From orthogneiss to migmatite: Geochemical assessment of the melt infiltration model in the Gföhl Unit (Moldanubian Zone, Bohemian Massif)

P. Hasalová a,b,⁎, V. Janoušek a,c, K. Schulmann b, P. Štípská b, V. Erban c

a Institute of Petrology and Structural Geology, Charles University, Albertov 6, 128 43 Prague 2, Czech Republic
b Université Louis Pasteur, CGS/EOST, UMR 7517, 1 rue Blessig, Strasbourg 67084, France
c Czech Geological Survey, Klárov 3, 118 21 Prague 1, Czech Republic

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Abstract

The Gföhl Unit is the largest migmatite terrain of the Variscan orogenic root domain in Europe. Its genesis has been until now attributed to variable degrees of in situ partial melting. In the Rokytná Complex (Gföhl Unit, Czech Republic) there is a well-preserved sequence documenting the entire migmatitization process on both outcrop and regional scales. The sequence starts with (i) banded orthogneiss with distinctly separated monomineralic layers, continuing through (ii) migmatitic mylonitic gneiss, (iii) schlieren migmatite characterised by disappearance of monomineralic layering and finally to (iv) felsic nebulitic migmatite with no relics of the original banding.

While each type of migmatite shows a distinct whole-rock geochemical and Sr–Nd isotopic fingerprint, the whole sequence evolves along regular, more or less smooth trends for most of the elements. Possible mechanisms which could account for such a variation are that the individual migmatite types (i) are genetically unrelated, (ii) originated by equilibrium melting of a single protolith, (iii) formed by disequilibrium melting (with or without a small-scale melt movement) or (iv) were generated by melt infiltration from external source. The first scenario is not in agreement with the field observations and chemistry of the orthogneisses/migmatites. Neither of the remaining hypotheses can be ruled out convincingly solely on whole-rock geochemical grounds. However in light of previously obtained structural, petrologic and microstructural data, this sequence can be interpreted as a result of a process in which the banded orthogneiss was pervasively, along grain boundaries, penetrated by felsic melt derived from an external source.

In terms of this melt infiltration model the individual migmatites can be explained by different degrees of equilibration between the bulk rock and the passing melt. The melt infiltration can be modelled as an open-system process, characterised by changes of the total mass/volume and accompanied by gains/losses in many of the major- and trace elements. The modelling of the mass balance resulted in identification of a component added by a heterogeneous nucleation of feldspars, quartz and apatite from the passing melt. This is in line with the observed presence of new albitic plagioclase, K-feldspar and quartz coatings as well as resorption of relict feldspars. At the most advanced stages (schlieren and nebulitic migmatites) the whole-rock trace-element geochemical variations document an increasing role for fractional crystallization of the K-feldspar and minor plagioclase, with accessory amounts of monazite, zircon and apatite.

⁎ Corresponding author. Université Louis Pasteur, CGS/EOST, UMR 7517, 1 rue Blessig, Strasbourg 67084, France.
E-mail address: hasalovap@seznam.cz (P. Hasalová).

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The penetrating melt was probably (leuco-) granitic, poor in mafic components, Rb rich, with low Sr, Ba, LREE, Zr, U and Th contents. It probably originated by partial melting of micaceous quartzo-feldspathic rocks.

If true and the studied migmatites indeed originated by a progressive melt infiltration into a single protolith resembling the banded orthogneiss, this until now underappreciated process would have profound implications regarding rheology and chemical development of anatectic regions in collisional orogens.

**Keywords:** Melt infiltration; Porous flow; Migmatites; Whole-rock geochemistry; Mass balance; Bohemian Massif (Gföhl Unit)

### 1. Introduction

The melt extraction from lower crustal rocks and its further transport to higher crustal levels are key issues of granite petrogenesis. Generally the melt fraction produced during anatexis is rather small; it is originally concentrated mainly at grain boundaries and along microfractures (Rushmer, 1995; Sawyer, 2000; Guernina and Sawyer, 2003). The mechanism of the melt segregation is still a controversial issue. It may be drained from the grain boundaries to melt-assisted sites, networks of leucosome-filled structures or dykes (see Lister and Kerr, 1991; Petford et al., 1994 for review) responsible for its further transport. The other, arguably more popular model involves pervasive flow through a porous space or fracture network (McKenzie, 1984; Wickham, 1987; Weinberg, 1999) and is strongly controlled by the permeability of the system.

Regardless of its exact mechanism, the segregation will be governed by the melt density and viscosity, which in turn reflect its composition, volatile contents and temperature (Brown et al., 1995; Petford et al., 2000). Important factor is also the thermal and mechanical state of the crust surrounding the melting zone (Weinberg and Searle, 1998). The efficiency of the melt extraction can be further boosted by active deformation in the source (McKenzie, 1984). At the grain-scale, the transport is controlled mainly by the melt geometry (wetting angle) (von Bargen and Waff, 1986) and its amount (Rushmer, 1995; Vigneresse et al., 1996). The grain-scale movement is possible as soon as the melt forms an interconnected network. Even when the wetting angle is low (von Bargen and Waff, 1986), an important prerequisite for the operation of pervasive porous flow is the lower solidus temperature of the country-rock compared to that of the penetrating melt.

The other passionately discussed problem is the further transport of the granitic melts. As an alternative to dyking (Petford et al., 1994, 2000), mesoscale pervasive magma migration, either channelized using the country-rock anisotropies (Collins and Sawyer, 1996; Weinberg and Searle, 1998; Brown and Solar, 1999; Vanderhaeghe, 1999; Weinberg, 1999; Leitch and Weinberg, 2002; Olsen et al., 2004), or penetrative through the whole-rock volume (Hasalová et al., in press-a) has been proposed for hot low-viscosity crustal rocks. For the latter model, the deformation is essential.

The whole-rock compositions of the migmatitic rocks, and the trace elements with radiogenic isotopes in particular, can be useful in constraining the nature of the source, residual mineralogy, as well as degree, mechanism and timescales of the crustal melting (see Whittington and Treloar, 2002 for review). For instance the water-saturated or dehydration melting driven by decomposition of individual hydrous phases (muscovite, biotite or amphibole) will each leave its own geochemical imprint. Moreover, the chemistry of the melt will be substantially different depending on whether its bulk equilibrated with the restite (batch melting) or the extraction was in small increments with no mutual homogenization (fractional melting). Apart from these two models for the equilibrium melting, variable degrees of disequilibria seem common-place. The disequilibrium melting takes over especially in collisional orogens, when the melt extraction was accelerated by deformation or the accessory phases were shielded by major phases in the source not participating in the melting reaction (e.g. Watt and Harley, 1993; Nabelek and Glascock, 1995; Bea, 1996; Harris and Ayres, 1998). Last but not least, the protolith is often not exposed, and thus its exact nature and composition are unknown.

The ideal elements for modelling of partial melting are those residing in the main rock-forming minerals. This is the case of LILE, whose main reservoirs are feldspars and micas (e.g. Harris and Inger, 1992). On the other hand, several trace elements are incompatible in the main rock-forming minerals but form essential structural components (ESC) in accessory phases (Hanson and Langmuir, 1978). The dissolution of such an accessory mineral at the contact with the melt will continue until either the appropriate saturation level is reached or the source is exhausted (Watson and Harrison, 1984). The saturation models are defined for the main accessories of interest in granitic magmas, zircon (Watson and Harrison, 1983), apatite...
(Harrison and Watson, 1984) and monazite (Montel 1993). In reality, the dissolution of the accessory minerals is governed by a number of factors, including the temperature, melt composition, crystal size, the water activity in the magma, diffusivity, absolute solubility of the ESC coupled with the degree of the melt undersaturation and its distribution within the rock volume.

Taken together, balancing the element fluxes during partial melting is not an easy task, as the granitic melts can result from a complex interplay of a number of processes, operative on various scales (e.g. Barbe et al., 1996). For open systems, a promising approach are mass transfer calculations, which take into account the changes in the total mass or volume of the system concerned (Grant, 1986; Olsen and Grant, 1991).

The Gföhl Unit of the Moldanubian Zone, Czech Republic, offers an exceptional opportunity to study the origin of migmatitic rocks, because it is possible to observe the spatial and structural relationships between the migmatites with different proportion of melt directly in the field. This means that one is able, for rocks with the same orthogneiss protolith and increasing degrees of migmatization, to investigate gradual changes in modal proportions of the rock-forming minerals, their chemistry, microstructures and whole-rock geochemical signature.

The present contribution aims to describe the major- and trace-element as well as Sr–Nd isotopic whole-rock geochemical variation in the individual gneiss and migmatite types in this unit. This, together with information on field relations, petrography and textures is used to evaluate the possible genetic hypotheses. Using the mass transfer calculations and other arguments we test the feasibility of the possible genetic scenarios. Included among them is the newly defined melt infiltration model (Hasalová et al., in press-a,b), in whose terms the various types of the Gföhl gneisses and migmatites were interpreted by a process in which the banded orthogneiss was pervasively, along grain boundaries, penetrated by felsic melt derived from an external source. Such a melt infiltration model would have potentially large consequences for generation of granitic magmas and crustal rheology in collisional orogens.

2. Regional setting

The Moldanubian Zone represents the orogenic root domain of the Bohemian Massif, which developed during the Variscan collision in the Devonian to Carboniferous times (Fig. 1a) (Matte et al., 1990; Dallmeyer et al., 1995; Schulmann et al., 2005 and references therein). At its eastern extremity, the Moldanubian rocks were thrust over the Cadomian Brunia basement along a so-called Moldanubian Thrust. The associated deformation produced a crustal-scale shear zone in the basement (the Moravian Zone, MZ), which crops out as three NE–SW trending tectonic windows (Suess, 1912; Urban, 1992; Schulmann et al., 1994; Fritz et al., 1996) (Fig. 1b).

The Moldanubian Zone is a tectonic assemblage of medium- to high-grade metamorphic rocks, intruded by numerous large, mostly Carboniferous plutons (Finger et al., 1997). The Moldanubian sequence has been subdivided into mainly metasedimentary and gneissic middle crustal rocks of Proterozoic and Lower Palaeozoic protolith ages assigned to the Drosendorf Unit by some authors (e.g. Tollmann, 1982; Franke, 2000) and the structurally upper and higher grade Gföhl Unit (Fuchs and Matura, 1976; Petrakakis, 1997 and references therein). The middle crustal rock assemblage consists of the Monotonous Series (mainly migmatitic Grt–Bt–Sil paragneiss with minor orthogneiss and amphibolite) and the Varied Series (paragneiss with intercalations of amphibolite, calc-silicate gneiss, marble, quartzite and graphite schist).

The Gföhl Unit is dominated by high-grade felsic gneiss and migmatite and layered migmatitic amphibolites. It also includes a high-pressure felsic Grt–Ky–Kfs granulite, which encloses bodies of garnet and spinel peridotites, pyroxenites and eclogites (Carswell, 1991; O’Brien and Carswell, 1993; Medaris et al., 1995). The exhumation of high-grade rocks and their juxtaposition to the middle crust within the orogenic root has been recently attributed to vertical extrusion of orogenic lower crust followed by horizontal channelized spreading at middle crustal levels, associated with retrogression and widespread melting (Schulmann et al., 2005; Tajčmanová et al., 2006).

The Gföhl gneiss and migmatite vary from banded orthogneiss without signs of melting towards migmatite with isotropic (nebulitic) structure without traces of earlier fabrics. The protolith is considered to be granitic (Dudek et al., 1974; Matějovská, 1975). The U–Pb SHRIMP dating showed that the protolith was Early Palaeozoic (488 ± 6Ma: Friedl et al., 2004). The age of Variscan metamorphism of the Gföhl gneiss has been estimated at 341 ± 4 and 337 ± 3Ma (conventional U–Pb ages for zircon and monazite, respectively: van Breemen et al., 1982) in Moravia and at 340 ± 10Ma (SHRIMP ages of outer growth zones of zircons: Friedl et al., 1998) together with 339.9 ± 0.9Ma (U–Pb monazite: Friedl et al., 1994) in Austria.

The Gföhl gneisses and HP felsic granulites show in many places intimate mutual association (Dudek et al., 1974; O’Brien and Rötzler, 2003), which led some workers to propose that the gneisses may represent
retrogressed granulites (Cooke and O’Brien, 2001). The peak conditions in the granulite bodies, estimated by many workers at 1000°C and 15kbar (see O’Brien, 2006 for review), have been recently constrained by others to c. 800–900°C and 18kbar (Medaris et al., 1998; Štípská and Powell, 2005; Tajčmanová et al., 2006). The peak pressure estimates have been interpreted as metamorphic conditions of the vertical extrusional fabric at c. 340Ma (Schulmann et al., 2005). The estimated pressure of re-equilibration associated with flat fabric that originated during the lateral spreading varies from 10kbar (Štípská et al., 2004), through 7kbar (Racek et al., 2006) to 4.0kbar (Tajčmanová et al., 2006), at temperatures between 700 and 800°C. In Austria the metamorphic conditions of the Gföhl gneiss and migmatite were estimated at 750°C.

Fig. 1. (a) Location of the Gföhl Unit in the context of the Central European Variscides. (b) Geological map of the eastern part of the Bohemian Massif with location of the studied area (outlined). (c) Schematic map of the Rokytná Complex, showing the distribution of whole-rock samples collected in the course of this work.
and c. 7kbar (Petrakakis, 1986a,b), and nearly identical values were obtained by Owen and Dostal (1996) from western Moravia.

3. Field relations and migmatite occurrence

The studied area (Fig. 1b, c), the so-called *Rokytná Complex* (Svoboda et al., 1966), is situated at the eastern extremity of the Gföhl Unit close to its contact with the Moravian Zone. It is bound by the rocks of the Moravian Zone (MZ) in the East, by the Náměšť granulite body (NGB) in the North and by the Třebíč durbachite Massif (TM) in the West (Fig. 1b). The main rock types in the Rokytná Complex are high-grade orthogneisses and migmatites, enclosing minor bodies of amphibolites, granulites and paragneisses (Matějovská, 1975). The migmatites of the Rokytná Complex are texturally highly variable. In order to refer to individual rock types, the current paper employs migmatite terminology of Mehnert (1971), based on their macroscopic appearance.

Two major deformation events were recorded in this gneiss-migmatite complex (Urban, 1992; Schulmann et al., 1994; Hasalová et al., in press-a). The D₁ event most likely corresponded to early stages of lower crust exhumation, triggered by shortening of the thickened orogenic root (Schulmann et al., 2005). The D₂ shearing

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**Fig. 2.** Sketch showing the individual gneiss and migmatite types and their relationships within an outcrop (the width of the figure is 5m; modified after Hasalová et al., in press-a). Banded orthogneiss with distinct S₁ compositional layering (a) is folded and transposed (b) to the stromatitic migmatite (c) that passes gradually to the schlieren migmatite (d) and finally to the completely isotropic nebulitic migmatite with no relics of gneissosity (e). Shown are typical macrophotographs of each rock type.
has been attributed to horizontal spreading of lower crust at mid-crustal levels (e.g. Tajčmanová et al., 2006).

The deformation phase D1 resulted in formation of steep, west dipping solid-state foliation S1, represented by compositional layering in the banded orthogneiss (Fig. 2a). The D2 deformation led to the development of a large crustal-scale shear zone and was associated with reworking and folding of S1 compositional layering that is locally preserved in elongated relict domains (Fig. 2). These relict domains with gently folded S1 fabric are surrounded by highly deformed zones with tightly folded S1 fabric. Locally the S1 fabric is completely transposed into the new S2 foliation dipping gently to the SW. The resulting composite S1–2 fabric is characterized by banded structure with polymineral K-feldspar- and plagioclase-rich domains resembling stromatitic migmatite (Fig. 2c).

Detailed field study revealed that, with increasing degree of deformation, the stromatitic migmatite gradually passes into more isotropic schlieren migmatite (Fig. 2d) still containing rootless folds modifying the relics of the S1 fabric. This rock type is alternating with irregular bodies or elongated lenses of felsic fine-grained nebulitic migmatite (Fig. 2e).

Such migmatite variations, which have originated through intense D2 deformation superimposed on early steep anisotropy, can be identified both on the outcrop and the regional scales. In the studied area, stromatitic migmatites generally prevail over schlieren migmatites; the banded orthogneisses and nebulitic migmatites are subordinate. Macroscopically visible melt accumulations or granitic veins parallel to S2 and tensional gashes perpendicular to S2 are locally present.

3.1. Definition of individual rock types

Hasalová et al. (in press-a) showed the intimate relationship between different migmatite types suggesting that they all originated from the same protolith and that the banded orthogneiss and nebulitic migmatite can be considered as end-members of a continuous structural evolution.

All the studied samples contain stable mineral assemblage Pl + Kfs + Qtz + Bt ± Grt ± Sil; common accessory phases are apatite, monazite, zircon and xenotime. The modal proportion of feldspars remains in all rock types nearly the same, only the quartz shows a marked increase. On the other hand, the biotite and garnet contents decrease nearly the same, only the quartz shows a marked increase. The modal proportion of feldspars remains in all rock types

<table>
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<tr>
<th>Mineral proportions (wt.%)</th>
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<tbody>
<tr>
<td>Qtz</td>
</tr>
<tr>
<td>Bt</td>
</tr>
<tr>
<td>Grt</td>
</tr>
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</table>

Grt garnet; Bt biotite; Pl plagioclase; Kfs K-feldspar; Qtz quartz; $X_{Fe}$ = Fe(Fe + Mg); anorthite = Ca / (Ca + Na + K); Or orthoclase = K / (Ca + Na + K). Mineral proportions were obtained by the constrained least-squares method (Albarède, 1995); shown are the ranges and (in brackets) average values; $R^2$ = goodness of fit (see text for details).

The stromatitic migmatite is marked by the onset of disintegration of the original monomineral banding and is composed of plagioclase and K-feldspar aggregates with subordinate quartz. These aggregates are rimmed by biotite locally overgrown by fibrolitic sillimanite.
The schlieren migmatite is made of K-feldspar–quartz-rich and plagioclase–quartz-rich aggregates. The original banding is distinguishable only from the modal content of the mineral phase dominant in these feldspar aggregates.

The nebulitic migmatite represents the most isotropic rock type, completely lacking relics of the original gneissosity. The migmatite occurs as irregular flat bodies or elongated lenses.

3.2. Mineral chemistry and microstructures

The banded orthogneiss always contains two chemically and microstructurally distinct plagioclase populations (Table 1 and Fig. 3a–d): (i) well-equilibrated grains with straight boundaries in plagioclase aggregates (An25–30) and (ii) newly-grown plagioclase (An10–20) forming interstitial grains or thin films coating the K-feldspar or plagioclase grains. In the stromatitic migmatite, the degree of albite coating increases and both the newly formed grains (An4–10) and the relict grains (An15–25) are more sodic (Table 1). In the schlieren migmatite large, corroded relict grains (An7–15) are resorbed by new interstitial plagioclase (An1–3) (Fig. 3c). This results, in the nebulitic migmatite, in

<table>
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<th>Table 1</th>
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<tbody>
<tr>
<td>Banded orthogneiss</td>
</tr>
<tr>
<td>$X_{Fe}$ (Grt)</td>
</tr>
<tr>
<td>$X_{Fe}$ (Bt)</td>
</tr>
<tr>
<td>An (Pl relict)</td>
</tr>
<tr>
<td>An (Pl new)</td>
</tr>
<tr>
<td>Or (Kfs)</td>
</tr>
<tr>
<td>Ti (Bt) (pfu)</td>
</tr>
</tbody>
</table>

Fe = Fe/(Fe + Mg); An anorthite = Ca / (Ca + Na + K); Or orthoclase = K / (Ca + Na + K). Mineral proportions were obtained by the constrained least-squares method (Albarède, 1995); shown are the ranges and (in brackets) average values; $R^2$ = goodness of fit (see text for details).
complete overgrowths of small cuspate plagioclase (An$_{0.4}$) accompanied by K-feldspar and quartz on residual, strongly irregular plagioclase (An$_{5-10}$) grains. The newly-grown albitic plagioclase is, in agreement with Sawyer (1999, 2001) interpreted as having crystallized from the former melt.

<table>
<thead>
<tr>
<th>Mineral Position</th>
<th>Banded orthogneiss</th>
<th>Stromatitic migmatite</th>
<th>Schlieren migmatite</th>
<th>Nebulitic migmatite</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Kfs core</td>
<td>Pl core</td>
<td>Pl rim/films</td>
<td>Kfs core</td>
</tr>
<tr>
<td>Wt. %</td>
<td>(1)</td>
<td>(2)</td>
<td>(3)</td>
<td>(7)</td>
</tr>
<tr>
<td>SiO$_2$</td>
<td>64.69</td>
<td>67.48</td>
<td>61.70</td>
<td>64.37</td>
</tr>
<tr>
<td>Al$_2$O$_3$</td>
<td>18.41</td>
<td>20.16</td>
<td>24.12</td>
<td>18.55</td>
</tr>
<tr>
<td>CaO</td>
<td>0.05</td>
<td>2.83</td>
<td>5.29</td>
<td>0.04</td>
</tr>
<tr>
<td>Na$_2$O</td>
<td>2.24</td>
<td>9.63</td>
<td>8.59</td>
<td>1.56</td>
</tr>
<tr>
<td>K$_2$O</td>
<td>13.96</td>
<td>0.12</td>
<td>0.43</td>
<td>15.04</td>
</tr>
<tr>
<td>BaO</td>
<td>0.39</td>
<td>0.00</td>
<td>0.00</td>
<td>0.13</td>
</tr>
<tr>
<td>Total</td>
<td>99.73</td>
<td>100.22</td>
<td>100.12</td>
<td>99.68</td>
</tr>
</tbody>
</table>

Fig. 3. Microscopic appearance and composition of feldspars in studied samples. (a–c) BSE images showing presence of melt and important changes in the feldspar textures within the studied sequence. (a) Typical microstructure of banded orthogneisses and stromatitic migmatites: recrystallized feldspar aggregate with numerous quartz (black arrows) and plagioclase (white arrows) interstitial grains/films tracing the K-feldspar boundaries (sample PH60/B). Relict plagioclase (An$_{24-30}$) is rimmed by more albitic plagioclase (An$_{10-20}$) and K-feldspar grains are traced by thin plagioclase (An$_{10-20}$) films (white arrows). This albitic plagioclase represents a component crystallized from the partial melt. (b) Typical appearance of schlieren and nebulitic migmatites featuring irregularly shaped feldspar and quartz grains. New interstitial plagioclase (white arrows), K-feldspar and quartz (black arrows) grains are tracing most of the feldspar boundaries. These newly crystallized feldspar and quartz resorb relict feldspar grains, causing their highly irregular shapes (sample PH59/C). (c) Detail of the irregular lobate K-feldspar grain completely embayed with newly crystallized quartz and plagioclase. Relict plagioclase grains are also surrounded by new albitic plagioclase (white arrows) (sample PH90). (d) Compositional map of the Ca distribution. It shows that a less sodic plagioclase is overgrown by albitic plagioclase crystallized from the melt (sample PH90). The feldspar compositions are summarized in the table. Note a continuous decrease in the Na$_2$O contents in each of the relict and newly crystallized plagioclases throughout the studied sequence.
Composition of K-feldspar is rather uniform (Table 1), however its microstructure evolves (Fig. 3). In the banded orthogneiss, recrystallized K-feldspar grains form almost monomineral bands (Fig. 3a). In the stromatitic migmatite, the K-feldspar bands start to be disintegrated with individual grains having slightly lobate boundaries traced by interstitial quartz and plagioclase. The schlieren migmatite shows K-feldspar relics that are highly irregular and surrounded by myrmekite and newly-grown interstitial quartz, plagioclase and K-feldspar (Fig. 3c). The degree of K-feldspar corrosion is highest in the nebulitic migmatite, where the residual grains are completely overgrown by cusprate plagioclase, accompanied by new K-feldspar and quartz (Fig. 3b).

Quartz occurs as recrystallized ribbons or interstitial grains and inclusions in the feldspar aggregates in the banded orthogneiss. Stromatitic, schlieren and nebulitic migmatite show gradual disappearance of the former quartz ribbons. Instead the quartz forms irregular polycrystalline aggregates with lobate grain boundaries or interstitial grains at feldspar boundaries, where it participates in resorption of the relict feldspar.

Biotite evolves texturally from elongated flakes in layers separating the feldspar-rich aggregates (banded orthogneiss) to grains dispersed in plagioclase-rich aggregates (stromatitic migmatite) or in the matrix composed of plagioclase, quartz and K-feldspar (schlieren and nebulitic migmatites). Cusprate biotite shapes, developed mostly in the nebulitic migmatite, were interpreted by Mehnert et al. (1973) and Büsch et al. (1974) as resulting from the reaction with the melt. The biotite chemistry evolves continuously, in the sequence from the banded orthogneiss to nebulitic migmatite, towards more Fe-rich and Ti-poor compositions (Table 1). Biotite is locally overgrown by fibrolitic sillimanite.

Garnet is unzoned almandine, with remarkable systematic increase in $X_F$ from the banded orthogneiss to nebulitic migmatite (Table 1). Garnet in banded orthogneiss and stromatitic migmatite occurs as small idiomorphic grains along plagioclase and biotite layers. In schlieren and nebulitic migmatite, rare large atoll-shaped garnet appears in the matrix.

The most abundant among accessory minerals is apatite; characteristic but less common are monazite, zircon and xenotime. The Th content in monazite is continuously increasing towards the nebulitic migmatite, documenting an increasing brabantite substitution (Fig. 4 and Table 2). Monazite in the nebulitic migmatites is characterised by the highest contents of Th (up to 24wt.% ThO$_2$), with elevated U (<4wt.% UO$_2$), Ca (up to 6wt.% CaO) and Y (<2.2wt.% Y$_2$O$_3$).

Hasalová et al. (in press-a) assessed and quantified the microstructural changes from banded orthogneiss to nebulitic migmatite, finding out that the grain size of all felsic phases decreases continuously in the studied sequence. Resulting grain size distribution was interpreted to result from increasing nucleation rate and decreasing growth rate in the textural sequence. This is in accordance with the resorption of old/relict minerals and crystallization of new minerals from a melt along the feldspar boundaries. For further details on mineralogy and microstructures in the migmatite sequence the reader is referred to this publication.

4. Whole-rock geochemistry

4.1. Major elements

The studied samples are all broadly granitic in composition, as demonstrated for instance by the Q'-ANOR diagram of Streckeisen and Le Maître (1979) based on the CIPW normative mineralogy (Fig. 5). Whereas the banded orthogneisses and stromatitic migmatites correspond mostly to granite, schlieren migmatites straddle the boundary of the alkali feldspar granite domain occupied by the nebulitic migmatites. Analogously, each of the groups shows a rather restricted range of major-element compositions. However there is a tendency for systematic changes in the whole migmatite sequence (Fig. 6 and Table 3). Thus the banded orthogneiss is are the least siliceous (SiO$_2$ = 68.09–72.93wt.%) and richest in most other...
oxides (TiO₂ = 0.29 – 0.56, Al₂O₃ = 11.50 – 15.40, FeO = 1.98 – 3.39, MgO = 0.32 – 1.04, CaO = 0.95 – 1.66). Convex downward trends with an initial sharp decrease and an inflection point at SiO₂ = 73.73–74% are characteristic also of the mafic components TiO₂, FeO, MgO, and CaO. The compositional ranges observed in the most siliceous nebulitic migmatite are SiO₂ = 73.82–78.10, TiO₂ = 0.10–0.19, FeO = 1.41–1.75, MgO = 0.15–0.30 and CaO = 0.95–1.66. On the other hand, the alkalis and P₂O₅ fail to define clear, non-scattered trends. All the studied samples are peraluminous; the values of the Shand’s index A/CNK (molar Al₂O₃ / (CaO + Na₂O + K₂O), Table 3), decrease from 1.13–1.20 in the banded orthogneiss to 1.09–1.16 in the nebulitic migmatite.

4.2. Trace elements

4.2.1. Overall variation

While some of the trace elements show a more or less monotonous decrease with increasing silica (Ba, La, Eu and Zr) the trends for the others are more complicated (Fig. 7 and Table 4). The comparisons of trace-element concentrations between individual samples and their groups are facilitated by spider plots normalized to the...
average composition of the Earth’s upper crust (Taylor and McLennan, 1985) (Fig. 8). The banded orthogneiss is characterised by having concentrations comparable with the upper crustal averages. Notable is only a slight (up to c. 2.5×) enrichment in the HREE, and remarkable troughs for Nb (0.3–0.5), Sr (0.18–0.42), Tb (0.40–0.54) and a less apparent one for Ti (0.48–0.93). The character of distribution patterns for other rock types is generally similar; however many of the trace elements show large variations and an overall tendency do decrease, especially in the more siliceous schlieren and nebulitic migmatites. The troughs for Sr, Tb and Ti progressively deepen. Later in the sequence conspicuous negative Ba, Th and, lesser, U, La, Ce, Nd, Sm, Zr and Hf anomalies also appear.

4.2. LILE

There is a marked Cs depletion in all samples from the Rokytná Complex. Without exception, the Rb/Cs ratios (Table 4) are significantly higher than in the average continental crust (Rb/Cs = 30; Taylor and McLennan, 1985 or 23; Wedepohl, 1995). The Rb/Cs ratios fall into a broad interval between 40 and 183, with the ratios increasing systematically in the sequence from the banded orthogneiss to nebulitic migmatite. On the other hand, the available analyses yield K/Rb ratios of 126–350 typical of ordinary crustal rocks (i.e. ranging between c. 120 and 500: Shaw, 1968; Rudnick et al., 1985) (Table 4). Therefore there is no evidence to support a notable Rb depletion.

4.2.3. REE

Like most major and other trace elements, the total REE contents also drop sharply in the sequence from banded orthogneiss to nebulitic migmatite, spanning a broad range (∑REE 36–220ppm, Table 4). At the same time the chondrite-normalized (Boydton, 1984) patterns (Fig. 9a–d) feature a strong progressive depletion in LREE and MREE; the HREE decrease is retarded. Thus a marked drop in LaN/YbN ratios (from 7.98 to 1.57; see also Fig. 9e) is accompanied with a rotation in the LREE segment (LaN/SmN 3.39 to 2.19; Fig. 9f). With rising degree of fractionation (expressed, for instance, by SiO2 contents) there is an increase in the magnitude of the negative europium anomaly (Eu/Eu⁎ dropping from 0.62 to 0.15). Banded orthogneiss, stromatitic and schlieren migmatites are fairly homogeneous in terms of their total REE contents and distributions. However, generally subparallel patterns for the samples of the nebulitic migmatite vary much in LREE and HREE contents, as they do in the magnitude of their negative Eu anomalies.

The REE patterns show a gradually developing M-type (or concave) lanthanide tetrad effect (Masuda et al., 1987). Its magnitude expressed as a parameter TE1–3 (Irber, 1999) ranges from negligible/none (banded orthogneiss, TE1–3 = 0.98–1.01, Table 4), through noticeable in stromatitic migmatite (1.02–1.04), schlieren migmatite (1.05–1.10) to pronounced in the nebulitic migmatite (1.04–1.17). In addition, the TE1–3 values correlate positively with some variables expressing fluid/melt or crystal/melt fractionation, such as Rb/Cs (Fig. 9g) and Eu/Eu⁎ (Fig. 9h).

Most of the workers nowadays agree that the occurrence of M-type lanthanide tetrad effect is connected to fluorine-rich environments, being caused by melt/fluid (Irber, 1999; Monecke et al., 2002; Zhao et al., 2002) or melt/melt (Veksler et al., 2005) fractionation. However, the fluorine contents are constantly low (0.10–0.13) regardless of the migmatite type (Table 3) and thus this element does not seems to play a major role. In our case the reason for the formation of the tetrad effect remains enigmatic.

4.3. Zircon and monazite saturation temperatures

The calculated zircon saturation temperatures (Watson and Harrison, 1983; Hanchar and Watson, 2003) range from c. 830°C to 710°C (Table 3), and decrease monotonously in the structural sequence from banded orthogneiss to nebulitic migmatite (Fig. 10). The same trend is seen for the monazite saturation temperatures (Montel, 1993) spanning a comparable interval (c. 800°C to 680°C, Table 3).
4.4. Radiogenic isotopes

Whole-rock Sr–Nd isotopic ratios for the four main varieties of the Rokytná gneisses and migmatites, age-corrected to 340Ma, are presented in the Table 5 and Fig. 11.

The two most basic banded orthogneisses (PG7 and PH60B) show an identical, most negative \( \varepsilon^{Nd}_{340} \) value of \(-6.7\). The rest of the data set seems to contain Nd significantly more radiogenic (\( \varepsilon^{Nd}_{340} = -5.6 \) to \(-5.2\)). Altogether the data correspond to fairly uniform two-stage Nd depleted-mantle model ages (\( T_{Nd}^{DM} = 1.46–1.58\)Ga).

There are, however, more pronounced differences in the Sr isotopic compositions (Fig. 11a). The Sr isotopic ratios correlate positively with silica spanning the broad range from 0.7153 for the most basic orthogneiss PG7 to 0.7332 and 0.7347 in the most siliceous nebulites PG1 and PG3 (Fig. 11b).

The Sr–Nd data for the Rokytná Complex resemble the previously published analyses from the Austrian outcrops of the Gföhl gneiss. Frank et al. (1990), as a part of their extensive study concerned with Rb–Sr whole-rock and thin slab dating, obtained a comparably broad range of Sr isotopic compositions (\(^{87}\)Sr\(^{86}\)Sr\(_{340} = 0.7192–0.7412\)).
The only three Sr–Nd isotopic pairs for Gföhl gneisses were reported by Vellmer (1992): $^{87}\text{Sr}/^{86}\text{Sr}$$_{340}$ = 0.7237–0.7367 and $\varepsilon_{\text{Nd}}$$_{340}$ = −5.4 to −6.7 ($T_{\text{Nd}}^{DM}$ = 1.47–1.55 Ga).

The $^{87}\text{Sr}/^{86}\text{Sr}$$_{340}$–$\varepsilon_{\text{Nd}}$$_{340}$ plot (Fig. 11a) shows clearly that the Sr–Nd isotopic compositions of the Rokytná gneisses fall within the compositional range of the typical Moldanubian granulites (Vellmer 1992; Valbracht et al., 1994; Becker et al., 1999; Janoušek et al., 2004). However they differ from the Czech Moldanubian metasediments (paragneisses and kinzigites; Table 3).
Janoušek et al., 1995 and unpublished data), which have epsilon Nd values significantly lower than the gneisses described in the present work (Fig. 11a).

5. Discussion

Among the Gfohl gneisses and migmatites, several macroscopically distinct types can be distinguished. As shown by Hasalová et al. (in press-a), they apparently form a continuous structural sequence related to a disintegration of the parental orthogneiss. This distinctly banded rock is gradually transposed to stromatitic and schlieren migmatites. The migmatites are characterised by abundance of feldspar- and quartz-rich aggregates, presumably formed by fragmentation of the original banding. Eventually, the monomineralic layering vanishes and schlieren migmatite develops into isotropic nebulitic migmatite with no relics of gneissic morphology. At the first glance, this sequence resembles evolution of migmatite types described in classical works (e.g. Mehnert, 1971; Brown, 1973).

Fig. 7. Selected binary plots of SiO₂ vs. trace elements (typical LILE, REE and HFSE, in ppm). Some of the trace elements (Ba, La, Eu and Zr) show a more or less continuous decrease throughout the studied sequence, the trends for the others are more complicated.
Partial melting of rocks with a broadly granitic composition is the generally accepted explanation for the petrogenesis of the Gföhl migmatites (Dudek et al., 1974; Matějkovská, 1975). Even though such a conclusion is generally in line with the results presented in the current paper, the exact cause for the chemical variability observed in the Gföhl migmatites was never seriously discussed. The possible mechanisms involve: (i) tectonic or intrusive juxtaposition of genetically unrelated migmatites that have originated from distinct protoliths, (ii) equilibrium melting of a single metagranitic protolith, (iii) disequilibrium melting (with or without a small-scale melt movement) and (iv) infiltration of a melt derived from an external source into the parental orthogneiss. These hypotheses are evaluated below.

5.1.1. Distinct protoliths to each of the migmatite types

The first and arguably the most unrealistic is a model in which each of the migmatite types had its own, genetically unrelated protolith and their juxtaposition was purely accidental. This possibility can be first of all ruled out because of the observed gradual transposition of the \( \Sigma_1 \) banded gneisses into the \( \Sigma_2 \) stromatitic and schlieren migmatites (Hasalová et al., in press-a).

<table>
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<tr>
<th>Sample</th>
<th>Banded orthogneiss</th>
<th>Stromatitic migmatite</th>
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Note: *values given as < mean that the concentration is below the respective detection limit.
pointed out that the structural data are compatible with progressive deformation within a ductile shear zone resulting in generation of a continuous spectrum of variably deformed rock types ranging from pristine banded orthogneiss to the most affected schlieren migmatite. This interpretation is based on observed progressive folding of early steep orthogneiss fabric, in more advanced stages with development of isoclinal folds, leading eventually to complete fabric transposition and formation of schlieren migmatite. At the first glimpse, the elongated bodies of nebulitic migmatite can be viewed either as veins of isotropic granite penetrating parallel to the main S₂ mylonitic anisotropy (e.g. Brown and Solar, 1998) or as tectonically expelled granitic liquid injected into country rocks, which were hot thus preventing freezing (Weinberg and Searle, 1998). However, the nebulitic migmatite can be also regarded as the most extreme end-member of the structural sequence, i.e. completely disintegrated parental orthogneiss. This possibility is strongly supported by the microstructures, mineral chemistry (Table 1) and whole-rock geochemical parameters (Figs. 6–9) including the Sr–Nd isotopic compositions (Fig. 11 and Table 5) which change continuously from the banded orthogneiss to the nebulitic migmatite. Additionally, the overall resemblance in the trace-element distribution patterns and only gradual changes in the elemental concentrations

Fig. 8. Spider plots normalized to the composition of the average upper crust (Taylor and McLennan, 1985). The overall shape of the individual patterns for all rock types is similar. Broad ranges in compatible elements for the schlieren and nebulitic migmatites are interpreted as being due to the K-feldspar dominated fractional crystallization. Shaded field corresponds to the total variation in the whole dataset.
Together with the monotonously decreasing zircon/monazite saturation temperatures (Fig. 10) argue against significant differences in the protoliths. In other words, the available data for four studied orthogneiss and migmatite types suggest that they all have to have originated from the same protolith.
5.1.2. Equilibrium melting

The equilibrium melting model rests upon an assumption that the anatexis was slow enough for the full equilibration between the solid residue and the in situ partial melt to be achieved. In the closed-system, equilibrium melting model, followed by homogenization of the individual melt batches, one would not expect any variations in their radiogenic isotope compositions (e.g. Briquet and Lancelot, 1979). However, the Sr–Nd isotope data in the Rokytná Complex display considerable variability and regular changes with independent geochemical parameters (see for instance the linear trend in the SiO₂–⁸⁷Sr/⁸⁶Sr plot: Fig. 11b). This may reflect the source isotopic heterogeneity, perhaps due to in situ growth in the pre-Variscan protolith to the Gföhl gneisses (Frank et al., 1990). Such an isotopic heterogeneity is frequently observed on grain to meter scale in many high-grade metamorphic terrains and anatectic granites worldwide (e.g. Barbero et al., 1995).

Even though the variation in radiogenic isotope compositions does not preclude small-scale equilibria, the evolution of mineral compositions (Table 1) provides an independent argument disproving the equilibrium melting hypothesis. Assuming a closed-system, equilibrium melting along a prograde path, the X_Fe in garnet and biotite should decrease in accord with the decreasing modal proportion of garnet. In addition, the plagioclase should exhibit an increase in anorthite component with increasing degrees of melting. Such a compositional evolution was reported in a number of field and experimental studies (e.g. Vielzeuf and Holloway, 1988; Gardien et al., 1995; Dallain et al., 1999). However the compositional changes in the Rokytná Complex are just opposite, the X_Fe in garnet and biotite increase and the basicity of plagioclase drops (Table 1). This evolution is thus incompatible with a hypothesis invoking increasing degrees of melting in a closed-system.

Another possible argument is the strong textural and chemical disequilibrium between the rims and cores of the plagioclase crystals, demonstrating that it were apparently only their rims that have been in equilibrium with the melt (Fig. 3). However, because of the commonly slow diffusion of major and trace elements in the plagioclase (e.g. Blundy and Shimizu, 1991), this observation cannot exclude the in situ partial melting completely.

On this basis, the equilibrium melting hypothesis can be discounted as the main process responsible for the origin of the studied sequence, leaving the disequilibrium partial melting and open-system interactions as the only viable alternatives.

5.1.3. Disequilibrium melting

There is a growing evidence from experimental (e.g. Johannes 1980; Hammouda et al., 1996; Knesel and Davidson 1996, 2002) as well as from field observations and whole-rock geochemistry (e.g. Sawyer, 1991; Barbero et al., 1995; Bea, 1996; Harris and Ayres, 1998) indicating that the equilibrium between the melt and the solid residue does not have to be always attained in course of the crustal anatexis.

Firstly, the disequilibrium may occur when some accessory phases have remained armoured by main rock-forming minerals that have not participated in the melting reaction(s) (e.g. Watson and Harrison, 1984; Watt and Harley 1993; Nabelek and Glascock, 1995; Bea, 1996). However in the high-grade metamorphic rocks, the great majority of the accessory mineral grains is thought to be located at newly formed (or migrated) grain boundaries of the main rock-forming minerals and thus probably in contact with the partial melt (Watson et al., 1989). In lower-grade rocks, the accessories are mostly included in biotite and/or hornblende that in the course of the...
dehydratation melting would release them (Clemens, 2003).

The more common causes of disequilibrium are rapid melt production, extraction and segregation, not allowing enough time for the diffusional equilibration between the residual minerals and the melt. In normal granitic magmas derived by dehydratation crustal melting, the dissolution of zircon is assumed to be fast enough for the equilibrium to be attained (Harrison and Watson 1983; Watson 1996). However, achievement of equilibrium is less likely in the case of monazite and apatite, whose dissolution is controlled mainly by sluggish LREE and phosphorus diffusion (Harrison and Watson, 1984; Rapp and Watson, 1986). Swift extraction of low to moderate melt fractions can be promoted by a feedback mechanism between increasing melt production and deformation or shear-enhanced compaction, as was demonstrated for instance in the Himalayas (D’Lemos et al., 1992; Rutter and Neumann, 1995; Ayres et al., 1997). Muscovite dehydration melting can as well cause high dilatation strain that may further boost melt segregation (Rushmer 1996, 2001).

Disequilibrium melting may have a profound influence on the radiogenic isotopic composition of granitic magmas. Their Sr isotopic signature is controlled mainly by the main rock-forming minerals in the source (principally the balance between feldspars with low Rb/Sr and micas with high Rb/Sr, and thus also high \(^{87}\text{Sr}/^{86}\text{Sr}\) isotopic ratios). Consequently, the melts produced by disequilibrium dehydration melting involving muscovite and/or biotite will tend to have a more radiogenic Sr than their source. On the other hand, the bulk of the Nd in the felsic magmas will be controlled mainly by monazite and apatite, and thus the Nd isotope ratios will be governed by dissolution kinetics of these minerals. As shown by modelling of Zeng et al. (2005), these two phases ought to show a contrasting behaviour. The apatite with high Sm/Nd ratio will develop, with time, a more radiogenic Nd. Thus the progressive dissolution of apatite will yield melts with Nd isotopic signature increasingly more radiogenic (higher \(\varepsilon_{\text{Nd}}\) values) than the source. Monazite, which has relatively low Sm/Nd values, will have an opposite effect.

Disequilibrium melting model can explain the observed variation in the \(^{87}\text{Sr}/^{86}\text{Sr}_{340}−\varepsilon_{\text{Nd}}\) as well as \(\text{SiO}_2−^{87}\text{Sr}/^{86}\text{Sr}_{340}\) plots (Fig. 11a, b). The trends may theoretically argue for an increasing role of biotite dehydration melting and apatite dissolution in the more siliceous samples. Two disequilibrium melting scenarios can be distinguished, one without, and one involving melt movement.

If there was no melt movement, the differences in the amount of melt should directly reflect the primary variations in the protolith fertility, starting from nearly zero in the most basic of the banded orthogneisses. This would imply that the more acid portions with presumably higher melt contents would correspond to more fertile protoliths. In the context of dehydration melting

Table 5

<table>
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<th>Sample</th>
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<th>Rb (ppm)</th>
<th>Sr (ppm)</th>
<th>(^{87}\text{Rb}/^{86}\text{Sr})</th>
<th>(^{87}\text{Sr}/^{86}\text{Sr}) a</th>
<th>(^{143}\text{Nd}/^{144}\text{Nd}) a</th>
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<td>52.4</td>
<td>18.400</td>
<td>0.823759 (9)</td>
<td>0.734707</td>
<td></td>
</tr>
<tr>
<td>PG3</td>
<td>Neublastic migmatite</td>
<td>331.0</td>
<td>57.4</td>
<td>16.873</td>
<td>0.814908 (14)</td>
<td>0.733248</td>
<td></td>
</tr>
</tbody>
</table>

a values in parentheses are errors on the last decimal place (2SE).
b isotopic ratios with subscript “i” were all age-corrected to 340 Ma.
c two-stage Nd model ages calculated after Liew and Hofmann (1988).
d duplicated measurement (including sample decomposition and column separation).
model, they would have to contain more micas compared to their more basic (and more refractory) counterparts. This is unlikely in a co-genetic granitic suite of comparable silica range. More importantly, the obtained zircon and monazite saturation temperatures are mutually comparable and reasonably high for biotite dehydration melting to be a feasible melting mechanism. This seems to indicate that there was enough time for equilibrium between the residual zircon/monazite and the liquid to be established.

The whole-rock geochemical data can also be used to prove or disprove whether the melts could have originated from protoliths resembling the banded orthogneiss. From the composition of the least siliceous nebulitic migmatite follows that the anatectic melt had to have been granitic, poor in most major- and minor-element oxides (TiO$_2$, MgO, CaO) albeit rather Fe- and K-rich (FeO$_t$ $\sim$ 1.6%; K$_2$O/Na$_2$O $\sim$ 1.9–2.3) (Table 3 and Fig. 6). As shown by Sylvester (1998), the CaO/Na$_2$O ratios in peraluminous magmas are controlled mostly by the plagioclase/clay ratio of the source. This enables the distinction between granites generated from plagioclase-poor (pelitic; low CaO/Na$_2$O) and plagioclase-rich (psammitic; high CaO/Na$_2$O) sources. The Rokytná migmatites are characterized by intermediate CaO/Na$_2$O (0.15–0.26) and high Al$_2$O$_3$/TiO$_2$ (61–127) ratios. The nebulitic migmatites are relatively rich in Rb (231–331ppm), as well as extremely poor in Ba (69–326ppm) and Sr (25–57ppm), translating to relatively high Rb/Sr (5.8–9.2) and Rb/Ba (1.0–3.4) ratios. This, together with the presence of marked negative Eu anomalies, are
compatible with generation via rather low-degree, dehy-
dration melting of muscovite–biotite-rich quartzo-feld-
spathic lithologies (Harris and Inger, 1992; Barbarin, 1996;
Sylvester 1998 and references therein). Thus the chemistry
of the nebulitic migmatite does not seem to match in situ
melts of the rocks similar to the banded orthogneiss.

The more plausible is the possibility that the melt was
drained out on distances not exceeding the outcrop
scale. If the melt extraction was rapid, there would not be
enough time for its full equilibration. Nowadays, the
most accepted model for melt migration (low to
moderate melt fraction) involves movement through
the network of interconnected pores, driven by defor-
mation. Ultimately the deformation results in pervasive
drift migration utilizing the main rock anisotropies such
as foliation planes, fold hinges and boudin necks as
suggested by many field studies (e.g. Collins and
Sawyer, 1996; Weinberg, 1999; Brown and Solar,
1999). Additionally, shear-enhanced compaction would
drive melt into a network of melt-filled vein-like
leucosomes. Porous flow through such a vein network
would transfer the melt rapidly to the higher structural
levels (Rutter and Neumann, 1995).

In this scenario, one would expect at least a
rudimentary preservation of the melt flow network
throughout the area affected. However no such
structures have been observed in the field. Moreover,
the distribution of migmatites in the studied region is not
homogenous; there exist large regions that are formed
by products of advanced migationation (schlieren and
nebulitic migmatites) that are however distributed rather
randomly and definitely not structurally above the other
two migmatite types as should be anticipated. On this
basis, a small (outcrop) scale melt movement can be
probably ruled out.

Additionally, we have calculated (using the whole-
rock and mineral chemistries) the composition of a melt
that would be in equilibrium with the stable mineral
assemblage of the banded orthogneiss (Pl–Kfs–Qtz–
Sil–Grt–Bt) at 6.5kbar and 740–760°C. The calcula-
tions were performed using THERMOCALC 3.25
(Powell et al., 1998) and the internally-consistent thermo-
dynamic dataset 5.5 (Holland and Powell, 1998) in the
Na2O–CaO–K2O–FeO–MgO–Al2O3–SiO2–H2O
(NCKFMASH) system. The obtained melt is granitic
(SiO2 = 68.3–69.3wt.%; alkali feldspar granite in CIPW-
based Q–ANOR plot of Streckeisen and Le Maître 1979),
slightly peraluminous (A/CNK ~ 1.1), CaO poor (0.32–
0.41%) and alkali rich (K2O + Na2O = 8.8–8.9wt.%),
with prevalence of potassium over sodium (K2O/Na2O ~
1.35). Simple comparison with the composition of
nebulitic migmatite reveals that the modelled melt has
clearly too low SiO2, CaO, MgO (0.07–0.09%), FeOt
(0.27–0.33%), and exceedingly high alkalis (Na2O = 3.7–
3.8% and K2O = 5.1%) (cf. Fig. 6). Even in this case,
the nebulitic migmatite seem to be too different to re-
present rocks dominated by partial melts of the banded
orthogneisses.

An alternative would be that the nebulitic migmatites
represent diatexites that left behind some of their
residue. In spite of its macroscopic appearance, the
nebulitic migmatite probably never contained higher
amounts of melt, regardless the fact that considerable
melt volume had to have passed through the system to
account for all the observed variations. Hasalová et al.
(in press-a) presented the AMS (anisotropy of magnetic
susceptibility) data on biotite that show strong degree of
anisotropy of magnetic susceptibility even in the
nebulitic migmatite. This is in contrast with AMS data
from other migmatitic terrains (e.g. Ferré et al., 2004)
and with results of numerical modelling (e.g. Blumen-
feld and Bouchez, 1988) that both yielded significantly
lower values of degree of magnetic anisotropy in
diabetes. The high degree of AMS in the Rokytá
Complex thus documents predominance of solid-state
deformation with only minor contribution from melt.
The system had to have been still matrix and not melt
supported, precluding free rotation of biotite in
viscously flowing melt.

5.1.4. Melt infiltration from an external source

By “melt infiltration” we mean a process, whereby
melt derived from an external source passes pervasively,
i.e. along grain boundaries, the whole rock volume. In
this model, the textural and geochemical variations can
be interpreted by different degrees of equilibration
between the bulk rock and the passing melt. Melt
infiltration is a well known process from studies
concerned with metasomatism in the Earth’s mantle
and contact aureoles around igneous intrusions. In the
mante, melt infiltration, loosely termed reactive flow,
corresponds to grain-scale porous flow along high
porosity dissolution channels (e.g. McKenzie, 1989;
Van der Wal and Bodinier, 1996; Kelemen et al., 1997;
Reiners, 1998). The effects of reactive flow can be
identified due to highly contrasting modal and chemical
composition of the peridotite and the percolating basaltic
melt (Godard et al., 1995). The large difference in
viscosities between the melt and host rock is the reason
why the porous flow in the mantle, albeit slow, does not
require deformation to occur. Even though this mecha-
nism has been invoked to explain the transport of basaltic
magma in the mantle (Kelemen et al., 1997 and
references therein), many workers still tend to prefer
fracture-controlled melt flow (e.g. Spiegelman and Kenyon, 1992; Harte et al., 1993).

However, pervasive melt infiltration is even further from being generally accepted for melt transport mechanism in the crust. Melt percolation from/along grain boundaries has until now been reserved only for the scale of several centimetres (e.g. Sawyer, 2001) and it has been considered only rarely as a serious possibility for large-scale melt transport in the crust (Hasalová et al., in press-a,b).

The main problem consists in the fact that this process is considered too sluggish to be efficient. McKenzie (1984)

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**Fig. 12.** (a) Modelling of chemical effects connected with the leucogranitic melt infiltration into the Rokytná Complex. As the melt passes through the system, K-feldspar, plagioclase and quartz crystallize/overgrow the corresponding phases in the matrix. The model treated various rock samples as mixtures of the banded orthogneiss and the crystallizing minerals from the melt. The compositions of the latter are listed in the table forming a part of the Fig. 3; as the starting material has been chosen the least evolved banded orthogneiss (PG7). Solid black arrows represent the tie line between the real compositions of the banded orthogneiss (PG7) and the nebulitic migmatite (PG1). The dashed arrows represent the outcome of the modelling. Note the good correspondence of the real and modelled trends. The stars portray the best-fit solution for the assemblage added. The results indicate that the whole compositional spectrum from the banded orthogneiss to the least siliceous nebulitic migmatites could have been produced by addition of up to c. 150 wt.% of the both feldspars and quartz combined (if the original rock is taken as 100%, i.e. corresponding to a relative gain of 60%). (b) Summary of the modelling for various migmatite pairs. Presented are proportions of the minerals (Kfs, Pl, Qtz) added into the system, total percentages of these newly crystallized phases (protolith gneiss PG7 = 100%) and sum of squared residuals ($R^2$) indicating the goodness of the fit.
showed that the pervasive porous flow in crustal rocks would be possible only in a matrix which is able to compact, in other words they concluded that such a flow is essentially a deformation-driven process. This could have been the case in the Rokytná Complex, where a large, crustal-scale shear zone played an important role in its exhumation.

During the penetration by externally-derived melt, the rock complex would clearly display open-system behaviour, with components having been lost to, or introduced from, the passing melt of unknown volume. The whole-rock composition would then represent a net result of the changing system mass/volume, the contributions from newly precipitated minerals as well as the mass balance of geochemical species exchanged with the melt/fluid phase.

We have attempted to model the process as an addition of K-feldspar, plagioclase and quartz crystallized from the melt passing through the banded orthogneiss. The exact composition of the pristine orthogneiss is unknown; therefore we have opted for the presumably least influenced sample PG7. The mineral compositions employed were those of the newly formed rims overgrowing the relict mineral grains in the matrix (Fig. 3). A graphical representation of the modelling approach is shown in Fig. 12a. The whole compositional spectrum from the banded orthogneiss to the least siliceous nebulitic migmatite can be explained by a relative gain up to 60% of K-feldspar, plagioclase and quartz. The proportion of the three minerals changes from roughly balanced to a strong prevalence of K-feldspar (∼ 40%) and quartz (∼ 40%) over plagioclase (∼ 20%).

Additionally, assessment of the mass balance in individual major- and trace elements during the orthogneiss transformation used the approach of Gresens (1967) in the form of isocon plots (Grant, 1986, 2005) (Fig. 13a). The results of individual isocon analyses were pulled together into binary plots of an immobile element vs. relative gains/losses (%) of individual elements (Fig. 13b). The most appropriate choice of immobile components seems to be high-field strength elements Zr, Hf, La, Sm, Nb and Th, as follows from examination of isocon plots as well as the ordered ratios of concentrations of individual elements in the ‘altered’ and ‘original’ rocks (Grant, 2005).

Slopes of the isocons fitted to these HFS elements are lower than unity in all cases, pointing to an increase in the overall mass of the system (an increase in volume if changes in density were negligible) (e.g. Fig. 13a). Many of the elements show coherent behaviour, with three groups emerging (Fig. 13b): (1) immobile elements that have been used in the construction of the isocons, oscillating around zero (e.g. Th, La, Sm and Nb), (2) hydrous fluid-mobile trace elements (LILE) showing extreme relative enrichments (Rb, Cs and U), and (3) significantly less affected major and minor elements Si, Al, K, Na and P. In accordance with the previous modelling (Fig. 12), the behaviour of the elements in the latter group (Si, Al, K, Na and P), can be interpreted as precipitation of new albitic plagioclase, K-feldspar, quartz and apatite. Variable but less pronounced depletion in Ca, Sr and Ba may point to instability and partial replacement of the original K-feldspar and more calcic plagioclase. This would explain the origin of the observed albitic plagioclase and K-feldspar overgrowths/films formed on older feldspars generation of the matrix (Fig. 3).

Taken together, both modelling approaches seem to support the model of melt infiltration, which is in accordance with the major- and trace-element variation and Sr–Nd isotopic data indicating open-system behaviour.

5.1.4.1. The possible role for fractional crystallization.

While some major elements display consistent trends over the whole silica range, the others feature remarkable horizontal shifts at the end of the sequence, i.e. in the more siliceous schlieren and nebulitic migmatites (Fig. 6). In addition, the spider plots (Fig. 8) show large ranges for compatible elements in these two migmatite types, with the contents decreasing rapidly in the most silica-rich samples. This seems to indicate that there was an additional process involved.

Theoretically such a differentiation to more siliceous compositions could have been driven by the restite unmixing (Chappell et al., 1987; Barbero and Villaseca, 1992; Chappell, 1996; Williamson et al., 1997). However this does not agree with the observed limited variations in the ferromagnesian components, nearly constant A/CNK and sharp decrease in feldspar-compatible elements, such as Ba, Sr and Eu. Moreover this scenario would require separation of large volumes of refractory garnet-bearing residua, whose occurrence is unknown from the Gföhl Unit.

The major- and trace-element signatures of the most siliceous samples of the nebulitic migmatite are remarkably similar to felsic, strongly fractionated granites. The broad variation and progressive decrease in Ba, Sr, Eu, LREE, Th (Figs. 6 and 7) and, to some extent, Pb (not shown) seem to bear a testimony to fractional crystallization, a process very efficient in removing compatible elements from an evolving magma. This mechanism is expected to have operated already during the ascent and, more importantly, at the terminal stages of the melt infiltration process.

This hypothesis is corroborated for instance by the binary plot involving Sr and Ba (Fig. 14). The diagram
Fig. 13. Mass balance calculations of individual major- and trace elements (after Gresens, 1967) for transformation of banded orthogneiss (PG7) into schlieren migmatite (PG4). The slope of the isocon has been obtained taking the presumably immobile elements Zr, Hf, La, Sm, Nb and Th (circled) into consideration. The sample plot with highlighted elements that are interpreted as having been added to the system due to growth of new mineral phases (Pl, Kfs, Ap) and brought by hydrous fluid (Rb, Cs and U). (b) Plots of Zr vs. relative gains/losses (%: Grant, 2005) of individual elements. As a starting material was taken the least evolved orthogneiss (PG 7) with the highest Zr contents. This element has been chosen as it varies greatly, decreases monotonously (see also Fig. 7) and seems immobile in the course of the progressive breakdown of the orthogneiss matrix.
demonstrates that the entire compositional spectrum of the schlieren and nebulitic migmatites can be modelled by reasonable amounts of K-feldspar > albite plagioclase and biotite fractionation.

In addition, some role for the accessory minerals is evident. The sharp decrease in LREE and Th with increasing silica (Figs. 7 and 8) points to monazite removal (Fig. 4), a phenomenon common in relatively LREE-poor granitic melts (Villaseca et al., 2003). Lastly, the drop in MREE can be accounted for by apatite fractionation.

5.1.4.2. P–T conditions of the orthogneiss-melt interaction. In order to understand better the mechanism of the migmatization, information on the P–T conditions is essential. Some constraints are provided by the saturation thermometry on accessory minerals. In slightly peraluminous, calcium-poor leucogranites the most important will be zircon (Watson and Harrison, 1983) and monazite (Montel, 1993).

Given that the amount of melt present in the banded orthogneiss was limited, the whole-rock geochemistry can be used as a proxy to assess the concentrations of many elements in its Early Palaeozoic protolith. Even though inherited Precambrian components are by no means rare in the Gfohl gneiss zircons, at least in Austria (Friedl et al., 2004), their proportion by volume is small. Consequently, their influence on the Zr budget and thus the calculated Zr saturation temperatures should be negligible. Then the zircon (c. 815 ± 24°C, median ± 2σ) and monazite (c. 800 ± 38°C) saturation temperatures yield the upper constraints on the liquidus temperature of the magma parental to the granitic protolith (e.g. Hanchar and Watson, 2003; Miller et al., 2003; Janoušek, 2006). This agrees with the coincidence of the two independent saturation temperatures derived from both accessory minerals.

On the other hand, the nebulitic migmatite may provide information on the maximum liquidus temperature of the Variscan anatectic melt. The temperature is near 740°C, derived from zircon (747 ± 53°C) and monazite (740 ± 90°C). If the melt infiltration has indeed taken place, the apparent gradual decrease in zircon and monazite saturation temperatures in the whole-rock sequence (Fig. 10) may be an artefact, resulting from mixing of two dissimilar components.

The second approach, using the THERMOCALC calculations, was adopted by Hasalová et al. (in press-b). These authors calculated equilibration temperature of the banded orthogneiss, stromatitic, schlieren and nebulitic migmatites to be 790–850°C at 7.5 kbar, 760–820°C at 6.5 kbar, 715–770°C at 5.5 kbar and 690–750°C at 4.5 kbar, respectively. Moreover, the P–T calculations clearly show that the infiltrating melt equilibrated with the host rock on the retrograde branch of the P–T path.

5.2. Broader implications for the petrogenesis of felsic migmatites

Several authors have recently proposed a new mechanism for transport of felsic magma through hot, mid-crustal rocks, termed pervasive flow (Weinberg, 1999; Olsen et al., 2004). In this model, foliation-parallel veins/sheets of granitic composition invade hot country rocks, whose low-viscosity inhibits hydrofracturing and dyking (Weinberg and Searle, 1998). Additionally, their high temperature (higher than the solidus of the invading melt) enables the magma to migrate upwards without crystallizing. This process should result eventually in a formation of up to several km thick injection complexes, common in hot crustal terrains (Weinberg, 1999; Leitch and Weinberg, 2002). Hasalová et al. (in press-a) argued that the large-scale pervasive flow does not require formation of channelized pathways but can also occur penetratively, along grain boundaries. These authors
presented microstructural observations indicating that the porous flow commonly described in mantle rocks (McKenzie, 1984; Reiners, 1998) can in fact play a significant role in the crust as well.

The model of pervasive flow still contains several aspects that remain to be clarified. Firstly, its physical background is still poorly understood. So far there are only several theoretical papers (Leitch and Weinberg, 2002; Olsen et al., 2004), accompanied by works providing field and petrographic evidence from a handful of terrains worldwide (e.g. Collins and Sawyer, 1996; Weinberg and Searle, 1998; Vanderhaeghe, 1999).

A key problem is the expected involvement of small volumes of penetrating acid melt, which should theoretically have high viscosities. The process of reactive porous flow is very slow even for mantle-derived melts (tens to hundreds centimetres per year; McKenzie, 1984) and is likely to be even more so in the crust. This would make the system vulnerable to freezing, and subsequently render the long-distance melt percolation impossible.

These problems can be possibly less important for volatile-rich melts, as water, F and B dramatically decrease the viscosity of granitic melts (Dingwell et al., 1996; Giordano et al., 2004; Whittington et al., 2004). This was probably not the case in the Rokyná Complex, as the whole-rock fluorine contents are uniformly low (Table 1) and tourmaline, the main boron host, is absent. However, the pervasive melt migration, even at small volumes, can occur as soon as a sufficient porosity is created. This is associated with dilatation related either to cavitation process, known well from the material science (Čadek, 1988), in conjunction with melt underpressure or to grain-scale hydrofracturing related to melt overpressure and high differential stress (Závada et al., in press). The most straightforward scenario in our case seems that the pervasive porous flow was a deformation-driven process associated with diffusion creep deformation mechanism. It could have been connected to a crustal-scale shear zone, described in many previous works (e.g. Schulmann et al., 2005; Racek et al., 2006; Tajčmanová et al., 2006).

Generally speaking, we have not found in the course of the current research any arguments invalidating the melt infiltration model. In fact the whole-rock geochemistry may provide important constraints on the nature and source of the protolith, the penetrating melt and the overall mass balance. However, the crucial point that ought to be stressed here is that the whole-rock geochemical data by themselves, even though not contradicting the melt infiltration model, do not provide an unequivocal proof for its operation. They cannot rule out completely some common scenarios, such as disequilibrium melting with small-scale melt movement or in situ equilibrium melting, in which the individual small melt batches failed to homogenize with each other, still reflecting the geochemical/isotopic variation of their protolith. In order to disprove the other genetic possibilities microstructural, petrological and field data are essential. In this context the reader is referred to the complementary work providing the necessary additional information (Hasalová et al., in press-a-b).

6. Conclusions

Four types of gneisses and migmatites (banded orthogneiss, stromatitic, schlieren and nebulitic migmatites) were defined previously in the Rokyná Complex (Gföhl Unit, Bohemian Massif). While each of the types shows a distinct geochemical and Sr–Nd isotopic fingerprint, the whole sequence evolves along regular, more or less smooth trends for most of the elements. This evolution of the Rokyná Complex has been interpreted by (Hasalová et al., in press-a-b) as a result of melt infiltration. The melt infiltration has been defined as a process in which the banded orthogneiss was pervasively, along grain boundaries, penetrated by felsic melt derived from an external source. In this context the individual migmatite types can be explained by different degrees of equilibration between the bulk rock and the passing melt.

The current study is in agreement with this model, however the available whole-rock geochemical data alone are not unequivocal and need to be supported by field, microstructural and petrological observations. Nevertheless, the major- and trace-element as well as Sr–Nd isotopic compositions yield some important constraints concerning the nature of the protolith, the composition and possible source of the penetrating melt and the overall mass balance.

If true, the melt infiltration can be modelled as an open-system process, characterised by changes of the total mass/volume, accompanied by gains/losses in many of the major- and trace elements. The numerical modelling of the mass balance resulted in identification of a component added by a heterogeneous nucleation of feldspars, quartz and apatite from the passing melt. This is in accord with presence of new albite plagioclase, K-feldspar and quartz coatings as well as resorption of relict feldspars. At the most advanced stages, the chemical variations in the schlieren and nebulitic migmatites require an increasing role for fractional crystallization of the K-feldspar and minor plagioclase, with accessory amounts of Th-rich monazite ± apatite.

In general, the melt infiltration model may potentially explain the origin of some of the felsic migmatites in other high-grade metamorphic terrains. If proven widespread, the process would have profound implications for
chemical development of large crustal segments overlaying anatectic regions and the melt transport therein. Melt infiltration would also strongly influence the rheology of large crustal domains, with potential consequences for deformation mechanism in collisional orogens. Clearly, much more work remains to be done before it could be truly established and well understood.

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Appendix 1. Analytical techniques

Four samples of each rock type were taken in order to document the compositional variation on both the outcrop as well as regional scale (Fig. 1c).

Minerals were analysed using the Cameca SX100 electron microprobe at the Czech Academy of Sciences in point beam mode at 15kV and 10nA. Representative analyses are summarized in the Table 2.

Following a lithium tetraborate fusion, major-element oxides and trace elements were determined by ICP MS VG PQ 2+ at the Centre de Géochimie de la Surface, Université Louis Pasteur, Strasbourg. The obtained results are listed in Tables 3 and 4. Fluorine was measured by an ion selective electrode in the laboratories of the Czech Geological Survey in Prague (Table 3). Data management, recalculation, plotting and statistical evaluation of the whole-rock geochemical data were facilitated using GCDkit (Janoušek et al., 2006).

For the isotopic study, samples were dissolved using a combined HF–HCl–HNO₃ attack. Strontium was isolated by exchange chromatography techniques on PP columns with Sr.spec Eichrom resin and bulk REE were isolated on PP columns filled with TRU.spec Eichrom resin (Pin et al., 1994). The Nd was further separated on PP columns with Ln.spec Eichrom resin (Pin and Zalduegui, 1997). Isotopic analyses were performed on Finnigan MAT 262 thermal ionization mass spectrometer in dynamic mode using a double Re filament assembly (CGS). The ¹⁴³Nd/¹⁴⁴Nd ratios were corrected for mass fractionation to ¹⁴⁶Nd/¹⁴⁴Nd = 0.7219, ⁸⁷Sr/⁸⁶Sr ratios assuming ⁸⁶Sr/⁸⁸Sr = 0.1194. External reproducibility is given by results of repeat analyses of the La Jolla (¹⁴³Nd/¹⁴⁴Nd = 0.511852 ± 14 (2σ; n = 23)) and NBS 987 (⁸⁶Sr/⁸⁸Sr = 0.710247 ± 26 (2σ; n = 25)) isotopic standards. The Rb, Sr, Sm and Nd concentrations were obtained by ICP-MS.

The decay constants applied to age-correct the isotopic ratios are from Steiger and Jäger (1977) (Sr) and Lugmair and Marti (1978) (Nd). The εNd values and single-stage CHUR Nd model ages were obtained using Bulk Earth parameters of Jacobsen and Wasserburg (1980), the two-stage Depleted Mantle Nd model ages (TDM) were calculated after Liew and Hofmann (1988). The Sr–Nd isotopic data for studied migmatites are listed in Table 5.

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