Zeitschrift: Schweizerische mineralogische und petrographische Mitteilungen = Bulletin suisse de minéralogie et pétrographie

Band: 81 (2001)

Heft: 2

Artikel: Nd, Sr, Pb isotope study of the Western Carpathians : implications for Palaeozoic evolution

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DOI: https://doi.org/10.5169/seals-61685

SCHWEIZ. MINERAL. PETROGR. MITT. 81, 159–174, 2001

Nd, Sr, Pb isotope study of the Western Carpathians: implications for Palaeozoic evolution

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Abstract

Nd, Sr and Pb–Pb whole rock isotope data from the Variscan basement of the Western Carpathians are presented in order to define the eastern continuation of the Variscides in Central Europe. Isotope analyses of migmatites, orthogneisses and granitoids from the Tatra Mountains (Upper unit) and two neighbouring areas (Velká Fatra and Lower Tatra Mountains) have been examined. The granites, migmatites and orthogneisses show mainly crustal characteristics with $\varepsilon Nd_{(330)}$ values ranging from –2 to –9 and $\varepsilon Sr_{(330)}$ values between 10 and 387. These results are in good agreement with the Pb–Pb whole rock data and document a crustal origin for the investigated rocks with only minor influence of an enriched mafic component such as recycled oceanic crust. REE and geochemical data are consistent with a collisional environment for the emplacement of the granitoids and the precursor of the orthogneisses. The isotope data confirm that the Western Carpathians own a quite similar isotopic composition than other Variscan regions in Europe.

On the basis of the data presented, in combination with our previous U–Pb zircon dating, we propose the following geological evolution for the Tatra Mountains. In Early Devonian time (~406 Ma) subduction-related magmatism led to the formation of the earliest granitoids, the precursors of the present-day orthogneisses. In Late Devonian and Carboniferous time (360–350 Ma), younger granitoids intruded, most probably as a result of collisional thickening. During this event the earlier granitoids were recrystallised and deformed, resulting in the formation of the orthogneiss fabric. The youngest granitic magmatism in the Carboniferous (315 Ma) is related to the Late Variscan orogenic collapse.

Keywords: Variscides, Western Carpathians, radiogenic isotopes, granitoids.

Introduction

The Variscides constitute the main orogenic belt of Europe resulting from the convergence between Gondwana and Laurasia during Palaeozoic time (e.g. MATTE, 1986; ZIEGLER, 1986; NEUGEBAUER, 1988; FRANKE 1989). The Carpathians are the key area for the study of the eastern and south-eastern continuation of the Variscan orogen in Central and Eastern Europe. Similar to the Alps, Variscan complexes form the pre-Mesozoic basement of the Carpathian belt (e.g. KRIST et al., 1992; NEUBAUER and VON RAUMER, 1993). However, the precise chronology and plate tectonic setting of the Western Carpathians are not well constrained.

This study is concerned with the Variscan rock complexes exposed in the Tatra Mountains, the Velká Fatra and the Lower Tatra Mountains (Fig. 1a), which are located in the northernmost sector of the Western Carpathians. They are representative of the so-called core mountains within the Tatric unit, a major tectonic unit of the Western Carpathians (ANDRUSOV, 1968; KRIST et al., 1992).

Nd, Sr and Pb–Pb isotopic data on orthogneisses, migmatites and granites are presented, most of them from the Western and the High Tatra Mountains (Fig. 1b). It is the aim of this paper to present these isotopic results, together with petrological and geochronological data (JANÁK et al., 1996, 1999, POLLER et al., 2000; POLLER and TODT, 2001), and to discuss them relevant to the

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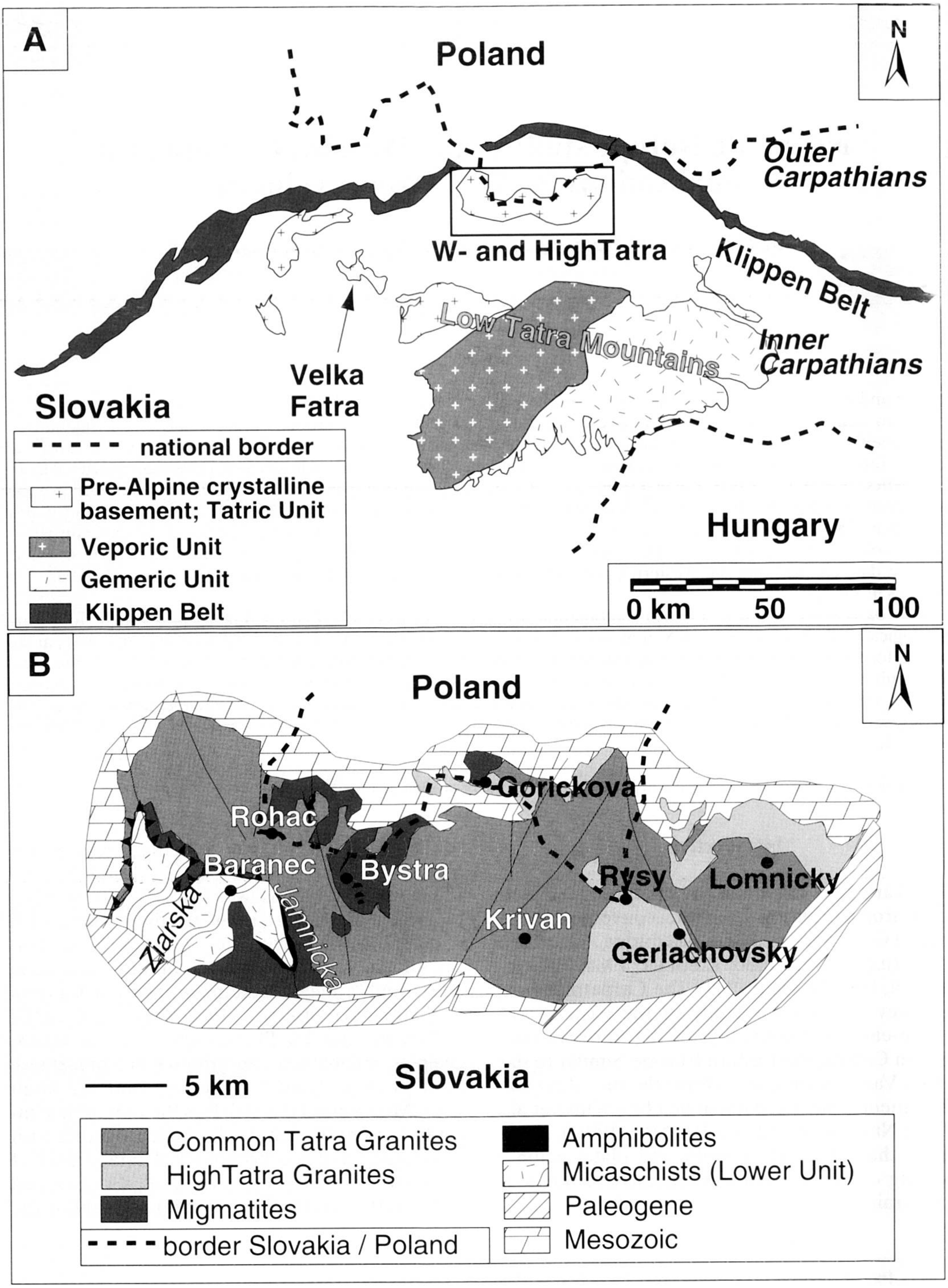

Fig. 1 Simplified geological maps (after KOHÚT and JANÁK, 1994). (A) Western Carpathians with Fatra, Low Tatra, and Tatra Mountains. (B) Detailed view to the Western and High Tatra Mountains.

geodynamic evolution of the pre-Alpine basement of the Western Carpathians and its relationship to other parts of the European Variscides.

Geological Setting

The crystalline basement of the Tatra Mountains (Fig. 1b) is composed of pre-Mesozoic metamorphic and granitoid rocks, overlain by Mesozoic and Cenozoic sedimentary cover sequences and nappes. Metamorphic rocks are abundant in the western part (Western Tatra Mountains), whereas in the eastern part (High Tatra Mountains) they occur only as xenoliths in granitoids. Within the basement, two superimposed tectonic units (lower and upper) have been distinguished, differing in lithology and metamorphic grade (JANÁK, 1994).

The *Lower Unit* is exposed in the Western Tatra in a tectonic window and is composed of micaschists. Kyanite-, staurolite-, fibrolitic sillimanite-, and garnet-bearing metapelites alternate with quartz-rich metapsammites, interpreted to represent former flysch sediments (KAHAN, 1969; GURK, 1999). Two metamorphic zones have been distinguished – the kyanite-staurolite and kyanite-sillimanite (fibrolite) zone (JANÁK, 1994).

The *Upper Unit* is composed of metapelitic paragneiss, amphibolite, and orthogneiss. At its base, kyanite indicates a pressure-dominated type of metamorphism. Metapelites contain kyanite and show migmatisation with formation of granitic leucosomes. Orthogneisses are porphyritic and coarse grained, sometimes also leucocratic and medium to fine-grained.

The banded amphibolites include layers of mafic (amphibolitic) and felsic (tonalitic to trondhjemitic) composition and enclose lenses of retrograded eclogites containing garnet and clinopyroxene (JANÁK et al., 1996). Higher levels of the Upper Unit belonging to the sillimanite zone, exhibit widespread migmatisation and are intruded by granitoids. Sillimanite, garnet, K- feldspar, and cordierite are diagnostic minerals in these high-grade metapelites. The granitoids comprise leucogranites to biotite tonalites as well as amphibole diorites, and form a sheet-like pluton (KOHÚT and JANÁK, 1994).

The crystalline basement of the Tatra Mountains was affected by Variscan and Alpine deformation under distinct P–T conditions and kinematics (KAHAN, 1969; FRITZ et al., 1992). The earliest Variscan deformation (D1) is related to south-east thrusting of the Upper Unit onto the Lower Unit. Subsequent deformation (D2) was the result of Variscan extension and mostly obliterated the compressional, thrust-related deformation fabrics. Alpine deformation (D3) under brittle conditions is manifested by north-west shear, resulting from Late Cretaceous compression. The last major deformation (D4), related to Tertiary extension, produced updoming and normal faulting in a north-south to NW–SE direction.

Rb–Sr whole-rock isochron dating (BURCHART, 1968) indicates Early Palaeozoic tectonometamorphic activities between 380–420 Ma. Granitoid magmatism, according to Rb–Sr isochrons, took place at 290–310 Ma (BURCHART, 1968) or at 340–350 Ma (GAWEDA, 1995).

Recent U–Pb single zircon dating resulted in a detailed chronology for the different magmatic events in the Tatra Mountains (POLLER et al., 2000; POLLER and TODT, 2001). The oldest granites are the present-day porphyric orthogneisses. These intruded 406 Ma ago, followed some 50 Ma later by the main metamorphic overprint and intrusion of granites of the Western Tatra 355 Ma ago. For the High Tatra Mountains this Late Devonian age is reflected only by concordant zircons in the migmatites. Moreover, migmatites show a second concordant zircon age at 335 Ma that is interpreted as the final migmatisation. The final granite magmatism at 315 Ma has been documented by zircons in the High Tatra Mountains only. Cooling ages from granitoids and migmatites range between 330 and 300 Ma according to ^{40}Ar–^{39}Ar white mica dating (MALUSKI et al., 1993; JANÁK and ONSTOTT, 1993; JANÁK 1994) and reflect the late Variscan exhumation.

There is no geochronological indication for a high-temperature Alpine rejuvenation of the basement. Apatite fission track data record the final uplift during the Late Tertiary, at 15–10 Ma (KOVÁC et al., 1994).

In this study, rocks from the Upper Unit – orthogneisses, migmatites, and granitoids – of the Tatra Mountains have been investigated using geochemical and radiogenic isotope methods, in order to constrain the nature of the source rocks involved and the geotectonic evolution of the Tatricum. Details of petrography and analytical procedures are given in the Appendices 1 and 2, respectively.

Results

ORTHOGNEISSES

The *porphyric, coarse-grained orthogneisses* show monzogranitic-granodioritic and peraluminous characteristics documented e.g. by the alumina

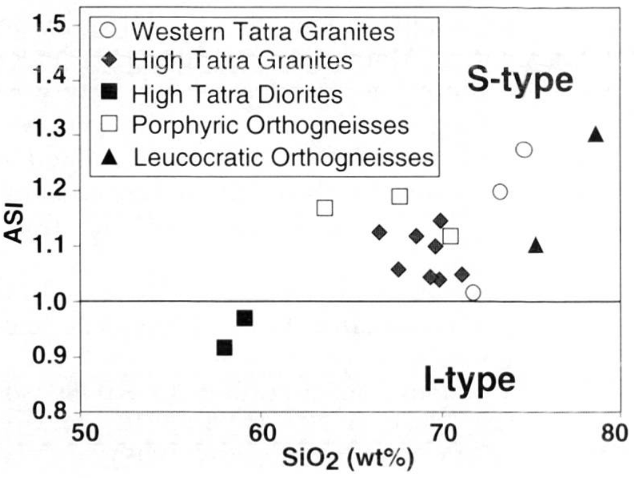

Fig. 2 Discrimination diagram using the alumina saturation index (ASI) versus SiO_2 for the characterisation of S- and I-type granitoids (after CHAPPELL and WHITE, 1974).

saturation index (ASI), which is above 1.12 for all investigated orthogneisses (Fig. 2). Their mineral (see Appendix 1) and chemical compositions reflect an S-type precursor. Considering the high ASI values, FeO_{tot} / (FeO_{tot} + MgO) ratios around 0.7, and the crustal-type $^{87}Sr/^{86}Sr$ values above 0.709 they are characterised as "MPG" granites after BARBARIN (1999). MPG stands for muscovite-bearing peraluminous granites and describes the precursor of the Tatra orthogneisses with respect to their petrographical and geochemical features. A continental collision is the typical geotectonic environment of these granitoids (BARBARIN, 1999). The main process responsible for the evolution of such a crustal-derived magma should be fractional crystallisation. The differentiation status of the Western Tatra orthogneisses is assumed to be rather evolved, visible in the high ASI values (BARBARIN, 1996) and the strong peralumineous character (CHAPPELL et al., 1998). Nevertheless, typical characteristics of highly fractionated granitoids such as increasing P, HREE , and Y contents are not always found. Analyses of the REE of the porphyric orthogneisses resulted in typical crustal chondrite-normalised patterns (Fig. 3), showing a negative Eu anomaly with Eu/Eu* ratios between 0.4 and 0.9.

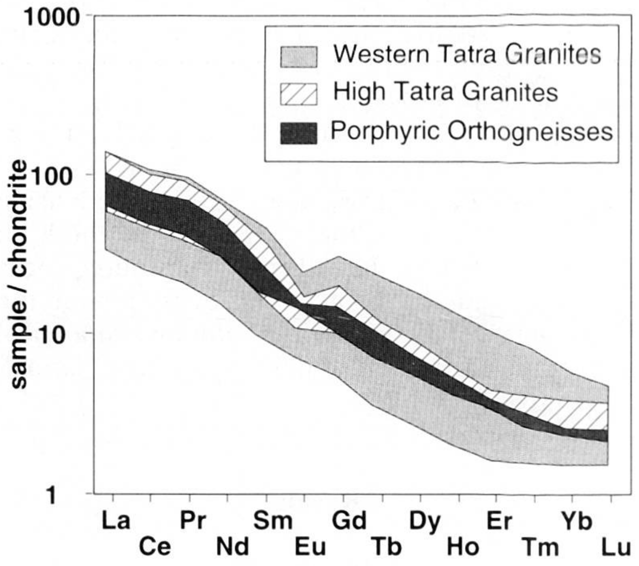

Fig. 3 Chondrite normalised rare earth element diagram for Western and High Tatra granitoids. Normalising data are from EVENSEN (1978).

The *leucocratic orthogneisses* are syenogranitic, with clearly higher SiO_2 contents (75–79 wt%) than the porphyric type. The ASI in sample UP 1005 is 1.3, which indicates the protolith as an S-type granitoid, whereas sample UP 1012 has an ASI of 1.1, indicating a slightly less peraluminous character of the precursor.

Trace elements Nb, Y, Rb, and Zr as well as $FeO_{tot}/(FeO_{tot}$ + MgO) ratios between 0.72 and 0.77 classify the precursors of these rocks as former MPGs that most probably intruded during the collision of two continents or continental fragments. The source material of the leucocratic orthogneisses again was crustal, but compared with the porphyric gneisses, the leucocratic gneisses contain less mafic minerals such as biotite. The leucocratic orthogneiss UP 1012 has an Eu/Eu* ratio of 0.7 defining only a slight negative Eu anomaly, and La_n/Lu_n ratio is around 4.

The geochemical differences (lower content of Al_2O_3, TiO_2, Fe_2O_3, K_2O, higher SiO_2 in the leucocratic orthogneisses) between the two orthogneiss groups of the Western Tatra Mountains are also reflected by the radiogenic isotope characteristics.

The $^{206}Pb/^{204}Pb$ ratios of the orthogneiss whole rock samples range from 17.99 to 18.89 and the $^{207}Pb/^{204}Pb$ ratios from 15.60 to 15.74 (Tab. 4), both of which are significantly higher than mantle or MORB values. The leucocratic sample UP 1012 shows lower Pb–Pb ratios than the porphyric orthogneisses. The $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ data of the Western Carpathian rocks (Fig. 4b) fit well within the continental crust fields of DOE and ZARTMAN (1979). Nevertheless, it is impossible to distinguish whether in addition to the continental crust component there is also a contribution from other sources such as recycled oceanic crust or an ocean island arc. Thus, there is evidently participation of continental-derived sediments and/or oceanic-pelagic sediments during the genesis of these rocks, as documented by the crust-dominated Pb–Pb ratios.

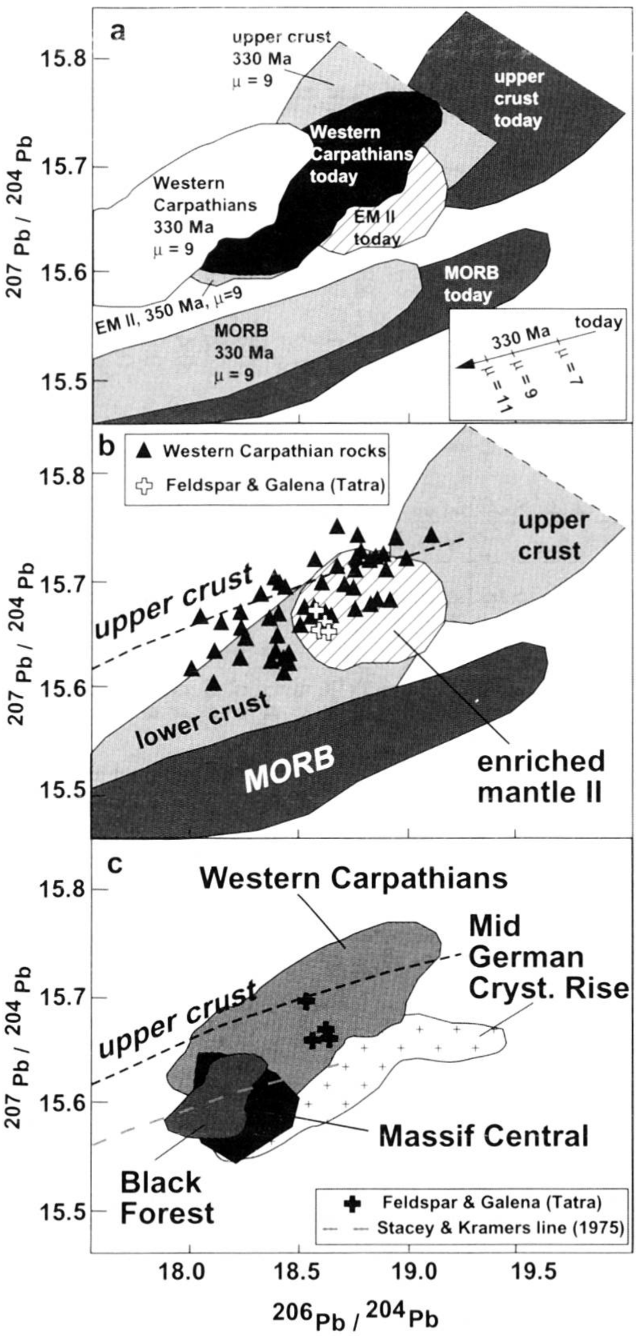

Fig. 4 $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagrams illustrate (a) evolution, (b) origin and (c) isotopic relationship of the Western Carpathian rocks towards typical Variscan areas. In figure 4a the Tatra rocks as well as reference fields for upper crust, enriched mantle II and MORB (after DOE and ZARTMAN, 1979) are shown with their present day position as well as recalculated to 350 Ma with a μ of 9. This illustration reflects the evolution of the Pb–Pb for the last 330 Ma. No essential change between the relative position of the Western Carpathians and the reference fields is documented. Therefore, in figures 4b and 4c the present day Pb–Pb values are used. In figure 4b the Tatra rocks are compared to several sources (after DOE and ZARTMAN, 1979; ZINDLER and HART, 1986). Figure 4c shows the good correspondence between the Tatra and several other Variscan orogenic areas (data from REISCHMANN and ANTHES, 1996; VIDAL and POSTAIRE, 1985). The reference line for upper crust refers to DOE and ZARTMAN (1979).

Sm–Nd whole-rock studies resulted in $\varepsilon Nd_{(330)}$ values between –2.8 and –9.4 (Fig. 5). The $^{87}Sr/^{86}Sr_{(330)}$ ratios show wide variation between 0.7080 (UP 1025) and 0.7322 (UP 1012). Similarly, the $\varepsilon Sr_{(330)}$ values range from 45 to 132 in the porphyric types, whereas for the leucocratic orthogneisses the values scatter between 205 and 387. Also, the Nd model ages of the two orthogneisses are different. Whereas the leucocratic orthogneisses have $T_{(DM)}$ ages between 1650 and 1780 Ma, interpreted by a two-stage evolution, the porphyric orthogneisses have mean crustal residence ages (one step model) ranging from 1220–1350 Ma.

GRANITOIDS AND MIGMATITES

Petrographically, all Tatra granites represent basically crustal-anatectic rocks with primary magmatic muscovite. Biotite is the dominant Fe–Mg mineral, whereas hornblende occurs only rarely in dioritic xenoliths. Prevalence of plagioclase ($An_{20–45}$) over K-feldspar is a typical attribute of the investigated granitoid rocks. The mafic and

Tab. 1 Coordinates of the sample locations from the Tatra Mountains.

samples	northern latitude	eastern longitude
UP 1002	49° 09'04"	19° 42'35"
UP 1005	49° 08'40"	19° 43'00"
UP 1012	49° 09'28"	19° 46'45"
UP 1014	49° 08'24"	19° 48'00"
UP 1015	49° 09'56"	20° 09'25"
UP 1016	49° 09'56"	20° 09'25"
UP 1017	49° 09'47"	20° 09'17"
UP 1018	49° 09'47"	20° 09'17"
UP 1023	49° 10'16"	19° 45'02"
UP 1025	49° 10'32"	19° 44'35"
UP 1026	48° 59'49"	19° 16'31"
UP 1027	48° 59'51"	19° 15'47"
UP 1028	48° 59'55"	19° 15'39"
UP 1030	48° 59'28"	19° 08'40"
UP 1031	49° 00'18"	19° 09'58"
UP 1036	49° 10'45"	19° 51'23"
UP 1038	49° 04'52"	19° 10'26"
UP 1039	49° 11'04"	19° 46'40"
UP 1040	49° 11'50"	19° 46'45"
UP 1044	49° 09'16"	19° 58'50"
UP 1049	49° 09'47"	20° 09'17"
UP 1050	49° 09'47"	20° 09'17"
UP 1051	48° 56'30"	19° 34'20"
UP 1052	49° 08'00"	20° 07'22"
UP 1053	49° 08'00"	20° 07'22"
UP 1055	49° 16'04"	19° 59'05"
UP 1057	49° 10'48"	20° 03'08"
UP 1063	49° 10'30"	19° 50'27"
UP 1088	48° 56'26"	19° 34'35"
UP 1090	48° 54'00"	19° 24'55"

Tab. 2a Major element (in wt%), trace element, and REE (in ppm) data for the investigated rocks of Western and High Tatra Mountains.

	leucocratic orthogneisses		porphyric orthogneisses			High Tatra diorites		High Tatra granitoids and migmatites												
sample	1005	1012	1002	1014	1025	1015	1016	1023	1036	1040	1017	1018	1044	1049	1050	1052	1053	1055	1057	1063
SiO_2	78.6	75.4	70.4	63.6	67.8	58.9	57.6	73.8	70.9	73.1	71.1	69.4	69.9	69.9	67.6	73.4	68.7	66.6	69.8	67.8
TiO_2	0.1	n.d.	0.3	0.8	0.6	0.8	0.9	0.1	0.4	0.2	0.3	0.4	0.3	0.4	0.5	0.6	0.4	0.6	0.3	0.5
Al_2O_3	13.2	14.5	16.3	17.2	16.9	17.1	16.8	14.7	14.9	15.1	15.0	15.6	15.6	15.0	16.2	12.3	16.2	16.7	15.9	16.7
Fe_2O_3	0.9	0.5	2.6	5.5	3.4	6.8	7.4	0.7	2.8	1.8	2.2	2.9	2.2	2.4	3.1	4.4	3.0	3.4	2.3	3.9
MgO	0.4	0.1	1.1	2.4	1.5	3.8	4.4	0.2	1.0	0.5	0.8	0.9	0.7	0.8	1.0	0.8	1.0	1.5	0.8	1.3
CaO	0.5	0.6	1.7	2.9	2.4	5.4	5.7	3.2	2.9	1.1	2.0	4.2	3.7	3.4	4.2	2.3	4.7	4.5	4.6	2.0
Na_2O	4.8	4.6	6.0	4.0	4.5	3.1	3.2	0.5	5.0	4.4	3.6	2.7	1.4	2.0	2.6	3.2	2.1	2.4	2.2	4.8
K_2O	1.3	4.3	1.4	2.6	2.2	2.4	2.4	4.9	1.0	3.1	4.3	2.6	4.3	4.5	3.3	1.9	2.1	2.4	2.2	1.8
P_2O5	0.1	0.2	0.1	0.1	0.2	0.4	0.4	0.1	0.4	0.1	0.2	0.1	0.2	0.2	0.2	0.2	0.1	0.2	0.1	<0.1
LOI	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	1.0	0.8	1.5	0.9	0.88	0.6	1.2	1.6	1.5	1.2
Ba	91	70	300	792	549	1317	1272	2481	307	826	1245	739	1183	1374	1129	286	545	578	879	324
Ga	12	15	15	22	20	21	22	12	16	18	17	20	19	18	20	18	22	21	21	18
Nb	6	5	5	16	10	10	11	3	7	7	8	12	8	9	10	9	10	9	7	8
Ni	5	2	9	28	14	10	12	6	5	2	5	3	3	5	5	2	3	7	4	19
Pb	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	40	n.d.	n.d.	n d.	35	20	47	41	17	26	18	24	28
Rb	37	136	57	86	75	97	90	75	28	71	102	80	99	110	95	66	76	79	53	66
Sr	109	52	300	429	515	1439	1394	312	357	310	553	556	352	554	620	189	412	568	562	281
Th	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	2	n.d.	n.d.	n.d.	18	5	28	27	3	13	9	7	7
U	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	1	1	2	1	4	2	1	1	n.d.
Y	<3	6	6	13	10	24	26	6	27	12	8	10	12	11	11	29	11	11	11	22
Zn	5	5	35	131	88	95	95	32	50	39	40	54	29	46	55	61	68	69	45	75
Zr	18	23	28	187	145	125	130	39	97	97	134	178	124	135	181	123	174	181	150	112
La	n.d.	2.5	3.9	41.1	25.3	56.6	n.d.	13.4	45.3	16.3	n.d.	25.0	22.1	34.6	45.5	8.0	38.1	31.0	28.1	21.0
Ce	n.d.	5.0	7.8	79.4	50.9	107.7	n.d.	25.5	92.4	32.3	n.d.	48.0	31.0	68.3	90.5	18.1	75.5	59.9	53.1	43.2
Pr	n.d.	0.5	1.0	9.7	6.4	13.6	n.d.	3.1	11.9	3.8	n.d.	n.d.	n.d.	8.7	11.6	n.d.	9.8	7.5	6.6	n.d.
Nd	n.d.	1.9	3.9	36.1	24.1	51.0	n.d.	11.6	43.6	14.1	n.d.	23.4	20.1	32.0	14.0	10.3	36.7	28.6	24.1	17.1
Sm	n.d.	0.6	1.1	6.5	4.8	8.3	n.d.	2.4	9.8	2.6	n.d.	6.0	5.2	6.6	8.2	5.0	7.1	5.2	4.6	n.d.
Eu	n.d.	0.1	0.9	1.5	1.3	1.9	n.d.	1.9	1.9	0.6	n.d.	n.d.	n.d.	1.1	1.3	n.d.	1.3	1.0	1.0	n.d.
Gd	n.d.	0.5	1.2	4.9	3.6	6.0	n.d.	1.8	8.5	2.0	n.d.	n.d.	n.d.	4.9	5.8	n.d.	5.2	3.5	3.1	n.d.
Tb	n.d.	0.1	0.2	0.6	0.5	0.8	n.d.	0.2	1.2	0.3	n.d.	n.d.	n.d.	0.6	0.7	n.d.	0.6	0.4	0.4	n.d.
Dy	n.d.	0.6	1.1	2.9	2.3	4.4	n.d.	1.1	6.4	1.6	n.d.	n.d.	n.d.	2.8	3.1	n.d.	2.8	2.2	2.1	n.d.
Ho	n.d.	0.1	0.2	0.5	0.4	0.8	n.d.	0.2	1.1	0.3	n.d.	n.d.	n.d.	0.4	0.5	n.d.	0.4	0.4	0.4	n.d.
Er	n.d.	0.3	0.5	1.1	0.9	2.3	n.d.	0.4	2.3	0.8	n.d.	n.d.	n.d.	0.8	0.9	n.d.	0.9	1.0	1.0	n.d.
Tm	n.d.	0.1	0.1	0.1	0.1	0.3	n.d.	0.1	0.3	0.1	n.d.	n.d.	n.d.	0.1	0.1	n.d.	0.1	0.1	0.1	n.d.
Yb	n.d.	0.5	0.3	0.7	0.7	2.1	n.d.	0.4	1.3	0.7	n.d.	n.d.	n.d.	0.6	0.5	n.d.	0.6	0.9	0.8	n.d.
Lu	n.d.	0.1	0.1	0.1	0.1	0.3	n.d.	0.1	0.2	0.1	n.d.	n.d.	n.d.	0.1	0.6	n.d.	0.1	0.1	0.1	n.d.

n.d. = not determined LOI = loss on ignition

Tab. 2b Major element (in wt%), trace element, and REE (in ppm) data for the investigated rocks of Fatra and Low Tatra.

	Fatra granitoids						Low Tatra granitoids		
sample	1026	1027	1028	1030	1031	1038	1051	1088	1090
SiO_2	73.8	65.6	67.1	70.4	69.0	76.0	70.4	71.3	68.4
TiO_2	0.1	0.8	0.7	0.4	0.6	0.1	0.4	0.4	0.6
Al_2O_3	15.0	16.8	16.3	15.6	16.0	14.1	14.4	14.6	17.1
Fe_2O_3	1.3	4.1	3.7	2.6	3.3	0.4	3.4	2.9	4.0
MgO	0.4	1.6	1.3	1.0	1.2	0.1	1.5	1.2	1.9
CaO	0.8	2.6	3.1	1.8	2.1	0.6	3.5	2.0	2.5
Na_2O	4.3	4.3	4.2	4.1	4.0	4.3	2.1	3.7	4.4
K_2O	3.3	2.7	2.4	2.9	3.1	3.4	2.9	3.6	2.9
P_2O_5	0.1	0.2	0.2	0.1	0.2	0.1	0.1	0.1	0.3
LOI	0.9	1.1	0.8	1.1	0.8	0.7	0.9	0.8	1.7
Ba	479	786	788	864	906	399	362	567	1498
Ga	16	23	20	21	21	17	18	16	21
Nb	8	9	9	8	8	5	10	9	11
Ni	4	4	2	3	4	1	15	13	11
Pb	27	17	18	28	28	17	19	22	15
Rb	102	100	70	83	98	104	116	114	83
Sr	137	604	570	436	547	151	154	163	548
Th	6	7	5	8	9	1	11	8	9
U	3	n.d.	n.d.	2	n.d.	n.d.	3	2	1
Y	8	7	7	13	13	5	41	29	9
Zn	42	85	76	62	62	11	42	38	69
Zr	58	214	192	162	191	50	108	100	178
La	14.0	28.4	18.1	28.2	28.3	7.1	24.3	14.0	37.0
Ce	21.1	55.3	36	59.0	59.2	n.d.	47.8	34.1	71.2
Pr	n.d.	5.2	n.d.	3.1	n.d.	n.d.	6.0	6.2	11.0
Nd	10.0	24.0	18.3	27.0	25.1	5.0	22.7	20.4	36.3
Sm	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	5.3	2.1	8.4
Eu	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.8	n.d.	n.d.
Gd	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	5.3	n.d.	n.d.
Tb	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.9	n.d.	n.d.
Dy	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	6.4	n.d.	n.d.
Ho	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	1.4	n.d.	n.d.
Er	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.4	n.d.	n.d.
Tm	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.6	n.d.	n.d.
Yb	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	3.6	n.d.	n.d.
Lu	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.5	n.d.	n.d.

n.d. = not determined
LOI = loss on ignition

the metamorphic xenoliths occur together, enclosed in rather homogenous granitoid rocks of the High Tatra. These granitoids have Sr contents above 300 ppm, and low $^{87}Rb/^{86}Sr$ ratios of 0.2 to 1.0. The $FeO_{tot}/(FeO_{tot}+MgO)$ ratios of all investigated granitoids lie between 0.69 and 0.76.

Similar to the orthogneisses, the granitoids of the Tatra Mountains also have a peraluminous character as reflected in alumina saturation values of 1.05 to 1.3, whereas the diorites have ASI values of 0.92 to 0.97 and are therefore meta-aluminous. The granitoids represent medium- to high-potassium magmatic rocks. According to CHAPPELL and WHITE (1974), the granitoids can be classified as S-type rocks: only the mafic dioritic xenoliths are clearly I-type derived (Fig. 2).

The chondrite normalised REE patterns (Fig. 3) of the granitoids exhibit only slight negative Eu anomalies and show uniform fractionation trends (La_n/Yb_n = 51–15).

Whole rock Pb–Pb analyses (Table 4) show a spread in $^{206}Pb/^{204}Pb$ of between 18.34 and 19.08 and in $^{207}Pb/^{204}Pb$ between 15.66 and 15.74. The granitoids occupy nearly the same position as the orthogneisses in figure 4; thus their crustal character is well constrained by the common Pb system.

Sm–Nd whole rock analyses resulted in $\varepsilon Nd_{(330)}$ values of –4 to –1, in accordance with preliminary data of KOHÚT et al. (1999). In combination with $\varepsilon Sr_{(330)}$ values ranging from 10 to 60 (Tab. 3, Fig. 5), they reflect a strong crustal influence for the granitoids. The Nd model ages for the granitoids range from 1250–1450 Ma (Fig. 6) and compare well with those of the porphyric orthogneisses.

The migmatites (UP 1052 and UP 1063) show nearly the same geochemical composition as the surrounding granites, reflecting a close relationship to the associated granites. Nevertheless, their isotopic composition is different. With $\varepsilon Sr_{(330)}$ values of 114 and 112, the migmatites show a stronger crustal influence than the surrounding granites. In contrast to the granites, the $\varepsilon Nd_{(330)}$ value of the migmatite UP 1052 is positive and its Nd model age is much younger (about 900 Ma), probably reflecting juvenile input. The study of STILLE and SCHALTEGGER (1996) postulated a lithospheric mantle, enriched in LREE and characterised by elevated Sm/Nd ratios up to 0.168 for Central Europe 900–700 Ma ago (so-called Central European Enriched Mantle). The Sm/Nd ratios of the Tatra migmatites range at 0.118 and are, thus, rather "normal" (i.e. typical for continental crust), but the high $\varepsilon Nd_{(330)}$ may indicate the involvement of such enriched lithospheric mantle material. Migmatisation during Variscan time might have caused mobilisation of parts of this old lithospheric mantle and of continental crust. This resulted in fractionation of the REE and finally shifted the Sm/Nd ratios to "normal" crustal values. Despite this possible explanation for the high Nd isotope ratios, it is not possible to rule out that the anatexis of the migmatites was too fast for a complete equilibration of the Sr and Nd isotopes. As noted by DAVIES and TOMMASINI (2000), equilibration of the Nd isotopes during anatexis of granitoid rocks may require several tens of millions of years. For the Tatra Mountains a rather

Tab. 3 Sm–Nd and Rb–Sr isotope data from the Western Carpathians (Western Tatra, High Tatra, Low Tatra, Fatra).

sample	Nd ppm	Sm ppm	$^{143}Nd/^{144}Nd$	± 2σ mean	$^{147}Sm/^{144}Nd$	εNd (330)	Sr ppm	Rb ppm	$^{87}Sr/^{86}Sr$	± 2σ mean	$^{87}Rb/^{86}Sr$	εSr (330)
W–Tatra Orthogneisses (porphyric)												
UP 1002	2.06	0.64	0.512178	33	0.12053	–5.7	236.0	14.8	0,715059	13	0.18049	132.1
UP 1014	28.92	5.40	0.512216	13	0.11286	–4.7	375.3	24.8	0,711120	41	0.19073	76.2
UP 1025	15.66	3.19	0.512337	26	0.12319	–2.8	484.6	53.4	0.709588	13	0.31982	45.2
W–Tatra Orthogneisses (leucocratic)												
UP 1005	1.80	0.46	0.512067	20	0.15448	–9.3	104.9	37.3	0.724154	12	1.02587	204.8
UP 1012	2.00	0.54	0.512184	21	0.16300	–7.4	50.9	117.6	0.763669	14	6.70700	387.2
High Tatra Diorites (Xenoliths)												
UP 1015	17.36	3.31	0.512491	20	0.11540	0.5	459.5	29.52	0.705037	79	0.18452	–9.7
UP 1016	10.40	2.27	0.512594	24	0.13213	1.8	345.9	32.35	0.705285	63	0.26798	–11.7
Western and High Tatra Granitoids / Migmatite												
UP 1023	9.88	2.00	0.512315	32	0.12203	–3.2	313.8	71.6	0.712230	20	0.65707	59.1
UP 1036	39.62	8.74	0.512369	23	0.13343	–2.6	364.1	24.1	0.708453	155	0.19131	38.3
UP 1040	13.00	2.51	0.512317	31	0.11694	–2.9	304.1	59.0	0.708852	23	0.55615	20.0
UP 1017	22.44	4.56	0.512393	44	0.12290	–1.7	556.3	100.3	0.708066	34	0.52306	10.7
UP 1018	35.66	6.74	0.512328	31	0.11431	–2.6	557.1	79.7	0.707641	26	0.41519	11.9
UP 1044	7.20	1.65	0.512366	37	0.13848	–2.8	123.3	27.7	0.711144	27	0.64280	46.4
UP 1049	13.08	2.93	0.512420	67	0.13548	–1.7	218.3	34.4	0.708517	65	0.45302	21.8
UP 1050	14.76	3.13	0.512312	50	0.12804	–3.5	284.8	40.8	0.707800	43	0.41405	14.2
UP 1052	32.81	6.44	0.512562	34	0.11889	1.8	230.2	82.9	0.717762	29	1.03287	114.3
UP 1053	33.60	6.45	0.512331	42	0.11612	–2.6	436.1	59.6	0.710662	36	0.39557	56.1
UP 1055	23.98	4.19	0.512376	23	0.10569	–1.3	491.2	46.7	0.707725	30	0.27256	22.6
UP 1057	19.96	3.77	0.512303	26	0.11407	–3.1	371.5	24.2	0.707698	20	0.18707	27.9
Fatra Granitoids												
UP 1026	11.43	2.51	0.512322	33	0.13268	–3.5	130.4	35.9	0.719053	57	2.05193	64.7
UP 1027	23.76	4.20	0.512301	53	0.10692	–2.8	506.3	10.6	0.708409	22	0.15532	40.1
UP 1028	21.36	3.96	0.512361	78	0.11202	–1.8	704.1	33.1	0.707838	37	0.13492	33.3
UP 1030	23.78	4.53	0.512333	36	0.11522	–2.5	401.8	29.3	0.709339	20	0.54365	27.4
UP 1031	26.51	4.93	0.512292	45	0.11245	–3.2	477.2	61.6	0.708925	49	0.37189	33.0
UP 1038	6.75	1.34	0.512301	44	0.11961	–3.3	132.6	99.5	0.716430	28	2.15152	20.9
Low Tatra Granitoids												
UP 1051	21.69	4.94	0.512144	55	0.13759	–7.1	194.2	135.1	0.723573	67	2.01800	131.1
UP 1088	16.90	2.99	0.512115	30	0.10715	–6.4	147.1	92.5	0.723157	19	1.80444	139.5
UP 1090	31.88	7.04	0.512379	22	0.13347	–2.4	447.4	36.4	0.707550	23	0.23283	43.3

2σ mean errors refer to mean of the blocks.

rapid anatexis is proposed on the basis of U–Pb zircon dating. The migmatisation took place 335 Ma ago, and the last collisional granites (not affected by the migmatisation) intruded only 20 Ma later (POLLER and TODT, 2001). Therefore, a complete equilibration of Sr and Nd isotopes in the migmatites in such a short time span is at least questionable.

Tab. 4 Common Pb data for the Western Carpathian rocks.

sample	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
W-Tatra Orthogneisses (porphyric)			
1002-1	18.2554	15.6482	38.4473
1002-2	18.9062	15.6837	38.4718
1014-1	18.7979	15.7202	39.1558
1014-2	18.8969	15.6859	39.9310
1025-2	18.7486	15.7470	38.8883
W-Tatra Orthogneisses (leucocratic)			
1005-1	18.0958	15.6363	38.2198
1005-2	18.1410	15.6647	38.2970
1012-1	17.9995	15.6193	37.9985
1012-2	18.1031	15.6065	37.9630
Western and High Tatra Granitoids			
1023-1	18.5946	15.6997	38.5827
1023-2	18.5675	15.6741	38.4885
1036-1	18.7033	15.6993	39.2420
1040-1	18.5368	15.6703	38.4764
1044-1	18.4125	15.6990	38.5024
1049-1	18.3485	15.6678	37.3983
1050-1	18.8499	15.6871	39.1130
1053-1	18.6178	15.6644	38.8183
1054-1	18.7587	15.7262	38.5105
1054-2	19.0834	15.7433	39.0427
1055-1	18.7476	15.7113	38.6565
1055-2	18.5029	15.6768	38.3930
1057-1	18.6883	15.7062	39.0218
1057-2	18.6503	15.7158	37.7916
1063-1	18.7123	15.7039	38.7899
Low Tatra Granitoids			
1088-1	18.8072	15.6829	38.5531
1090-1	18.6096	15.6754	38.5588
1092-1	18.4679	15.6676	38.5429
1094-1	18.7333	15.6979	39.4536
1095-1	18.6890	15.6727	38.4324
Tatra Galenas			
1089	18.4892	15.6738	38.4759
1092	18.4679	15.6676	38.5429
Tatra Feldspars			
1044-F	18.4125	15.6990	38.5024
1056-F	18.4925	15.6591	38.3344

All data are corrected for fractionation.
The 2σ external reproducibility is about 300 ppm.

Discussion

Two main questions can be addressed with the data presented: (1) What source material is incorporated in the Tatra rocks? (2) Is there a close isotopic relationship between the Western Carpathian basement and the Variscan orogen? Finally, the geotectonic environment and evolution of the Tatra can be constrained.

SOURCE ROCK OF THE TATRA GRANITOIDS

According to the granite classification of BARBARIN (1999) the granitoids have most characteristics of the MPGs; only the An content of the plagioclases (20–45%) seems to be too high for typical crustal derived granodiorites and syenogranites. This is possibly due to less differentiation and fractional crystallisation in comparison to the orthogneisses.

However, granites with intermediate Sr and Nd isotopic composition and crustal Pb–Pb isotopic ratios usually are envisaged to be the result of mixing with more primitive material through direct input or by a heat source from the underplating basaltic magma (e.g. PIN and DUTHOU, 1990; COCHERIE et al., 1994). We interpret this feature as being inherited from repeatedly melted and recycled crust. Additionally, the granitoids are assumed to originate not only from old continental

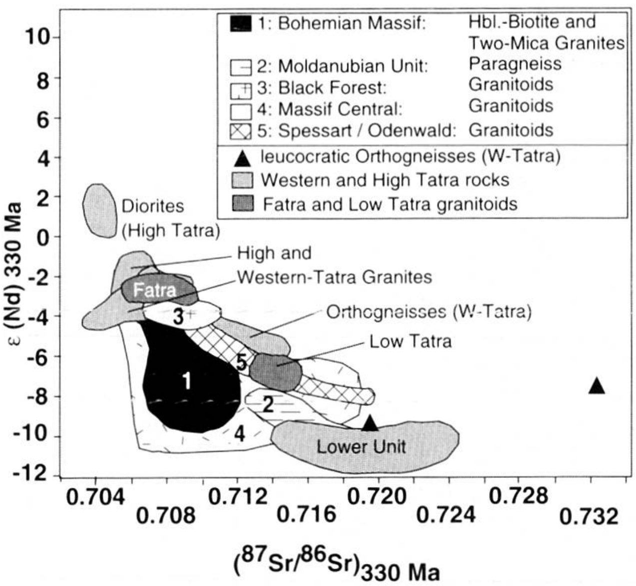

Fig. 5 $\varepsilon Nd_{(330)}$ versus $^{87}Sr/^{86}Sr_{(330)}$ data of the different Western Carpathian areas (Tatra and Fatra) are compared to also recalculated (to 330Ma) data from Bohemian Massif, Moldanubian unit (JANOUSEK et al., 1995), Black Forest, Spessart, Odenwald (LIEW and HOFMANN, 1988), and Massif Central (BEN OTHMAN et al., 1984; PIN and DUTHOU, 1990).

sediments but also may have incorporated mafic crustal material such as reworked amphibolites.

Even a minor contribution from a primitive source (such as ocean-derived amphibolites or even small amounts of lithospheric mantle material) is possible, mainly because of the presence of infracrustal mixed dioritic magma, which reached the anatectic zone during the collisional processes and caused the high isotopic ratios (the most negative $\varepsilon Nd_{(330Ma)}$ is –3.7) of the Tatra Mountain granitoids.

The origin of the Tatra rocks can be characterised in more detail by the radiogenic isotopes. As documented by the Pb–Pb data, the orthogneisses of the Tatra Mountains were generated mainly from upper crustal material. The Nd and Sr isotopic results indicate that a more mafic component was also involved, e.g. lithospheric mantle or recycled oceanic crust.

In the $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ diagram (Fig. 4b) of DOE and ZARTMAN (1979), the present Pb–Pb data from the Western Carpathians overlap both with the upper and lower crustal field. Nevertheless, a small amount of mafic, but mainly crustal, material should be responsible for the present Pb composition of Tatra rocks. This is supported by the recalculated values and fields shown in figure 4a. The Pb data were calculated back in time to 330 Ma, using several μ-values. This is a key age for the Western Carpathians (this is the time of emplacement of the dioritic xenoliths and final remelting of the migmatites). In figure 4a the positions of present-day isotopic compositions (same as in Figs 4b and 4c) are shown, together with those of 330 Ma for $\mu = 9$. The result is clear in that, the position of the Western Carpathians relative to the reference fields of upper crust, mid ocean ridge basalts (MORB) or enriched mantle II (EM II = recycled oceanic crust) does not change significantly. Consequently, the present-day Pb isotope data as shown in figures 4b and 4c reflects the position of the Western Carpathian rocks.

The leucocratic orthogneisses show significant input from reworked crust, which seems to be reasonable considering the crustal residence ages (~1750 Ma) of these orthogneisses (Fig. 6). Because of the high Sm/Nd ratios (> 0.155), these model ages were calculated using a two-step evolution. The leucocratic orthogneisses seem to be the most evolved rocks analysed in the Western Tatra. Obviously, their Sm/Nd ratios were changed by fractional crystallisation during the later metamorphic overprinting and the following granite intrusion.

The upper intercept U–Pb zircon ages of orthogneisses and granites, because of inherited cores, also reflect the influence of old crust. Consequently, the generation of the Tatra granitoids and gneisses seems to be essentially crustal, probably involving continentally collision and crustal thickening.

RELATIONSHIPS BETWEEN THE WESTERN CARPATHIANS AND THE VARISCIDES

The samples from the Tatra Mountains and from the neighbouring granitoid plutons of Velká Fatra (KOHÚT, 1992) and the Lower Tatra Mountains show close similarities in their isotopic composition. The Velká Fatra samples show $\varepsilon Sr_{(330)}$ values

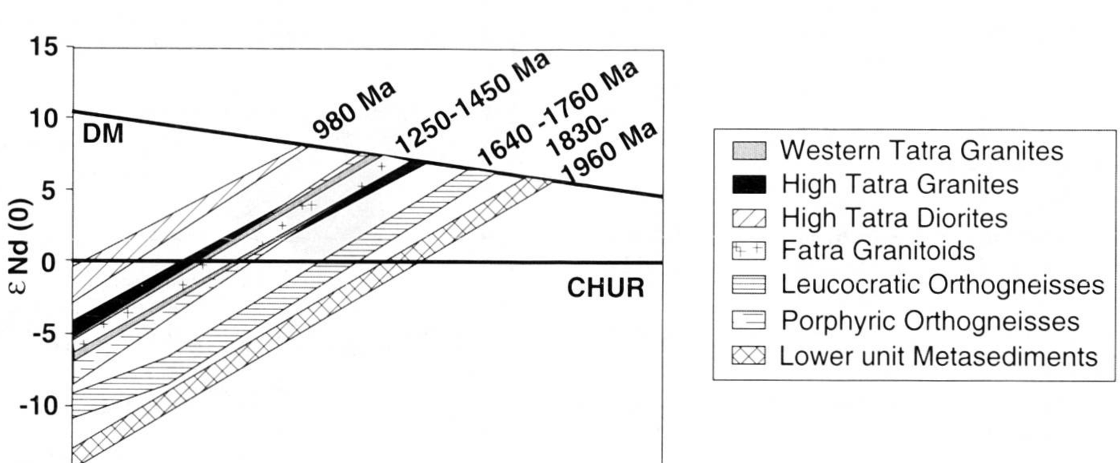

Fig. 6 Nd model ages calculated as one-step evolution (only for leucocratic orthogneisses: two step evolution after LIEW and HOFMANN, 1988). The Tatra rocks show 3 different groups of mean crustal residence ages.

between 21 and 65, depending on the amount of juvenile components. The corresponding $\varepsilon Nd_{(330)}$ values of the Fatra rocks range from –2 to –4. Granodiorites and syenogranites from the Lower Tatra Mountains show a general overlap with the Tatra Mountains with regard to Sr and Nd, as well as to Pb–Pb isotopic composition.

Comparisons with the Bohemian Massif, the Moldanubian units (JANOUSEK et al., 1995) and some regions from the western Variscides such as Massif Central, Black Forest, Spessart and Odenwald (LIEW and HOFMANN, 1988), indicate a close isotopic relationship to the Western Carpathians (Fig. 5; all Sr and Nd literature data also were re-calculated to 330 Ma).

The metasediments from the Lower Unit of the Western Tatra Mountains (GURK, 1999) show good isotopic correspondence with Moldanubian paragneisses. The isotopic character of the granites and orthogneisses of the Tatra Mountains and the Velká Fatra is consistent with several Variscan granitoids of France and Germany.

Only the diorites of the Tatra Mountains with high εNd (330) values (above zero) have no

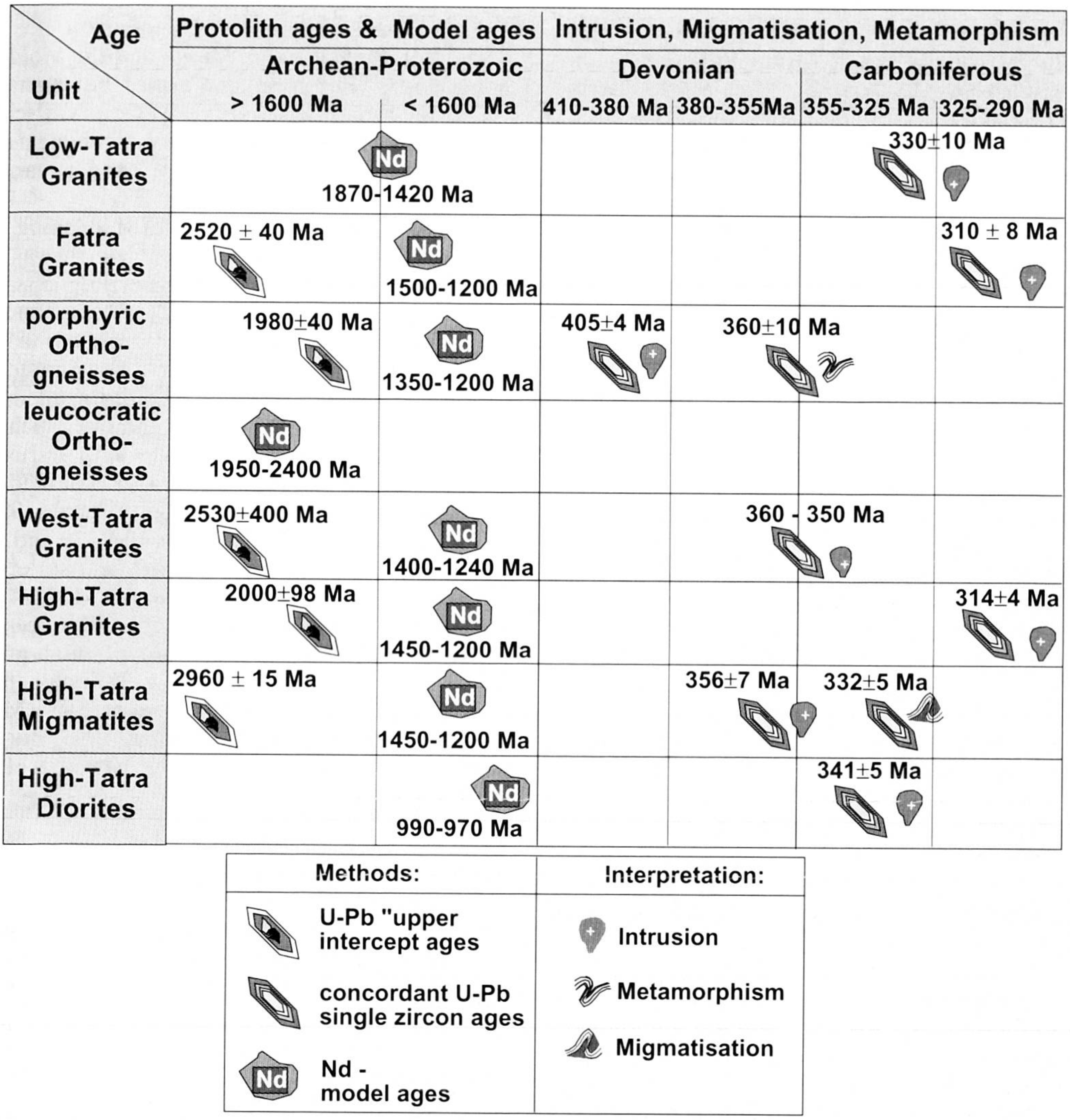

Fig. 7 Summary diagram for several rock series of the Western Carpathians including Nd model ages, U–Pb upper intercept ages indicating the protolith ages, and several concordant U–Pb ages defining intrusion, metamorphism, and migmatisation.

equivalent rock series with the same emplacement age. However, similar εNd values between +2 and –8 are reported by FINGER et al. (1997) for post-orogenic granitoids from the Saxothuringian (Reichenberger granite), the Moldanubian (Freistadt granite) or even from the Alps (Düssi diorite and Venediger metatonalite). Also, geochemically similar rocks have been reported from the Limousin, the Massif Central (SHAW et al., 1993) and Corsica (COCHERIE et al., 1994).

Nd model ages for most of the investigated Western Carpathian granitoid rocks are between 1250 and 1450 Ma. Together with an average $^{147}Sm/^{144}Nd$ ratio of 0.12 they represent a crustal composition that occurs elsewhere in the European Variscides. Several Variscan regions, such as the Pyrenees (BEN OTHMAN et al., 1984) or the Mid German Crystalline Rise (REISCHMANN and ANTHES, 1996), show the same features. The Pb–Pb isotopes also confirm the isotopic correspondence between the Western Carpathian rocks and granitoids from the Black Forest, the Massif Central, and the Mid German Crystalline Rise (Fig. 4b). An overview of all isotopic data including the U–Pb zircons ages of the Tatra Mountains is given in figure 7.

GEODYNAMIC IMPLICATIONS FOR THE WESTERN CARPATHIANS

Based on the presented isotopic and geochemical data and U–Pb zircon ages mentioned here, we propose the following model for the geotectonic evolution of the Tatra Mountains.

The intrusion of the orthogneiss precursor 405 Ma ago probably was triggered by the subduction of oceanic crust under a continental plate at the northern border of Gondwana (POLLER et al., 2000). The orthogneisses in the Tatra Mountains are closely associated with partially melted banded amphibolites containing eclogite relics as well as kyanite-bearing migmatites, which were affected by medium- to high-pressure (HP) and high-temperature (HT) metamorphism reaching ~12 kbar and 750 °C (JANÁK et al., 1996; JANÁK et al., 1999). Therefore, the overprinting of the orthogneisses seems to have occurred under medium- to high-pressure and HT conditions in deep crustal levels. The absence of significant amounts of juvenile material in the orthogneisses suggests that the medium- to high-pressure and HT conditions were the result of crustal thickening related to continental collision. It is assumed that two microplates at the northern border of Gondwana were involved in this process (POLLER et al., 2000). Such a scenario accounts for the observed petrographic, geochemical, and isotopical results. Similar ages as for the intrusion of the orthogneiss precursor (405 ± 4 Ma) are reported from rocks of the Mid German Crystalline Rise which are generally considered to be related to subduction zones (ALTENBERGER et al., 1990; ALTENBERGER and BESCH, 1993; REISCHMANN and ANTHES, 1996).

High pressure metamorphism analogous to that of the Tatra Mountains between 360 Ma and 350 Ma is known e.g. from the Erzgebirge Crystalline Complex. SCHMÄDICKE et al. (1995) reported Sm–Nd ages between 360±7 Ma and 333±6 Ma for eclogitic rocks of the Erzgebirge. Synmetamorphic granite intrusion (SCHULMANN et al., 1991), followed by exhumation and extension, are reported for the Bohemian Massif and the Moldanubian zone. Therefore, similarities between the Western Carpathians and the Western Variscides are documented not only by the geochemical and isotopic composition, but in ages and metamorphic evolution as well.

Immediately after the crustal thickening, resulting from microplate collision which was responsible for the overprint of the orthogneiss, the Western Tatra granites intruded (350 Ma, POLLER et al., 2000). The geotectonic evolution of the Tatra Mountains then underwent extension for a short time. Finally, the last granitoids intruded 315 Ma ago in the High Tatra Mountains. Such an Early Carboniferous magmatism with intrusion of large amounts of granitoids is documented for the Bohemian Massif as well (FRIEDL et al., 1994).

Final cooling took place between 330 and 300 Ma as recorded by Ar–Ar data (MALUSKI et al., 1993; JANÁK and ONSTOTT, 1993; JANÁK 1994). A similar evolution with collision-related processes, involving tectonic exhumation of high-grade units, has been described in well-documented parts of the Variscan belt (BURG et al., 1984; MATTE, 1986; O'BRIEN and CARSWELL, 1993; BROWN and DALLMEYER, 1996; ESCUDER VIRUETE et al., 1997; VON RAUMER, 1998).

Acknowledgements

We are very grateful to G. Feyerherd and I. Bambach for the preparation of the drawings and I. Raczek for her help with Nd and Sr analyses as well as constructive discussions and remarks. U.P. thanks A. Hofmann for the possibility to work in his institute, and for useful comments and careful reading of an early version of this manuscript. We thank C. Goodrich for the correction of the style. Discussions with F. Finger, A. Kröner and P. Uher contributed to the final version. Comments of H.J. Förster improved a previous version of the manuscript. Reviews of M. Thöni and M. Engi are acknowledged. This work was supported by the Max-Planck-Gesellschaft and the DFG (PO608/1-1 and PO608/1-3).

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Manuscript received March 13, 2001; revision accepted June 6, 2001.

Appendix 1

SAMPLE DESCRIPTIONS

The investigated *orthogneisses* are represented by five samples. Samples UP 1002 (Ziarska valley), UP 1014 (Jamnická valley), and UP1025 (Baranec) are coarse grained, with porphyric, augen-like K-feldspar or plagioclase of 2–3 cm size, exhibiting a mylonitic S–C fabric (LISTER and SNOKE, 1988). Feldspars are elongated and show dynamic recrystallisation with characteristic tails. K-feldspar show typical microcline twinning. Plagioclase compositions range from albite to andesine, locally, myrmekite developed. Micas form characteristic "mica-fish" porphyroblasts. Muscovite is slightly phengitic, biotite is Fe-rich and usually replaced by chlorite. Quartz shows undulatory extinction with elongated and recrystallised grains, forming aggregates and ribbons. Minor garnet corresponds to almandine-spessartine, with up to 15% of pyrope component. However, garnet is strongly retrogressed and partly replaced by chlorite.

Samples UP 1005 (Ziarska valley) and UP 1012 (Jamnická valley) are leucocratic, fine-grained rocks, composed of plagioclase (dominantly albite), K-feldspar (microcline), and quartz. Muscovite is more abundant than biotite. Minor garnet was mostly replaced by chlorite. All investigated orthogneisses were affected by a solid-state deformation at ductile to brittle conditions (PATTERSON et al., 1989) and experienced strong regression.

The *granites* and *migmatites* are represented by three granitic samples from the Western Tatra and nine from the High Tatra. Additionally, two dioritic xenoliths of the High Tatra Mountains and two migmatites were sampled. The granites of the Tatra generally are divided into two groups: the so-called High Tatra granitoids and the so-called common Tatra granitoids, occuring mostly in the Western Tatra (KOHÚT and JANÁK, 1994).

The "High Tatra granites" are granodioritic to tonalitic in composition. They contain plagioclase, quartz, rare K-feldspar, biotite, and rarely muscovite. In the High Tatra near Gerlachovsky peak, contacts between the High Tatra granitoids and some tonalitic to dioritic enclaves as well as contacts to garnet-sillimanite-bearing metapelites were observed. These metapelites may be deformed and show transition to migmatitic rocks (samples UP 1052; UP 1063). These migmatites were studied recently under petrological aspects and the initial migmatisation is suggested to have taken place above 730 °C and 11–12 kbar (JANÁK et al., 1999).

The "common Tatra granitoids" are more intermediate in composition and can be characterised as monzogranites. They surface not only in the High Tatra, but also in the Western Tatra. They are composed of plagioclase, K-feldspar, biotite, and muscovite. Only sample UP 1036 (Bystrá) is free of muscovite and plagioclase is more abundant than K-feldspar.

The fabric of all analysed granites is homogeneous. The texture is holocrystalline, euhedral to subhedral feldspars are mostly randomly oriented and partly replaced by sericite. Micas are locally deformed into kink-bands, and quartz exhibits undulatory extinction. In general, the granitoids show only brittle and no penetrative deformation.

Appendix 2

ANALYTICAL TECHNIQUES

For the geochemical and geochronological investigation 10 to 20 kg of fresh material were collected in the field. A list with the exact sample locations is given in table 1. The samples were crushed, and approximately 300 g of the resulting material were pulverised using an agate mill. For mineral separation, the crushed material was ground using a rotary mill to a grain size < 500 µm.

Zircons were separated using a Wilfley table. The heaviest fraction was processed with heavy liquids and a Frantz magnetic separator to obtain the zircon fraction. Suitable grains for U–Pb measurements and cathodoluminescence (CL) documentation were selected by hand picking.

For Pb–Pb whole rock analyses, fresh splits of the rocks were taken and dissolved in Savilex beakers during 3 days on the hotplate with HF and HNO_3. For Sm–Nd and Rb–Sr analyses as well as for XRF and ICP-MS measurements, the fine grained powder (<50 µm) from the agate mill was used.

Isotopic measurements were carried out at the Max-Planck-Institut für Chemie in Mainz using a Finnigan MAT 261 mass spectrometer.

Pb, Rb–Sr and Sm–Nd were measured with multiple collectors operating in static mode. Errors on isotopic ratios are given as 2σ errors of the block mean (10–25 blocks and 10 Scans each block). Blanks for Nd, Sm, Sr, and Rb analyses were below 100 pg and thus not significant. All Nd and Sr measurements were done as IC and ID runs. The Rb–Sr chemistry was performed by standard ion exchange methods. For Sm–Nd chemistry HDHP columns were used following the chemical procedures described in WHITE and PATCHETT (1984). For the Rb–Sr measurements a mixed ^{84}Sr – ^{85}Rb spike was used. The Sm–Nd spike was enriched in ^{150}Nd and ^{149}Sm. For both spikes the concentrations are known up to 0.1 % precision. Therefore the ^{87}Rb/^{86}Sr and ^{147}Sm/^{144}Nd ratios are limited to 0.1% precision also. The εNd and εSr values are calculated for 330 Ma in table 3; figure 5 uses also recalculated (330 Ma) values.

The Pb–Pb procedure was done by HBr-chemistry on microcolumns (see ARNDT and TODT, 1994).

For Pb measurements, two NBS 982 or NBS 981 were loaded on each turret and measured at the beginning and at the end of the sample analyses. The fractionation was 1.45 ± 0.3‰ per amu. The blanks were less than 50 pg Pb total and therefore insignificant.

Geochemical analyses were performed by XRF at the Institut für Mineralogie, Johannes-Gutenberg Universität Mainz. The measurements were controlled with respect to international standards.

The REE were analysed in the Department of Earth Sciences, the Memorial University of Newfoundland, St. Johns (Canada). Both, the major and trace element as well as the REE data are given in table 2.

The Pb data refer to present-day (Tab. 4). Recalculations of initial values for reasonable ages (e.g. 330 Ma, intrusion of diorites; migmatisation of migmatites) gave no essential changes (see Fig. 4a). These calculations were done with several μ-values (ranging from 7 to 15) which resulted in different changes of the distance between the present day and the 330 Ma position. In all situations, the relative position of the Western Carpathian field and the reference fields of upper crust, MORB and enriched mantle II (after DOE and ZARTMAN, 1979) did not change significantly. Therefore, all values and fields in the ^{207}Pb/^{204}Pb vs. ^{206}Pb/^{204}Pb plots of figures 4b and 4c refer to present-day.

Additionally, the Pb whole rock analyses were controlled by feldspar and galena Pb analyses (crosses in Figs 4b and 4c). These mineral analyses show good correspondence with the whole rock data of the Western Carpathians.