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Deciphering the petrogenesis of deeply buried granites: whole-rock geochemical constraints on the origin of largely undepleted felsic granulites from the Moldanubian Zone of the Bohemian Massif

Vojtěch Janoušek, Fritz Finger, Malcolm Roberts, Jiří Frýda, Christian Pin and David Dolejš

ABSTRACT: The prominent felsic granulites in the southern part of the Bohemian Massif (Gföhl Unit, Moldanubian Zone), with the Variscan (~340 Ma) high-pressure and high-temperature assemblage garnet + quartz + hypersolvus feldspar ± kyanite, correspond geochemically to slightly peraluminous, fractionated granitic rocks. Compared to the average upper crust and most granites, the U, Th and Cs concentrations are strongly depleted, probably because of the fluid and/or slight melt loss during the high-grade metamorphism (900–1050 °C, 1·5–2·0 GPa). However, the rest of the trace-element contents and variation trends, such as decreasing Sr, Ba, Eu, LREE and Zr with increasing SiO_2 and Rb, can be explained by fractional crystallisation of a granitic magma. Low Zr and LREE contents yield ~750 °C zircon and monazite saturation temperatures and suggest relatively low-temperature crystallisation. The granulites contain radiogenic Sr ($^{87}Sr/^{86}Sr_{340}$ = 0·7106–0·7706) and unradiogenic Nd (ε^{340}_{Nd} = −4·2 to −7·5), indicating derivation from an old crustal source. The whole-rock Rb–Sr isotopic system preserves the memory of an earlier, probably Ordovician, isotopic equilibrium.

Contrary to previous studies, the bulk of felsic Moldanubian granulites do not appear to represent separated, syn-metamorphic Variscan HP–HT melts. Instead, they are interpreted as metamorphosed (partly anatectic) equivalents of older, probably high-level granites subducted to continental roots during the Variscan collision. Protolith formation may have occurred within an Early Palaeozoic rift setting, which is documented throughout the Variscan Zone in Europe.

KEY WORDS: Austria, Czech Republic, high-grade metamorphism, Nd isotopes, partial melting, Sr isotopes, Variscan orogeny.

Felsic garnet ± kyanite-bearing high-pressure and high-temperature (HP–HT) granulites with broadly granitic compositions are prominent constituents in the central part of the European Variscides (e.g. Pin & Vielzeuf 1983; O'Brien & Rötzler 2003). These light-coloured, SiO_2–rich rocks are known from many places of the Bohemian Massif, namely from the Moldanubian (Lower Austria and southern Czech Republic; Fig. 1) and the Saxothuringian zones (the Saxonian Granulite Massif and Erzgebirge; Romer & Rötzler 2001; Rötzler & Romer 2001; O'Brien & Rötzler 2003). Since the Moldanubian granulites may provide insights into processes related to deep subduction and subsequent exhumation of the continental crust, research over the past 40 years has focused mainly on constraining their precise pressure–temperature–time (P–T–t) paths (for reviews, see O'Brien 2000; O'Brien & Rötzler 2003). However, their pre-metamorphic history has been largely neglected.

Concerning the protolith, the felsic granulites were considered as former rhyolites or granites which acquired their high-grade metamorphic character during the course of the Variscan collision (Fiala *et al.* 1987a, b; Vellmer 1992). Based on existing P–T estimates and fluid-absent melting models for crustal rocks, Roberts & Finger (1997) proposed that they probably experienced a certain degree of partial melting and melt persisted during much of their metamorphic history. Other authors (Vrána & Jakeš 1982; Vrána 1989; Jakeš 1997; Kotková & Harley 1999) went further and suggested that the granulites actually represent high-pressure granitic magmas which formed during the Variscan metamorphic cycle.

Clearly, the problem of granulite genesis cannot be resolved without involving additional aspects of granite petrology and geochemistry, constraining the protolith formation and metamorphic reworking. The present paper characterises the compositional variation of the felsic Moldanubian granulites and provides their petrogenetic interpretation. The authors demonstrate that most, if not all, of their whole-rock geochemical signature may be related to a pre-existing Early Palaeozoic igneous-rock suite, whereas the Variscan high-pressure metamorphism could have been nearly isochemical.

1. Geological setting and petrology

1.1. Moldanubian Zone

The Moldanubian Zone represents remnants of the crystalline core of the Variscan orogen in central Europe (e.g. Dallmeyer *et al.* 1995). In the southern part of the Bohemian Massif, shared by the Czech Republic and Austria, the Moldanubian Zone consists of several crustal segments with contrasting ages and complex polyphase deformational histories, intruded by numerous, mainly post-tectonic granitic masses (Dallmeyer *et al.* 1995; O'Brien 2000 and references therein). It has been

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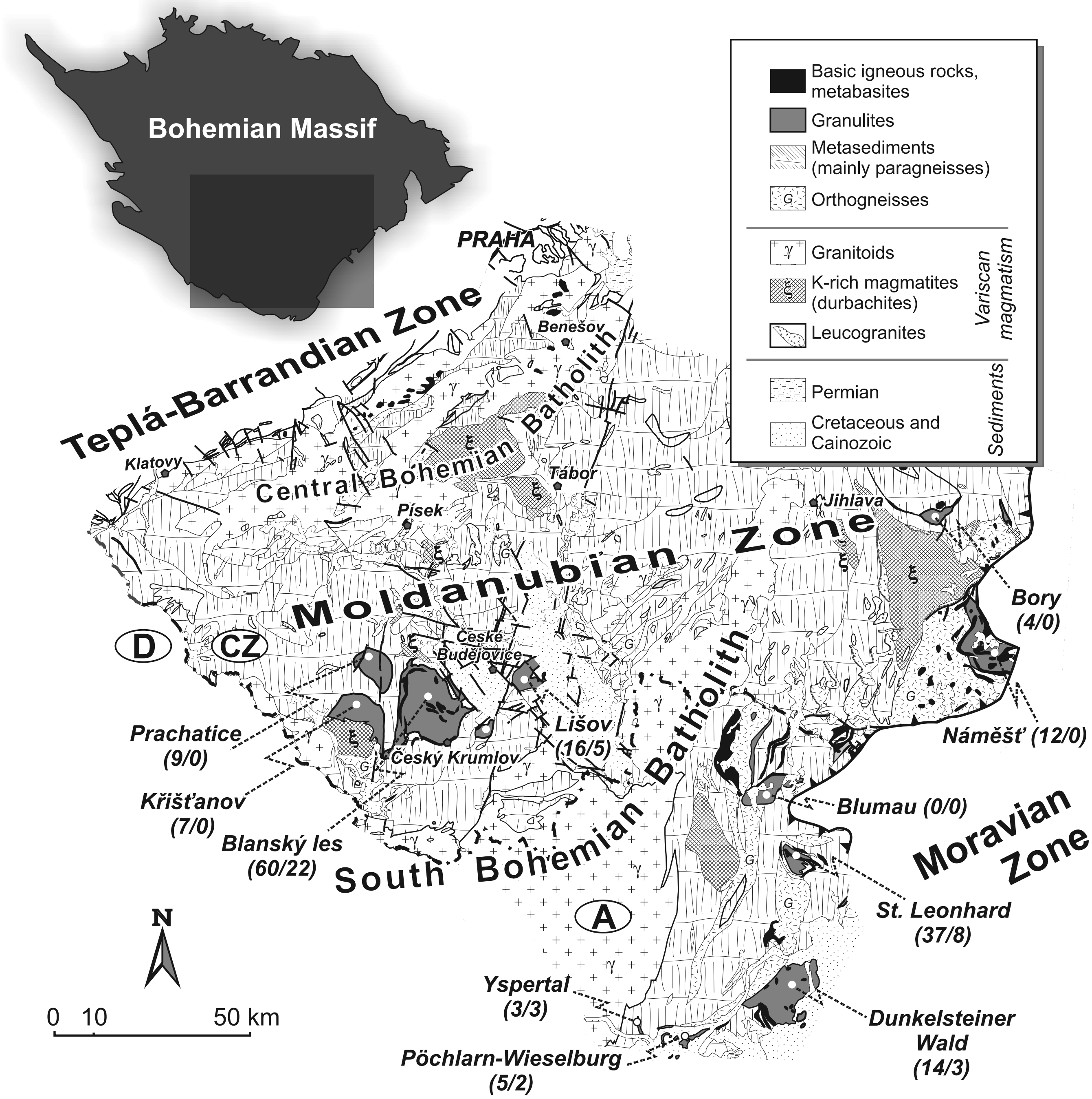

Figure 1 Geological sketch of the Moldanubian Zone in southern Bohemia, south-western Moravia and Lower Austria (simplified from the Czech Geological Survey map 1:500 000). The numbers of felsic granulite samples from larger massifs (total/new determinations) are shown.

subdivided into a structurally lower, mainly metasedimentary, amphibolite-facies Drosendorf Assemblage and an overlying, higher-grade, allochtonous Gföhl Assemblage (e.g. Matte *et al.* 1990; Fiala *et al.* 1995; Franke 2000).

The lower part of the Drosendorf Assemblage consists of the Monotonous Unit (Ostrong Unit) made up mainly of cordierite–biotite–sillimanite paragneisses, which are partly migmatitic and contain intercalations of orthogneisses and minor amphibolites. The upper part, termed the Varied (or Variegated) Unit, consists of paragneisses, orthogneisses, amphibolites, carbonate rocks, calc-silicate gneisses, quartzites and graphite schists.

The lower parts of the Gföhl Assemblage include mainly anatectic orthogneisses (Gföhl gneiss). The higher are dominated by granulites, associated with minor bodies of garnet and spinel peridotites, pyroxenites, dunites and eclogites (Fuchs & Matura 1976; Fiala *et al.* 1995).

1.2. Granulite bodies

Several granulite massifs occur in the Waldviertel region of Lower Austria. These are interpreted as the isolated klippen of an original regionally extensive nappe (Tollmann 1995). From SW to NE, these are the Ysperta, Pöchlarn–Wieselburg, Dunkelsteiner Wald, St. Leonhard and Blumau massifs (Fuchs & Matura 1976, Fig. 1). In the southern Czech Republic, the granulite bodies tend to be roughly oval-shaped and dome-like. The largest is the Blanský les Massif (24 × 4 km), followed by Křišťanov, Prachatice, Lišov, Krasejovka and other numerous small occurrences (Fig. 1). In southwest Moravia,

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the most prominent are the Náměšt' and Bory bodies (Fiala *et al.* 1987a, b).

1.3. Petrology

The individual massifs consist of a heterogeneous blend of high-pressure crustal and upper mantle rocks assembled and exhumed in course of the Variscan orogeny. Felsic granulites with granitic compositions are by far the most abundant and occur in all the main massifs, accounting for more than 75–80 vol.% (Fiala *et al.* 1987a; Vellmer 1992). In comparison, mafic–intermediate pyroxene-bearing and sedimentary-derived types are rather rare.

The typical felsic granulites are characterised by the high-pressure assemblage of garnet, kyanite, quartz and hypersolvus feldspar. Rutile, zircon, apatite, ilmenite ± monazite are the common accessories (Fiala *et al.* 1987a; Carswell 1991; Vellmer 1992; O'Brien & Rötzler 2003). The estimated peak metamorphic conditions are 900–1050°C and 1·5–2·0 GPa (O'Brien & Rötzler 2003 and references therein).

Following the high-pressure metamorphic peak, the rocks were exhumed along a steep decompression path to mid-crustal levels and subjected to a medium-pressure overprint (0·8–1·2 GPa; 800–900°C: Cooke 2000). Further evolution was characterised by near isobaric cooling (0·8–0·5 GPa), variable retrogression of the original assemblages and intense deformation, reflecting the effects of stacking and subsequent exhumation of the Moldanubian nappe pile (Carswell & O'Brien 1993; Cooke 2000). These processes resulted in the breakdown of some garnet to a secondary biotite, kyanite to sillimanite or garnet with kyanite to hercynite–plagioclase coronas (Owen & Dostal 1996; O'Brien & Rötzler 2003). Most of the granulites were mylonitised and recrystallised, with pronounced foliations of platy quartz and fine K-feldspar–plagioclase mosaics. Regional differences in the metamorphic style do exist; in southern Bohemia (e.g. Prachatice and Blanský les massifs), the effects of a low-P, high-T overprint that led to the crystallisation of cordierite and the formation of partial melt patches can be observed (O'Brien 2000).

1.4. Geochronology

1.4.1. Carboniferous granulite-facies metamorphism, subsequent cooling and uplift. The numerous U–Pb zircon and monazite ages clustering around 340 Ma are interpreted by most authors as reflecting the granulite-facies metamorphic climax (van Breemen *et al.* 1982; Schenk & Todt 1983; Kröner *et al.* 1988; Aftalion *et al.* 1989; Wendt *et al.* 1994; Kröner *et al.* 2000; Friedl *et al.* 2003; Kotková *et al.* 2003). In contrast, Roberts & Finger (1997) proposed that much of the zircon could have crystallised only on a retrogression path. Svojtka *et al.* (2002) concluded, based on Sm–Nd ages for garnet cores with prograde zoning, that the HP–HT metamorphic peak may have been even c. 10–15 Ma older than the retrograde event.

However, it is worth noting that many of the dated zircons were enclosed in high-pressure phases (garnet and former hypersolvus feldspar; Aftalion *et al.* 1989; Kröner *et al.* 2000). The metamorphic peak was probably short-lived, as demonstrated by diffusion modelling of garnet zoning (Medaris *et al.* 1990, 2003). The exhumation rate was high since the granulite pebbles already occur in Upper Visean conglomerates (several mm year^{-1}: Vrána & Novák 2000; Kotková *et al.* 2003). Rapid uplift and cooling is compatible with the fact that zircons enclosed in late cordierite-bearing, decompression melt pockets also gave ages of c. 340 Ma (Kröner *et al.* 2000). Taken together, the age of the high-grade metamorphism is not likely to differ substantially from 340 Ma.

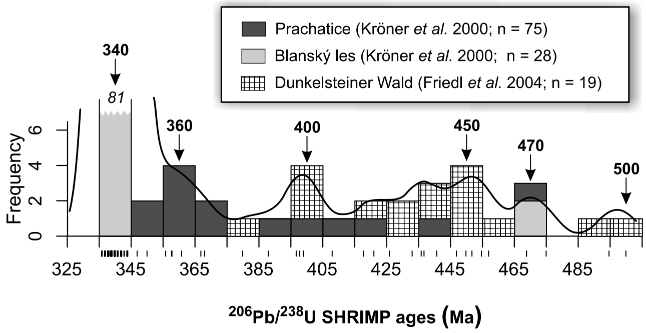

Figure 2 Histogram of frequencies for the $^{206}Pb/^{238}U$ SHRIMP zircon ages from the Moldanubian granulites. Data taken from Kröner *et al.* (2000; five samples from Prachatice and two from Blanský les massifs) and Friedl *et al.* (2004; sample *St*, Dunkelsteiner Wald).

1.4.2. Pre-Carboniferous ages. Devonian ages (c. 360–370 Ma) were obtained from some prismatic zircons as well as cores of multifaceted metamorphic grains (Wendt *et al.* 1994; Kröner *et al.* 2000). However, the SHRIMP dating of Kröner *et al.* (2000) and Friedl *et al.* (2003, 2004) revealed abundant oscillatory zoned zircon cores with concordant Ordovician (Fig. 2), as well as less prominent older ages. Friedl *et al.* (2003, 2004) assumed a c. 450-Ma-old igneous protolith for their sample from the Dunkelsteiner Wald.

The early Rb–Sr whole-rock dating of Austrian felsic granulites also yielded Ordovician ages (486 ± 11 and 450 ± 30 Ma; Arnold & Scharbert 1973, recalculated with the accepted decay constants of Steiger & Jäger 1977). Similar whole-rock Rb–Sr ages (~470 Ma) were obtained for felsic granulites from Blanský les (present work and A. Arnold, unpublished data, quoted by Kodym *et al.* 1978). Rb–Sr thin slab dating of the St. Leonhard Massif yielded a somewhat younger age of 428 ± 16 Ma (Frank *et al.* 1990). The significance of these pre-Variscan ages will be addressed below.

2. Whole-rock geochemistry

2.1. The database

The present authors have assembled a database in which new analyses supplement critically selected literature data in order to assess the overall whole-rock geochemical variability of the Moldanubian granulites. Data management, recalculation, plotting and statistical evaluation were facilitated using *GCDkit* (Janoušek *et al.* 2003a).

In this database, over 20 new samples complemented 93 previously published Austrian whole-rock analyses (Vellmer 1992; Becker 1997; Becker *et al.* 1999; Cooke 1999). Several granulite quarries show a striking alternation of darker and lighter bands, sometimes tightly folded (Fig. 3a, b). The banding could have formed by the Variscan metamorphic segregation or migmatisation. Alternatively, it may represent older, pre-Variscan heterogeneities (enclaves, dykes?) flattened and obscured by the strong deformation. In order to verify the presence of isotopic (dis-) equilibria, sample pairs were collected from adjacent darker (denoted by suffix *D*) and lighter (*L*) bands (*Popp-D*/*L*, *Ost-D*/*L* and *Ysp-D*/*L*). Petrographic sample descriptions are given in Appendix 1.

The database includes 50 new and ~180 old Czech whole-rock analyses (Kodym *et al.* 1978; Slabý 1983; Čadková *et al.* 1985; Strejček 1986; Fiala *et al.* 1987a; Gürtlerová *et al.* 1997). Peculiar K-feldspar–garnet perpotassic granulites, documented from a single small body within the Blanský les Massif (Vrána 1989), were not considered. Many of the new samples came

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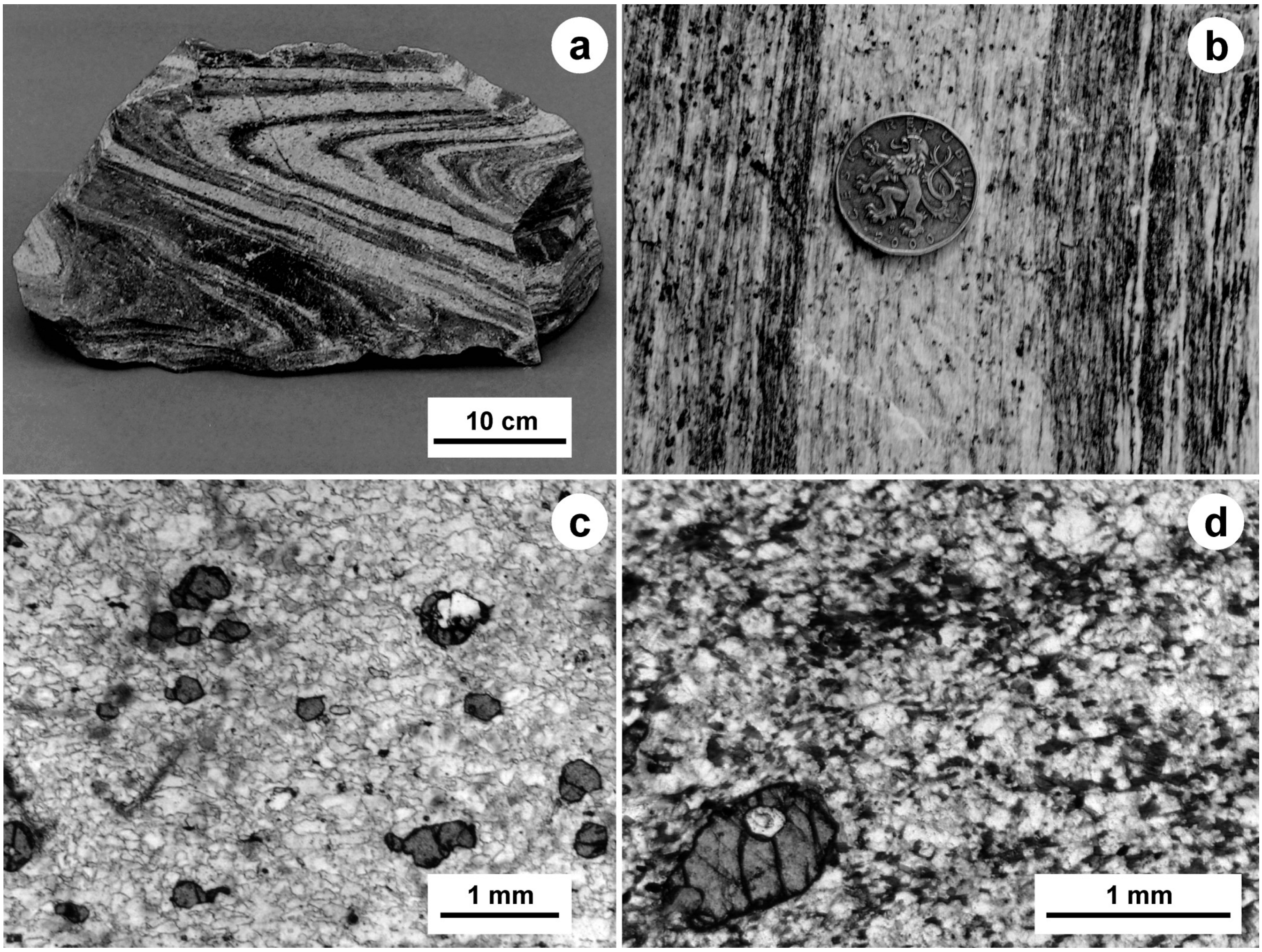

Figure 3 (a) Folded distinctly banded felsic granulite, Yspertal quarry. (b) Detail of the hand specimen of a felsic granulite with dark bands rich in biotite and strong mylonitic fabric, Plešovice quarry, Blanský les Massif. (c) Photomicrograph of felsic granulite *Popp-L* (Krug, St. Leonhard Massif). Garnet porphyroclasts are set in a fine-grained feldspar–quartz matrix. Plain polarised light. (d) Photomicrograph of dark facies of the sample from Osterberg, Pöchlarn–Wieselburg (*Ost-D*). A large garnet porphyroclast encloses a subhedral grain of mesoperthite. The strongly recrystallised feldspar–quartz matrix contains streaks of biotite. Plain polarised light.

from the borehole H-1 drilled in the Blanský les Massif (Appendix 1).

As noted above, caution should be taken to separate individual granulite groups with potentially distinct origin (Fiala *et al.* 1987a). For instance, the major-element binary plots do not follow simple patterns, but are characterised by inflections or discontinuities at SiO_2 ~60 and/or SiO_2 ~70 wt.% (Vellmer 1992). The present paper focuses on a rather coherent group of most common felsic garnet ± kyanite granulites, defined on the basis of geochemical variation diagrams as having SiO_2 >70 wt.%. In addition, they needed to be devoid of orthopyroxene, since the rare charnockitic types are suspected of forming a distinct geochemical and petrogenetic suite. In total, some 200 analyses of the felsic granulites were evaluated (Fig. 1).

2.2. Major elements

All Moldanubian granulites span a wide range of felsic (granitic) to mafic (gabbroic) compositions, as shown by the multicationic diagram P–Q (Debon & Le Fort 1983; Fig. 4). It is worth stressing that the sampling is strongly biased towards the less-common basic types. All the rocks are subalkaline and define a calc-alkaline trend in the Na_2O+K_2O–FeOt–MgO ternary diagram (AFM; Irvine & Baragar 1971; Fig. 5a). As noted already by Fiala *et al.* (1987b), the granulites can be characterised as medium- (mafic and transitional) to high-K (felsic) calc-alkaline rocks on the basis of the SiO_2–K_2O plot of Peccerillo & Taylor (1976) (Fig. 5b).

The major-element composition of the felsic granulites is very silicic (SiO_2=74·4 ± 2·3 wt.%, median ± 1σ; Table 1), slightly peraluminous (A/CNK=1·11 ± 0·08), K-rich (K_2O/Na_2O=1·54 ± 0·37) and with Fe prevailing strongly over Mg (FeOt=1·94 ± 0·71 wt.%, MgO=0·47 ± 0·35 wt.%, mg# ~30). At given SiO_2 contents, CaO (1·03 ± 0·48 wt.%) is rather high while Al_2O_3 is low (13·28 ± 0·91 wt.%). The Harker variation diagrams (Fig. 6) show strong negative correlations of SiO_2 with TiO_2, Al_2O_3, FeOt, MgO, CaO and mg#. In contrast, alkalis and the alumina saturation index (A/CNK, not shown) are scattered, potentially implying some metamorphic mobility.

2.3. Trace elements

2.3.1. LILE. All the Moldanubian granulites, including the mafic types, have typical crustal K/Rb ratios (i.e. between c. 120 and 500; Shaw 1968; Rudnick *et al.* 1985) (Fig. 7a; Table 1), and thus, are not significantly depleted in Rb (see also Fiala *et al.* 1987a, b; Vellmer 1992). The felsic types show wide range of Rb contents (28–362 p.p.m., 164 ± 59 p.p.m.) and Rb/Sr ratios (0·5–16·5, median=2·9). The SiO_2–Rb plot is characterised by a scattered positive trend (Fig. 6). The Rb/Cs

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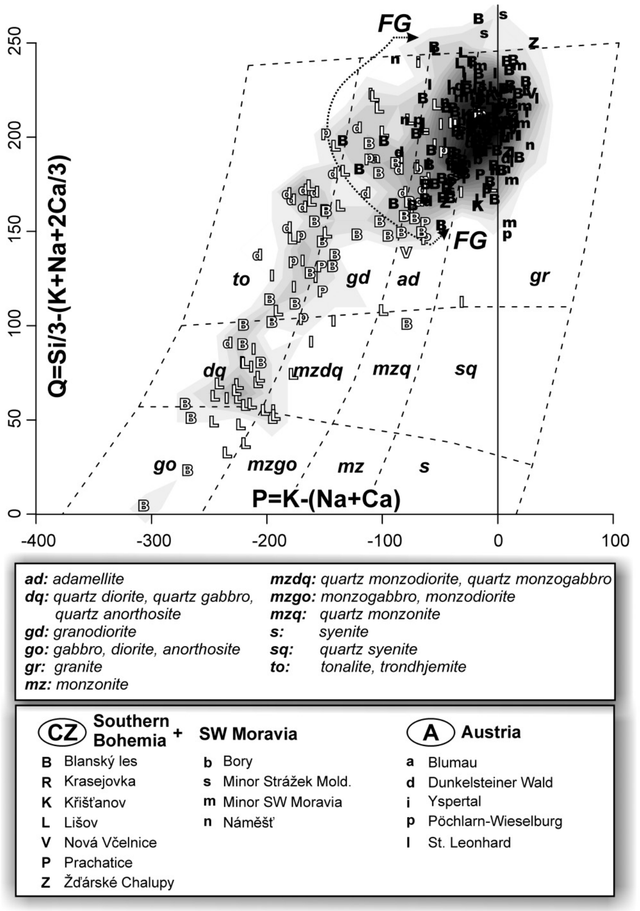

Figure 4 Frequency plots of Moldanubian granulites in the multicationic diagram P–Q (Debon & Le Fort 1983). *P* represents the proportion of K-feldspar to plagioclase and *Q* the quartz content. Plotting symbols represent individual granulite massifs. Filled letters stand for felsic granulite (*FG*) samples dealt with in detail in the text. The frequency of all available analyses is expressed by various shades of grey.

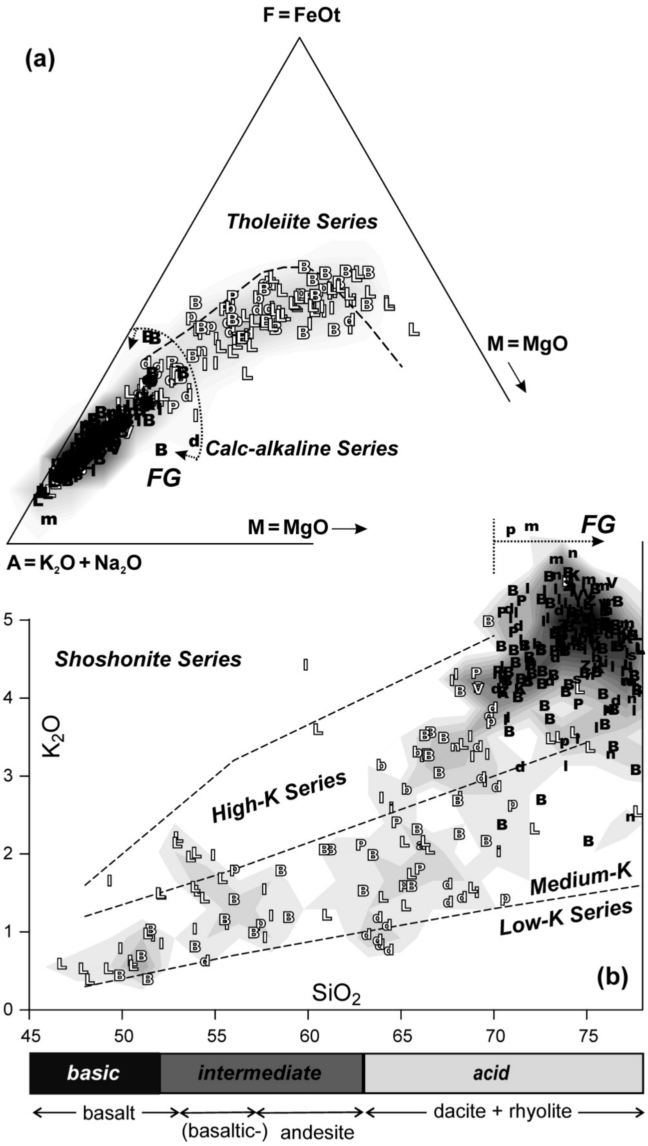

Figure 5 (a) AFM diagram ($A = Na_2O + K_2O$, $F = FeOt$, $M = MgO$: Irvine & Baragar 1971) illustrating the calc-alkaline trend shown by the Moldanubian granulites. (b) SiO_2–K_2O plot with the discrimination boundaries between the tholeiitic, calc-alkaline, high-K calc-alkaline and shoshonitic rocks after Peccerillo & Taylor (1976). Symbols as in Figure 4.

ratios for the felsic granulites (11–1100, median = 122) are higher than the average continental crust (Rb/Cs = 30; Taylor & McLennan 1985; or even 23; Wedepohl 1995). Strontium, and in particular Ba, vary over a wide range. Both show good covariation and sharp decrease with increasing silica (Figs 6 & 7 b–c).

2.3.2. U and Th. While many of the felsic Moldanubian granulites have Th/U (0·2–39, median = 3·1), corresponding to normal igneous rocks (Rudnick & Presper 1990), a significant overall depletion of both elements exists (Fiala *et al.* 1987a), down to less than 0·05 times the concentration of average upper crust (Taylor & McLennan 1985).

2.3.3. REE. Chondrite-normalised REE plots resemble typical patterns of fractionated granitic rocks, showing moderate to weak enrichment in LREE ($Ce_N/Yb_N = 1{\cdot}5$–$7{\cdot}2$, $Ce_N/Sm_N = 1{\cdot}3$–$3{\cdot}4$; Fig. 8) and negative Eu anomalies ($Eu/Eu^* = 0{\cdot}14$–$0{\cdot}62$, $Eu^* = \sqrt{Sm_N Gd_N}$), the magnitude of which increases with increasing silica (see also Vellmer 1992; Jakeš 1997). There is a sharp decrease in LREE and MREE contents with increasing SiO_2, while the HREE contents remain relatively constant (Figs 6 & 8). Such variation is a common feature in felsic granitic magmas which have undergone fractional crystallisation, particularly at monazite saturation (Miller & Mittlefehldt 1982, 1984; Ayres & Harris 1997). Given the siliceous and slightly peraluminous nature of the studied rocks, the presence of monazite as a main LREE carrier is likely (Watt & Harley 1993; Broska *et al.* 2000). If one assumes that the whole-rock compositions correspond to those of monazite-saturated melt, the calculated crystallisation temperatures (Montel 1993) would be close to 750 °C (Fig. 9a).

2.3.4. Zr. Zirconium concentrations in the felsic granulites are generally low (86 ± 50 p.p.m.). These decrease with increasing SiO_2 to less than 50 p.p.m. (Fig. 6), which is compatible with zircon fractionation. The zircon saturation temperatures (Watson & Harrison 1983) of 740 ± 48 °C are comparable to those obtained from the monazite saturation model (Fig. 9b).

2.3.5. Spiderplots. Spiderplots for the felsic granulites (Fig. 10a) normalised to the average upper crust are characterised by conspicuous troughs in Nb, Sr, Hf, Zr and Ti. A similar feature is also seen for U and Th in most cases. Some elements tend to be variously depleted, starting with apparent crustal values (e.g. Ba, P and Nd). Rubidium and K seem to be in the range of the upper crust (if not somewhat higher) together with HREE and Y. In contrast, the LREE tend to be slightly lower than upper crustal values. With notable exception of U and Th, the patterns resemble those of felsic granites, variably modified by fractional crystallisation.

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Table 1 Selected whole-rock geochemical data for the (mostly felsic) Moldanubian granulites

	Popp-D	Popp-L	Ost-D®	Ost-L	St	Du©	Ysp-D	Ysp-L	1059	1062	1064	1074	1076	1078
Locality*	1	1	2	2	3	4	5	5	1146	1178	1312	1372	1381	1405 m
Massif†	StL	StL	PW	PW	DW	DW	Yt	Yt	BL	BL	BL	BL	BL	BL
SiO_2	71·69	74·82	63·90	77·55	76·94	70·14	76·03	75·84	70·43	73·87	75·54	76·54	76·02	70·78
TiO_2	0·47	0·21	0·78	0·09	0·12	0·50	0·20	0·15	0·35	0·27	0·12	0·07	0·08	0·32
Al_2O_3	14·53	12·95	17·89	11·90	12·30	13·97	12·66	12·36	14·36	13·53	12·58	12·10	12·16	13·92
FeOt	2·95	1·72	3·66	1·52	1·66	4·05	1·84	1·64	4·01	2·22	1·72	1·56	1·54	2·79
MnO	0·05	0·04	0·03	0·04	0·04	0·07	0·05	0·03	0·05	0·03	0·02	0·03	0·02	0·04
MgO	0·83	0·51	2·82	0·17	0·21	0·94	0·40	0·24	1·54	0·55	0·58	0·28	0·31	0·93
CaO	1·63	0·84	3·02	0·69	0·73	3·22	0·82	0·67	2·50	1·69	1·41	0·78	0·95	1·68
Na_2O	3·12	2·37	5·20	3·49	3·20	3·69	2·95	3·16	3·94	3·64	4·21	2·66	2·78	3·34
K_2O	4·65	5·40	2·74	3·82	4·32	2·85	4·45	4·98	2·36	4·06	3·13	4·85	4·94	4·31
P_2O_5	0·18	0·16	0·21	0·10	0·15	0·12	0·16	0·14	0·08	0·07	0·03	0·13	0·15	0·18
H_2O^+/LOI	0·57‡	0·20‡	0·51‡	0·22‡	0·44‡	0·27‡	0·49‡	0·30‡	0·27	0·08	0·27	0·12	0·14	0·64
F	–	–	–	–	–	–	–	–	0·05	0·03	0·06	0·05	0·04	0·03
Total	100·67	99·22	100·76	99·59	100·11	99·82	100·05	99·51	99·94	100·04	99·67	99·17	99·13	98·96
mg#	33·4	34·6	57·9	16·6	18·4	29·3	27·9	20·7	40·7	30·7	37·6	24·3	26·4	37·3
A/CNK	1·11	1·15	1·05	1·07	1·09	0·93	1·13	1·05	1·06	1·01	0·98	1·10	1·04	1·05
K_2O/Na_2O	1·49	2·28	0·53	1·09	1·35	0·77	1·51	1·58	0·60	1·12	0·74	1·82	1·78	1·29
(p.p.m.)														
Rb	145	168	201	331	288	48	220	180	66	126	92	241	216	153
Sr	96	73	158	35	34	71	61	58	79	70	73	35	46	108
Ba	612	398	402	58	139	565	371	357	496	601	476	114	190	586
Nb	7·5	4·9	11·1	b.d.	4·8	4·3	6·3	4·1	7·4	5·1	2·9	b.d.	3·3	8·9
Zr	115	105	120	16	38	108	102	74	135	112	29	50	66	106
Hf	3·8	1·5	2·4	b.d.	2·1	2·1	2·7	1·0	2·1	1·7	0·9	b.d.	0·6	1·7
Ga	19	18	26	20	18	18	18	17	19	17	18	17	16	19
Th	2·1	2·8	b.d.	2·6	3·7	b.d.	3·2	b.d.	5·0	3·8	b.d.	3·4	2·4	5·6
U	2·2	2·9	b.d.	2·3	2·7	b.d.	5·2	3·0	3·8	1·6	1·0	3·0	2·2	2·7
Cr	41	61	120	11	11	75	22	18	–	–	–	–	–	–
Co	5·9	2·4	7·1	1·2	2·2	13·5	4·6	2·7	10·7	4·8	5·4	3·4	3·4	6·8
Pb	1·3	6·7	b.d.	b.d.	4·1	b.d.	5·5	4·7	b.d.	0·9	4·4	4·3	7·2	4·3
Zn	48	23	24	15	17	59	37	26	59·8	27·6	28·2	22·5	23·4	43·3
La	32·2	22·0	22·0	9·0	11·6	37·0	17·4	14·1	21·1	13·5	–	9·4	12·1	24·1
Ce	69·3	50·4	48·6	19·3	27·0	74·6	40·0	32·1	47·0	39·4	–	23·6	30·5	56·0
Pr	–	–	–	–	–	–	–	–	6·02	4·86	–	2·38	3·82	6·67
Nd	28·9	19·2	22·7	6·3	11·5	29·2	17·4	13·3	24·2	20·8	–	10·6	14·2	26·8
Sm	6·67	4·42	5·51	1·61	3·05	5·35	4·16	3·48	5·05	4·72	–	2·67	3·41	5·67
Eu	0·91	0·42	0·93	0·09	0·17	1·00	0·34	0·23	0·89	0·54	–	0·25	0·38	0·89
Gd	7·30	4·72	5·35	2·28	3·58	4·88	4·30	3·70	5·43	5·93	–	3·15	3·92	6·38
Tb	–	–	–	–	–	–	–	–	1·2	1·2	–	0·83	0·96	1·24
Dy	7·93	6·07	5·68	4·38	5·76	3·71	6·06	6·00	5·5	7·84	–	5·47	6·73	7·51
Ho	–	–	–	–	–	–	–	–	0·99	1·73	–	1·15	1·50	1·45
Er	4·43	3·92	3·47	3·25	3·51	1·98	3·8	3·81	3·53	5·54	–	4·07	4·92	4·35
Tm	–	–	–	–	–	–	–	–	0·33	0·55	–	0·52	0·60	0·49
Yb	4·21	4·06	3·94	3·82	3·78	1·89	3·97	4·12	3·25	4·84	–	4·11	4·57	3·31
Lu	0·62	0·57	0·60	0·52	0·49	0·28	0·52	0·56	0·5	0·7	–	0·58	0·65	0·44
Y	45·8	38·7	34·7	30·3	35·6	19·9	36·8	38·0	34·1	50·2	11·2	36·2	44·9	48·6
Eu/Eu*	0·40	0·28	0·52	0·14	0·16	0·60	0·25	0·20	0·52	0·31	–	0·26	0·32	0·45
Ce_N/Yb_N	4·3	3·2	3·2	1·3	1·9	10·2	2·6	2·0	3·7	2·1	–	1·5	1·7	4·4
Ce_N/Sm_N	2·5	2·8	2·1	2·9	2·1	3·4	2·3	2·2	2·2	2·0	–	2·1	2·2	2·4
K/Rb	265·5	266·5	113·3	95·9	124·4	490·8	167·6	229·3	299·1	266·9	284·0	167·0	189·9	233·4
Rb/Sr	1·5	2·3	1·3	9·5	3·6	8·5	3·1	0·7	0·8	1·8	1·3	7·0	4·7	1·4

*Localities: (1) Naturstein Popp, Krug: (2) Osterberg: (3) Steinaweg: (4) Lindwurmkreuz: (5) Gleisen. The rest: Holubov, borehole H-1 (with depth indicated). ®Mafic sample. ©Charnockite.

†Massifs (Fig. 1): (StL) St. Leonhard: (PW) Pöchlarn–Wieselburg: (DW) Dunkelsteiner Wald: (Yt) Yspertal: (BL) Blanský les. ‡ Loss on ignition: (b.d.) below detection limit; (–) not determined.

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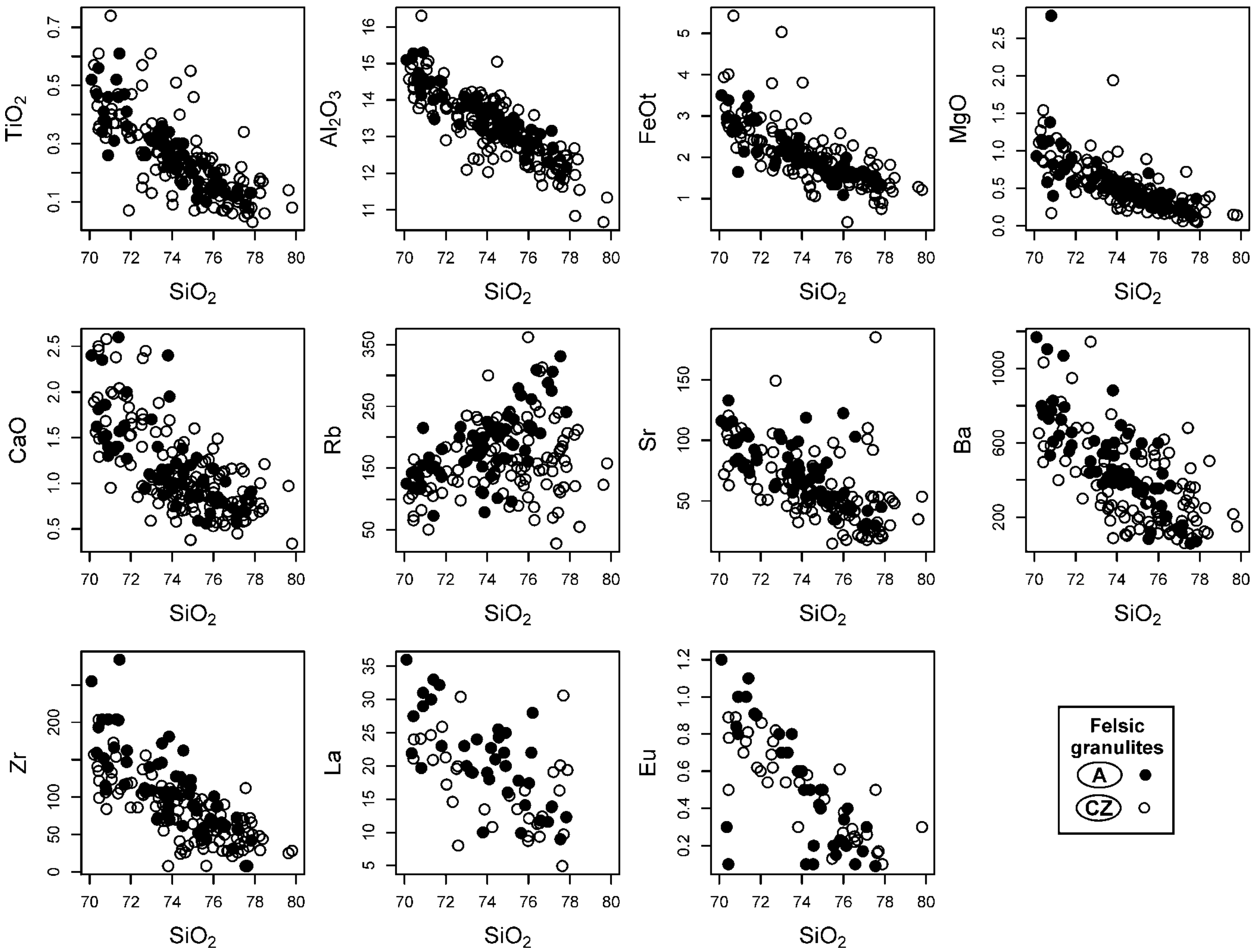

Figure 6 Binary plots of SiO_2 versus major-element oxides (wt.%), mg# and selected trace elements (p.p.m.) for felsic Moldanubian granulites from Austria (filled symbols) and the Czech Republic (empty symbols).

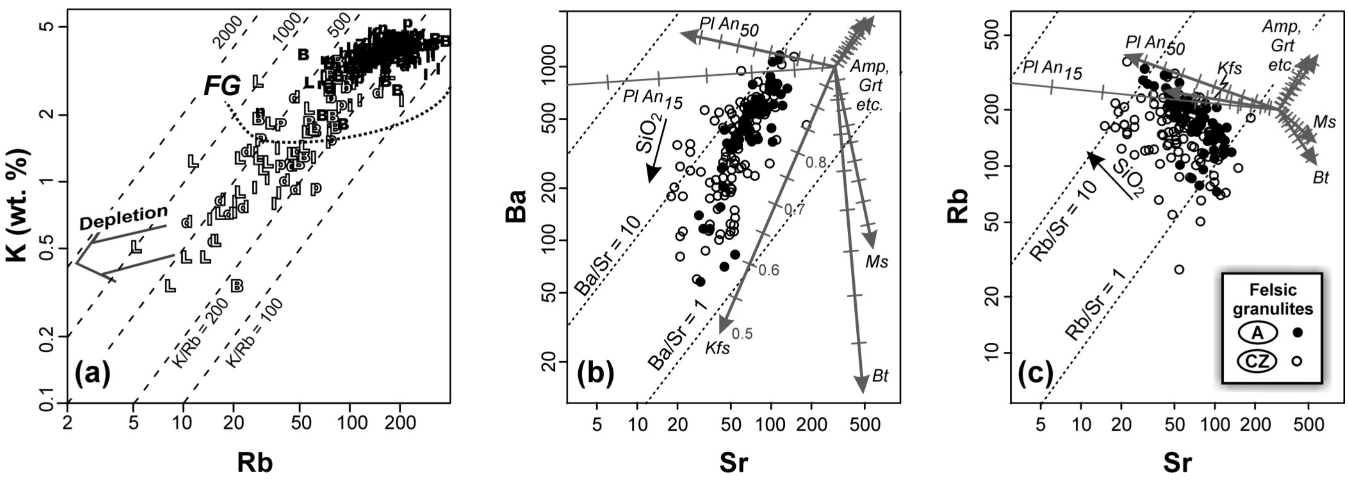

Figure 7 (a) Rb (p.p.m.) versus K (wt.%) diagram for all the Moldanubian granulites. Symbols as in Figure 4. *FG* felsic granulites. Depletion trend for granulites with K/Rb > 500 is taken from Jahn (1990). (b) Sr versus Ba and (c) Sr versus Rb plots (p.p.m.) for the felsic granulites from Austria (filled symbols) and Czech Republic (empty symbols). Labelled vectors correspond to up to 50% fractional crystallisation of the main rock-forming minerals. Trace-element distribution coefficients are from Hanson (1978; Kfs, Amp, Rb in Pl), Icenhower & London (1995; Bt and Ms) and Blundy & Shimizu (1991; Sr in Pl, assuming 750°C, An_{15} and An_{50}).

2.4. Radiogenic isotopes

2.4.1. Neodymium The ε^{340}_{Nd} values for the felsic granulites are distinctly negative ($-4{\cdot}2$ to $-7{\cdot}5$, median $= -6{\cdot}1$; Table 2 & Fig. 11a). This is reflected by rather high mean crustal residence ages (T^{DM}_{Nd}: two-stage depleted-mantle Nd model ages; Liew & Hofmann 1988) of 1·39–1·64 Ga (Fig. 11a), which corresponds to derivation from mature continental crust. The systematic shift in Nd isotopic compositions

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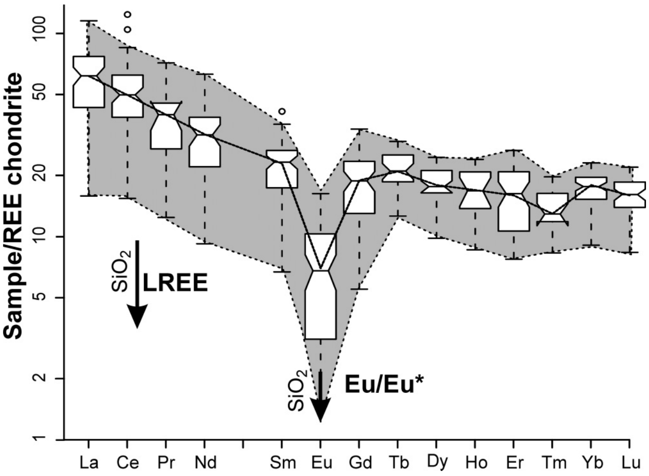

Figure 8 Chondrite-normalised plot for REE in felsic granulites. This new diagram ('spider boxplot') represents a combination of the more common chondrite-normalised REE plot with box and whiskers plots (boxplots). For each element, the box represents 50% of the population (being delimited by two quartiles); the horizontal line inside is the median and the 'whiskers' show the total range without outliers (denoted by small circles). The solid black arrows indicate trends with rising silica, if applicable. Normalisation values are from Boynton (1984).

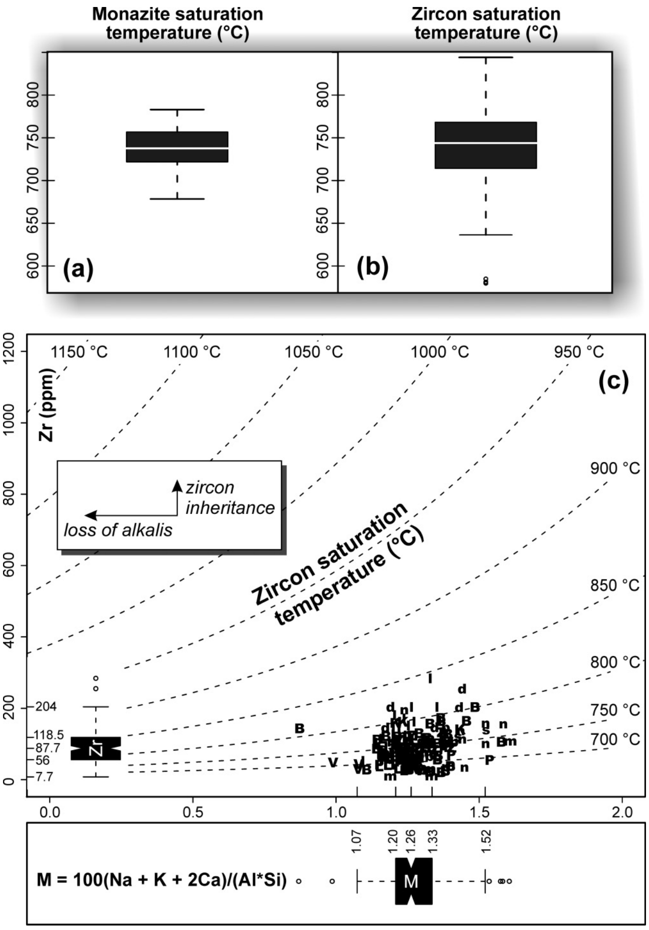

Figure 9 Monazite and zircon saturation calculations for felsic Moldanubian granulites. (a) Boxplot of monazite saturation temperatures (Montel 1993). (b) Boxplot of zircon saturation temperatures (Watson & Harrison 1983). (c) Binary plot of M=100 (Na+K+2Ca)/Al*Si versus Zr (p.p.m.) for the felsic granulites (see Fig. 4 for legend). Boxplots at the axes portray the statistical distribution of both parameters. The isotherms of Zr saturation levels were calculated using the model of Watson & Harrison (1983). The arrows indicate the effects of increasing zircon inheritance and (hypothetical) alkali loss, both of which would produce slight temperature overestimates.

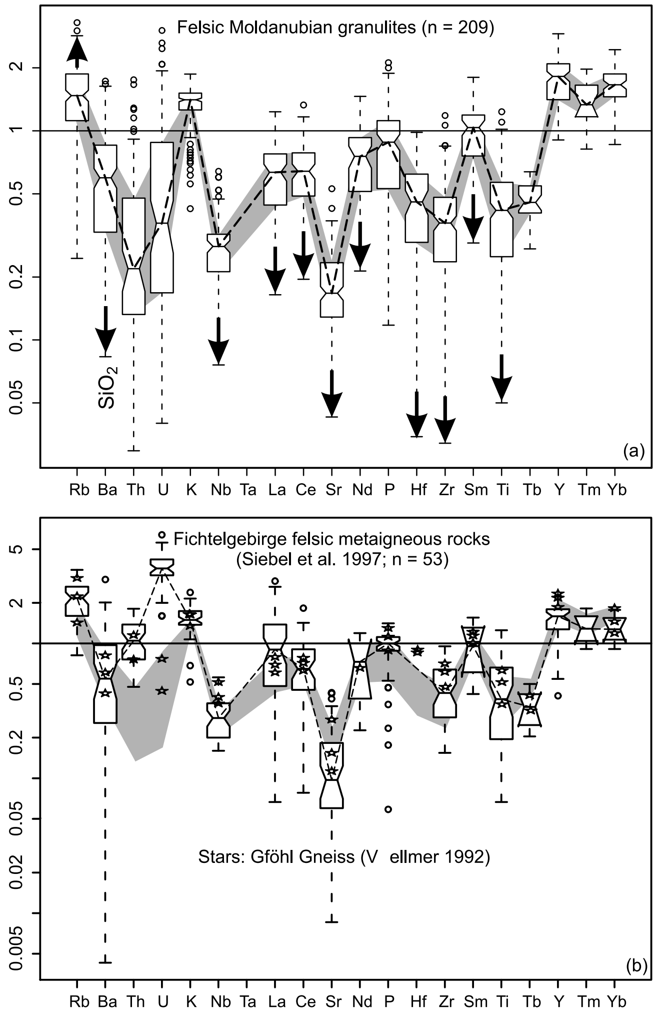

Figure 10 Average upper crust normalised (Taylor & McLennan 1985) 'spider boxplots' for felsic granulites (a) and acid metaigneous rocks from Fichtelgebirge (SiO_2 >70%; Siebel *et al.* 1997) (b). In the latter plot, stars denote three samples of Gföhl gneiss given by Vellmer (1992). An identical field corresponding to 50% of the felsic granulites is outlined in both diagrams. The solid black arrows indicate trends with increasing SiO_2.

between adjacent lighter (more negative ε^{340}_{Nd}) and darker facies (less negative ε^{340}_{Nd}) of the samples *Popp* (−7·0/−6·1) and *Ost* (−7·5/−5·7) rules out closed-system differentiation during the Variscan metamorphism. Small-scale isotopic disequilibrium was also evidenced by previous Sm–Nd (and U–Pb) garnet–whole-rock ages which show considerable scatter (470 to <330 Ma: Janoušek *et al.* 1996; Becker 1997; Chen *et al.* 1998; Prince *et al.* 2000). Conversely, granulites *Ysp-D* and *Ysp-L* have nearly the same ε^{340}_{Nd} values (−6·2/−6·4), in accordance with their mutually comparable major- and trace-element compositions.

2.4.2. Strontium. The newly measured Sr isotopic ratios show pronounced variations ($^{87}Sr/^{86}Sr_{340}$ ~0·711–0·770; Fig. 11b). The ratios demonstrate strong isotopic disequilibrium between adjacent lighter and darker samples *Popp* (0·730/0·719), *Ost* (0·753/0·715) and *Ysp* (0·771/0·738) at 340 Ma (Fig. 11b).

If one assumes an initial equilibrium state, the individual pairs would all give Early Palaeozoic ages in line with the results of the previous Rb–Sr dating (Arnold & Scharbert 1973; Frank *et al.* 1990). Similarly, the Blanský les granulites define an Ordovician isochron (469 ± 8 Ma; 2σ; $^{87}Sr/^{86}Sr_i$ ~0·706; Fig. 12a). Sample 1078, which is not included in the regression, could have been derived, along with 1076, from a

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Table 2 Selected whole-rock Sr–Nd isotopic data for the (mostly felsic) Moldanubian granulites

Laboratory†	Popp-D CNRS	Popp-L CNRS	Ost-D® CNRS	Ost-L CNRS	St CNRS	Du© CNRS	Ysp-D CNRS	Ysp-L CNRS
Rb (p.p.m.)	162	209	212	288	260	72	249	284
Sr (p.p.m.)	97	65	137	37	34	99	43	30
$^{87}Rb/^{86}Sr$	4·85	9·36	4·49	22·9	22·5	2·11	16·9	27·9
$^{87}Sr/^{86}Sr$	0·74240	0·77558	0·73679	0·86389	0·86499	0·72075	0·81994	0·90574
2 s.e.	3	3	3	3	3	3	3	3
$^{87}Sr/^{86}Sr_{340}$*	0·71893	0·73026	0·71506	0·75322	0·75625	0·71055	0·73796	0·77059
Sm (p.p.m.)	7·64	5·53	5·6	1·88	3·29	6·34	4·59	3·53
Nd (p.p.m.)	34·0	23·8	24·8	7·03	12·1	35·2	18·5	13·6
$^{147}Sm/^{144}Nd$	0·1358	0·1402	0·1363	0·1617	0·1640	0·1090	0·1501	0·1571
$^{143}Nd/^{144}Nd$	0·512142	0·512200	0·512119	0·512266	0·512291	0·512073	0·512218	0·512224
2 s.e.	7	7	11	8	9	8	8	11
$^{143}Nd/^{144}Nd_{340}$*	0·511840	0·511888	0·511816	0·511906	0·511926	0·511830	0·511884	0·511874
ε_{Nd}^{340*}	− 7·0	− 6·1	− 7·5	− 5·7	− 5·4	− 7·2	− 6·2	− 6·4
T_{DM}^{Nd} (Ga)	1·60	1·53	1·64	1·50	1·47	1·62	1·54	1·55
$^{87}Sr/^{86}Sr_{470}$*	0·70993	0·71287	0·70672	0·71076	0·71453	0·70664	0·70651	0·71875
ε_{Nd}^{470*}	− 6·0	− 5·2	− 6·5	− 5·2	− 4·8	− 5·8	− 5·4	− 5·7

Laboratory†	1059 CGS	1062 CGS	1064 CGS	1074 CGS	1076 CGS	1078 CGS
Rb (p.p.m.)	72	134	99	254	230	158
Sr (p.p.m.)	72	63	66	26	39	102
$^{87}Rb/^{86}Sr$	2·65	5·84	4·04	26·4	16·1	4·30
$^{87}Sr/^{86}Sr$	0·72398	0·74516	0·73349	0·8784	0·81717	0·73802
2 s.e.	20	5	8	11	8	8
$^{87}Sr/^{86}Sr_{340}$*	0·71117	0·71691	0·713932	0·75045	0·73926	0·71722
Sm (p.p.m.)	6·7	5·1	0·7	2·8	4·7	4·3
Nd (p.p.m.)	33·0	20·9	2·20	10·0	20·8	19·1
$^{147}Sm/^{144}Nd$	0·1227	0·1475	0·19235	0·1693	0·1366	0·1361
$^{143}Nd/^{144}Nd$	0·512087	0·512206	0·512335	0·512301	0·512277	0·512199
2 s.e.	21	14	40	19	15	19
$^{143}Nd/^{144}Nd_{340}$*	0·511814	0·511878	0·511907	0·511924	0·511973	0·511896
ε_{Nd}^{340*}	− 7·5	− 6·3	− 5·7	− 5·4	− 4·4	− 5·9
T_{DM}^{Nd} (Ga)	1·64	1·55	1·50	1·48	1·4	1·52
$^{87}Sr/^{86}Sr_{470}$*	0·70626	0·70607	0·70643	0·70137	0·70936	0·70924
ε_{Nd}^{470*}	− 6·3	− 5·5	− 5·7	− 4·9	− 3·4	− 4·9

*Subscripts indicate the age to which were isotopic ratios corrected: 340 Ma (granulite-facies metamorphism: Kröner *et al.* 2000 and references therein) and 470 Ma (estimated protolith age): ε_{Nd}^{i} values were calculated using the Bulk Earth parameters given by Jacobsen & Wasserburg (1980); T_{DM}^{Nd} are two-stage Nd model ages based on Liew & Hofmann (1988). ®Mafic sample (SiO_2=63·90 wt.%; see Table 1). ©Charnockite.
†CNRS: Université Blaise Pascal, Clermont-Ferrand; analytical techniques described by Pin *et al.* (1990) and Roberts *et al.* (2000). CGS: Czech Geological Survey, Prague–Barrandov; Sm and Nd concentrations determined by isotope dilution, Rb and Sr by XRF (Harvey & Atkin 1981). Isotopic ratios were measured in static mode on a Finnigan MAT 262 spectrometer in the CGS. In both laboratories, the $^{143}Nd/^{144}Nd$ ratios were corrected for mass fractionation by normalisation to $^{146}Nd/^{144}Nd$=0·7219, while $^{87}Sr/^{86}Sr$ ratios were normalised to $^{86}Sr/^{88}Sr$=0·1194.

domain with a somewhat higher $^{87}Sr/^{86}Sr_i$ (~0·709; Fig. 12a). Alternatively, it may have been partly influenced by the nearby Variscan granitic dykes (Appendix 1). Nevertheless, the bootstrap spectrum for all the samples seems to confirm the Ordovician age (Fig. 12b).

Five thin-slab and one whole-rock Rb–Sr isotopic analyses from the St. Leonhard Massif were re-evaluated for comparison (Frank *et al.* 1990, without feldspar fraction and sample 1, rejected by the original authors because of substantial alteration; M. Thöni, pers. comm., 2004). The age obtained (425 ± 19 Ma, MSWD=5·8) seems to be distinctly younger than that for the Blanský les set. Again, the bootstrap spectrum (Fig. 12c), despite skewness, shows no ages lower than 400 Ma and the ~430 Ma maximum remains rather well defined.

Furthermore, on the $^{87}Sr/^{86}Sr_{340}$–ε_{Nd}^{340} plot (Fig. 11c), the granulites from the Czech Republic and Austria form broadly subparallel, partly overlapping fields. They run diagonally from less to more radiogenic Sr–Nd isotopic compositions, reaching $^{87}Sr/^{86}Sr_{340}$ values >0·77, which are clearly above the initial ratios typical of common granitic rocks. The easiest explanation is that the data are not sufficiently age-corrected and that isotopic equilibrium during the Variscan was not attained. This agrees with the fact that the ε_{Nd}^{340} values become less negative with increasing silica along with the Sm/Nd ratios.

The significance of the individual Rb–Sr whole-rock ages should not be overrated since the system is known to be prone to resetting by metamorphism. Nonetheless, it is clear that most of the Sr isotopic ratios of the felsic granulites when corrected to Early Palaeozoic ages (see also Fig. 2) would yield $^{87}Sr/^{86}Sr_i$ isotopic ratios expected for normal, slightly-peraluminous, crustally-derived granites (Fig. 11b; Table 2).

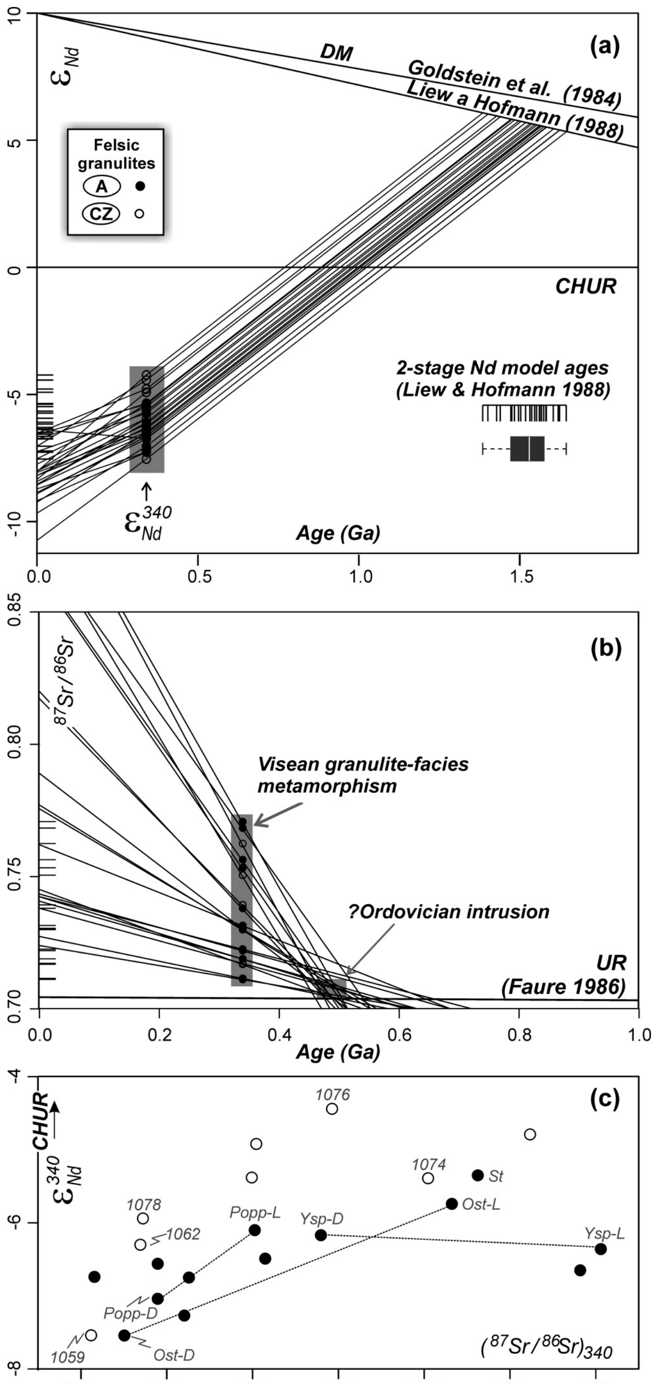

Figure 11 (a) Two-stage Nd development diagram for the felsic (except for *Ost-D*) Moldanubian granulites. DM = Depleted Mantle evolution lines after Goldstein *et al.* (1984) and Liew & Hofmann (1988). (b) Strontium isotopic plot. UR = Uniform (chondritic) Reservoir evolution line after Faure (1986). (c) Diagram of $(^{87}Sr/^{86}Sr)_{340}$ versus ε^{340}_{Nd} for felsic granulites. Tie lines join adjacent sample pairs from banded granulites (L = lighter; D = darker). The plot also includes data from Vellmer (1992), Valbracht *et al.* (1994) and Becker *et al.* (1999); new analyses are labelled by sample names.

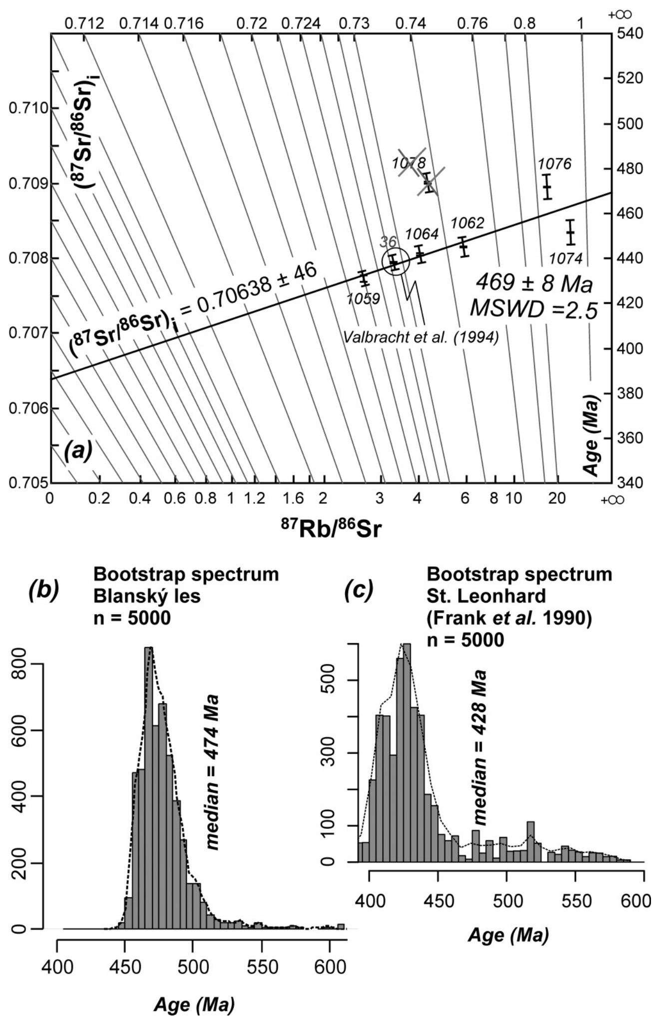

Figure 12 (a) Improved isochron diagram (Provost 1990) for Rb–Sr isotopic compositions of felsic granulites from the Blanský les Massif (this work and sample 36 of Valbracht *et al.* 1994). (b) Bootstrap spectrum (Diaconis & Efron 1983; Kalsbeek & Hansen 1989) obtained on the basis of 5000 randomly chosen data sets for seven points from Figure 12a (i.e. including 1078). (c) Bootstrap spectrum for six Rb–Sr isotopic analyses from the St. Leonhard Massif (Frank *et al.* 1990; feldspar fraction and sample 1 were omitted). The individual ages in both cases were calculated using the algorithm of York (1969) and the unpublished programme *BTRAP*.

3. Discussion

3.1. Pre-Variscan metagranites or Variscan high-P melts?

The CIPW-normative compositions of the felsic granulites fall close to a low-pressure minimum in the Ab–Qz–Or ternary diagram (Fiala *et al.* 1987a; Vrána 1989; Kotková & Harley 1999), and this demonstrates that they represent magma compositions with little cumulate or restite component. Indeed, most of the authors agreed that they were granitic or rhyolitic magmas at some stage of their evolution (Fiala 1987a, b; Carswell 1991; Vellmer 1992; Jakeš 1997; Kotková & Harley 1999). Even though there is near consensus on igneous heritage of the felsic Moldanubian granulites, it is important to discuss whether: (a) the bulk formed and segregated as high-pressure melts during Variscan subduction (Kotková & Harley 1999); or (b) they are merely Early Palaeozoic granites which were nearly isochemically metamorphosed during the Variscan collision, undergoing no or a negligible degree of partial melting.

In the present authors' view, there are several reasons why the latter model is more realistic:

3.1.1. Preservation of the pre-Carboniferous Rb–Sr whole-rock ages. As shown by the variation in the $^{87}Sr/^{86}Sr_{340}$ ratios and the preserved Ordovician/Silurian isochron ages, the whole-rock Rb–Sr system failed to re-equilibrate, even on a local scale, during the Variscan high-grade metamorphism. This leaves little scope for models which interpret the rocks as segregated Variscan granites and indicates that any granulite-facies partial melting did not lead to larger-scale chemical equilibrium. Given the correspondence of the Rb–Sr systematics with the SHRIMP ages for many inherited zircon cores,

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the Rb–Sr isotopes seem to have largely maintained memory of an older igneous protolith. The preservation of pre-metamorphic Sr isotopic signatures in high-grade meta-igneous rocks is by no means rare (e.g. Hradetzky & Lippolt 1993; Kühn *et al.* 2000). It could have been facilitated by the rarity of hydrous fluids, or melt immobility and isolation during the HP–HT stage, coupled with a presumably short duration of the metamorphic peak.

3.1.2. Zircon and monazite saturation temperatures. If the felsic granulites represent Variscan hot, restite-poor, high-PT melts, then they should have acquired Zr contents several times higher than observed (Fig. 9c). To invoke dramatic Zr-undersaturation (Kotková & Harley 1999) seems unrealistic because all felsic Moldanubian granulites contain abundant pre-Variscan, mainly Ordovician, zircon relics (Pöschl-Otrel 1995; Kröner *et al.* 2000; Friedl *et al.* 2003, 2004). For such low Zr concentrations, the saturation temperatures tend to be robust to changes in both the bulk magma composition and errors in the Zr determinations (Fig. 9c; see also Miller *et al.* 2003).

The fact that older zircon grains are usually overgrown by Variscan shells, interpreted as having crystallised from melt under granulite-facies conditions (Roberts & Finger 1997; Friedl *et al.* 2003), rules out the possibility that the inheritance is merely a result of shielding by restitic minerals. Thus, any partial melting during the Variscan high-grade metamorphism must have been volumetrically limited since considerably more restitic zircons should have dissolved otherwise.

Conversely, the observed strong negative SiO_2–Zr correlation (Fig. 6) seems to be indicative of zircon saturation in course of the melt fractionation (Chappell *et al.* 1998; Hoskin *et al.* 2000). As shown by the zircon saturation calculations, this must have been at temperatures much lower than 900–1000 °C, and thus, most likely a feature of the protolith stage. As the volume proportion of pre-Ordovician cores is rather limited (Kröner *et al.* 2000; Friedl *et al.* 2003, 2004), the equivalent zircon and monazite saturation temperatures ($\sim$750 °C) can be interpreted as a rough estimate of the crystallisation temperatures of the Early Palaeozoic protolith.

3.1.3. The HREE and Y signature. If the felsic granulites were essentially restite-depleted HP melts, then restite would inevitably be rich in garnet and kyanite. In such case, a significant HREE+Y depletion should be observed in the granulite samples, but this does not seem to be the case (Figs 8 & 10). The data of Kotková & Harley (1999), indicating that the granulite garnets typically show a decrease of garnet-compatible elements (HREE+Y) from core to rim, are indeed compatible with a Rayleigh-type fractionation during the crystal growth. However, this observation does not constrain the proportion of the melt involved (Otamendi *et al.* 2002).

3.2. Nature and origin of the igneous protolith to the Moldanubian granulites

3.2.1. Granite typology. The protolith to the Moldanubian granulites has probably corresponded to low-temperature granites (Miller *et al.* 2003). The geochemical characteristics resemble fractionated I-type granite suites, as defined in the Lachlan Fold Belt in eastern Australia. The similarities include high concentrations of Y and HREE at low-to-moderate Zr, a strongly negative Eu anomaly increasing with SiO_2 and generally low P_2O_5 contents (Chappell & White 1992; Chappell 1999). Compared to A-types, the Fe:Mg ratios and HFSE contents ($Zr+Nb+Ce+Y \sim 200 \pm 60$ p.p.m.) of the granulites are too low (cf. Whalen *et al.* 1987; Eby 1990). The low Nb contents (mostly <10) of the granulites are a typical feature for granites derived from rather primitive, arc-type crustal sources (Pearce *et al.* 1984).

3.2.2. Possible source. With the assumption of granulite samples at SiO_2 $\sim$70 wt.% to approximate the least differentiated granitic melts, this indicates relative paucity in Rb ($Rb^{[70]}$ $\sim$100 p.p.m., superscript = approximate silica contents by weight) and Sr ($\sim$100 p.p.m., $Rb/Sr^{[70]}$ $\sim$1), together with a high Ba content (600–800 p.p.m.).

The whole-rock geochemical character (A/CNK $\sim$1·1, K_2O/Na_2O $\sim$1·5, high Ba, low Rb, strongly negative ε^{470}_{Nd} at relatively unevolved $^{87}Sr/^{86}Sr_{470}$) argues for a partial melting of a quartzo–feldspathic source (orthogneiss or greywacke; Barbarin 1996 and references therein) fairly immature in terms of its major- and trace-element composition and with relatively low time-integrated Sm/Nd and Rb/Sr ratios. In addition, metapelitic parentage may be ruled on the basis of the high $CaO/Na_2O^{[70]}$ ratios ($\sim$0·6–0·8), fairly low $Al_2O_3/TiO_2^{[70]}$ ($\sim$25–40), $Rb/Ba^{[70]}$ ($\sim$0·1–0·3) and $Rb/Sr^{[70]}$ ($\sim$1) (Sylvester 1998). Apart from low Rb/Sr ratios, any significant role for muscovite dehydration melting can be discounted based on low $Sr/Ba^{[70]} = 0{\cdot}1$–$0{\cdot}2$ (Harris & Inger 1992).

The distinction between melts produced from metapsammitic parents and derived by anatexis of K-rich orthogneisses, respectively, remains fairly difficult using geochemical characteristics (e.g. Barbarin 1996; Sylvester 1998).

3.2.3. Magmatic evolution. Increasing Rb and decreasing Ba and Sr with increasing SiO_2 of the Moldanubian granulites (Figs 7b, c) are best explained by fractional crystallisation and are inconsistent with partial-melting models. Mass-balance calculations (Janoušek *et al.* 1997) indicate that unreasonably high degrees of partial melting ($\geqslant$80%) would be required to yield a comparably broad compositional range of partial melts.

An origin by large-scale fractionation of more mafic (dioritic–tonalitic) parents is unlikely given the generally crustal-like Sr and Nd isotopic compositions of the felsic granulites. Indeed, the much rarer more mafic types have noticeably more primitive isotope compositions (Becker 1997; Janoušek *et al.* 2003b). In addition, mafic granulite types are fairly subordinate at the present surface level.

The trends seem to be compatible with fractional crystallisation involving K-feldspar>plagioclase from rather siliceous (SiO_2 $\sim$70 wt.%) magma. Some role for biotite $\pm$ accessory oxide minerals (ilmenite?) is apparent in view of falling MgO, FeOt (Fig. 6), Nb and Zn contents (not shown). The drop in Zr and LREE can be accounted for by zircon and monazite fractionation, respectively (Fig. 6). The negative correlation of Sr, Ba, Zr, LREE and Eu/Eu* with the SiO_2, along with enrichment in Rb, are considered to be characteristic of evolved acid igneous rocks, crystallising feldspar, zircon and monazite dominated assemblages (Miller & Mittlefehldt 1984; Chappell & White 1992; Hoskin *et al.* 2000).

3.3. Geological correlation

It follows from the above that a specific, felsic igneous suite of Early Palaeozoic age should have been an important constituent of the subducted Gföhl terrane. In support of this, the associated Gföhl orthogneisses (e.g. Dudek *et al.* 1974; Cooke & O'Brien 2001) also show Ordovician protolith ages (Friedl *et al.* 2004) and whole-rock geochemical characteristics strongly resembling the felsic granulites (Fig. 10b; Vellmer 1992).

Geochemically similar felsic meta-igneous rocks can be found in the Saxothuringian Zone; for example, orthogneisses in the Orlica–Śnieżnik dome in Poland (Franke & Żelaźniewicz 2000; Turniak *et al.* 2000; Bröcker *et al.* 2003), or metagranites and metarhyolites in the Fichtelgebirge in Germany (Siebel *et al.* 1997; Wiegand 1997). Regrettably, detailed comparisons

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are hampered by a lack of reliable and sufficiently large whole-rock geochemical datasets.

Of the available data, the best characterised seem to be the meta-igneous rocks from the Fichtelgebirge (Siebel *et al.* 1997; Wiegand 1997). As follows from Figure 10b, their analyses with SiO_2 >70% match well the overall compositional spectrum for the felsic granulites in most of the elements save U and Th. Unfortunately, there is a dearth of data on Cs contents: the median of four samples analysed by Wiegand (1997) was 4·8 p.p.m., i.e. slightly lower than the upper crustal average of 5·8 p.p.m. (Wedepohl 1995). The most reliable data from felsic granulites (Vellmer 1992; Becker *et al.* 1999) seem to show that these are markedly Cs-impoverished (0·1–2·1 p.p.m., median = 0·9).

The meta-igneous rocks from Fichtelgebirge were dated at 455–480 Ma (Siebel *et al.* 1997; Wiegand 1997 and references therein), which falls within the spectrum of inherited ages observed in the granulites. Despite considerable disturbance of the Rb–Sr isotopic system in many of the Fichtelgebirge samples, some still yield Ordovician ages and initial ratios (Siebel *et al.* 1997) comparing favourably with the mean of the felsic granulites at that time (~0·709). Moreover, the ε^{480}_{Nd} values (orthogneisses: $-3{\cdot}5 \pm 0{\cdot}4$, metarhyolites: $-4{\cdot}8 \pm 1{\cdot}1$) also match the mean ε^{480}_{Nd} ($-4{\cdot}9 \pm 1{\cdot}9$) for the felsic granulites and overlap with the data for the Gföhl gneiss ($-4{\cdot}4$ to $-6{\cdot}1$; Vellmer 1992).

3.4. Geochemical changes during the granulite-facies metamorphism

Most of the felsic Moldanubian granulites show undepleted LILE contents. Unlike in many other granulite terrains around the world (Rudnick *et al.* 1985; Rudnick & Presper 1990), prograde metamorphism appears to have occurred largely isochemically. This implies that high-pressure melting in felsic granulites was widely suppressed and small-fraction partial melts were (mostly?) incapable of segregating from their source.

Nevertheless, a chemical feature that needs explanation is the distinct U, Th and Cs depletion compared to the average upper crust or the otherwise geochemically similar Saxothuringian meta-igneous rocks. As proposed already by Fiala *et al.* (1987a, b) and Vellmer (1992), these elements could have been stripped out by fluid or small-scale melt during the high-pressure metamorphism. The strong Cs impoverishment (Vellmer 1992; Becker *et al.* 1999) is a well-known feature of many granulite-facies terrains (Rudnick & Taylor 1987; Rudnick & Presper 1990). This effect has been ascribed to the fractionation between K-feldspar and melt/fluid (Rudnick & Presper 1990; Hart & Reid 1991; Morgan & London 2003).

Minor fluid or melt loss is compatible with the preservation of the nearly pristine high-pressure mineralogy (garnet + kyanite + rutile) in many granulites. Otherwise, as argued by Powell & Downes (1990), Clemens & Droop (1998), Brown (2002) and White & Powell (2002), the retention of the aqueous fluids (or hydrous melt) should have, upon retrograde cooling, led to the back-reaction of alumosilicates and garnet to biotite and muscovite.

3.4.1. Role of fluids. On the basis of the experimental work by Keppler & Wyllie (1990), a major role for Cl- and CO_2–rich fluids can be probably discounted, since U and Th are almost equally depleted but undecoupled (Fig. 10a). Interestingly, rare glimmerite veins, occurring in Austrian peridotite bodies adjacent to granulite massifs, show high concentrations of LILE, U and Th, accompanied by low Rb/Cs and K/Rb ratios (Becker *et al.* 1999). These contain Sr and Nd with isotopic compositions which overlap with the granulites at c. 340 Ma. According to Becker *et al.* (1999), high Zn and F, low Sr, negative Eu anomaly, as well as low HFSE, all point to origin of glimmerites through crystallisation from a F-bearing aqueous-carbonic fluid released during the high-temperature, high-pressure devolatilisation in the granulite massifs.

3.4.2. Melt loss. A melt would be two orders of magnitude more efficient in partitioning U and Th than a pure H_2O-rich fluid (Keppler & Wyllie 1990). The present authors have constrained potential metamorphic assemblages, partial melting and volatile regimes by pressure–temperature and temperature–X(H_2O) pseudosections (Powell *et al.* 1998), calculated using the internally consistent thermodynamic dataset (Holland & Powell 1998) and the associated melt model (Holland & Powell 2001; White *et al.* 2001). Solid-solution models were taken from the following sources: (feldspar) Fuhrman & Lindsley (1988) and Wen & Nekvasil (1994); (biotite) Powell & Holland (1999); (garnet and phengite) Coggon & Holland (2002) and (clinopyroxene) Holland & Powell (1996); hydrous cordierite was considered ideal. Computations were performed with the Perplex computer software (Connolly & Petrini 2002) for average granulite composition of 81·03 mol.% SiO_2, 0·209 mol.% TiO_2, 8·54 mol.% Al_2O_3, 1·89 mol.% FeO_t, 0·868 mol.% MgO, 1·113 mol.% CaO, 3·17 mol.% Na_2O and 3·19 mol.% K_2O, at variable H_2O concentrations.

The representative felsic granulite has a relatively monotonous and modally insensitive pre-peak mineral assemblage: biotite (4·5–5·5 vol.%), white mica (5·4–6·4 vol.%), plagioclase (27·7–28·3 vol.%), K-feldspar (22·2–23·0 vol.%) and quartz (37·7–38·1 vol.%) with accessory garnet (0·2–0·3 vol.%), ilmenite or rutile (0·1–0·3 vol.%). The bulk-rock H_2O budget, prior to the mica dehydration reactions, is 0·49–0·51 wt.%. During prograde burial, breakdown of micas leads to the assemblage feldspar + garnet + kyanite + quartz + rutile + granitic melt. However, the fertility of the protolith is severely restricted by the low modal abundances of micas. The calculated isopleths of melt fraction, produced by dehydration melting, are illustrated in the pressure–temperature space together with the available P–T estimates for the Moldanubian granulites (Fig. 13).

The overall *dP/dT* slope of the melt-fraction isopleths is approximately parallel to the fluid-absent haplogranitic melting curve (Holtz *et al.* 2001) and shifts to progressively higher temperatures with increasing pressure. This explains both the infertility of the granulites at the presumed metamorphic peak conditions and the irregular distribution of the melt-fraction isopleths (Fig. 13). Initial melting, limited by mica availability, does not exceed 10 vol.% for any of the granulite decompression paths. Rapid increase of the melt fraction would occur only at about 1100 °C and 1 GPa and 1200 °C and 2 GPa because of the closely approaching fluid-absent haplogranitic solidus.

These results are in a very good agreement with estimated melt production at the peak P–T conditions (1000 °C and 1·6 GPa) of 5–15 vol.% (Roberts & Finger 1997). Such low degrees of partial melting suggest that granulites have undergone very limited anatexis. It can be argued that the partial melt would start to escape only following the breakdown of the framework of residual minerals, after reaching the Melt Escape Threshold (>20–25 vol.% melt) (Vigneresse *et al.* 1996; Sawyer 1999). However, some authors have suggested that as little as 5 vol.% of melt can be efficiently separated if deformation is involved (Rutter & Neumann 1995). On this basis, some degree of melt loss from the felsic granulites cannot be ruled out.

The present authors have estimated the effect of extraction of a small-degree partial melt on the LILE budget of an average felsic granulite. To their knowledge, no suitable K_ds exist for hypersolvus feldspar, and hence, they have employed

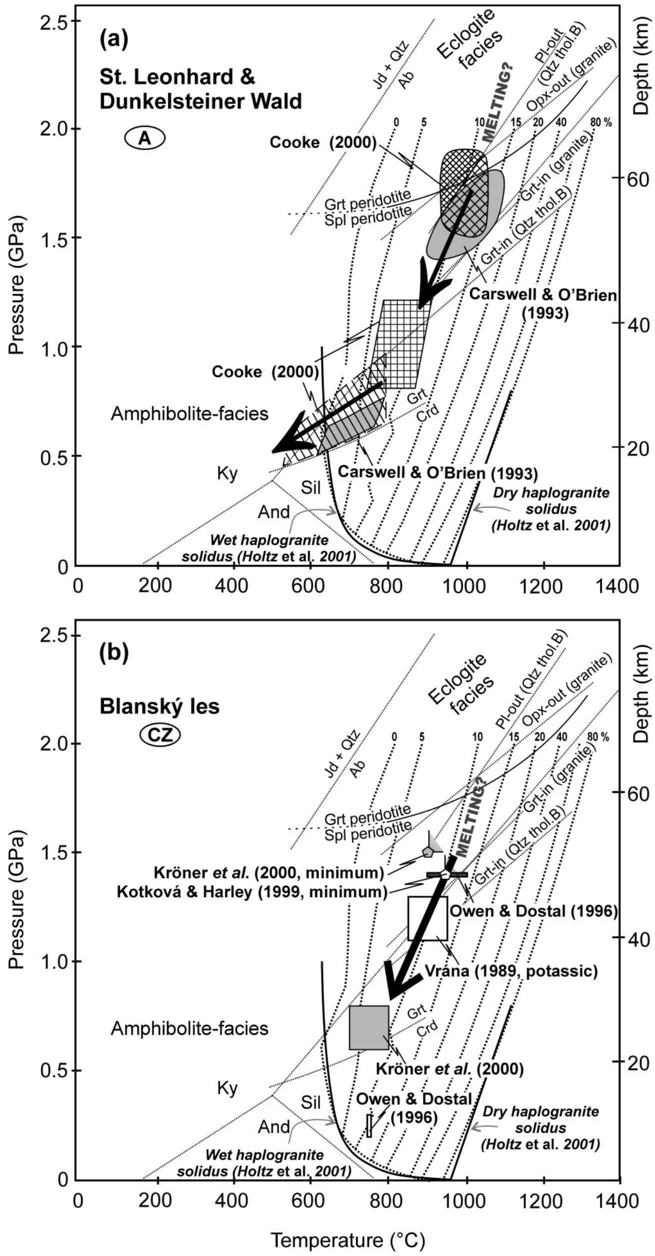

Figure 13 Pressure–temperature diagram for the Moldanubian granulites (including both felsic and mafic types). The water-saturated and dry haplogranite solidi are from Holtz *et al.* (2001), and the petrogenetic grid from Cooke (2000). Data sources: (a) Lower Austria (Carswell & O'Brien 1993; Cooke 2000); and (b) Blanský les Massif, Southern Bohemia (Vrána 1989; Owen & Dostal 1996; Kotková & Harley 1999; Kröner *et al.* 2000). Isopleths represent the volume fraction of partial melt produced by mica dehydration melting (see text for calculation procedure).

Element	K_d (Kfs/rhyolite)	References*	Depletion factor† (5 % melting)	Depletion factor† (10 % melting)
Cs	0.002-0.061	1-3	0.02-0.36	0.01-0.22
Rb	0.11-0.74	1-4	0.51-0.91	0.34-0.83
Sr	2-22.1	3-4	0.99-1.05	0.99-1.10
Ba	1-44	1-4	0.94-1.05	0.89-1.10

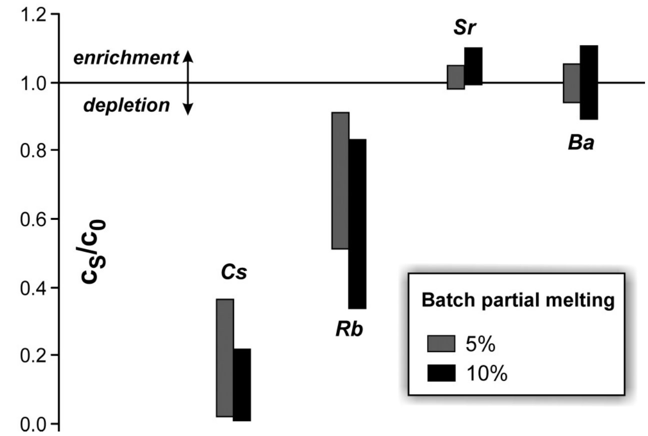

Figure 14 Modelling effects of slight melt loss on LILE budget (c_S/c_0=change in element concentration of the residue relative to the original source composition). Batch melting, estimated residual mineralogy 45% alkali feldspar+45% Qtz+10% Grt. Distribution coefficients K-feldspar/rhyolitic melt are from: (1) Leeman & Phelps (1981), (2) Mahood & Hildreth (1983), (3) Stix & Gorton (1990) and (4) Ewart & Griffin (1994); garnet and quartz were assumed to contain no appreciable concentrations of the modelled elements. †The concentration in the residue=concentration in the source × depletion factor.

those for K-feldspar/rhyolitic melt (Fig. 14). Taking into account the considerable scatter in the K_d values, some 0·4–12·5% batch melting would be sufficient to deplete Cs from 4·8 p.p.m. (an average Fichtelgebirge orthogneiss, see preceding section) to 0·9 (felsic granulite). On the other hand, Sr and Ba concentrations would be nearly unaffected by comparably small degrees of partial melting (Fig. 14). Even for Rb, published K_ds (Leeman & Phelps 1981; Mahood & Hildreth 1983; Stix & Gorton 1990; Ewart & Griffin 1994) suggest moderately incompatible behaviour. Thus, Rb depletion would be imperceptible at lower degrees of melting. In this respect, it is worth noting that several samples seem to have slight excess radiogenic Sr (Fig. 11b) compatible with a minor Rb loss in course of the Variscan high-grade metamorphism. However, the real significance of this observation might be doubtful since these are the most felsic samples, with $^{87}Rb/^{86}Sr$ ratios sensitive to small errors in Sr determinations (Jahn *et al.* 2000).

If a protolith composition similar to Fichtelgebirge meta-igneous rocks (Siebel *et al.* 1997) is assumed, depletion in a similar fashion to Cs would need to be invoked for Th (11→2·1 p.p.m.) and U (9→0·95 p.p.m.) as well. Rudnick & Presper (1990) demonstrated that both elements can strongly partition into melt/fluid, especially if the role of U–Th bearing accessory phases was limited.

In summary, additional work is needed to shed more light on the mechanism of the striking U, Th and Cs depletion in the Moldanubian granulites. Given the results of the thermodynamic modelling, only a small melt fraction was likely to be present at the high-grade metamorphic peak and any melt loss would be rather unlikely. This still would have little effect on the whole-rock geochemical signatures, including the LIL elements other than Cs.

3.5. Geotectonic setting

Taken together, the Ordovician meta-igneous crust with a composition resembling that preserved in the Fichtelgebirge could have yielded rocks similar to felsic Moldanubian granulites upon subduction to depths of at least 50–60 km (O'Brien 2000). This process was accompanied by surprisingly negligible geochemical modification.

The most appropriate geotectonic setting for the Ordovician igneous activity widespread throughout Variscan Europe (e.g. Pin & Marini, 1993) seems to be an intracontinental or back-arc rift, where crustal thinning and heating by upwelling mantle could have led to melting of various crustal lithologies (Siebel *et al.* 1997; Linnemann *et al.* 2000; Franke 2000). The high Nd model ages testify that the source of the Ordovician igneous rocks parental to the felsic granulites could have been a pre-existing crust, possibly Cadomian I-type granodiorites or geochemically undistinguishable immature psammitic (meta-) sediments. Such material was readily available at the active Gondwana margin, from which the Moldanubian and Saxothuringian crust were derived (e.g. Nance & Murphy 1996; Linnemann *et al*. 2000; Winchester 2002).

4. Conclusions

1. The majority of the felsic Moldanubian granulites do not seem to represent separated, high-pressure Variscan syn-collisional melts. The main arguments include:
   - concentrations of Zr far too low compared to calculated saturation levels at $T \geq 900\,°C$, coupled with presence of significant, mainly Ordovician, zircon inheritance;
   - consistently low zircon and monazite saturation temperatures;
   - preservation of Ordovician/Silurian whole-rock and thin slab Rb–Sr ages; and
   - high HREE+Y, ruling out the presence of larger amounts of garnet in the residue.
2. Instead, the felsic granulites are interpreted as metamorphosed (partly anatectic) equivalents of an older igneous complex.
3. Protoliths for the felsic granulites and Gföhl gneiss were compositionally similar to some meta-igneous rocks preserved in the Saxothuringian Zone; for instance, Ordovician–Silurian metarhyolites–orthogneisses from Fichtelgebirge. These were probably rift-related rocks which could have originated by remelting of Cadomian crust, such as I-type granodiorites or immature psammitic (meta-) sediments.
4. Trends of rapidly falling Sr, Ba, Zr, LREE and P with slightly increasing Rb and SiO_2 characteristic of evolved granites crystallising K-feldspar>plagioclase+biotite>zircon+monazite seem to have been inherited from the igneous stage.
5. During the Variscan orogeny, the Early Paleozoic acid meta-igneous crust was subducted to depths of at least 50–60 km. A moderately small melt fraction was presumably present during much of the metamorphic history. Most of the geochemical signature was conserved despite the HP–HT metamorphism, which led only to the loss of H_2O, U, Th and Cs.
6. The ultimate mechanism of the U, Th and Cs depletion is uncertain. It could have been caused by fluid loss or by removal of a small-degree (<10%) partial melt. While loss of such a small-degree partial melt would strip off most of the Cs, and presumably also U and Th, other LILE (in particular Rb, Ba and Sr) would be little affected.
7. In favourable cases (e.g. the absence of fluids and the short duration of the event), the Rb–Sr whole-rock system might be rather resistant to resetting during granulite-facies metamorphism.
8. Deep burial and HP–HT metamorphism of infertile crustal lithologies, such as felsic meta-igneous rocks, does not need to cause extensive partial melting. The resulting granulites would have largely undepleted geochemical signature and preserve much of geochemical information from their protolith.

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6. Appendix 1. Petrography of the studied samples

Popp-D/L: *Quarry of Natursteine Popp E of Ramsau, St. Leonhard Massif (GR 6892 3894)*
Ost-D/L: *Roadside exposure at Osterberg, on the Pöchlarn–Zelking road, Pöchlarn–Wieselburg Massif (GR 6873 3399)*

Sample pairs *Ost-L/D* and *Popp-L/D* represent adjacent lighter and darker granulite bands. Characteristic for the darker facies is the presence of considerable amounts of small biotite flakes (3–10%), which often define a weak foliation within the fine-grained (<0·1 mm) quartz–orthoclase–plagioclase matrix. Additionally, there is always garnet (<4 mm) and often some kyanite (<1 mm) in the darker bands.

The lighter varieties are typical felsic, massive granulites with generally very little biotite. They contain c. 2–5% garnet (0·2–4 mm across), the bigger of which often enclose inclusions of relict mesoperthite and sometimes show fine, exsolved rutile needles. The fine-grained (<0·1 mm) matrix consists of roughly 30–40% quartz, 25–30% K-feldspar and 25–30% sodic plagioclase. A few single, elongate kyanite crystals up to 4 mm are present, as well as accessory brown euhedral rutile (<1 mm).

Ysp-D/L: *Roadside quarry 'an der Gleisen', Yspertal (GR 6530 3422)*

The granulite at this locality is well banded and associated with a serpentinite body. Although sample *Ysp-D* is of clearly darker appearance than *Ysp-L*, there is no large compositional difference between the two. The former contains a small percentage of biotite (small, evenly distributed flakes) and has a finer grained, mylonitic groundmass. On the other hand, *Ysp-L* is a typical felsic granulite with rare garnets set in a quartzo–feldspathic matrix. Both contain some kyanite grains.

Du: *Small abandoned quarry west of Lindwurmkreuz, Dunkelsteiner Wald (GR 6903 3549)*

Sample *Du* is a dark-coloured granulite with a biotite-poor, charnockitic mineral paragenesis (Opx+Cpx+Grt) coming from a megaboudin c. 1–1·5 km long and up to 9 m thick. The sample contains large amounts of plagioclase (40%) and quartz (30%) and some K-feldspar (10%). All these minerals form a fine-grained mosaic.

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St: *Road cut near Steinaweg, Dunkelsteiner Wald (GR 6953 3593)*

Leucocratic, fine- to medium-grained, quartz-rich felsic granulite with alternating Grt-free and Grt-bearing layers. Within the weakly folded matrix, numerous dark pink/red garnets are visible, up to 5 mm in diameter. Particularly striking features are the widespread pale blue kyanite grains up to 4 mm in length, often aligned to the foliation defined by quartz ribbons and sparse biotite. Compared to most of the acidic rocks found in the Austrian granulite occurrences, the peak metamorphic assemblage of Grt + Ky + hypersolvus feldspar + Qtz is remarkably preserved in this sample, a reason why it was chosen for a detailed geochronological study (Friedl *et al.* 2003, 2004). Within the matrix, abundant mesoperthitic K-feldspar porphyroclasts up to 3 mm across are visible with only minor recrystallisation around their margins to granoblastic fine-grained K-feldspar and subordinate plagioclase. Accessory phases are rutile, apatite, zircon and ilmenite.

1059–1078: *Borehole H-1 at Holubov, 8 km N of Český Krumlov, Blanský les Massif*

The borehole drilled in 1965 penetrated mainly granulitic rocks, with leucogranulites prevailing over more mafic types. The rest was accounted for by several tectonic slices of ultrabasic rocks and the whole assemblage was cut by infrequent dykes of Variscan aplites–pegmatites. While systematic petrographic account (including 50 modal analyses of stained granulite samples) was published by Kodym *et al.* (1978), Fediuková (1978) focused solely on mineralogy and chemistry of mafic phases.

The felsic granulites with peak metamorphic assemblage hypersolvus feldspar + quartz + garnet + rutile ± kyanite show effects of variable retrogression and mylonitic deformation. These processes resulted mainly in breakdown of garnet to a secondary biotite, as well as recrystallisation of the mesoperthite into a K-feldspar and plagioclase mosaic. The present samples coming from near the bottom of the borehole (c. 1100–1400 m; Fig. 15) were selected so that they would be as little affected as possible, all having inequigranular granoblastic texture with only indistinct plane-parallel fabric. However, differences do exist. Some (1074 and 1076) contain practically no biotite, still preserving large mesoperthite augen. The others were nearly completely recrystallised to an equilibrated mosaic of two feldspars and quartz (1062), and contain a large proportion of secondary biotite, sometimes altered to chlorite (1064 and 1078). The replacement of garnet porphyroclasts by lath-shaped biotite started at the margins and proceeded along cracks (1064). Rarely tiny flakes concentrated into thin stripes within the matrix, defining a weak foliation (1062).

In all cases, the mineral assemblage is dominated by quartz, K-feldspar (mesoperthitic microcline) and plagioclase ($\sim An_{10-20}$, in less felsic types ranging to $\sim An_{30}$). These minerals occur in about equal proportions in more felsic types, but the plagioclase content increases at the expense of Kfs to c. 40 vol.% in the less felsic ones (1059). Garnet porphyroclasts (up to 7 vol.%), which are often poikilitic, are conspicuous enclosing mainly subhedral perthite grains. The contents of garnet and rutile drop sharply from less to more silica-rich samples. The rutile forms both euhedral prismatic crystals in the matrix as well as, in less felsic samples (1059), a plethora of unmixed oriented needles in the garnet porphyroclasts (e.g. Zhang *et al.* 2003). In the two most felsic granulites, there were rather abundant kyanite porphyroclasts (1074 and 1078: up to 1 vol.%), some elongate and subhedral, others lobate, enclosing perthite grains. On the other hand, the least felsic sample, 1059, contained kyanite with hercynite coronas, the host mineral having been sometimes completely consumed. In all cases, common accessories were ilmenite, zircon and apatite.

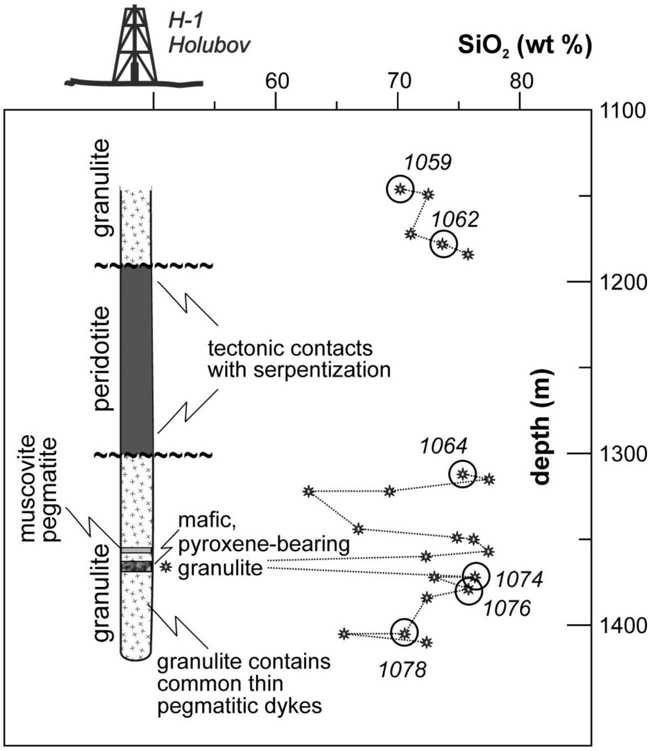

Figure 15 Simplified petrographic profile for the sampled part of borehole H-1, Holubov, Blanský les Massif. The location of the samples mentioned in the text as well as SiO_2 variations with depth are shown. In part after Kodym *et al.* (1978).

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