its Permian-Mesozoic cover suffered regional Alpine metamorphism. Dominant Veporic macrostructure, the low-angle normal ductile shear zone, is interpreted as a master detachment fault that exhumed the Veporic core by east-vergent unroofing (Plašienka et al. 1999).

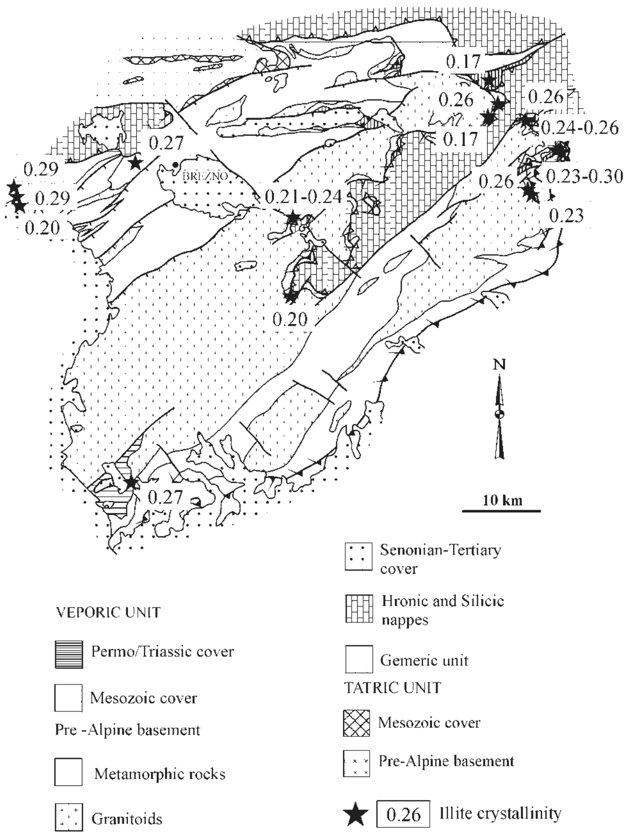

Fig. 1. Simplified geological map of the Veporic unit with locations of investigated samples and illite crystallinity data.

Table 1: Chemical composition of white mica and chlorite.

Sample Analysis #	DOPO2 Phn1	DOPO4 Phn3	Sample Analysis #	VAL1 Chl11	ZE1 Chl7
SiO_2	47.58	50.88	SiO_2	30.86	25.41
Al_2O_3	26.84	28.56	Al_2O_3	25.19	23.41
TiO_2	1.12	0.21	TiO_2	0.00	0.00
MgO	2.23	4.37	MgO	21.99	16.66
FeO	6.36	0.07	FeO	10.09	22.25
MnO	0.00	0.00	MnO	0.04	0.19
CaO	0.00	0.10	CaO	0.01	0.04
Na_2O	0.15	0.13	Na_2O	0.19	0.06
K_2O	11.33	11.68	K_2O	0.91	0.17
Total	95.61	96.00	Total	89.27	88.20
Si	6.530	6.709	Si	5.827	5.230
Aliv	1.470	1.291	Al	5.607	5.681
Alvi	2.872	3.149	Ti	0.000	0.000
Ti	0.116	0.021	Mg	6.190	5.111
Mg	0.456	0.859	Fe^{2+}	1.593	3.830
Fe^{2+}	0.730	0.008	Mn	0.006	0.033
Mn	0.000	0.000	Ca	0.002	0.009
SumOct	4.174	4.037	Na	0.070	0.024
Ca	0.000	0.014	K	0.219	0.045
Na	0.040	0.033			
K	1.983	1.965	Fe/Fe+Mg	0.205	0.428
SumA	2.023	2.012			
Fe/Fe+Mg	0.616	0.009			
X(Ca)	0.000	0.007			
X(Na)	0.020	0.016			
X(K)	0.980	0.977			

Structural formulae calculated on the basis of 22 oxygen atoms for white mica and 28 for chlorite

Triassic metaquartzites, metacarbonates and metapelites are the main lithological varieties of studied metasedimentary cover rocks. These rocks display a penetrative, flat-laying mylonitic foliation and stretching, and/or mineral lineation trending generally W-E, formed during the Alpine D_1 deformation stage.

Analytical methods

Metamorphic conditions were evaluated by measuring the Kübler index (crystallinity) of white mica and chlorite. Samples were crushed in Sieb Mill for less than 30 s. Carbonate was removed by treatment with 5 % acetic acid. Clay fraction was received by settling method in a water column. X-ray diffractograms were made from < 2 μm clay fraction of illite-muscovite ± chlorite concentrates as oriented slides (Siemens D-5000 diffractometer at 40 kV, 30 mA and CuKα radiation, from 2^o–21^o 2θ and scan rate 0.03^o/20 s, Univ. Basel). Illite and chlorite crystallinity data — half-height width of the first illite basal reflection and second chlorite basal reflection (IC, ChC) were calculated from more than 30 samples using the EVA MFC Application, DIFRACPlus evaluation program and Profile plus v. 1.06, DIFRACPlus software.

Chemical compositions of phyllosilicates were investigated by electron microprobe JEOL 8600 (Univ. Basel) at operating conditions of 10 nA and 15 kV. The data were reduced by the ZAF and PROZA correction method.

Results

Clay mineral crystallinity data for the basal reflections of illite-muscovite are shown in Fig. 1. Diffraction maximum for the first basal reflex of white mica is well developed with a narrow width, documenting high crystallization degree of mica crystals. Correlation between white mica full width at half maximum of diffraction maximum 001 (FWHM) and chlorite 002 FWHM, in samples containing both phases, points to epizonal (subgreenschist to greenschist facies) metamorphic conditions.

Authigenic white mica shows considerable Tschermak substitution and high Si content (up to 6.7 Si atoms per formula unit). In the metacarbonates of dolomitic origin, phyllosilicates are represented mainly by newly formed greenish chlorite. Representative chemical compositions of metamorphic white mica and chlorite are shown in Table 1.

Conclusions

Our results suggest that Mesozoic cover rocks in the Veporic unit have been metamorphosed at upper anchizonal and epizonal (greenschist facies) metamorphic conditions. Alpine metamorphism caused total recrystallization of former clay minerals and the growth of newly formed white mica and chlorite. These data point to Alpine regional metamorphism increasing from the Mesozoic cover to underlying Upper Paleozoic and basement rocks in the Veporic unit (Lupták et al. 1998; Janák et al. 1999; Plašienka et al. 1999).

References

Janák M., Plašienka D., Lupták B. & Frey M., 1999: Alpine metamorphism in the Central Western Carpathians: relationships to Cretaceous continental collision. *J. Conf. Abstr.,* 4, 1, *EUG 10 Abstr.,* 74.

Lupták B., Janák M., Plašienka D. & Frey M., 1998: Chloritoid-kyanite schists from the Veporicum: reaction textures, phase equilibria and implications for the Alpine (Cretaceous) metamorphism of the Western Carpathians. *CBGA XVI Congress Vienna, Abstr.*, 348.

Plašienka D., Janák M., Lupták B., Milovský R. & Frey M., 1999: Kinematics and metamorphism of a Cretaceous core complex: the Veporic unit of the Western Carpathians. *Phys. Chem. Earth* (in print).

PHOSPHATE ACCESSORY MINERALS IN HIGH TATRA GRANITOIDS

M. MICHALIK[1] and Ł. SKUBLICKI[2*]

[1]Institute of Geological Sciences, Jagiellonian University, ul. Oleandry 2a, 30-063 Kraków, Poland; michalik@ing.uj.edu.pl
[2]Department of Geological Sciences, University of Saskatchewan, 114 Science Place, Saskatoon, SK S7N 5E2, Canada; us821@mail.usask.ca

(*previously: Institute of Geological Sciences, Jagiellonian University, ul. Oleandry 2a, 30-063 Kraków, Poland)

Key words: granitoids, phosphorus in magma, monazite, apatite, High Tatra, Carpathians.

Introduction

Phosphorus in granites is incorporated in phosphate accessory minerals or in alkali feldspars and acid plagioclases (London 1992; Cerny 1995; Fryda & Breiter 1995). In Ca-poor granites alkali feldspars are the major or principal reservoir of phosphorus (London 1992). Phosphorus plays an important role in controlling trace element abundance and their partitioning between crystals, melt and vapour phases, in depolymerization of aluminosilicate melts, in lowering the liquidus temperature, and lowering viscosity (Mysen 1988; Bea et al. 1992; Dingwell et al. 1993). Different types of granitoids (A-, felsic I- and felsic S-types) differ in P_2O_5 content (Bea et al. 1992). The enrichment in phosphorus is typical of peraluminous and strongly fractionated granite melts of S-type (London 1992; Breiter et al. 1997).

Monazite and apatite are the commonly noted accessory minerals in granitoids. Recognition of relations among these phosphate accessory minerals grown from magmas and of restitic origin is important for understanding magma evolution. Understanding of the processes of monazite growth and breakdown in granitoids under metamorphic conditions or during hydrothermal alteration is also important because monazite crystals contain a significant part of the LREE of the rock (40–80 %; Finger et al. 1998), such that their presence or destruction greatly influences REE and characteristics mobility. Recognition of different types of monazite in granitic rock is also important because monazite incorporates large quantities of Th and U and can be used for chemical Th-U-Pb dating (e.g. Suzuki et al. 1991; Suzuki & Adachi 1994; Montel et al. 1996; Cocherie et al. 1997).

Geological setting

The study area is situated in the northern part of the High Tatra chain. Granodiorites and tonalites predominate in this area. Granites are scarce and occur mostly as pegmatitic bodies. Most rocks are peraluminous; metaluminous ones occur subordinately (Degenhardt et al. 1996). Apatite, zircon, monazite are common accessory minerals; titanite and xenotime occur rarely (Woldańska et al. 1998). Most crystals of allanite and epidote are secondary. Hematite, magnetite, and other opaque minerals are also present (Woldańska 1995; Woldańska et al. 1998). The High Tatra granitoids exhibit numerous alterations — albitization and sericitization of feldspars, growth of secondary muscovite, chloritization of biotite, growth of epidote (and/or allanite), carbonatization and prehnitization (Michalik et al. 1998). Breakdown of monazite and development of irregular coronas of apatite in reaction zone is also present (Broska & Siman 1998; Michalik & Skublicki 1998).

Methods of study

Samples were studied using both transmitted and reflected light optical microscopy, electron microscopy combined with X-ray energy dispersive analysis, X-ray diffraction, and various chemical analyses.

Phosphorus in High Tatra granitoids

The content of P_2O_5 in the study area falls into a range of 0.0–0.09 wt. %; only in a few samples higher values were noted (up to 0.46 wt. %); (Turnau-Morawska 1974). Michalik and co-workers obtained similar values (0.05–0.09 wt. %), (Michalik et al., unpublished data). A positive correlation between P_2O_5 and CaO content is apparent.

Occurrence of apatite and monazite in High Tatra granitoids

Apatite

Apatite, which occurs usually in close association with biotite and opaque minerals, is the most frequent accessory mineral in the High Tatra granitoids. Two types of apatite can be distinguished. The bigger ones are rounded, almost isometric, often fractured and greyish. Smaller apatite crystals, which form euhedral prisms or rods, are colourless or yellowish. All apatites are fluorapatites; chlorine is present only in a few of them. In numerous apatite crystals very low amounts (< 0.1 wt. %) of Mn were determined. REE were measured in a few crystals of apatite that form intergrowths with monazite.

Monazite

Monazite in the High Tatra granite is a relatively common accessory mineral. Most often monazite is present in clots with Fe-Ti oxides (in Fe-Ti oxides or in their rims) and biotite (or more rarely with feldspars). Monazite associated with Fe-Ti oxides is rounded or irregular in shape. Monazite, which occurs together with biotite, is also rounded or rarely subhedral.

Monazites from the High Tatra granitoids differ in chemical composition. Besides different amounts of LREE (Ce and La >>Nd, Sm), the content of Th, U, Ca and Si is also variable. It indicates the presence of substitution Th^{4+} – REE^{3+} compensated by PO_4^{3-} – SiO_4^{4+} and Ca^{2+} + Th^{4+} – $2REE^{3+}$.

Reaction zones are present around monazites associated with biotite and/or feldspars. The dominant process is the transformation of monazite to apatite. Apatite coronas are usually discontinuous and irregular in shape. Contacts between monazite and apatite in coronas are complex. Apatite veins crosscutting monazite grains and apatite patches in monazite are present. In a few cases small allanite crystals occur in close proximity to partly transformed monazites.

The occurrence of transformations of this type has not been noted in cases of monazites incorporated within Fe-Ti oxides.

Discussion and conclusions

The common occurrence of apatite as an accessory mineral in High Tatra granitoids and correlation between P_2O_5 and CaO indicate that apatite is the main reservoir of phosphorus. The typical P_2O_5 content in the studied granitoids is below the apatite saturation value in felsic magmas of both I- and S-types granites, which occurs at a P_2O_5 concentration of 0.14 wt. % (Watson & Copabianco 1981), suggesting that apatite crystallized from magma. Two different types of apatite crystals are probably related to different conditions of growth. The presence of apatite xenocrysts can be excluded.

Monazite is a relatively rare mineral in the High Tatra granitoids. It is difficult to interpret the meaning of differences in chemical composition, morphology of crystals, and associations. Preliminary age determinations (Michalik & Paszkowski 1997) suggest the presence of at least two generations of monazites.

Finger et al. (1998) and Broska & Siman (1998) have described transformation of monazite to apatite and allanite in granitoids under amphibolite facies conditions. Transformations of this type have been described earlier in the High Tatra granitoids (Broska & Siman 1998; Michalik & Skublicki 1998) however. The High Tatra granitoids have never been subjected to amphibolite facies conditions. Initial monazite-apatite transformation in the Tatra granitoids was probably induced by fluids active during alteration of feldspars and biotites. The limited activity of Ca^{2+} supplied during albitization and sericitization of plagioclases was probably the factor which controlled degree of monazite-apatite transformation. Monazites incorporated within Fe-Ti oxides have not been subjected to these transformations, Fe-Ti oxides having acted as a barrier prohibiting the access of fluids to monazite crystals.

Allanite only rarely occurs in close vicinity to transformed monazite. It never forms a corona around altered monazite as was described by Finger et al. (1998) and Broska & Siman (1998). It is possible that the monazite-apatite transformation reaction was important in mobilization of REE-containing fluids. Allanite, which is scarce in High Tatra granitoids, usually occurs independently of transformed monazites. Although it is difficult to estimate the real scale of REE-containing fluids mobilization, it is likely that this mobilization was not merely local.

References

Bea F., Fershtater G. & Corretge L.G., 1992: The geochemistry of phosphorus in granitic rocks and the effect of aluminium. *Lithos,* 29, 43–56.

Breiter K., Seltman R. & Thomas R., 1997: The phosphorus-rich rare metal granite system of Podlesi, Czech Republic. In: H. Papunen (Ed.): *Mineral Deposits*, Balkema, Rotterdam, 617–620.

Broska I. & Siman P., 1998: The breakdown of monazite in the West-Carpathian Veporic orthogneisses and Tatric granites. *Geol. Carpathica*, 49, 161–167.

Cerny P., 1995: Advances in Mineralogy — The key to petrologic and genetic problems. In: Pasava J. et al. (Eds.): *Mineral Deposits*, Balkema, Rotterdam, 427–430.

Cocherie A., Legendre O., Peucat J.J. & Kouamelan A., 1997: *In situ* Th-U-Pb dating using an electron microprobe: a powerful tool for complex polygenic monazites. *IX EUGS Meeting, Abstr., Terra Nova,* 9, 441.

Degenhardt J., Michalik M., Skublicki Ł. & Woldańska B., 1996: Petrology and geochemistry of granitoids from Polish High Tatra Mts. *Pol. Tow. Miner., Prace Spec.*, 7, 44–47 (in Polish).

Dingwell D.B., Knoche R. & Webb S.L., 1993: The effect of P_2O_5 on viscosity of haplogranitic liquid. *Eur. J. Mineral.*, 5, 133–140.

Finger F., Broska I., Roberts M.P. & Schermaier A., 1998: Replacement of primary monazite by apatite-allanite-epidote coronas in an amphibolite facies granite gneiss from the eastern Alps. *Amer. Mineralogist*, 83, 248–258.

Fryda J. & Breiter K., 1995: Alkali felspars as a main phosphorus reservoirs in rare-metal granites: three examples from the Bohemian Massif. *Terra Nova*, 7, 315–320.

London D., 1992: Phosphorus in S-type magmas: the P2O5 content of feldspar from peraluminous granites, pegmatites and rhyolites. *Amer. Mineralogist*, 77, 126–145.

Michalik M. & Paszkowski M., 1997: Dating of minerals based on SEM-EDS analyses (example of the High Tatra monazites). *IV Ogólnopolska Sesja Naukowa "Datowanie Minerałów i Skał", UMCS Lublin, 11–12 grudnia 1997,* 56–67 (in Polish).

Michalik M. & Skublicki Ł., 1998: Breakdown of monazite during alterations of the High Tatra granitoids. *Pol. Tow. Miner., Prace Spec.,* 11, 145–147 (in Polish).

Michalik M., Skublicki Ł. & Woldańska B., 1998: Postmagmatic alterations of the High Tatra granitoids (Tatra Mts., Poland). In: *Carpathian-Balcan Geological Association XVI Congress, Abstr., Vienna,* 381.

Montel J.M., Foret S., Veschambre M., Nicollet Ch. & Provost A., 1996: Electron microprobe dating of monazite. *Chem. Geol.,* 131, 37–53.
Mysen B.O., 1988: Structure and properties of silicate melts. *Elsevier*, 354.
Watson E.B. & Copabianco C.J., 1981: Phosphorus and rare earth elements in felsic magmas: An assessement of the role of apatite. *Geochim. Cosmochim. Acta*, 45, 2349–2358.
Suzuki K. & Adachi M., 1994: Middle Precambrian detrital monazite and zircon from the Hida gneiss on Oki-Dogo Island, Japan: their origin and implications for correlation of basement gneiss of Southwest Japan and Korea. *Tectonophysics,* 235, 277–292.
Suzuki K., Adachi M. & Tanaka T., 1991: Middle Precambrian provenance of Jurassic sandstone in the Mino Terrane, central Japan: Th-U-total Pb evidence from an electron microprobe monazite study. *Sed. Geol.,* 75, 141–147.
Turnau-Morawska M., 1948: Microgeological research in the central part of the crystalline Tatra. *Kosmos, Ser. A., Rozpr.*, 65, 59–100 (in Polish).
Woldańska B., 1995: Selected accessory minerals in granitoids from NE part of the High Tatra Mts. *Pol. Tow. Miner., Prace Spec.*, 6, 54–56 (in Polish).
Woldańska B., Skublicki Ł., Michalik M. & Degenhardt J., 1998: Primary and secondary accessory minerals in the High Tatra granitoids (Poland). In: *Carpathian-Balcan Geological Association XVI Congress, Abstr., Vienna,* 641.

GARNET IN PEGMATITES FROM ROMANIA AS A METALLOGENETIC INDICATOR

T. MURARIU, S. RĂDĂŞANU and D. ANDRONE

Department of Mineralogy-Geochemistry, Faculty of Geography and Geology, "Al. I. Cuza" University of State, B-dul Carol 20 A, 6 600 Iaşi, Romania; tmurariu@vela.uaic.ro

Key words: garnet, major and minor elements, metallogenetic potential, pegmatites.

Introduction

In the Romanian Carpathians there is a large pegmatitic Province (PPC), consisting of Preluca (I), Rodna (II), Gilău-Muntele Mare (III) and Getic (IV) (Mârza 1980) subprovinces. The PPC pegmatites have a granite-type composition and are considered to be of metamorphic origin (Mârza 1980; Murariu 1980; Hann 1987 etc.). On the basis of mineralogical and geochemical features, these pegmatites belong to the following groups: (1) feldspar pegmatites (quartz+feldspar +muscovite+biotite+tourmaline+garnet±zircon±apatite±beryl ±columbite±tantalite); (2) muscovite pegmatites (muscovite+feldspar +quartz+biotite+tourmaline+garnet±zircon±beryl±apatite) and (3) rare-element pegmatites, which include two types: (3A) beryl-phosphate type (beryl+columbite+triphylite) and (3B) albite-spodumene type (spodumene+montebrasite+tantalite+ cassiterite). The beryl-phosphate type pegmatites occur in the Apuseni Mts. (Gilău-Muntele Mare Subprovince) and in the Southern Carpathians (Getic Subprovince), while the albite-spodumene pegmatites occur only in the latter. According to Uher (1994), the 3A type and the 3B type pegmatites could be correlated, by their mineralogical composition, with the West-Carpathian Moravany granitic pegmatites, which are beryl-bearing pegmatites without or with scarce Nb-Ta minerals, occurring mainly in the Bratislava Massif. The pegmatites from the Romanian Province are hosted by amphibolite facies Precambrian metamorphosed rocks.

This paper shows the role of garnet in the assessment of metallogenetic potential for rare elements in the PPC pegmatites. The almandine-spessartine garnet in the Romanian pegmatites is an accessory mineral whose chemical composition varies (Murariu 1980; Murariu et al. 1986).

Methods

Manual sorting under a binocular microscope and Frantz magnetic separator were used to extract garnets. Chemical analyses were performed by atomic absorption spectroscopy (AAS), with a Perkin Elmer 2380 apparatus.

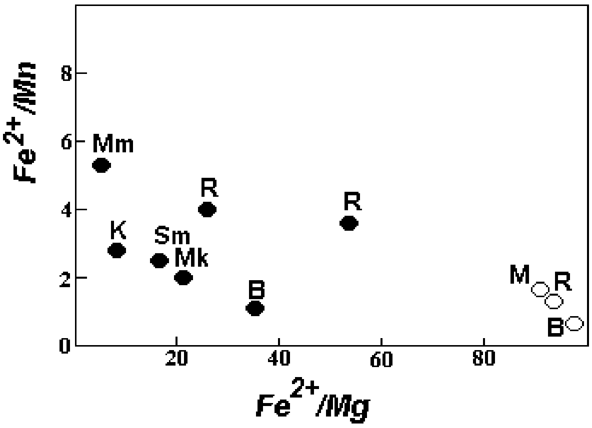

Fig. 1. Plot of (Fe^{2+}/Mn) vs. (Fe^{2+}/Mg) for garnets of barren pegmatites (•) and rare-element productive pegmatites (o).

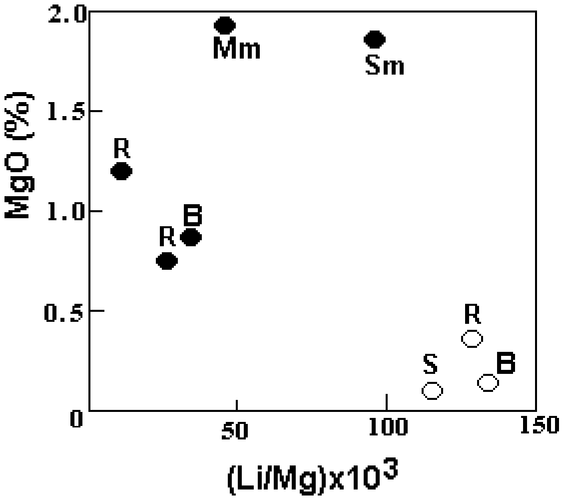

Fig. 2. Plot of MgO vs. (Li/Mg)$\times10^3$ for garnet of barren pegmatites (•) and rare-element productive pegmatites (o).

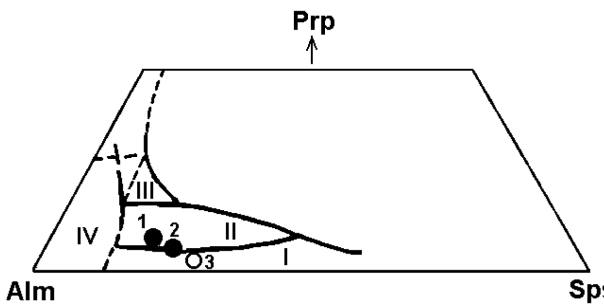

Fig. 3. Plot of some of garnet samples from PPC on diagram Prp vs. Alm vs. Sps: 1 — garnet from feldspar pegmatites, 2 — garnet from mica-bearing pegmatites, 3 — garnet from rare-element pegmatites; I — the field of garnet from rare-element pegmatites, II — the field of garnet from mica-bearing pegmatites, III — the field of garnet from pegmatites without micas, IV — the field of garnet from the amphibolite facies.

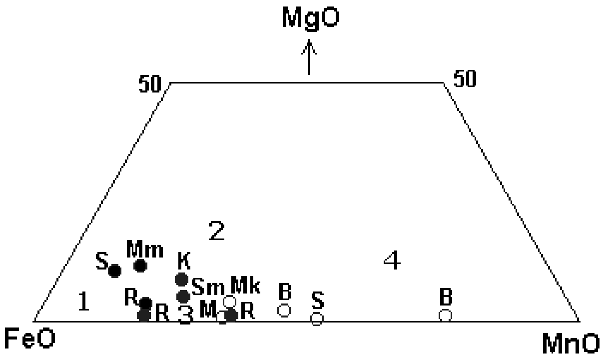

Fig. 4. Compositional fields of garnet from different pegmatite types: 1 — muscovite formation A, 2 — muscovite formation B, 3 — muscovite + rare-element formation, 4 — rare-element formation (after Salye 1973).

Table 1. Chemical compositions of pegmetite garnets from Carpathian Province (Romania).

	Pegmatites		
wt.%	Feldspar	Muscovite	Rare-element
SiO_2	37.94	37.82	37.41
TiO_2	0.08	0.10	0.08
Al_2O_3	18.67	19.25	18.50
Fe_2O_3	3.91	2.21	2.42
FeO	26.88	27.88	25.78
MnO	6.65	7.21	15.28
MgO	1.21	0.67	0.32
CaO	4.78	4.11	1.11
Fe^{2+}/Mg	28.62	54.18	100.20
Fe^{2+}/Mn	4.00	3.80	1.70
F	96.80	98.30	99.10
M	3.50	1.80	0.80
Y (ppm)	700.00	720.00	2200.00
Li (ppm)	82.00	112.00	250.00
$(Y/Si)\times10^3$	3.90	4.07	12.58
$(Li/Si)\times10^3$	0.50	0.63	1.43
$(Li/Mg)\times10^3$	11.23	27.71	129.53
End-member proportion (mol. %)			
Pyrope	5.10	2.87	1.48
Almandine	64.17	66.95	66.65
Spessartite	16.08	17.54	23.43
Andradite	12.74	7.16	3.68
Grossularite	1.85	5.48	4.76

Results and discussion

The garnet chemical composition shows large contents of Mn, Y and Li in rare-element pegmatites and of Ca and Mg in feldspar and muscovite pegmatites as well (Table 1). The garnets in the three types of pegmatites differ from each other by the following geochemical parameters (Table 1):

$$F = 100\ Fe_{tot}/(Fe_{tot}+Mg) \text{ and } M = 100\ Mg/(Fe^{2+}+Mg)$$

F represents the highest and M the lowest contents in the rare-element pegmatites. At the same time, the Fe^{2+}/Mg ratio is higher and the Fe^{2+}/Mn ratio is lower in the rare-element pegmatites, than are the same ratios in the feldspar and muscovite pegmatites (Fig. 1).

In the rare-element pegmatites the Li and Y contents in the garnet significantly increase. The same behavior has the values: $(Li/Si)\times10^3$, $(Li/Mg)\times10^3$ and $(Y/Si)\times10^3$ (Table 1).

In the MgO:$(Li/Mg)\times10^3$ diagram the barren pegmatites are separated from the rare-element ones (Fig. 2).

The metallogenetic potential of pegmatite can be evaluated using some ternary diagrams: Prp:Alm:Sps (Sokolov et al. 1962; Fig. 3); FeO:MgO:MnO (Salye 1975; Fig. 4).

To confirm the importance of pegmatite garnet as a metallogenetic indicator, some comparative garnet chemical analyses presented by Lazarenko et al. (1960), Dávidová (1968), Magakon (1971), Černý (1971), Salye & Glebovitzky (1976), Zagorski & Kuznetzova (1990), Černý et al. (1995), Sekiranov et. al. (1995) were also used: A=Ardino (Bulgaria); B=Baltic Shield; K=Karelia; M=Manitoba; Mk=Malé Karpaty; Mm=Mamsk; Mr=Maršiko-Moravia; S=Siberia; Sm=Smrček; V=Volînia; R=Romania (Figs. 1, 2, 4).

Acknowledgements: This research was supported by the type C Grant, no. 39698 (C.N.C.S.I.S. — Romania).

References

Černý P., 1971: A calcium rich pegmatite garnet from Czechoslovakia. *Neu Jb. Mineral., Abh., Mh.,* 511–514.
Černý P. & Hawthorne F.C., 1982: Selected peraluminous minerals. *Short Course Handbook,* Winnipeg, 8, 163–187.
Černý P., Novák M. & Chapman R., 1995: The Al (Nb,Ta)Ti_{-2} substitution in titanite: the emergence of new species? *Mineral. Petrol.,* 52, 61–73.
Dávidová Š., 1968: Granáty z pegmatitov Malých Karpát. *Acta Geol. Geogr. Univ. Comen., Geol.,* 13, 117–128.
Hann H.P., 1987: Pegmatitele din Carpaţii Meridionali. *Edit. Acad. RSR.,* 141.
Lazarenko K.K., Matkovski O.F., Vinar O., Saşkina V.P., 1960: Mineraloghiia viverjenih kompleksiv zahidnoi Volîni. *Lvov,* 235.
Magakon V.M., 1971: O temperaturah obrazovaniia i poslemagmaticeskaya izmeneniia sliudosnîh pegmatitov Mamskovo raiona. *V kn: Gheochimiia pegmatitov Vostochinoi Sibiri. Nauka,* Moskva, 48–71.
Mârza I., 1980: Considérations génétiques sur les pegmatites du cristallin de Gilău (Mts. Apuseni) et la Province pegmatitique Carpatique. *An I.G.G. LVII,* Bucureşti, 423–432.
Murariu T., 1980: Contribuţii la cunoaşterea granaţilor din pegmatite. *An. Univ. "Al. I. Cuza" Iaşi, XXVI, s. II.b, II-18.*
Murariu T., Ţabliovschi V., Zamârcă A., 1986: Trace elements in Romanian pegmatitic garnets. *An. Univ. "Al. I. Cuza" Iaşi, XXXII, s.II.b, 5–9.*
Salye M.E., 1975: Metalloheneticeskie formaţii pegmatitov dokembriia (Vostochinaia časť Baltiiskovo Ščita). *V kn: Muskovitovie pegmatiti SSSR, Nauka,* Leningrad, 15–36.
Salye M.E. & Glebovitzky V.A., 1976: Metallogenetičeskaia specializaciia pegmatitov. *Nauka,* Leningrad, 174.
Sekiranov A., Peytcheva I. & Shopov Y., 1995: Pegmatites from Ardinoregion in Central Rhodopes, Bulgaria. *Carpathian-Balkan Geol. Assoc. XVI Congress,* Vienna, 541.
Sokolov I.M., Bykova V.S. & Manuylova M.M., 1962: Granatî pegmatitovo jil Severo-Baikalskovo pegmatitovo poiasa. *Zap. VMO,* 91, 5, 537, 549.
Uher P., 1994: The Varistical West-Carpathian granitic pegmatites: Mineralogy, petrogenesis and relationship to pegmatite population in the Eastern Alps and Romanian Carpathians. *Geol. Carpatica,* 45, 5, 313–318.
Zagorski V.E. & Kuznetzova L.G., 1990: Geochimiia spodumenovich pegmatitov i šceločino redkometalnich metasomatitov. *Nauka,* Novosibirsk, 137.

TEPHROSTRATIGRAPHY OF NEOGENE VOLCANICLASTICS (MORAVIA, LOWER AUSTRIA, POLAND)

S. NEHYBA[1], R. ROETZEL[2] and M. ADAMOVÁ[3]

[1]Department of Geology and Paleontology, Faculty of Sciences, Masaryk University, Kotlářská 2, 611 37 Brno, Czech Republic; slavek@gap.muni.cz
[2]Geologische Bundesanstalt, Rasumofskygasse 23, A-1031 Wien, Austria
[3]Czech Geological Survey, Geologická 6, 152 00 Praha-Barrandov, Czech Republic

Key words: Neogene, tephra correlations, source area.

Introduction

Tephrostratigraphy is an effective tool for correlations extending across various depositional environments or basins. Extended tephra beds can play an important role in the definition of "key surfaces" (especially in basins without extensive outcrops and geophysical data) and also in basin analyses. Lower Miocene (Eggenburgian, Ottnangian, Karpatian ?) volcaniclastics have been studied in the area of the Carpathian Foredeep (CF), the Molasse Zone (MZ), the Vienna Basin (VB) and the Polish Lowland (PL). Middle Miocene (Lower Badenian) volcaniclastics have been studied in the area of the CF. More details can be find in Adamová & Nehyba (1998), Nehyba (1997), Nehyba & Roetzel (1999).

Methods

The reliable correlation of extended tephra beds requires a multiple criteria approach to tephra characterization. For that reason grain discrete methods (the physico-chemical properties of glass shards and pyrogennic minerals — biotite, zircon) were combined with bulk rock study (geochemistry, grain-size studies). Study of volcanic glass was oriented towards the chemistry and morphology of glass shards. Biotite chemistry was studied on selected grains. Zircon studies were more extensional. Outer morphology, presence of zonality, older cores and inclusions were observed in the whole zircon spectra (volcanic and nonvolcanic). Typology (Pupin 1980), elongation and fission-track dating were done only for the volcanic zircons.

Results

The studied Lower Miocene (LM) and Middle Miocene (MM) volcaniclastics are products of Plinian or Phreatoplinian eruptions of acidic character (rhyolite, rhyodacite) and represent the distall fallout tephra. The source of volcanic material was from the calc-alkaline volcanic suite of volcanic arc. The distance of the source volcanism is considered to be several hundred kilometres. The LM source was more distant than the MM one. The source of the LM volcaniclastics is most probably located in Northern Hungary and they are the product of the Neogene aerial type of dacite and rhyolite volcanic activity in the Carpatho-Pannonian region (Lower Rhyolite Tuff). The MM volcaniclastics are most probably the product of the differentiation of the aerial type of andesite volcanic activity in the same region (Middle Rhyolite tuff ?). The important role of postdepositional history on the composition, texture and structure of volcaniclastics was proved. The principal results for tephrostratigraphy were obtained from the study of volcanic zircon and content and distribution of Rare Earth elements (REE).

The substantial differences were recognized between LM and MM (Lower Badenian) volcanic zircons. Lower Badenian zircons originated from magma with a higher alkaline content and a lower temperature of crystallization than the parental magma for the LM (Eggenburgian, Ottnangian) zircons. Both LM and MM source magma had a hybrid character. However, the source magma of LM tephra was closer to crustal origin while the source of MM tephra was closer to mantle origin. The results of typology agree well with the uranium content in zircons.

Samples of MM tephra generally have a higher content of REE, clearly defined Eu-anomaly and generally a steep REE profile compared to LM ones. This can be explained by a source of volcanic material from a more differentiated magmatic source than for LM volcaniclastics.

Two horizons (horizon I and horizon II) were recognized within the studied LM tephra. The existence of these horizons is explained by several eruptional phases of the same source (a stratified magmatic chamber). These horizons were effectively used for solving stratigraphic problems of the region (Nehyba in press). The lower horizon (horizon I) has been recognized in southern part of the CF and is of Upper Eggenburgian age with an average radiometric age of 20.3 ± 2.4 Ma. The tephra of the upper horizon (horizon II) recognized in the CF (Czech Republic) can be correlated with tephra layers in the deposits of the Zellerndorf Formation and Langau-Formation in MZ (Austria). These rocks are of Ottnangian age. Volcaniclastics from the locality Herrnbaumgarten (VB) formerly assumed to be correlative with above mentioned tephra layers have different charactericties of volcanic zircons. For this reason, the volcanic material for the tephra in this locality most probably has a source different from that of the other volcaniclastics.

Volcanic zircons from MM (Lower Badenian) volcaniclastics form one uniform group although several individual tephra layers were recognized in core profiles. The average radiometric age for this tephra is 16.2 ± 2.1 Ma. Some differences within the Lower Badenian volcaniclastics were recognized in the content and distribution of REE. Samples from the lowest studied tephra horizons have different ratios of REE (Sm/Eu, Eu/Cd, La/Yb, La/Sm, LREE/HREE), a generally flatter REE profile and an Eu anomaly not as clearly developed. These data indicate a less differentiated magmatic source for the lower horizons than for the upper ones. Further confirmation of this result can be used for the very detailed correlational and basinal studies of the almost uniform Lower Badenian "tegel" deposits. The tephrostratigraphy of these deposits can produce a more precise stratigraphic scale than biostratigraphy.

Two tuffaceous horizons (tonsteins) from Belchatów (PL) were also included in the study. The FT dating of these volcaniclastics brought an average age around 17.0 Ma. The Karpatian age and a source from the northern Transylvanian Basin (Dej tuff) were assumed (Lorenc & Zimmerle 1993). But these volcaniclastics are not homogenous. One horizon has identical zircon characteristics to the Lower Badenian tephra of CF. The results from the second horizon are different. This results, together with an almost complete absence of the Karpatian tephra in the CF and the position of the Dej tuff, can indicate different sources of volcanic componenta and also different ages (Lower Badenian ?) for some of these horizons.

Conclusions

Lower Miocene and Middle Miocene acid tephra from various sedimentary basins of Central Europe can be effectivelly used for correlational and basin analys studies. Correlations of the obtained data with tepra from the possible source areas (Lower and Middle Rhyolite Tuff in Northern Hungary) are necessary for the further advance in the topic. Correlational studies with tephra deposits within Krosno Beds (Radziszów Tuff, Bandrów Tuff, Krzywe tuffs), tuffaceous horizons of the Konin region in Poland, tephra beds in Ukraine and also with selected tephra occurences in Styrian Basin (Schaufelgraben-Gleichenberg and Lobminger Member) and Bavaria could also bring important results (Ebner et al. 1998; Tkačuk et al. 1958; Unger et al. 1990).

Aknowledgements: Study was sponsored by grant of Grant Agency of Czech Republic 205/98/0694.

References

Adamová M. & Nehyba S., 1998: Geochemistry of volcaniclastic rocks of the Carpathian Neogene Foredeep. *Zem. plyn nafta*, Hodonín, 43, 2, 303–312.

Ebner F., Mali H., Obenholzner J.O., Vortisch W. & Wieser J., 1998: Pyroclastic Deposits from the Middle Miocene Stallhofen Formation. *Jb. Geol. B-A,* Wien, 140, 4, 425–428.

Lorenc S. & Zimmerle W., 1993: Miocene Kaolin Tonsteines from the Lignite Openpit Mine of Belchatów, SW of Lódz.. *Z. Dtsch. Geol. Gesell.,* Hannover, 144, 187–223 (in German).

Nehyba S., 1997: Miocene volcanics of the Carpathian Foredeep in the Czech Republic. *Věst. CGÚ* , Praha, 72, 4, 311–328.

Nehyba S. & Roetzel R., 1999: Lower Miocene volcaniclastics in South Moravia and Lower Austria. *Jb. Geol. B-A*, Wien (in press).

Nehyba S., in press: The cyclicity of the Lower Miocene deposits of the Carpathian Foredeep as the depositional response to sediment supply and sea-level changes. *Geol. Carpathica.*

Pupin J.P., 1980: Zircon and Granite Petrology. *Contr. Mineral. Petrology*, Berlin, 73, 207–220.

Tkačuk L.G., Kudrin L.N. & Ripun M.B., 1958: Neogene volcanic tuffs of the western part of USSR. *Vopr. Miner. Osad. Obraz.,* 5, Lvov (in Russian).

Unger H.J., Fiest W. & Niemeyer A., 1990: The Bentonites of the Molasse of Eastern Bavaria and their Relationship to the Volcanic Rocks of the Pannonic Basin. *Geol. Jb.,* 96, 67–112, Hannover (in German).

SOURCE ROCKS OF WESTERN CARPATHIAN GRANITOIDS IN THE LIGHT OF Rb/Sr AND Sm/Nd DATA

I. PETRÍK

Geological Institute of the Slovak Academy of Sciences, Dúbravská cesta 9, 84226 Bratislava, Slovak Republic; geolpetr@savba.savba.sk

Key words: granitoids, source rock, Rb/Sr, Sm/Nd, Western Carpathians.

The large number of Rb/Sr data gathered during last fifteen years (see Cambel et al. 1990) was recently completed by the first Sm/Nd determinations (Kohút et al. 1995; Hraško et al. 1998; Kohút et al. 1999). On the basis of the Sm/Nd data the latter authors stated that Western Carpathian granitoids were derived mainly from recycled Proterozoic crustal rocks with significant Paleozoic additions. In present work a systematic reinterpretation of older Rb/Sr data combined with published Sm/Nd data enabled the author to define several protolith end members in terms of their $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios which cover all common granite types in the Tatric and Veporic units.

Rb/Sr system. The discordancy between the Rb/Sr and U/Pb (zircon) datings recognized long ago has been interpreted as a result of initial slope giving too high Rb/Sr "ages" (Kráľ 1994; Petrík et al. 1994). This slope could have originated due to imperfect homogenization of protolith producing a mixing line between two end members with different initial Sr ratios (Kráľ l.c.). In this work, the measured Sr isotopic ratios were reduced according to the known U/Pb ages and the initial slopes of the data from Tribeč, Strážovské vrchy, Nízke Tatry, Slovenské rudohorie Mts. (Sihla type) granitoids were inspected. Surprisingly, no mixing line was identified for any of these granite types, instead a single or couple of outlying samples were found to have increased the pseudoisochron slope. These outliers are (1) leucocratic granites often remote from the main body as veins or sills in the metamorphic complex (Strážovské vrchy), (2) leucocratic veins within granite (Tribeč Mts.), (3) granites with obvious postmagmatic overprint (Strážovské vrchy) or (4) samples belonging to a different granite series (Nízke Tatry Mts.). These granites with high Rb/Sr ratios were originally included by the authors in measured sample sets in order to get better isochron statistics. For example omitting a sample of leucocratic vein (SR6) from the overlying gneiss complex in the Suchý Mts. decreases the age from 393 ± 6 Ma to 350 ± 50 Ma giving a "correct age" but with a much larger error. The reinterpretation of isochrons for the Suchý, Tribeč, Ďumbier, Kralička, Hrončok, and Sihla granite types in this way showed that they are within the error identical with corresponding U/Pb ages with slightly different initial ratios. These new ratios are used in the following model.

The Western Carpathian granitoids show a considerable span in new values of I_{Sr} framed by two extremes: the gabbro from Rochovce (Kohút et al. 1999) with I_{Sr} = 0.70232 ± 20 and the Kralička type granite with 0.7157 ± 6. All Tatric and Veporic massifs with both I and S type characteristics and I_{Sr} = 0.7055–0.7085 lie between them, the A-type Hrončok granite having rather high I_{Sr} = 0.7117 ± 24 (not shown in Fig. 1 because of lacking Nd data).

Sm/Nd system. Sm/Nd ratio changes less than Rb/Sr during melting and therefore the system may keep an older record. Kohút

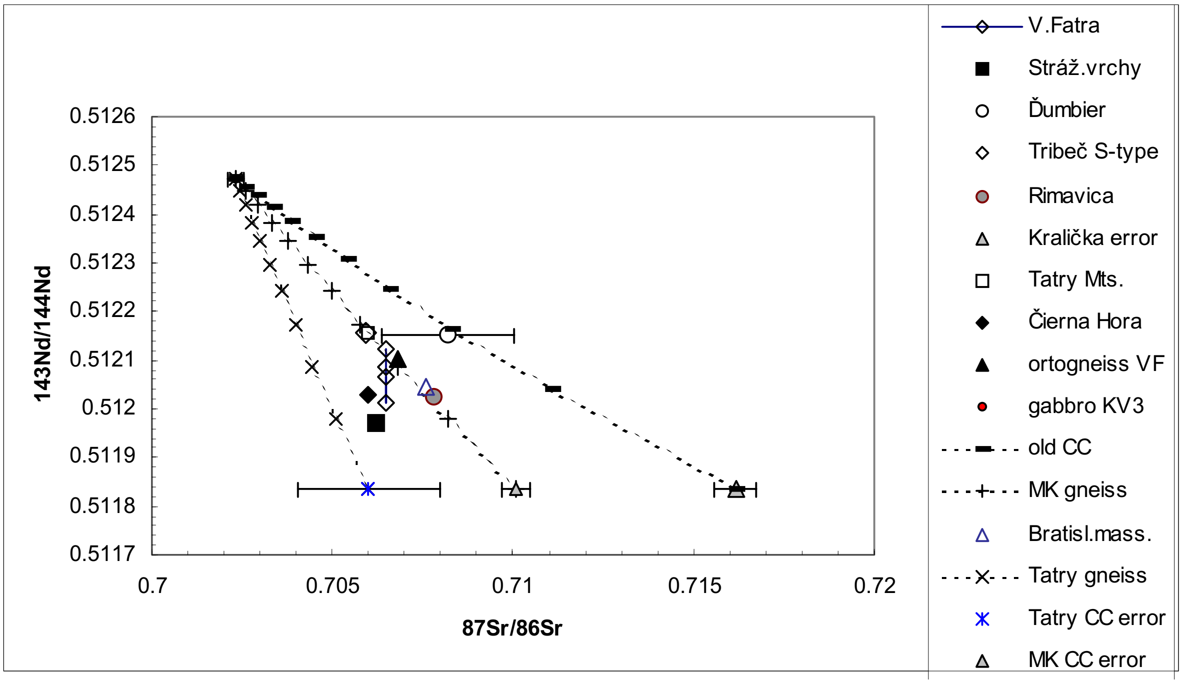

Fig. 1.

et al. l.c. have already stated that, in contrast to the Rb/Sr system, their Sm/Nd data do not show homogenization at the time of formation and suggested three components to explain their data, DM (depleted mantle), EM (enriched lower crust) and CC (continental crust). Based on the correlation of Sr and Nd data it is argued below that three types of CC member along with a DM member are necessary to cover the Sr-Nd data of all the S and I-type granitoids of the Tatric and Veporic units. Another two precursors are necessary for A-type granites and Gemeric specialized granites. The $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios in Fig. 1 were calculated at t = 350 Ma (Sm/Nd data are taken from Kohút et al. 1999).

Single samples from various granite massifs are completed by series of samples from one core — the Veľká Fatra Mts. The example of this comagmatic series shows the contrasting behaviour of the two systems — homogenization of Sr isotopes after mixing of both components while preserving the original Nd isotopic inhomogeneity. A gabbro from Rochovce is taken as the first end member. It comes from the Veporic-Gemeric boundary and has extremely low Sr and high Nd initials (0.7023, 0.51247, respectively). This rock type has numerous analogues in other granite cores (with no Sr-Nd data available) and may represent a juvenile mantle derivative (T_{DM2st}(t) = 621 Ma). The second end member is represented by an orthogneiss complex occurring extensively in the Nízke Tatry Mts. The Kralička type granite originating by partial anatexis of the orthogneiss shows the highest Sr and lowest Nd initial values, 0.71616 and 0.51183, respectively. Kohút et al. showed that this granite type has a very high T_{DM2st} of 1.6 Ga years. Zircons from similar orthogneisses from the Západné Tatry Mts. have upper intercepts around 2 Ga (Poller oral comm.). The protolith of the orthogneisses represents therefore an old, several times recycled, crustal component. Other two end members are represented by metasedimentary rocks from the Tatry and Malé Karpaty gneisses (see Cambel et al. 1990) with I_{Sr} = 0.706 and 0.7101 and ages of metamorphism of 400 and 380 Ma, respectively. This is a ubiquitous rock type forming the wall rock of granitoid intrusions (because their Nd isotopic composition is unknown, the Kralička value was used). Fig. 1 shows that almost all the analysed granitoids follow broadly the mixing line between two members (gabbro and gneiss) requiring approximately 10–30 % of basic end member. These granitoids come from both the Tatric and Veporic units. Only one point falls on the line between the gabbro and orthogneissic end members: the Ďumbier type granodiorite. This rock intrudes the orthogneiss and partial involvement of the old component in its protolith explains the high I_{Sr} (0.7082 ± 18) otherwise hardly understandable. In view of Sm-Nd isotopic characteristics it seems that a high proportion of the crustal components (70–90 wt. %) in the protolith is responsible for the prevailing S-type features of Western Carpathian granitoids. A slightly higher basic member proportion would enhance I-type features. Nd data for the Hrončok A-type granite is lacking, however its high I_{Sr} suggests a different, higher Rb/Sr precursor. This is supported by zircon upper intercept giving ca. 1 Ga (Putiš et al. 1999). It is noted that the extremely high Sr initials of Gemeric granites (ca. 0.750) would require still another source component for these mineralized S-type granites from the Gemeric unit. In summary, (1) common Western Carpathian I- and S-type granitoids require one mantle and a range of gneiss components, (2) the orthogneisses (partly also the Ďumbier granite) require a dominant proportion of an old crustal component, (3) A-type granite (Hrončok) requires a higher Rb/Sr source and (4) Gemeric granites an extremely high Rb/Sr source.

References

Cambel B., Kráľ J. & Burchart J., 1990: Isotope geochronology of the crystalline basement of the Western Carpathians. *Veda, Publ.*, Bratislava, 1990, 183 (in Slovak with English summary).

Hraško Ľ., Kotov A.B., Salnikova E.B. & Kovach V.P., 1998: Enclaves in Rochovce granite intrusion as indicators of the temperature and origin of the magma. *Geol. Carpathica,* 49, 125–138.

Kohút M., Kotov A.B., Salnikova E.B., Kovach V.P. & Michalko J., 1995: Hercynian granitic rocks of the Western Carpathians: products of crustal reactivation. In: Brown M. & Piccoli Ph.M. (Eds.): *The origin of granites and related rocks. USGS Circular,* 1129, 81–82.

Kohút M., Kotov A.B., Salnikova E.B. & Kovach V.P., 1999: Sr and Nd isotope geochemistry of granitic rocks from the Western Carpathians — implications for granite genesis and crustal evolution. *Geol. Carpathica* (in press).

Kráľ J., 1994: Strontium isotopes in granitic rocks of the Western Carpathians. *Min. Österr. Geol. Gesell.,* 86, 75–81.

Kráľ J., Goľcman Y.V. & Petrík I., 1987: Rb-Sr whole rock isochron data of granitic rocks from the Strážovské vrchy Mts.: The preliminary report. *Geol. Zbor. Geol. Carpath.,* 38, 171–179.

Petrík I., Broska I. & Uher P., 1994: Evolution of the Western Carpathian granite magmatism: age, source rock, geotectonic setting and relation to Variscan structure. *Geol. Carpathica,* 45, 283–291.

Putiš M., Kotov A.B., Uher P., Salnikova E.B. & Korikovsky S.P., 1999: Early Triassic age of the Hrončok A -type granite related to continental rifting: a new result of U-Pb zircon dating (the Veporic unit, Central Western Carpathians, Slovakia). *Geol. Carpathica* (in press).

THE GEODYNAMIC EVOLUTION OF THE TATRA MOUNTAINS CONSTRAINED BY NEW U-Pb SINGLE ZIRCON DATA ON ORTHOGNEISSES, MIGMATITES AND GRANITOIDS

U. POLLER[1], W. TODT[1], M. JANÁK[2] and M. KOHÚT[3]

[1]Max-Planck-Institut für Chemie, Abt. Geochemie, Postfach 3060, D-55020 Mainz, Germany; poller@mpch-mainz.mpg.de; todt@mpch-mainz.mpg.de

[2]Geological Institute, Slovak Academy of Science, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic

[3]Geological Survey of Slovakia, Mlynská dolina 1, 817 04 Bratislava, Slovak Republic

Key words: Western Carpathians, Tatra Mountains, U-Pb single zircon ages, Variscan Orogeny, geodynamic.

Introduction

The Western Carpathians belong to the Alpine-Carpathian belt. However, their pre-Mesozoic basement may be the key area for understanding of the eastern continuation of the Variscides in the Central Europe. The aim of this study was to constrain the pre-Alpine evolution of the Western Carpathian´s basement by means of precise single-grain zircon dating.

The investigated area — the Tatra Mountains, are representative of to the so-called "core mountains" within the Tatric unit of the Western Carpathians. In general, the crystalline basement of the Tatra Mountains is composed of pre-Mesozoic metamorphic rocks and granitoids, overlain by Mesozoic and Cenozoic sediments. The metamorphic rocks are most abundant in the western part (Western Tatra Mts.), whereas in the eastern part (High Tatra Mts.) the granitoids are more common. The basement belongs to two tectonic units, differing in metamorphic grade and lithology that are separated by major Variscan thrust fault (Janák 1994). The lower unit is exposed only in the Western Tatra Mts. It is composed of micaschists, exhibiting a medium grade metamorphism. Two metamorphic zones, the kyanite-staurolite zone and the kyanite-sillimanite zone have been distinguished (Janák 1994). The upper unit shows high-grade metamorphism and migmatization. The base of the unit is formed by orthogneisses, kyanite bearing paragneisses and banded amphibolites with garnet and clinopyroxene-bearing eclogitic relicts (Janák et al. 1996). They belong to the kyanite zone of high-pressure and high-temperature metamorphism. Higher levels are occupied by sillimanite, K-feldspar and cordierite bearing migmatites. They belong to the sillimanite zone, characterised by medium to low pressure and high temperature metamorphism and anatexis (Janák et al. 1999). The granitoids form a sheet like pluton, ranging from leucogranites to biotite tonalites and amphibole diorites (Kohút & Janák 1994).

Such inverted metamorphic sequence is a consequence of south-vergent thrusting during Variscan continental collision (Fritz et al. 1992; Janák 1994).

Methods

Isotopic analyses were carried out at the Max-Planck-Institute of Chemistry, Dept. of Geochemistry in Mainz. All U-Pb measurements were done on single zircon grains using either the vapour digestion method (Wendt & Todt 1991) or the CLC-method (Poller et al. 1997). For all U-Pb analyses a ^{205}Pb-^{233}U mix-spike was used. The analyses were performed in ion counting mode by peak hoping on a Finnigan MAT 261 mass spectrometer. Several standards (NBS 981, NBS 982, U nat) were measured and all samples were corrected for blank, common Pb (obtained by Pb-Pb analyses of cogenetic galena) and fractionation.

Results

The oldest age obtained from zircons of the orthogneisses is ca. 406 Ma. This age is interpreted as the crystallization of the orthogneiss precursor, most probably a peraluminous hybridic granite. The orthogneisses are spatially associated with high pressure metamorphic relics (Janák et al. 1996). Therefore, they may indicate a subduction-related melting event at an active continental margin during Upper Silurian/Lower Devonian time.

Recrystallization (gneissification) of the orthogneiss precursores in the Western Tatra Mountains, took place in Middle to Upper Devonian time (360 ± 10 Ma). This high-temperature metamorphism overlapped with intrusion of the Western Tatra granitoids in the Roháč and Bystrá area during Upper Devonian and Lower Carboniferous (360–350 Ma). Last magmatic event in the Western Tatra Mountains was detected by granite from Baranec, at ca. 340 Ma (Todt et al. 1998; Poller et al. 1998), during Early Carboniferous. Granitoid magmatism were most probably related to continental collision processes. Partial melting might be facilititated by heating from the upwelling mantle after the detachment of the lithospheric slab (Blankenburg & Davies 1995).

Concordant ages of at 355 ± 4 Ma detected in the migmatites near Končistá record the earliest anatectical event or migmatization in the High Tatra area (Janák et al. 1999), which seems to be coeval with similar event in the Western Tatra. Two events in late Visean time, at around 335 Ma are indicated by concordant zircon ages is are: a) the intrusion of the most mafic granitoid rocks in the High Tatra (diorites in Velická valley), and b) the recrystallization of zircons in the High Tatra migmatites that may record the second, low-pressure migmatization stage (Janák et al. 1999; Ludhová & Janák 1999).

The final magmatic stage in the High Tatra Mountains was the intrusion of two mica granodiorites to granites at around 315 Ma.

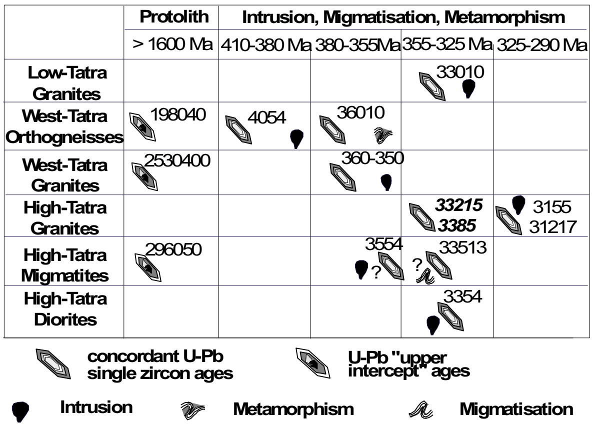

Fig. 1. Compilation of all U-Pb zircon ages of several rocks of the Tatra Mts.

Some granitoid samples from the contact with diorites and migmatites contain also 332–338 Ma old concordant zircons.

Discussion

Presented age data (Fig. 1) allow the comparison of the Tatra Mountains with other Variscan regions and a preliminary reconstruction of the geodynamic evolution of the investigated area during Late Silurian/Early Devonian and Carboniferous time.

Whereas in other parts of the Variscides like the Mid German Crystalline Rise (MGCR; Reischmann & Anthes 1996) or the Schwarzwald (Lippolt et al. 1986; Todt & Büsch 1981) Cambrian ages were detected for granitoid rocks, this age is almost lacking in the Tatra Mountains. Only some U-Pb ages of zircons from the micaschists of the "lower unit" define a minimum sedimentation age at around 550 Ma (Gurk & Poller 1999).

Recrystallization (gneissification) of the orthogneisses due to continental collision metamorphism in Devonian time has been reported also from e.g. the northern part of the Moldanubian unit (Teufel et al. 1986; Teufel 1987), the Massif Central and Brittany (Matte 1986). Granitoid magmatism in the Western Tatra at around 350 Ma corresponds with that in other segments of the Western Carpathians (Kohút et al. 1997; Petrík & Kohút 1997), or the Bohemian Massif (Klötzli & Parrish 1994). Younger ages of ca. 335 Ma, as recorded in the High Tatra granitoids have been also found in zircons from the MGCR by Reischmann & Anthes (1996). Finally, the last magmatic event at around 315 Ma, revealed by granitoids in the High Tatra Mountains, corresponds to collision-related magmatism in the more western parts of the Variscides. It is inferred that metamorphism and granitoid magmatism in the Tatra Mountains might have been related to rather long-lasting (Late-Devonian to Early Carboniferous) amalgamation and assembly of the Gondwana-derived microplates, preceding the final collision between the Gondwana and Laurentia/Baltica in the Late Carboniferous time.

Refering to the presented data the following scenario is proposed:

1. In Upper Silurian/Lower Devonian time, the granite precursor of the orthogneisses was generated due to subduction at an active continental margin or volcanic arc. During this time the Tatra Mountains were probably situated at the northern part of Gondwana, south-east to the Saxothuringian and Moldanubian units.

2. During subsequent compression triggered by continental collision in Late Devonian time, sedimentary flyschoid sequences were accreated and a granitic orthogneiss precursor underwent metamorphic recrystallization.

3. In Carboniferous time, the granitoids of the Western Tatra intruded. Coeval anatexis and migmatisation is recorded in the High Tatra.

4. After the main collisional stage, thermally weakened crust most probably collapsed and more mafic (dioritic) magma intruded in the High Tatra Mountains 335 Ma ago. This event may have triggered a HT/LP recrystallization in the migmatites.

5. Final magmatic event at ca. 315 Ma is documented by two mica granodiorite intrusion in the High Tatra Mountains.

References

Blanckenburg F. & Davies J.H., 1995: Slab breakoff: A model for syncollisional magmatism and tectonics in the Alps. *Tectonics,* 14, 120–131.

Fritz H., Neubauer F., Janák M. & Putiš M., 1992: Variscan mid-crustal thrusting in the Carpathians II: Kinematics and fabric evolution of the Western Tatra basement, *Terra Abstr., Suppl. 2,* 4, 24.

Gurk C. & Poller U., 1999: Petrography, geochemistry and geochronology of the Lower Unit, Western Tatra Mts., Slovakia. *J. Conf. Abstr*., 4, 809.

Janák M., 1994: Variscan uplift of the crystalline basement, Tatra Mts., Central Western Carpathians: evidence from $^{40}Ar/^{39}Ar$ laser probe dating of biotite and P-T-t paths. *Geol. Carpathica*, 45, 5, 293–300.

Janák M., O'Brien P.J., Hurai V. & Reutel Ch., 1996: Metamorphic evolution and fluid composition of garnet-clinopyroxene amphibolites from the Tatra Mts., Western Carpathians. *Lithos,* 39, 57–79.

Janák M., Hurai V., Ludhová L., O´Brien P.J. & Horn E.E., 1999: Dehydration melting and devolatilization of high-grade metapelites: the Tatra Mountains, Western Carpathians. *J. Metamorphic Geol.* (in print).

Klötzli U. & Parrish R.R., 1994: Zircon Pb-Pb and U-Pb geochronology of the Rastenberg granodiorite (Lower Austria): evidence for the incorporation of Cadomian and possibly Archean crust into Variscan granitoids of the South Bohemian Pluton. *Mitt. Österr. Miner. Gesell*., 139, 68–70.

Kohút M. & Janák M., 1994: Granitoids of the Tatra Mts., Western Carpathians: Field Relations and Petrogenetic Implications. *Geol. Carpathica*, 45, 5, 301–311.

Kohút M., Todt W., Janák M. & Poller U., 1997: Thermochronometry of the Variscan basement exhumation in the Velká Fatra Mts. (Western Carpathians, Slovakia). *Terra Abstr.*, 9, 494.

Lippolt H.J., Kober B., Hradetzky H. & Mertz D.F., 1986: Schwarzwald: Geochronologische Studien an Metamorphiten und Mineralisation (abs.); *KTB Kolloquium Seeheim, Poster Programm*, 41.

Matte P., 1986: Tectonics and Plate Tectonics Model for the Variscan Belt of Europe. *Tectonophysics,* 126, 329–374.

Ludhová L. & Janák M., 1999: Phase relations and P-T path of cordierite-bearing migmatites, Western Tatra Mountains, Western Carpathians. *Geol. Carpathica* (in print).

Petrík I. & Kohút M., 1997: The evolution of granitoid magmatism during the Hercynian orogen in the Western Carpathians. In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological Evolution of the Western Carpathians. *Miner. slovaca-Monogr.*, 235–252.

Poller U., Liebetrau V. & Todt W., 1997: U-Pb single-zircon dating under cathodo-luminescence control (CLC-method): application to polymetamorphic orthogneisses. *Chem. Geol.,* 139, 287–297.

Poller U., Janák M., Kohút M. & Todt W., 1998: Pre-Variscan evolution of the Western Tatra Mts (Slovakia): A Sm-Nd, Rb-Sr and Pb-Pb isotope study. *Chin. Sci. Bull.,* 43, 103.

Poller U., Todt W., Janák M. & Kohút M., 1999: U-Pb single zircon TIMS data in combination with CL and EMP Studies on granitoids of the High Tatra Mountains (Slovakia). *J. Conf. Abstr.,* 4, 669.

Reischmann T. & Anthes G., 1996: Geochronology of the mid-German crystalline rise west of the River Rhine. *Geol. rdsch.,* 85, 761–774.

Teufel S., 1987: Vergleichende U-Pb und Rb-Sr-Altersbestimmungen an Gesteinen des Übergangsbereiches Saxothuringikum/Moldanubikum, NE Bayern. *PhD thesis*, *Univ. Göttingen,* 110.

Teufel S., Ahrendt H. & Hansen B.T., 1986: Die Beziehung der Zone von Erbendorf-Vohenstrauß zum Moldanubikum aufgrund radiometrischer Altersbestimmungen (Abstr.) *76. Jahrestagung der Geologischen vereinigung*, Giessen, 75.

Todt W. & Büsch W., 1981: U-Pb investigations on zircons from pre-Variscan gneisses; A study from the Schwarzwald, West Germany. *Geochim. Cosmochim. Acta*, 45, 1798–1801.

Todt W., Poller U., Kohút M. & Janák M., 1998: U-Pb zircon dating of metamorphic rocks of the Western Tatra using single zircon under cathodoluminescence control. *Chin. Sci. Bull*., 43, 130.

Wendt J.I. & Todt W., 1991: A vapour digestion method for dating single zircons by direct measurements of U and Pb without chemical separation. *Terra Abstr*., 3, 507–508.

THE RELATIONSHIPS BETWEEN THE VARISCIDES AND THE WESTERN CARPATHIAN BASEMENT: NEW Sr, Nd AND Pb-Pb ISOTOPE DATA FROM THE TATRA MOUNTAINS

U. POLLER[1], W. TODT[1], M. JANÁK[2] and M. KOHÚT[3]

[1]Max-Planck-Institut für Chemie, Abt. Geochemie, Postfach 3060, D-55020 Mainz, Germany; poller@mpch-mainz.mpg.de; todt@mpch-mainz.mpg.de

[2]Geological Institute, Slovak Academy of Science, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic

[3]Geological Survey of Slovakia, Mlynská dolina 1, 817 04 Bratislava, Slovak Republic

Key words: Western Carpathians, Tatra Mountains, Sr-Nd-Pb isotopes, Variscan orogeny.

Introduction

The Western Carpathians belong to the Alpine-Carpathian orogenic belt, which evolved as a classical area of the Alpine orogeny during Mesozoic-Cenozoic times (Plašienka et al. 1997). Their pre-Mesozoic rock complexes, however, belong to the Variscan basement within the Alpine-Carpathian orogenic belt (Neubauer & von Raumer 1993). Consequently, reliable isotopic and geochronological data may significantly contribute to the reconstruction of eastern and south-eastern continuation of the Variscan belt in Central Europe, at the boundary between the Bohemian Massif and the Alpine-Carpathian orogen. This study deals with the Variscan basement exposed in the Tatra Mountains. The Tatra Mountains are a typical "core mountains", belonging to the Tatric unit, a major tectonic unit in the Central Western Carpathians.

The crystalline basement of the Tatra Mountains is composed of pre-Mesozoic metamorphic rocks and granitoids, overlain by Mesozoic and Cenozoic sediments. The metamorphic rocks are most abundant in the western part (Western Tatra Mts.), whereas in the eastern part (High Tatra Mts.) the granitoids are more common. The basement belongs to two tectonic units, differing in metamorphic grade and lithology that are separated by major Variscan thrust fault (Janák 1994). The lower unit is composed of micaschists, exhibiting a medium grade metamorphism. Two metamorphic zones, the kyanite-staurolite zone and the kyanite-sillimanite zone have been distinguished (Janák 1994). The upper unit shows high-grade metamorphism and migmatization. The base of the unit is formed by orthogneisses, kyanite bearing paragneisses and banded amphibolites with eclogitic relics of high pressure metamorphism (Janák et al. 1996). Higher levels are occupied by sillimanite and cordierite bearing migmatites exhibiting medium to low pressure and high temperature metamorphism and anatexis (Janák et al. 1999). The granitoids form a sheet like pluton, ranging from leucogranites to biotite tonalites and amphibole diorites (Kohút & Janák 1994).

The main topic of this study is to document close relationships between the Western Carpathian´s basement and the Variscides based on isotopical data from the Tatra Mountains. These first Nd, Sr and Pb-Pb whole rock isotopic results have been compared with the more western Variscan regions. They suggest that the Tatra Mountains basement rocks can be well correlated with those in the Bohemian Massif (Janoušek et al. 1995), Schwarzwald, Spessart, Odenwald (Liew & Hofmann 1988; Reischmann & Anthes 1996) and Massif Central (Pin & Duthou 1990; Turpin et al. 1990).

Methods

Several orthogneisses, micaschists, amphibolites and granitoids from the Western Tatra as well as granitoids and migmatites from the High Tatra Mountains were sampled and analysed using different methods. All isotopic measurements were performed done at the Max-Planck-Institute of Chemistry, Dept. of Geochemistry in Mainz. The Sr and Nd analyses were carried out on whole rock powders, whereas the Pb-Pb isotopes were measured on whole rock splits. After standard chemistry separation the isotopes were measured on a Finnigan MAT 261 mass spectrometer in multicollector mode. Detailed information about the chemistry procedure is given in Patchett & Bridgwater (1984) and in Liew & Hofmann (1988) for Sr and Nd, and in Arndt & Todt (1994) for Pb-Pb. The used spikes were enriched in ^{85}Rb and ^{84}Sr, respectively in ^{150}Nd and ^{149}Sm. For all Rb-Sr and Sm-Nd measurements an IC and an ID run was executed. The common Pb was only measured as IC. All analyses were controlled by standard measurements (NBS 987, La Jolla, NBS 981, NBS 982, La Jolla) and were corrected for fractionation and blank.

Results

The investigated orthogneisses and most of the granitoids are peralumineous with granodioritic to syenogranitic chemistry. They show the S- or H-type granitoids characteristics according to Castro et al. (1991). Only the diorites from the Velicka valley in the High Tatra are more mafic and they can be classified as I-type granitoids. All analysed granitoid rocks show an affinity towards the volcanic arc and collisional granite fields after Pearce et al. (1984). The whole rock Pb-Pb analyses (Fig. 1) of the granitoids and orthogneisses show only small spread in ^{206}Pb/^{204}Pb, from 18.04 to 18.85 and ^{207}Pb/^{204}Pb, from 15.53 to 15.66. Only the amphibolites reach the values up to 20.63 for the ^{206}Pb/^{204}Pb, and 15.87 for the ^{207}Pb/^{204}Pb ratios. The granitoids plot close to the Stacey & Kramers (1975) evolution line, whereas the porphyric orthogneisses plot more towards the upper crust line (evolution lines after Doe & Zartman 1979), indicating a small amount of mantle influence for the granitoids. Figure 1 shows a clear overlap of the Tatra field (black) with the upper crust field (fields for crust and MORB after Zindler & Hart 1986). MORB influence can be excluded by the presented Pb-Pb isotopes for the acidic granitoid rocks as well as the amphibolites with even higher values. The lower crust (represented by the Eifel xenoliths with data from Rudnick & Goldstein 1990, Fig. 1), had only small influence on the Tatra Mts. samples. The crustal character is well constrained by the common Pb system. Nevertheless, the Pb isotopes of enriched mantle II (equivalent to recycled oceanic crust) occupy an intermediate position between the upper and lower crust. Therefore a slight influence from such a recycled source can not be excluded. Sm-Nd whole rock analyses resulted in εNd(0) values from −4.8 to −9.3 for most granites and orthogneisses. The ^{87}Sr/^{86}Sr values ranging from 0.7076 to 0.7150 suggest that most granitoids are hybridic in composition. The Nd model ages for the granitoids are within the range of 1300–1500 Ma, the same as for the porphyric orthogneisses.

Discussion

In addition to the characterisation of the investigated rocks, the isotopic results were compared with Sr, Nd and Pb-Pb data of similar rock series from several Variscan regions. As first, the Sr and Nd isotopic compositions were recalculated for the age of 330 Ma (presumed model age of the Variscan granitoid magmatism). Figure 2 shows the εNd(330 Ma) versus ^{87}Sr/^{86}Sr (330 Ma) plot with different Tatra Mts. rocks (black fields) in comparison with

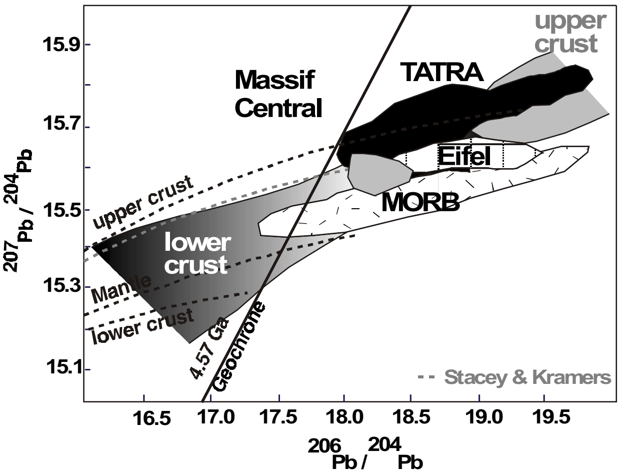

Fig. 1. ^{207}Pb/^{204}Pb vs. ^{206}Pb/^{204}Pb plot with analyses of the Tatra in comparison with other regions and Pb-reservoirs. Details see text.

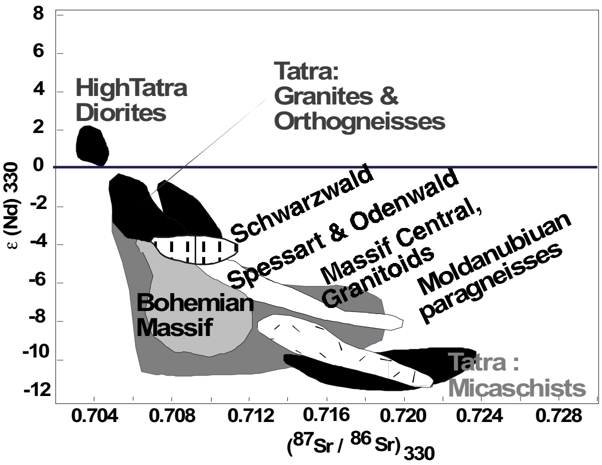

Fig. 2. εNd(330 Ma) vs. ^{87}Sr/^{86}Sr (330 Ma) plot of the Tatra rocks in comparison with other Variscan regions. Details see text.

several other Variscan regions. The Tatra Mts. micaschists of the lower unit correspond to paragneisses of the Moldanubian unit (data from Janoušek et al. 1995). The granites and orthogneisses of the Tatra show higher εNd(330 Ma) values, nevertheless they overlap two mica granites from Bohemian Massif (Janoušek et al. 1995). Granitoids from the Schwarzwald, Spessart and Odenwald (Liew & Hofmann 1988) as well as granitoids and leucogranites from the Massif Central (Pin & Duthou 1990; Turpin et al. 1990) show good correspondence with the Tatra granitoids and orthogneisses. Only the High Tatra diorites with positive εNd(330 Ma) values and rather low Sr isotope ratios seem to be not equivalent to other Variscan regions.

The similarities in Sr, Nd and Pb-Pb isotopic characteristics of the investigated samples with those from the several Variscan regions, point to the linkage between the Western Carpathian´s basement and the western and central European Variscides before the break-up of Pangea.

References

Arndt N. & Todt W., 1994: Formation of 1.9 Ga old Trans-Hudson continental crust: Pb isotopic data. *Chem. Geol.*, 118, 9–26.

Castro A., Moreno-Ventas I. & de la Rosa J.D., 1991: H-type (hybrid) granitoids: a proposed revision of the granite-type classification and nomenclature. *Earth Sci. Rev.*, 31, 237–253.

Doe B.R. & Zartman R.E., 1979: Plumbotectonics, The Phanerozoic. In: Barnes H.L. (Ed.): Geochemistry of hydrothermal ore deposits, *Wiley Intersci. Publ*., New York, Chichester, Brisbane, Toronto, 22–70.
Janák M., 1994: Variscan uplift of the crystalline basent, Tatra Mts., Central Western Carpathians: evidence from 40Ar/39Ar laser probe dating of biotite and P-T-t paths. *Geol. Carpathica,* 45, 293–300.
Janák M., O´Brien P.J., Hurai V. & Reutel C., 1996: Metamorphic evolution and fluid composition of garnet-clinopyroxene amphibolites from the Tatra Mountains, Western Carpathians. *Lithos,* 39, 57–79.
Janák M., Hurai V. Ludhová L., O´Brien P.J. & Horn E.E., 1999: Dehydration melting and devolatilization of high-grade metapelites: the Tatra Mountains,Western Carpathians. *J. Metamorphic Geol.* (in print).
Janoušek V., Rogers G. & Bowes D.R., 1995: Sr-Nd isotopic constraints on the petrogenesis of the Central Bohemian Pluton, Czech Republic. *Geol. Rdsch.,* 84, 520–534.
Kohút M. & Janák M., 1994: Granitoids of the Tatra Mts., Western Carpathians: field relationships and petrogenetic implications. *Geol. Carpathica,* 45, 301–311.
Liew T.C. & Hofmann A.W., 1988: Precambrian crustal components, plutonic associations, plate environment of the Hercynian Fold Belt of Central Europe: Indications from a Nd and Sr isotopic study. *Contr. Mineral. Petrology,* 98, 129–138.
Neubauer F. & von Raumer J.F., 1993: The alpine basement-linkage between variscides and East-Mediterranean mountain belts. In: von Raumer J.F. & Neubauer F. (Eds.): *Pre-Mesozoic geology of the Alps,* Springer Berlin - Heidelberg - New York, 641–663.
Patchett P.J. & Bridgewater D., 1984: Origin of continental crust of 1.9-1.7 Ga age defined by Nd isotopes in the Kedilidian terrain of south Greenland. *Contr. Mineral. Petrology,* 87, 311–318.
Pearce J.A., Harris N.B.W. & Tindle A.G., 1984: Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. *J. Petrology*, 25, 956–983.
Pin C. & Duthou J.L., 1990: Sources of Hercynian granitoids from the French Massif Central: inferences from Nd isotopes and consequences for crustal evolution. *Chem. Geol*., 83, 281–296.
Plašienka D., Grecula P., Putiš M., Hovorka D. & Kováč M., 1997: Evolution and structure of the Western Carpathians: an overview. In: Grecula P., Hovorka D. & Putiš M. (Eds.): Geological Evolution of the Western Carpathians. *Miner. slovaca-Monogr.,* 1–24.
Reischmann T. & Anthes G., 1996: Geochronology of the mid-German crystalline rise west of the River Rhine. *Geol. Rdsch.,* 85, 761–774.
Rudnick L.R. & Goldstein S.L., 1990: The Pb isotopic compositions of lower crustal xenoliths and the evolution of lower crustal Pb. *Earth Planet Sci. Lett.,* 98, 192–207.
Stacey J.S. & Kramers J.D., 1975: Approximation of terrestrial lead isotope evolution by two-stage model. *Earth Planet Sci. Lett.,* 26, 207–221.
Turpin L., Cuney M., Friedrich M., Bouchez J.L. & Aubertin M., 1990: Meta-igneous origin of Hercynian peralumineous granites in N.W. French Massif Central: implications for crustal history reconstruction. *Contr. Mineral. Petrology*, 104, 163–172.
Zindler A. & Hart S., 1986: Chemical Geodynamics. *Ann. Rev. Earth Planet Sci.,* 14, 493–571.

RELATIONSHIPS BETWEEN LITHOLOGY AND CHEMICAL COMPOSITION OF GROUNDWATER FROM MOLDAVIAN PLATEAU (ROMANIA)

GH. POPA

"Al.I.Cuza" University, Faculty of Geography and Geology,
Department of Mineralogy-Geochemistry, B-dul Copou 20A, Iassy,
Romania; gpopa@uaic.ro

Key words: Moldavian Plateau, lithological structures, terrace deposits, kaolinit.

Introduction

This work proposes to point out the possible correlations between the chemical composition of the groundwaters and the specific rocks alteration reactions. The hydrogeochemical approach of this kind of problems implies the application of the paternal test of ions present in the water. The Moldavian Plateau was a lacustrine and coastal plain, hydrogeologically important only the Cretaceous-Eocene sedimentation cycles. The vegetation is well developed in this area and it is known that in such cases only 15 % of K, resulting from chemical weathering of rocks, reaches the groundwaters, the rest being included in the vegetation cycle.

Methods

The research methodology consisted of three stages: a) The field work stage during which the geological and lithological elements were identified, and samples of waters, snow and rocks were taken, too. In this way, through a punctiform verification, proofs both from the recent rocks as well as from the chemically weathered ones were drawn in order to express the chemical weathering processes. The water samples were taken from springs, open pits and drillings in accordance with the Romanian standards regarding the preservation, gathering and analysis of waters. The atmospheric input was determined by chemical analyses of snow. The snow samples were taken by the punctiform method. b) The laboratory stage in which the mineralogical compostion of rocks and the chemical composition of waters and snow (main constituents) were determined. The samples of waters were analysed at the Environment Protection Agency — Iassy, and the samples of rocks and snow in the Mineralogy Laboratory of the University — Iassy. c) The processing of the chemical analyses results (XCEL-97). (Dragomir 1998).

Results

About 6000 chemical determinations were performed (5750 for groundwaters and 190 for snow). The average chemical composition of water of hydrostructures categories is illustrated in the Table 1 and the correlation coefficients in Table 2.

Discussion

The following hydrogeological characteristics are specific of this area: a) Critical hydrological balance having a high index of aridity, with rivers frequently drying up. b) The lack of the postvolcanic eruption and of the posttectonic depression. c) The lack of the metamorphic and eruptive deposits; d) the law value of the relief intensity. e) The presence of the inner lowlands and of the erosion depression. f) The alternative stratification at the level of the Sarmatian cover through impermeable and permeable deposits, with an important role of collecting rocks and accumulations for the groundwaters. g) The unloading of the water strata through contact springs at the level of the Quaternary deposits and through springs slope, being consequent on the valley slopes whose strata incline in conformity with the slope. h) The terraces are well developed and present hydrogeological importance. i) The carst does not show a hydrogeological importance. j) The feeding of rivers is by rains and snows. These aspects correlated with the geological features (especially tectonic) led to the following types of hydrostructures: 1) unloaded hydrostructures (cracks in the water multistratum and in the interstream area); 2) alluviation hydrostructures (of dejection cones and terraces); 3) hydrostructures of depth (Dragomir 1998; Ionesi 1994; Pascu 1983).

Table 1: The average chemical composition of waters samples (meq/l).

IONS	Waters from interfluve deposits	Waters from terrace deposits	Waters from flood-plain deposits	Waters from calcareous deposits
Na^+	7.60	3.70	5.62	2.50
K^+	0.24	0.14	0.26	0.17
Ca^{2+}	7.52	4.92	4.91	5.59
Mg^{2+}	5.14	5.44	4.29	2.57
Cl^-	2.04	1.94	1.90	0.84
SO_4^{2-}	7.40	2.28	3.29	2.39
HCO_3^-	10.81	9.84	9.88	7.43
NO_3^-	0.12	0.15	0.16	0.06

Table 2: Correlation coefficients: waters from terrace deposits.

	Ca^{2+}	Mg^{2+}	Na^+	K^+	Cl^-	NO_3^-	SO_4^{2-}	HCO_3^-
Ca^{2+}		0.46	0.48	0.58	0.92	-0.35	0.17	0.50
Mg^{2+}			0.63	0.10	0.66	0.02	0.01	0.86
Na^+				0.36	0.57	0.04	0.28	0.78
K^+					0.34	-0.20	0.86	-0.01
Cl^-						-0.30	-0.05	0.67
NO_3^-							0.02	-0.05
SO_4^{2-}								-0.19

The saline efflorescences which appear in the Bahlui Basin were interpreted as proceeding from the remanent salts which were kept by the deposits formed (by evaporation and colmatage) during the retirement of the Sarmatian Sea. The salts deposited as follows: carbonates, sulphates (gypsum, mirabilite, glauberite), rock salts. The salty character of the Moldavian Plateau waters is due to the exchange of ions. It is supposed that the former clays present in this area retained Na^+ (becoming sodium clays) and Cl^- combined with Ca^{2+} and Mg^{2+} forming chlorides which were leached by the infiltration waters. At present the clays fix Ca^{2+} and isolate Na^+ (Pricajan 1972).

For the hydrogeochemical balance the Garrels and Mackenzie's procedure (Drever 1988, pg. 144–156) was used starting from two elements: a) the geological and hydrogeological characteristics of this area; b) the chemical composition of the analysed waters. In order to draw the balance the sources of waters were selected located in the areas with a minimum pollution or even absent and with an invariable chemical composition for long periods of time (5 years). The series of mobility Ca-Na-K-Mg was accepted. The cations and anions in snow were subtracted from groundwater (the atmosphere input) and after that Cl^- with Na^+ and SO_4^{2-} with Ca^{2+} were compensated. This procedure is suggested because as a rule Na^+ and Cl^- appear as a result of rendering soluble NaCl and SO_4^{2-} and Ca^{2+} as a result of rendering soluble the gypsum. Then HCO_3^- was corrected in order to equilibrate Na, K, Ca, Mg, to improve in a way the biotic factor (CO_2 resulted from the vegetation activity) (Popa 1997).

It is supposed that the following chemical reactions should be accepted in order to explain the hydrogeochemical balance of the terrace cover waters (in all cases the final results are clay minerals):

1) The change of kaolinite back into plagioclase:

$$2.26Al_2Si_2O_5(OH)_4 + 0.48Na^+ + 2.00Ca^{2+} + 4.52HCO_3^- + 0.95SiO_2 = 2.50Na_{0,19}Ca_{0,81}Al_{1,81}Si_{2,19}O_8 + 4.52CO_2 + 0.38H_2O$$

2) The change of kaolinite back into biotite:

$$0.05Al_2Si_2O_5(OH)_4 + 0.10K^+ + 0.75HCO_3^- + 0.30Mg^{2+} + 0.20SiO_2 = 0.10KMg_3AlSi_3O_{10}(OH)_2 + 0.38H_2O + 0.75CO_2$$

3) The precipitation:

$$1.36Mg^{2+} + 2.72HCO_3^- = 1.36MgCO_3 + 1.36CO_2 + 1.36H_2O$$

Such balances seem logical but they are far from being perfect. At least two parameters are ignored: a) The biotic factor (which is an important source for CO_2 leading to the formation of H_2CO_3 and consequently the aggressiveness of water towards the minerals). b) The exchange of ions (some research workers assume the presence of Na in the groundwaters in the Moldavian Plateau especially due to the exchange of ions the Sarmatian clays made). It is difficult to settle the values of this exchange of ions however, at any rate, it should not be ignored.

References

Dragomir S. 1998: Geochimia apelor subterane de pe teritoriul judeţului Iaşi. Thesis. Universitatea “Al.I.Cuza”, Iaşi.

Drever J.I. 1988: The geochemistry of natural water. Prentice Hall, New Jersey.

Ionesi L. 1994: Geologia unităţilor de platformă şi a orogenului nord Dobrogean. Ed. Teh. Bucureşti.

Pascu M. 1983: Apele subterane din Romania. Ed. Tehnica, Bucuresti.

Pricajan A. 1972: Apele minerale si termale din Romania. Ed. Tehnica, Bucuresti.

Popa Gh. 1997: Some hydrogeochemical considerations on the Sarul Dornei region (Romania). Romanian Journal of Mineralogy, 78, 77-78.

MESOZOIC PICRITES FROM THE WESTERN CARPATHIANS

J. SPIŠIAK

Geological Institute of the Slovak Academy of Sciences, Severná 5, 974 01 Banská Bystrica, Slovak Republic; spisiak@gu.bb.sanet.sk

Key words: picrites, mineralogy, geochemistry, Cretaceous, subtratric nappe, Western Carpathians.

Geology

In the Mesozoic of the Western Carpathians there occur several typical volcanic formations: the teschenite-picrite formation in the Cretaceous flysch, the formation of alkaline basalts/basanites in the Cretaceous of cover and Subtatric units, and the thoeliite (MORB) formation in the Triassic of the Meliaticum ophiolite complex. The first two formations are very close to each other as for age and geotectonics and they both contain picrite rocks. In the Outer Carpathians volcanites vary quite a lot (from picrites through syenites), whereas in the Central Western Carpathian there are mostly basanite, seldom picrite, rocks and/or their volcanoclasts.

The region with the most abundant occurrences of hypabyssal as well as superficial bodies is the Silesian unit of the Western Carpathians flysch belt, where picrites are part of the already classic teschenite-picrite formation (Pacák 1926; Smulikowski 1929 in Hovorka & Spišiak 1988). The rocks belong to the Teschen-Hradište formation and the volcanic activity is Upper-Valanginian through Lower-Aptian. The bodies occur in various forms. Picrites usually form intrusive bodies that are thick tens of metres.

In the Mesozoic of the Central Western Carpathians, picrites have been reported from broader area of Banská Bystrica (Predný diel and Poníky, Slavkay 1979; Hovorka & Slavkay 1966; Hovorka & Spišiak 1988). Picrites have only been detected in boreholes. The bodies are several metres thick. The host rocks are represented by Middle-Triassic limestones of the basalt part of the Krížna nappe (tegument) (l.c.), and/or by breccia dolomites. The picrites from around Banská Bystrica are strongly deformed, cataclysed and intensively altered. Due to intensive cataclasis the contact zones with the surrounding rocks were not found and, consequently, the mode of deposition of picrite bodies and the relationship of picrites to the surrounding carbonates are not unambiguous.

On the whole, it is possible to say that the mineralogical composition and petrological characteristics are similar to those of the Outer Western Carpathians.

Mineralogy and petrology

At sight, there are observable light-green porphyric olivines of lenticular cross-section, with sizes between 2–6 mm, short dark Cpx columns (2–5 mm), dark mica shales (2–3 mm) and products of younger hydrothermal (pyrite, carbonate), or tectonic processes. The matrix of volcanic rocks is of deep-green colour and macroscopically aphanitic. Here and there it is possible to see signs of mutual poikilic overgrowing of pyroxenes, amphiboles and dark micas of the second generation. The basic structure here is porphyric with olivine and clinopyroxene phenocrysts. In general, the olivine volumes in the rock is higher than that of Cpx.

Olivines (20–50 vol. %) are present only as porphyric phenocrysts in this rock. Mostly, they are intensively serpentinized creating typical loop textures. The cores of individual loops are made of serpentinized olivine, but also talc, calcite and low-birefringent through isotropic serpentine of chrysotile-lizardite group. Olivines are strongly magnesian with prevailing forsterite component (Fo 86, 4–87, Kuděláskova 1987). The olivines in the picrites from around Banská Bystrica are fully altered.

Clinopyroxenes occur in columnar shapes of three generations, or types:

— *Clinopyroxenes I* (xenocrysts) make columns of sizes between 1–3 mm. They occur in close association with olivines, and/or Cr-spinels. They display no zonality.

— *Clinopyroxenes II* — porphyric phenocrysts of crystals, big up to 10 mm. They have idiomorphic shape and pink-lilac through brown-lilac colour. They show distinctive pleochroism. Part of the porphyric Cpx phenocrysts display zonal structure which can be seen from gradual ("undulose") extinguishing of crystals in cross-sections perpendicular to the c axis. Cpx are typical by oscillatory and sector zoning, the latter being more rare.

— *Clinopyroxenes III* — matrix clinopyroxenes. They form small columns of 0.1–0.4 mm, or clusters of irregularly bordered grains. They show similar optical and chemical features as the rims of porphyric phenocrysts and are likely to have crystallized simultaneously.

The compositions of picrite Cpx correspond to diopside (Fig. 1) with Cpx I tending to have lower Fe contents and/or higher Mg contents. Cpx II from picrites have similar projections to those from alkaline basalts of cover units and the Krížna nappe. The former are more homogenous with a weaker zonality as a result of slower cooling of picrite melts. In the picrites from the Outer Western Carpathians only Cpx II have been analyzed. In different discrimination diagrams (Le Bas 1962; Leterrier et al. 1982 and others) clinopyroxenes from the studied picrites lie in the field of Cpx from alkaline rocks.

Al^{IV} (tetrahedral Al) and Ti contents in clinopyroxenes are closely interrelated. Jagi & Onuma (1967) supposed that ions of these elements are present in a hypothetical molecule of Ti-pyroxene $CaTiAl_2O_6$ with considerably decreased solubility in diopside at high pressures. If these findings are applied to the studied pyroxenes, then Cpx I crystallized under high pressures (most likely together with olivine) and Cpx II represents a product of differentiation and fractionation of the primary magma during its ascent to the surface (decompression).

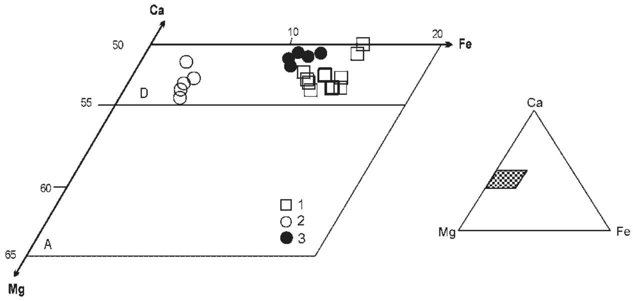

Fig. 1. Classification diagram (after Morimoto et al. 1988) of picrite Cpx from the vicinity of Poniky with fields: A — augite, D — diopside, 1–2 = analyses of picrite Cpx from wider surroundings of BB (original analyses), 3 = analyses of picrite Cpx from the Outer Western Carpathians (Šmíd 1978 in Hovorka & Spišiak 1988)

A large part of analyses coincide with the field of alkaline basalt clinopyroxenes from the other tectonic units of the Central Western Carpathians which, at the same time, explains the similarity of the compositions of picrite and alkaline basalt clinopyroxenes in this zone.

Dark micas are represented by deep-brown through brown-red (Z) flakes, 0.1–3 mm in size. In detail, they are irregularly distributed — here and there they form irregularly bordered clusters of flakes. The percentage of their volume in the rock is 5 %. According to their chemical compositions picrite dark micas fall in the field of phogopite in Deer's et al. (1962) classification diagram. They have characteristic high MgO (approx. 22 weight %) and TiO_2 (approx. 4.2 weight %) contents and zonality.

Amphiboles are present in picrites as accessoric minerals of typical brown colour and with intensive pleochroism. They occur rarely and form idiomorphic columns, but mostly irregular grains. Owing to their colour, optical properties, overall mineral association, as well as their analogy with alkaline basalt amphiboles of the Krížna nappe they are likely to be amphiboles of kaersutite type.

Apatite is a typical accessory mineral of picrites. It forms long column-like through spicular crystals.

A typical ore mineral is Fe-Al-Cr-spinel of snuff-brown colour which has a lobate, skeleton, very irregular bordering. Most grains of spinel-group minerals have narrow rims formed by opaque ore phase. Its composition (Stevens's classification) corresponds to Cr-spinel.

The space between minerals of generations I and II are in some places filled by brownish glassy matrix, locally with a developed microspherolitic structure, irregular clusters of small ore minerals and clusters of chlorite flakes. In the glassy aggregate a fine-flake low-birefringent serpentine, probably lizardite sporadically occurs. The product of different types of post-magmatic alterations is the association of calcite, talc, pyrite, chlorite and limonite.

Geochemistry

In spite of a rather high degree of alteration we tried to study the picrite chemical composition. For geochemical evaluation we used namely the group of REE (Fig. 2). Then we compared the compositions of the picrites under study with those of alkaline basalts of oceanic islands and an average teschenite. The examined basalts show a very similar character of normalization curve with a typical sharp inclination to heavy REE and a hardly observable Eu anomaly.

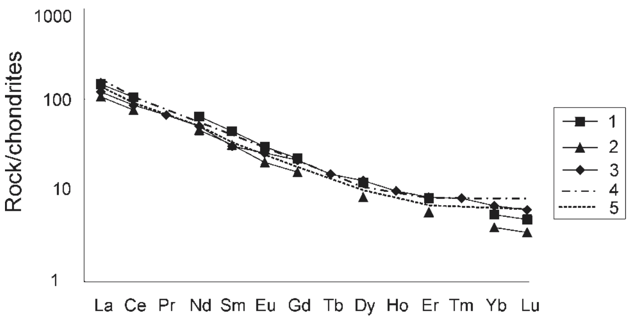

Fig. 2. Normalized curve of REE distribution in the picrites of the Central Western Carpathians 1, 2 = analyses from Table 3), 3 = average composition of the picrites of the Outer Western Carpathians (analyses from Kudělásková 1987 and Dostál & Owen 1998), 4 = average composition of ocean-island alkaline basalts (OIB, Sun & McDonough 1989), 5 = average composition of teschenites (Rosi et al. 1992).

Conclusions

— The picrites from the Silesian nappe of the Outer Western Carpathians and those from the wider surroundings of Banská Bystrica are similar in their petrographic character. Likewise, a close similarity is displayed by the compositions of basic rock-forming minerals.

— The occurrence of picrite eruptives points to a fast penetration of primary upper-mantle melt, not influenced by fractional crystallization, to shallow levels of the crust. Such a penetration is possible only in a zone of great inhomogeneity — in a zone of dilatation which enabled the fast (very likely single-act) penetration of the melt into the upper mantle. Such a genesis is supported by the presence of two very different Cpx generations and other minerals (olivine, spinel, phlogopite).

— On the basis of present knowledge we can classify the picrites from around Banská Bystrica with the province of Mesozoic (Lower-Cretaceous) alkaline basalts/basanites.

This paper is a partial output of the grant projects No. 2/4077/99 and 1/6000/99 VEGA. We wish to express our thanks to Mrs. Halašiová for computer-processed pictures.

References

Deer W.A., Howie R.A. & Zussman J., 1962: Rock-forming minerals, *Longmans Green and Co*., London, III, 270.

Dostál J. & Owen J.V., 1998: Cretaceous alkaline lamprophyres from northeastern Czech Republic: geochemistry and petrogenesis.

Hovorka D. & Slavkay M., l966: Pikrit od Poník. Geol. práce, Spr., 39, *GúDŠ,* Bratislava, 4l-5l.

Hovorka D. & Spišiak J., 1988: Mesozoic volcanism of the Western Carpathians (in Slovak with English abstract). *Veda*, Bratislava, 263.

Kudělásková J., l987: Petrology and geochemistry of selected rock types of teschenite association, outer Western Carpathians. *Geol. Zbor. Geol. Carpath*., 38, 545-573.

Le Bas M.J., l962: The role of aluminium in igneous clinopyroxenes with relation to their parentage. *Amer. J. Sci.,* 260, 267-288.

Leterrier J., Maury R.C., Thonon C., Girard I. & Marche L.M., l982: Clinopyroxene composition as a method of identification of the magmatic affinities of paleo-volcanic series. *Earth Planet. Sci. Lett*., 59, 139-154.

Morimoto N., Fabries J., Ferguson A.K., Ginzburg I.V., Ross M., Seifert F.A., Zussman J., Aoki K. & Gottardi G., 1988: Nomenclature of pyroxenes. *Amer. Mineralogist*, 73, 1123-1133.

Slavkay M., l979: Dalšie výskyty ultrabázických efuzív v chocskom a krížnanskom príkrove pri Banskej Bystrici. *Miner. slovaca*, ll, 3, 239-245.

Spišiak J. & Hovorka D., 1997: Petrology of the Western Carpathians Cretaceous primitive alkaline volcanics. *Geol. Carpathica*, 48, 2, 113-121.

Stevens R.E., 1944: Composition of some chromite of the western hemisphere. *Amer. Mineralogist*, 29, 1-34.

Sun S.S. & Mc Donough W.F., 1989: Chemical and isotopic systematic of oceanic basalts: implication for mantle composition and processes. In: Saunders A.D. & Norry M. J. (Eds.): Magmatism in the oceanic basins. *Geol. Soc. Lond., Spec. Publ*., 42, 313-345.

PEGMATITE POTASSIUM FELDSPARS FROM SOMEŞUL RECE-VALEA IARA, PEGMATITE DISTRICT (APUSENI MTS., ROMANIA)

D. STUMBEA

Department of Mineralogy-Geochemistry, University “Al. I. Cuza”, 20A Bd. Copou, 6600 Iaşi, Romania; dstumbea@vela.uaic.ro

Key words: potassium feldspars, geochemistry, geothermometry, pegmatites, Carpathians Mts.

Introduction

The Someşul Rece-Valea Iara Pegmatite District belongs to the Gilău-Muntele Mare Pegmatite Subprovince, a part of the so-called Carpathian Pegmatite Province. In this district, pegmatite bodies are hosted by medium grade metamorphic rocks (the Someş series), built up of gneisses, migmatites, leptynites, and rarely quartzites and amphibolites. The Muntele Mare Granite body encloses the metamorphic formations of this series in the W and S.

The mineralogy of pegmatites is simple, most abundant are *quartz* (several generations), *potassium feldspars* (generally maximum and intermediate microcline) and *Ca-Na feldspars* (low albite and albite-oligoclase), subordinate minerals are *biotite*, *tourmaline* and *garnets* and *beryl*, *cordierite*, *staurolite*, *sillimanite* and *amphiboles* are scarce. This simple mineralogy is typical of pegmatites that formed during metamorphic processes. Despite their metamorphic genesis, the pegmatite bodies from Someşul Rece-Valea Iara District (Mârza 1980; Stumbea 1998a) have a granite-like composition.

Methods

The following methods were used to better characterize the potassium feldspars from the pegmatite bodies: microscopical description of mineralogical associations, chemical analysis of major elements, electron microprobe analysis of major elements (by means of a CAMECA SX 50 analyser), emission spectrophotometry analysis of minor elements and X-ray powder diffractometry (by means of a PHILIPS PW 1730 diffractometer).

Results

The study of potassium feldspars under microscope confirmed the abundance of microcline; in the neighbouring of the Muntele Mare granite body especially the presence of orthoclase has been found too. Macroscopically, but also microscopically, perthitic intergrowths and graphic structures have been observed, but perthitic-like and graphic-like structures that developed during substitution processes also occur.

Using Barth’s (1969) correlation diagrams, between the chemical composition of K-feldspars and their optical and physical properties, the following parameters have been determined (Table 1).

Besides the parameters listed in Table 1, *2V angle* (81°53’-82°47’), *birefringence* (0.00685-0.00715) as well as *abrasion hardness* — 250-300 on (001) crystallographic face, 380-415 on

(010), 530–595 after a perpendicular plane in relation to [100] axis have been determined.

To characterize the order–disorder grade, the *triclinicity* (Δr), the *degree of order* ($\Delta\rho$) and the *position of aluminium in silica tetrahedra* (T_1o, T_2o, T_1m, T_2m) have been determined (Table 2).

Generally, the abundances of major elements in K-feldspars from the Someşul Rece-Valea Iara Pegmatite District are similar to occurrences from the rest of the Carpathian Pegmatite Province. However, in terms of alkali oxides, we have an evidence that the correlation between K_2O and Na_2O changes within the alignment *Tina Dumitreasa (in the SW)–Corabia Hill (in the NE)*. While a positive correlation exists in SW perimeter, a negative one prevails in NE perimeter. In addition, $K{:}X_{deficiency}/X_{excess}$ and $Al_T{:}T_{deficiency}/T_{excess}$ diagrams show the mean K and Al_T cations fulfilling the X and T sites, in the K-feldspar structural formula.

The distribution of minor elements in potassium feldspars is also comparable to the K-feldspars from the rest of the Carpathian Pegmatite Province. However, we also observed positive correlations between Ba and K_2O, and between Sr and K_2O, respectively, as well as Ba prevailing over Sr and a decreasing content of Pb in the sequence *K-feldspars in granites–K-feldspars in pegmatite hosted by granite bodies–K-feldspars in pegmatite hosted by metamorphic rocks*. In addition, K-feldspars from pegmatite bodies hosted by granites contain more Ba and Sr than the K-feldspars in the metamorphic rocks from the Someş series.

The temperatures calculated for the K-feldspars range between 545–575 ^oC (for K-feldspars from pegmatite bodies lying unconformably to the schistosity of metamorphic rocks) and 430–475 ^oC, respectively (for feldspars generated by substitution processes). The main range of temperatures for K-feldspars from the Someşul Rece-Valea Iara Pegmatite District is between 250 and 600 ^oC, with a maximum frequency between 308 and 367 ^oC and an average of about 400 ^oC.

Discussion

K-feldspars from the Someşul Rece-Valea Iara Pegmatite District are represented by low temperature variants — *maximum* and *intermediate microcline*. This complies with similar results published in the Romanian geological literature concerning pegmatites of metamorphic genesis from the Carpathian Pegmatite Province (Murariu 1979).

The analysis of spatial distribution of triclinicity in the Someşul Rece-Valea Iara Pegmatite District reveals three ranges of equal values; their shape is elongated and follows the direction of an anticlinal fold axis (SW-NE) (Dimitrescu 1994). On the other hand, the change of sign of correlation between K_2O and Na_2O from WNW to ESE (central parts of Şoimului, Negrii and Galbena valleys) indicates where the associated factors, such as *depth/oriented pressure/temperature*, change. Our correlation of Na-K, K-Ca, Na-Ca cations in the pegmatite muscovite, as well as the change from Al excess to an Al deficiency in Y structural position of muscovite support this finding.

The K-feldspar temperature isolines for the Someşul Rece-Valea Iara Pegmatite District are similar to the lines of equal triclinicity (Stumbea 1997).

The lithium and rubidium contents in the potassium feldspars, from Someşul Rece-Valea Iara District pegmatites indicate a mica-bearing province in which pegmatites develop in the high grade metamorphic areas (after Trueman & Černý's 1982 classification). In these areas, pegmatite bodies should have a metamorphic genesis (see also Mârza 1980; Stumbea 1998a) and — in our case — their origin seems to be related to the folding processes that generated a SW-NE striking anticline (Stumbea 1998b).

Table 1: Refractive index, α and γ angles and the specific gravity (G) of K-feldspars from the Someşul Rece-Valea Iara Pegmatite District.

	FK 53	FK 60	FC 30	FC 43	FC 35	FC 36	FC 38	FC 39
n(α)	1.5170	1.5195	1.5217	1.5220	1.5205	1.5200	1.5215	1.5190
n(β)	1.5220	1.5220	1.5260	1.5265	1.5245	1.5240	1.5255	1.5235
n(γ)	1.5230	1.5250	1.5270	1.5290	1.5270	1.5265	1.5280	1.5250
α(°)	90°40'	90°54'	90°44'	90°52'	90°37'	90°42'	90°38'	90°36'
γ(°)	87°67'	87°34'	87°37'	87°37'	87°37'	87°37'	87°38'	87°42'
G	2.5265	2.5650	2.5810	2.5810	2.5745	2.5735	2.5765	2.5665

Table 2: Triclinicity (Δr), degree of order ($\Delta\rho$) and position of aluminium in silica tetrahedra (T_1o, T_2o, T_1m, T_2m) for K-feldspars from some pegmatite bodies of the Someşul Rece-Valea Iara Pegmatite District as compared to standard values.

Sample	Δr	Δρ	Aluminium in silica tetrahedra			
			T_1o	T_2o	T_1m	T_2m
F17	0.971	0.96	0.990	0.975	0.015	0.010
F19	0.995	0.94	0.977	0.960	0.017	0.012
F20	0.961	0.90	0.050	0.935	0.024	0.017
Maximum microcline*	0.98	0.90	0.97	0.94	0.03	0.03
Intermediate microcline*	-	0.42	0.96	0.69	0.27	0.04

* Standard values from Deer et al. (1982)

References

Barth T.F.W., 1969: Feldspars. *Wiley-Interscience*, New York-London-Sydney-Toronto, 1–261.

Deer W.A., Howie R.A. & Zussman J., 1982: An Introduction to the Rock-Forming Minerals. *Longman,* Essex, 1–528.

Dimitrescu R., 1994: Structura părţii centrale a munţilor Gilăului. *Stud. Cerc. Geol.,* 39, 3–6.

Mârza I., 1980: Considérations génétiques sur les pegmatites du cristallin de Gilău (Monts Apuseni) et la province pegmatitique carpatique. *Anuarul I.G.G., (Bucureşti),* LVII, 423–432.

Murariu T., 1979: Studiul mineralogic, geochimic şi structural al pegmatitelor din Munţii Rodnei. *Stud. Tehn. Econ., Ser. I, Mineral. Petrogr., (Bucureşti),* 15, 3–272.

Stumbea D., 1997: La température d'engendrement des pegmatites de Monts Apuseni. *An. Univ. "Al. I. Cuza", XLII-XLIII (1996–1997), Geologie, Iaşi,* 77–85.

Stumbea D., 1998a: Pegmatitele din cristalinul Gilăului (Munţii Apuseni). *Ph. D. thesis, Univ. "Babeş-Bolyai",* Cluj-Napoca.

Stumbea D., 1998b: The Nature of relationship between Muntele Mare Granite and Pegmatite bodies hosted by neighbouring metamorphic rocks in Apuseni Mountains (Romania). *Carpathian-Balkan Geological Association XVI Congress. Abstr.,* Viena, 580.

Trueman D.L. & Cerný P., 1982: Exploration for rare-element granitic pegmatites. *Short Course Handbook, 8,* Winnipeg, 463–493.

VOLCANISM AND SEDIMENTATION IN THE NORTH-WESTERN PART OF THE CENTRAL SLOVAK NEOGENE VOLCANIC FIELD

L. ŠIMON

Geological Survey of the Slovak Republic, Mlynská dolina 1,
817 04 Bratislava, Slovak Republic; simon@gssr.sk

Key words: Volcanism, Sedimentation, Central Slovakia, Western Carpathians.

Introduction

The Central Slovak volcanic field is situated in the inner side of the Carpathian arc. The volcanic rocks are Badenian to Pannonian in age (16.5–8.5 Ma). The volcanic activity was contemporaneous with the evolution of a horst and graben structure induced by backarc extension (Konečný et al. 1995).

The north-western part of the Central Slovak Neogene volcanic field (Fig. 1) is maily represented by region of the Vtáčnik Mts. and Hornonitrianská kotlina Depression (Šimon et al. 1997). The aim of this abstract is to present the main geological and volcanological aspects of the area. Most of the information obtained during 12 years of systematic geological mapping and related volcanological and petrographic research is summarized in unpublished reports of the Geological Survey of the Slovak Republic, Bratislava (Šimon 1991; Šimon et al. 1991, 1994, 1997, 1998).

An overview of the structure and development of the region

The geological structure of the region incorporates Neogene sedimentary and volcanic rocks (Fig. 2). Owing to fault tectonics the geological structure of the region is relatively complicated and characterized by a horst–graben pattern of Neogene age. The faults with large vertical throws divide the region into the main blocks, which are mostly tilted (rotated) and segmented by other faults with smaller vertical throws. The Neogene fill of the region proper may be generally visualized as a block of higher order of a westward dipping horst-graben structure that was segmented by a NE and NW running fault into partial blocks.

Generally, the volcanic rocks in the region lie on a smoothed and flattened surface made up of Mesozoic and Paleogene rocks. This indicates that the erosion and peneplainization took place due to regional uplift at the end of Early Miocene and preceded the first manifestations of Early Badenian extension and volcanism.

The first manifestations of Early Badenian extension and subsidence were associated with paleogeographical changes which provoked a transport of clastic material from the eroded pre-Tertiary units (Kordíky Formation and lower part of Kamenec Formation) and, at the same time, an encroachment of the sea as far upon this land as to form an embayment (Early Badenian marine sediments). In other parts of the region the sedimentation had a fluvial-limnic character. Approximately coeval with these event was the Early Badenian activation of andesitic volcanism in the Central Slovak volcanic region (Complex of andesites with garnet). South of Handlová and near Cígeľ it is represented by explosive-effusive volcanism which produced independent extrusive domes. The products of the garnetiferous andesite volcanic suite became later sources of the material for the Kamenec and Kordíky Formations.

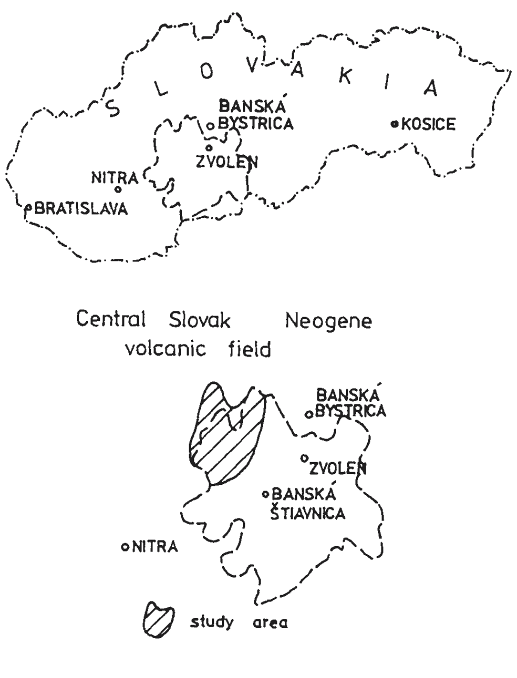

Fig. 1. Location of the study area in the Central Slovak Neogene volcanic field.

Badenian volcanic activity continued with the formation of large stratovolcanoes — the Štiavnica stratovolcano and Kremnica stratovolcano. The Štiavnica stratovolcano (with its centre near Banská Štiavnica) is a product of explosive-effusive activity, and its periferal zone with the Prochoť intrusive Complex encroaches upon the region (SE part of the Vtáčnik Mts.). Transitional zone of the Kremnica stratovolcano (Zlatá Studňa Formation — with its centre near Kremnica) reaches as far as the region under study. The development of the volcano, especially its beginning, took place in a subaqueous, but later in a terrestrial environment.

The above mentioned stratovolcanoes were subsequently, during the Middle and Late Badenian stages, intensively denuded and their material is found in the Kamenec Formation of the Vtáčnik Mts. and of the Hornonitrianska kotlina Depression, where a subsidence initiated later development of swampy and lacustrine environments and coal seam formation (the Handlová-Nováky Formation and Koš Formation). The subsidence was locally accompanied by acidic volcanism of extrusive-explosive type which was responsible for the formation of independent volcanic domes in the area of the Vtáčnik Mts. (Nová Lehota and Plešina Formations).

Sudden geological changes occurred in the area under study at the end of the Badenian and during the Early Sarmatian stages. An extensive caldera developed in the area of the Štiavnica volcanic apparatus which encroached upon the southern part of the region. The Kremnica graben formed in the north-eastern part of the region. A rapid subsidence of the Kremnica graben and of the Žiarska kotlina Depression took place. The initial stage of subsidence was compensated by intense effusive and explosive volcanic activity represented by basalts, basaltic andesites, pyroxenic andesites and leucocratic andesites of the Kľakovská dolina Formation and of the Turček Formation which developed partly in subaqueous environment, and which forms the basal part of the Kremnica graben and Žiarska kotlina Depression fill. The subsidence was later accompanied by predominantly effusive activity represented by hornblende-pyroxene andesites of the Stráň and Kremnický štít Formations. The subsidence of the graben also caused the paleogeographical changes, such as the erosion of its surroundings and change of stream courses towards the Žiarska kotlina Depression followed by the development of fluvial gravels of the Lehota Formation (as much as 300 m thick). During the Early Sarmatian stage the volcanic activity of pyroxenic andesites resumed and the stratovolcanoes formed along the marginal fault of the Kremnica graben, in the Kľakovská dolina valley and near Remata (Vtáčnik and Remata Formations). The pyroxenic andesites of the 4th stage of the Štiavnica stratovolcano reached as far as the southern part of the region.

Younger tectonic movements at the end of Sarmatian stage caused relative uplift and the volcanic rocks became subject to rapid erosion. An extensive rhyolitic volcanism (Jastrabá Formation) was activated at the end of Sarmatian and at the beginning of Pannonian stage. As a response to rhyolitic masses progressing upwards an enormous fault belt (the Vyhne-Ihráč zone) and the marginal faults of the Žiarska kotlina depression developed.

The closing stage of the volcanic development was characterized by the consolidation movements of blocks and by the subsid-

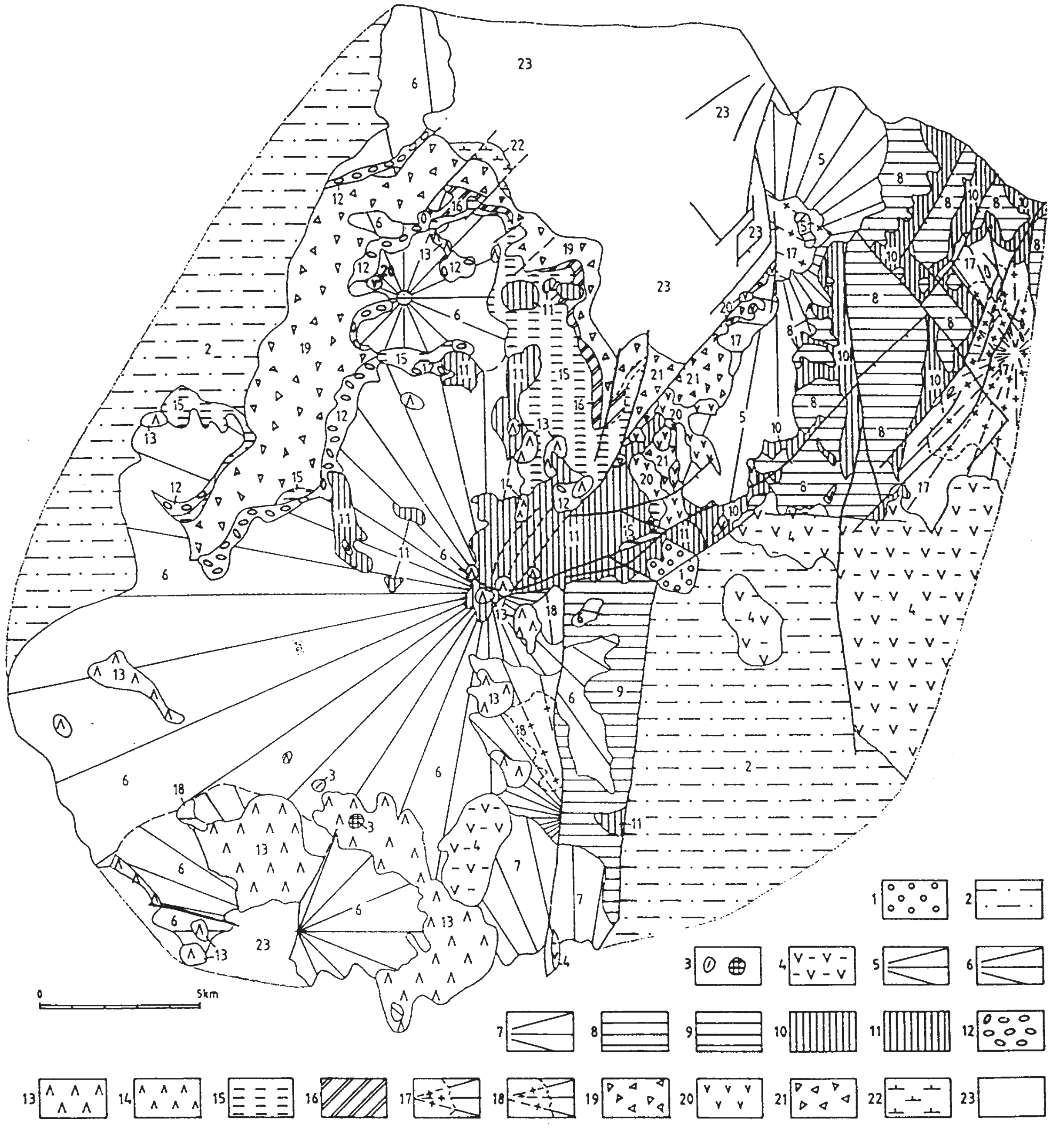

Fig. 2. Layout of formations and complexes in the north-western part of the Central Slovak Neogene volcanic field (Šimon 1999 complied according to Šimon et al. 1997 and Lexa et al. 1998). 1 — Pliocene sediments, 2 — Pannonian-Pontian sediments, 3 — Ostrovica dykes and neck, 4 — Jastrabá Formation, 5 — Remata Formation, 6 — Vtáčnik Formation, 7 — upper structure of the Štiavnica Formation, 8 — Kremnický štít Formation, 9 — Stráň Formation, 10 — Turček Formation, 11 — Kľakovská dolina Formation, 12 — Lehota Formation, 13 — Plešina Formation, 14 — Nová Lehota Formation, 15 — Koš Formation, 16 — Handlová and Nováky Formation, 17 — Zlatá studňa Formation: a) intrusive complex, b) stratovolcanic complex, 18 — The first stage of Štiavnica stratovolcano: a) intrusive complex, b) stratovolcanic complex, 19 — Kamenec Formation, 20 — Complex of garnet-bearing andesites, 21 — Kordíky Formation, 22 — Early Miocene sediments, 23 — Pre-Neogene rocks.

ence of partial segments. This brought about a revival of volcanic activity. As a results there formed the Ostrovica dykes and a neck Complex — the youngest (Pannonian in age) manifestations of volcanism in the Vtáčnik Mts., as well as in the whole region.

References

Konečný V., Lexa J. & Hojstričová V., 1995: The Central Slovakia Neogene volcanic field: a review. *Acta Vulcanol.*, 7, 63–78.

Lexa J., Halouzka R. & Havrila M., 1998: Geological map of the Kremnické vrchy Mts. *GSSr*, Bratislava.

Šimon L., 1991: Faciálny model formácie Kľakovskej doliny južne od Handlovej. *Manuskript – archív GSSR*, Bratislava, 57.

Šimon L., Lexa J., Halouzka R., Macinská M., Jánová V., Vranovská A., Stolár M., Ďurkovičová J., Vozár J., Novosad P., Sládková M. & Wiegerová V., 1991: Vysvetlivky ku geologickej mape 36-311 Janova Lehota. *Manuskript – archív GSSR*, Bratislava, 107.

Šimon L., Elečko M., Gross P., Kohút M., Miko O., Pristaš J., Lexa J., Mello J., Hók J., Macinská M., Köhler E., Jánová V., Raková J., Snopková P., Samuel O., Stolár M., Vozár J., Kováč P., Vass D., Marcin D., Ďurkovičová J., Sládková M. & Wiegerová V., 1994: Vysvetlivky ku geologickým mapám 36–133 (Handlová), 35–244 (Prievidza-4), 36–131 (časť Ráztočno). *Manuskript – archív GSSR*, Bratislava.

Šimon L., Elečko M., Lexa J., Konečný V., Kohút M., Gross P., Pristaš J., Halouzka R., Mello J., Polák M., Vozárová A., Vozár J., Havrila M., Kolerová M., Stolár M., Jánová V., Marcin D., Szalaiová V. & Lanc J., 1997: Vysvetlivky ku geologickej mape Vtáčnika a Hornonitrianskej kotliny v mierke 1:50,000. *GsSr*, Bratislava.

Šimon L., Verbich F., Beck J., Fazekaš J., Sámal J., Kováč J., Marcin D., Elečko M. & Lexa J., 1998: Handlová-juh, hnedé uhlie, regionálna geológia — štúdia. *Manuskript – archív GSSR*, Bratislava.

CLASTS OF TOURMALINE-RICH ROCKS IN LOWER TRIASSIC QUARTZITES, THE TATRIC UNIT, CENTRAL WESTERN CARPATHIANS: TOURMALINE COMPOSITION AND PROBLEM OF SOURCE AREAS

P. UHER

Geological Institute, Slovak Academy of Sciences, Dúbravská cesta 9, 842 26 Bratislava, Slovak Republic; geoluher@savba.savba.sk

Key words: Western Carpathians, Tatric Unit, Lúžna Fm., Lower Triassic, tourmaline-rich rocks, schorl, dravite, uvite, foitite, electron microprobe, source areas.

Introduction

Tourmaline-rich rocks (TRR) belong to uncommon but distinctive and exotic lithologic type of Lower Triassic (Scythian) conglomerate intercalations in the Lúžna Fm. quartzite sequence. The Lúžna Fm. (Fejdiová 1980) is typical member at a base of Tatric Mesozoic cover, its thickness ranges 50 to 200 m in average, the sequence continues to the Unterostalpin of the Eastern Alps (Semmering quartzite in Austria). The conglomerates to breccias form irregular layers and erosive channel bodies up to several dm thick in psammitic quartzites; the whole clastic sequence was interpreted as a product of continental ephemeral stream sedimentation in (semi)arid environment (Mišík & Jablonský 1978). Clasts in the conglomerates are well to poorly rounded, up to 15 cm in size. Vein quartz is the most widespread rock type of clasts, TRR, rhyolites, lydites, metaquartzites, metapelites and other rocks are subordinate to rare (0–10 %, Mišík & Jablonský 1978). TRR clasts were found only in SW part of the Tatric Unit: mainly in the Malé Karpaty and adjacent Hundsheim Mts., rarely in the Považský Inovec, Tribeč and Strážovské vrchy Mts., as well as in the Unterostalpin of Wechsel area of the Eastern Alps (Mišík & Jablonský, this vol.). Tourmalinised quartzites, greywackes, phyllites and pyroclastic rocks were recognized as principal lithologic types of TRR (Mišík & Jablonský 1978).

Tourmaline description and composition

TRR are generally fine-grained to cherty dark grey to black rocks without or with apparent schistosity, tourmaline (Tur) content is variable, usually between ca. 30 to 90 vol. %. Anhedral quartz is other common mineral of TRR, fine-grained muscovite, biotite, chlorite and feldspars are rare to subordinate, zircon, monazite-(Ce), xenotime-(Y), hematite, rutile, titanite and epidote belong to accessory phases. Tur forms clusters of acicular to columnar crystals (0.03 to 3 mm, commonly ~0.1–0.4 mm long), randomly oriented or as radiating aggregates which apparently penetrated into quartz. Tur crystals are unzoned to strongly zoned, locally with fine oscillatory zoning and weak to strong absorption: ε — pale yellow to green, ω — dark khaki green to deep blue, rarely dark brown.

36 electron microprobe analyses of 2 TRR samples from the Malé Karpaty Mts. show narrow to wide compositional variability of Tur, even within a single crystal or sample (Table 1, Fig. 1). *TQ-1* Tur reveals X-site vacant rich schorl to foitite compositions slightly enriched in Mg and commonly also in ${}^{X}\square$ (X-site vacancy) toward the rim of crystals: Fe/(Fe+Mg) = 0.70 to 0.53, ${}^{X}\square$ = 0.46 to 0.65 pfu (per formula unit). Al > 6 apfu (atoms pfu) indicates a strong prevalence of Fe^{2+} over Fe^{3+}, Cr locally elevated to 0.22 apfu (1.7 wt. % Cr_2O_3), Ti and Mn are low to moderate (0.0–0.05 apfu). Ca is commonly low (up to 0.12 apfu; 0.7 wt. % CaO), K is almost absent. In contrast, *TQ-2* Tur shows distinctive and more complex zonality within a crystal (Table 1). Central parts have schorl>dravite compositions, they exhibits a wide variability in Al and ${}^{X}\square$: Fe/(Fe+Mg) = 0.53–0.58, Al = 5.10–6.36 apfu, ${}^{X}\square$ = 0.09–0.36 pfu, whereas rims belong to Fe>>Mg schorl to foitite (or buergerite ?), often Al-poor, K- and ${}^{X}\square$-rich: Fe/(Fe+Mg) = 0.72–0.97, Al = 4.46–5.49 apfu, K = 0.04–0.07 apfu (0.2–0.3 wt.

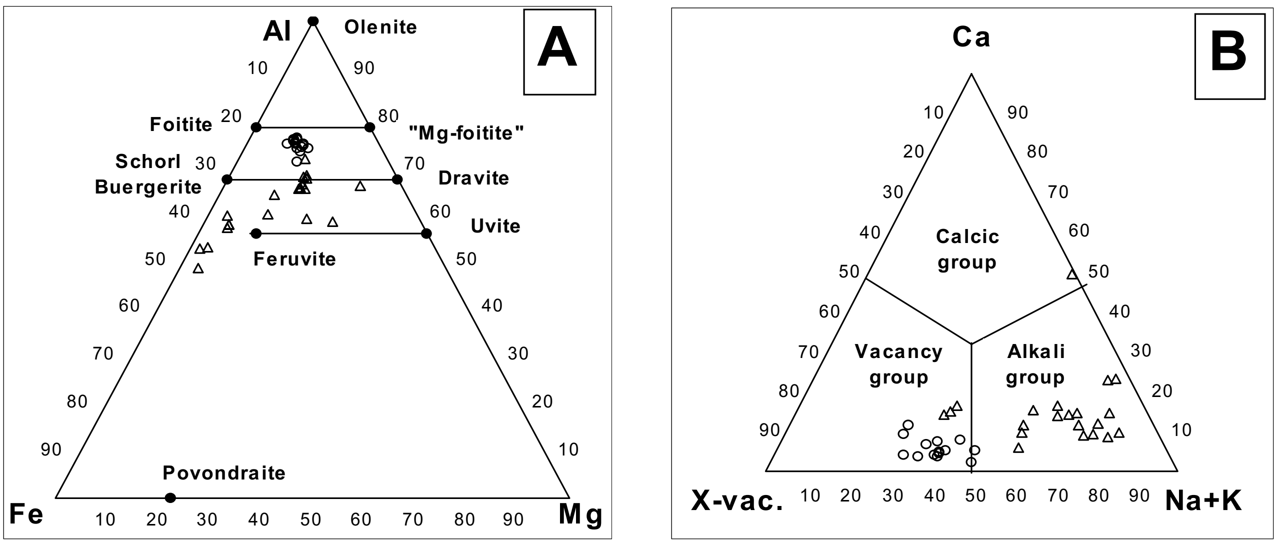

Fig. 1. Compositions of tourmaline from TQ-1 (open circles) and TQ-2 (open triangles) tourmaline rich rocks, the Malé Karpaty Mts.

Table 1: Composition of tourmaline-rich rock (TJ-5) and tourmalines (oxides in wt.%).

Locality	Traja jazdci Hill			Devínska Kobyla Hill			
Sample#	TJ-5	TQ-1	TQ-1	TQ-2	TQ-2	TQ-2	TQ-2
Position	Rock	Center1	Rim1	Center1	Rim1	Center2	Rim2
SiO_2	81.58	35.93	36.10	35.22	29.43	35.05	36.12
TiO_2	0.30	0.16	0.09	0.37	0.35	1.98	0.74
B_2O_3*	2.25	10.46	10.63	10.32	9.71	10.26	10.62
Al_2O_3	8.52	33.18	34.96	29.38	23.70	25.50	30.37
Cr_2O_3	n.a.	1.66	0.14	0.00	0.60	0.21	0.52
FeO_{tot}	2.64	9.20	10.67	12.85	29.28	10.59	5.15
MnO	0.03	0.20	0.14	0.17	0.41	0.00	0.00
MgO	1.52	3.96	3.51	5.38	0.59	8.68	9.73
CaO	0.71	0.37	0.23	0.49	0.75	2.73	1.32
Na_2O	0.35	1.10	0.99	2.41	0.89	1.44	2.21
K_2O	0.22	0.00	0.00	0.00	0.23	0.11	0.06
H_2O*	1.16	3.61	3.67	3.56	3.35	3.54	3.67
Total	99.28	99.82	101.12	100.15	99.29	100.09	100.51
Formulae based on the sum of T + Z + Y = 15 cations							
Si^{4+}		5.973	5.902	5.934	5.267	5.937	5.909
Al^{3+}T		0.027	0.098	0.066	0.733	0.063	0.091
Total T		6.000	6.000	6.000	6.000	6.000	6.000
B^{3+}		3.000	3.000	3.000	3.000	3.000	3.000
Al^{3+}Z		6.000	6.000	5.768	4.266	5.028	5.764
Fe^{3+}Z		0.000	0.000	0.232	1.734	0.972	0.236
Total Z		6.000	6.000	6.000	6.000	6.000	6.000
Ti^{4+}		0.020	0.011	0.047	0.047	0.252	0.091
Al^{3+}Y		0.474	0.637	0.000	0.000	0.000	0.000
Cr^{3+}		0.218	0.018	0.000	0.085	0.028	0.067
Fe^{2+}, $^{3+}$Y		1.279	1.459	1.578	2.649	0.528	0.469
Mn^{2+}		0.028	0.019	0.024	0.062	0.000	0.000
Mg^{2+}		0.981	0.856	1.351	0.157	2.192	2.373
Total Y		3.000	3.000	3.000	3.000	3.000	3.000
Ca^{2+}		0.066	0.040	0.088	0.144	0.495	0.231
Na^{+}		0.355	0.314	0.787	0.309	0.473	0.701
K^{+}		0.000	0.000	0.000	0.053	0.024	0.013
Total X		0.421	0.354	0.875	0.506	0.992	0.945
Vac. X		0.579	0.646	0.125	0.494	0.008	0.055
Total Cat.		18.420	18.354	18.876	18.505	18.992	18.945
OH^{-}		4.000	4.000	4.000	4.000	4.000	4.000
O^{2-}		31.095	30.987	30.879	29.681	30.992	31.049
Fe/(Fe+Mg)		0.565	0.630	0.573	0.965	0.406	0.229

*B_2O_3 and H_2O contents of the minerals calculated by ideal stochiometry.

% K_2O) and $^{X}\square$ = 0.17–0.49 pfu. High total Fe: Al atom. ratio (0.65–0.99) could indicate high Fe^{3+} content (buergerite to povondraite mole fraction). Locally, Ti-rich calcic dravite to magnesian uvite compositions occur (Table 1).

In addition, several microprobe analyses of Tur from one TRR clast in Lower Triassic near Trattenbach (Wechsel area) reveal nearly unzonal schorl compositions with Fe/(Fe+Mg) = 0.54 and $^{X}\square$ = 0.20 pfu (Aubrecht & Krištín 1995).

Possible origin and source areas of tourmaline-rich rocks

The origin of tourmaline-rich rocks (tourmalinites) has been a controversal subject for over three decades, the theories have focused on (1) premetamorphic replacement, (2) syngenetic-exhalative, (3) colloidal and/or gel related, (4) evaporitic, (5) contact metasomatic, and (6) regional metasomatic origin (Slack 1996). TRR are also notable for their common close association with a variety of stratabound or skarn ore deposits, e.g. Au, Ag, Cu-Pb-Zn, Fe, Sn, W, Co and U (Kebrt et al. 1984; Slack 1996).

Fine-grained aggregates of Tur penetrating the matrix, remnants of clastic sedimentary textures and minerals of TRR (quartz, micas, chlorite) and absence of higher-grade index metamorphic minerals (e.g. garnets, Al_2SiO_5-polymorphs), indicate a low to medium temperature origin for TRR clasts in the Lúžna Fm. quartzites. Tur crystallized probably during extensive replacement of clastic (rarely volcanoclastic) quartz-alumosilica rocks by B-rich exhalative volcanic/hydrothermal fluids in submarine conditions, the process could takes place under low to medium-grade metamorphic conditions at t~200–400 ^{o}C. However, at least a part of TRR, could be originated at higher temperatures near contacts of magmatic rocks under conditions of amphibolite grade. Chemical composition of TRR indicates a quartz-rich nature of original rock (Table 1) and it is similar to some other Tur-Qtz TRR, e.g. from the Jánov Grúň Complex, Veporic Unit, Slovakia (Miko & Hovorka 1978), or from Mt. Isa region, Australia (Plimer 1988 in Slack 1996).

A source area of TRR is another complex problem. Lower Triassic paleogeographic position of the Tatric Unit as well as whole Central Western Carpathians was very different from that in present due to the large left lateral strike slip movements during Alpine evolution. On the basis of recent data, the Alpine-Carpathian units were derived from the margin of the North European shelf and the Tatric Unit occupied an area in SE continuation of Armorician continent (Michalík 1994). Consequently, TRR clasts of the Lúžna Fm. could be derived from the metamorphic complexes of internal Hercynian orogenic zones, probably from the French Massif Central.

References

Aubrecht R. & Krištín J., 1995: Provenance of detrital tourmaline in the Lower Jurassic of the Malé Karpaty Mts. *Miner. slovaca*, 27, 37–44.

Fejdiová O., 1980: The Lúžna Formation: a formal Lower Triassic lithostratigraphic unit. *Geol. Práce, Spr.*, 74, 95–102 (in Slovak).

Kebrt M., Lhotský P., Pertold Z. & Adam J., 1984: Tourmalinites and tourmaline quartzites in the metamorphic rocks of the Bohemian Massif. In: *Correlation of proterozoic and Paleozoic stratabound deposits.* 7. *ÚÚG & ÚGV PřF UK,* Prague, 85–102.

Michalík J., 1994: Notes on the paleogeography and paleotectonics of the Western Carpathian Area during the Mesozoic. *Mitt. Österr. Geol. Gesell.*, 86, 101–110.

Miko O. & Hovorka D., 1978: Quartz-tourmalinic rocks of the Veporide crystalline complex in the Nízke Tatry Mts. *Záp. Karpaty, Sér. Miner. Petrogr. Geoch. Metalogen.*, 5, 7–28 (in Slovak).

Mišík M. & Jablonský J., 1978: Lower Triassic quartzites and conglomerates of the Malé Karpaty Mts. (pebble analysis, transport directions, genesis). *Acta Geol. Geogr. Univ. Comen.*, 33, 5–36 (in Slovak).

Mišík M. & Jablonský J., 1999: Lower Triassic quartzites of the Western Carpathians: source of clastics, transport directions. *Geol. Carpathica* (this volume).

Slack J.F., 1996: Tourmaline associations with hydrothermal ore deposits. In: Grew E.S. & Anovitz L.M. (Eds.): *Boron. Mineralogy, petrology and geochemistry. Rev. Miner.,* 33, *Miner. Soc. Amer*., 559–643.