# The Effect of Melt Infiltration on Metagranitic Rocks: the Snieznik Dome, Bohemian Massif

P. Stípská[1,2]*, P. Hasalová[1], R. Powell[3], P. Závada[1,4], K. Schulmann[1,2], M. Racek[5], C. Aguilar[1] and F. Chopin[2]

[1]Center for Lithospheric Research, Czech Geological Survey, Kla´rov 3, Prague 1 118 21, Czech Republic;
[2]Institut de Physique du Globe, UMR 7516, Universite´ de Strasbourg, 1 rue Blessig, Strasbourg 67084, France;
[3]School of Earth Sciences, University of Melbourne, Melbourne, VIC 3010, Australia;
[4]Institute of Geophysics, Czech Academy of Sciences, Bocn II 1a/1401, Prague 141 31, Czech Republic;
[5]Institute of Petrology and Structural Geology, Charles University, 12843, Prague 2, Czech Republic

*Corresponding author. Present address: Institut de Physique du Globe, UMR 7516, Université de Strasbourg, 1 rue Blessig, Strasbourg 67084, France. Telephone: .420257089492. Fax: .420 257 089 500.
E-mail: stipska@gmail.com

Received July 28, 2018; Accepted January 27, 2019

## ABSTRACT

Highly deformed banded phengite–biotite metagranite from the Snieznik dome in the Bohemian Massif has been modified locally to have stromatic, schlieren or nebulitic textures typical of migmatites. This occurred mostly along subvertical deformation zones at eclogite-facies conditions, at a scale of several centimetres to several metres, mostly parallel to the foliation. The transition from banded to migmatite types of orthogneiss is marked by an increase in the amount of phases interstitial along grain boundaries in the dynamically recrystallized monomineralic feldspar and quartz aggregates, and by increasing consumption of recrystallized K-feldspar grains by fine-grained plagioclase and quartz, as well as myrmekite (intergrowth of Pl_Qz). The new minerals are in textural equilibrium with phengite. The myrmekite, quartz and feldspars can be coarse-grained (grain size 0.5 cm). These features are considered to be the result of grain-scale melt infiltration that caused dissolution–reprecipitation along grain boundaries in the presence of phengite. The infiltration was pervasive at the grain scale, but localized at hand-specimen to outcrop scales. All the rock types have the same mineral assemblage of Grt_Ph_Bt_Ttn_Kfs_Pl_Qz6Rt6Ilm; they have similar garnet, phengite and biotite compositions, and based on mineral equilibria modelling we infer equilibration at a pressure of 1.5–1.7 GPa and a temperature of 690–740

°C. Because the rocks are inferred to be $H_2O$-undersaturated and above the temperature conditions of the wet solidus, infiltration must have involved a hydrous melt, as opposed to an $H_2O$ fluid. Stability of melt-bearing mineral assemblages and mineral compositions are almost independent of the melt proportion in the system, thus explaining the identical assemblage and mineral compositions observed in all the migmatite types. This precludes the estimation of the amount of melt infiltrated. Migmatite textures, however, suggest that variable degrees of melt–rock interaction occurred, being low in the banded migmatite types and higher in the nebulitic and schlieren types. Retrograde equilibration was largely restricted to retrograde zoning in phengite, garnet and plagioclase, and crystallization of biotite around phengite and garnet, presumably in a continuous reaction consuming melt. This may have occurred down to 0.7–1.0 GPa. According to Sr–Nd isotope data, the infiltrating melt is probably derived from similar rocks, structurally beneath the observed ones. The infiltration may have facilitated exhumation of a 2km wide structural domain from 1.7 to 0.7 GPa, within which are the subvertical deformation zones along which the infiltration occurred.

Key words: orthogneiss migmatite; fluid-assisted melting; eclogite facies; melt percolation; exhumation

## INTRODUCTION

Classically, partial melting processes have been studied predominantly in metasediments. However, recently, partial melting processes in granitoids are receiving increasing attention, especially in areas of subduction to (U)HP conditions. Of particular interest is the role of melt in strain localization in granitoid crust (Rosenberg & Handy, 2005; Labrousse et al., 2015; Závada et al., 2018), thus allowing trans-lithospheric exhumation of subducted crust (Maierová et al., 2018). Detailed petrological and geothermobarometric studies of granite migmatites are rare (Hasalová et al., 2008b, 2008c), and are missing for HP conditions. Recognition of the former presence of melt in rocks relies on macroscopic and microscopic observations (e.g. Brown, 2007; Holness & Sawyer, 2008; Sawyer, 2010). Such observations in a field-based study led Hasalová et al. (2008a, 2008b, 2008c) to suggest pervasive grain-scale melt migration as the main melt transport mechanism in metagranitoid rocks, a model recently applied in other field areas (Stuart et al., 2017; Závada et al., 2018). Based on these studies, grain-scale melt migration is thought to be focused in localized zones of deformation, thus not necessarily affecting the whole volume of rocks. This is in contrast to a generally accepted melt transport paradigm for the crust, based mainly on observations in metasedimentary rocks (e.g. Brown, 2007). This model involves melt drainage to sites of accumulation, and their interconnection into a network of leucosomes, sills and dykes through which melt is transported to higher crustal levels to form plutons by channelized flow (e.g. Brown, 2007). In this model, grain-scale migration is considered only to explain melt moving from melt-production reaction sites to sites of accumulation (e.g. Sawyer, 2001).

To understand where melting occurred on the path of the rocks—on their burial and/or on their ascent-requires evidence for the presence of a melt to be found and P–T conditions to be estimated. Such a combined study has rarely been undertaken on metagranitoid rocks (e.g. Hasalová et al., 2008c; Korhonen et al., 2010). But, for example, mineral equilibria modelling has been used to trace heating (Schulmann et al., 2008) and exhumation (Hasalová et al., 2008c) in medium-pressure metagranitic migmatites, or burial and/or exhumation in high-pressure metagranitic rocks (e.g. Chopin et al., 2012b), or to estimate melting conditions based on inclusion assemblages in magmatic zircon (Hopkins et al., 2010). Experimental re-homogenization of former melt inclusions in garnet allowed estimation of trapping conditions of melt at near UHP peak (Ferrero et al., 2015). The role of fluids, especially H2O, is an important aspect for the melting of rocks. Three major types of melting have been recognized, leading to melting starting at different P–T conditions and potentially to very different amounts of melt produced. In the presence of fluid, crustal rocks will undergo an abrupt transition from fluid-present to melt-present at the wet- or H2O-saturated solidus, through a process of fluid-present melting (e.g. Huang & Wyllie, 1981; Hermann & Spandler, 2008; Hermann & Rubatto, 2014). In this development, fluid is used to mean an H2O-rich hydrous phase, not including melt. The amount of melt produced is only a few per cent, as the amount of fluid, for example in the pores of a rock, is limited (e.g. White et al., 2001). If fluid is not present, melting will occur at appropriate elevated temperatures by the breakdown of hydrous minerals in reactions referred to as dehydration melting (e.g. Clemens & Vielzeuf, 1987; Auzanneau et al., 2006; Masotta et al., 2018). The most important dehydration reactions in quartzo-feldspathic rocks involve white mica and biotite (Huang & Wyllie, 1981; Auzanneau et al., 2006; Hermann & Rubatto, 2014), and the amount of melt that can be formed is proportional to the volume of these minerals consumed.

Fluid-present and dehydration melting are effectively closed-system processes, with the melt produced without addition of fluid or melt from outside the rock being melted. Open-system processes can also be involved in melting, with addition of fluid or melt to a rock (for granitoid examples see Collins et al., 1989; Pognante, 1992; Jung et al., 2009; Sawyer, 2010; for a review see Weinberg & Hasalová, 2015). These are termed fluid fluxed melting and melt-fluxed melting, respectively. Fluid-fluxed melting is associated with near wet-solidus conditions where fluid can gain access to rocks from or via rocks that are below their solidus (e.g. White et al., 2005). The amount of melt produced depends on the amount of H2O added and can therefore be very high (e.g. White et al., 2005; Hasalová et al., 2008c; Ganzhorn et al., 2016). This process is recognized by high degrees of melting in otherwise near wet-solidus rocks. At temperatures above the wet solvus it is difficult to envisage fluid addition. If open-system processes are involved then melt addition and/or melt advection are required (e.g. Greenfield et al., 1996; Sawyer, 1998; Milord et al., 2001; Hasalová et al., 2008c; Collins et al., 2016; Stuart et al., 2017), giving rise to melt-fluxed melting. Here, we study granitic migmatites cropping out in the vicinity of eclogite boudins and HP metagranitic rocks within the continental subduction wedge of the Bohemian Massif. This is an ideal site for studying meltrelated processes in granitoid rocks at high pressure. Field studies document evidence for the transition to migmatite and the relation of migmatite formation to deformation fabrics. Grain-scale textural characteristics are used to consider the possible melt migration process. The P–T conditions of melting are estimated. Forward-modelling of equilibrium mineral assemblages is used to assess whether the migmatite formation occurred through dehydration melting, or fluid-fluxed or melt-fluxed melting.

## GEOLOGICAL SETTING

### Tectonic setting

The Orlica–Snieznik dome (OSD) in the NE part of the Bohemian Massif has a general structure of a mantled dome, occurring in the Orlica and Snieznik mountain ranges (in Czech, the Orlické and Kralicky Snezník mountain ranges; Fig. 1) (Chopin et al., 2012a; Mazur et al., 2012). The dome is composed of a granitoid derived gneissic core containing boudins of eclogite and high-pressure granulite with Cambro-Ordovician protoliths (Kröner et al., 2001; Bröcker et al., 2010; Mazur et al., 2010) and belts of metamorphosed volcano-sedimentary rocks with Neoproterozoic to Ordovician ages (Jastrzebski et al., 2010; Mazur et al., 2012). The high-grade gneissic core is mantled by medium- to low-grade metamorphic rock units composed of Proterozoic schists, slates, volcanic rocks and Cambro-Ordovician rift sequences (Kröner et al., 2000; Mazur & Aleksandrowski, 2001; Štípská et al., 2001). The compilation of pressure–temperature–time–deformation (P–T–t–D) data allows formulation of an evolutionary model (Chopin et al., 2012a). This involves Late Devonian subduction of the Cambro-Ordovician oceanic crust and passive margin followed by underthrusting of buoyant felsic crust that resulted in its stacking beneath dense upper plate rocks (Mazur et al., 2012), a process called 'relamination' (e.g. Hacker et al., 2011; Maierová et al., 2014, 2018). The continental rocks were subducted to asthenospheric depths as testified by evidence for (U)HP eclogite- and granulite-facies conditions (Bröcker & Klemd, 1996; Kryza et al., 1996; Klemd & Bröcker, 1999; Jedlicka & Faryad, 2017), whereas the middle crustal rocks achieved amphibolite facies conditions (Murtezi, 2006; Jastrzebski, 2009; Skrzypek et al., 2011a, 2011b, 2014; Jastrzebski et al., 2019). The subduction–relamination stage was followed by crustal-scale Early Carboniferous doming when the lower crustal rocks were exhumed to middle crust levels along vertical deformation zones (Štípská et al., 2004, 2012) in the cores of antiformal sub-domes. The growth of the dome, 60km across, was associated with unroofing and sliding of the upper plate crustal rocks along localized extensional detachments, thus enhancing final exhumation of the buoyant HP core (e.g. Skrzypek et al., 2011b; Maierová et al., 2014). The vertical extrusion of the dome was accompanied by intrusions of kilometer-scale granitoid dykes emplaced along both axial-planar zones of sub-domes and detachment fabrics (Parry et al., 1997; Lehmann et al., 2013; Závada et al., 2017). The dynamics of doming is attributed to the combined effect of continuous accumulation of crustal material above a west-dipping subduction zone (Chopin et al., 2012a; Mazur et al., 2012), and opposite indentation of a continental buttress from the east (Schulmann & Gayer, 2000; Štípská et al., 2006; Košuličová & Štípská , 2007).

The subduction to exhumation processes operated in the period between 360 and 330Ma (Steltenpohl et al., 1993; Lange et al., 2005a, 2005b; Anczkiewicz et al., 2007; Bröcker et al., 2009, 2010; Skrzypek et al., 2017).

## The metagranitic rocks in the Orlica–Snieznik dome

The core of the dome is dominated by metagranitic rocks, classically subdivided into two types, named according to the type localities: (1) the coarse-grained augen to banded Snieznik type; (2) the mylonite, ultramylonite and migmatite Gieraltow type (Fig. 1) (e.g. Fischer, 1936; Don et al., 1990; Chopin et al., 2012a, 2012b). Although there is discussion about the origin and age of these rock types (Redlińska-Marczyńska et al., 2016), most of the studies interpret the textural variations as resulting from different degrees of deformation, metamorphism and anatexis of the same Cambrian granite protolith (Kröner et al., 2001; Lange et al., 2002, 2005a; Bröcker et al., 2009; Chopin et al., 2012b). The mylonitic and migmatitic varieties are associated with eclogite and granulite, whereas the augen-type dominated domains were deformed and metamorphosed together with adjacent amphibolite-facies metasedimentary rocks. Consequently, the metamorphic conditions of different types of gneiss reveal contrasting P–T conditions. For instance, Bröcker & Klemd (1996) inferred conditions of 2.8 GPa for a gneiss at a contact with eclogite. Combined petrological and microstructural study interpreted a sample sequence from a strain gradient close to the eclogite belt as reflecting different stages of prograde metamorphism (Chopin et al., 2012b), as follows. The least deformed augen orthogneiss represents a solid-state low-strain domain equilibrated at 1.2–1.5 GPa, 600–670°C. Progressive localization of strain in the mylonitic gneiss during burial, possibly accompanied by partial melting, was coeval with equilibration at 1.9 GPa, 700–770C. A melt inclusion study from a felsic granulite showed trapping conditions for melt at 2.7 GPa, 875°C, suggesting near UHP conditions of melting (Ferrero et al., 2015). In contrast, the orthogneiss in direct ductile structural continuity with hanging-wall paragneiss was metamorphosed at 0.8 GPa, 600°C (Jastrzebski et al., 2019). These petrological studies indicate that the orthogneiss experienced a variable degree of deformation and burial within a continental subduction wedge.

The metamorphic conditions and the relation of migmatite formation to the structural succession remain controversial. For example, Turniak et al. (2000) considered the migmatite formation as a HT–LP event incompatible with eclogite boudins. Similarly, Lange et al. (2002) suggested that the migmatite formation overprinted the strain gradient and therefore is syn- to posttectonic. In contrast, Chopin et al. (2012b) interpreted melting as syntectonic with continental subduction. The age of anatexis of _340–360Ma was determined from zircon rims in metagranitic migmatites (Turniak et al., 2000; Lange et al., 2005a) and on zircon in leucosomes and granitoid mobilizates (Štípská et al., 2004; Bröcker et al., 2009).

# ANALYTICAL METHODS

## Analytical methods for Sr–Nd isotopic composition

Strontium and neodymium isotopic compositions were obtained at the Czech Geological Survey. Samples were dissolved using a combined HF–HCl–HNO3 digestion. Strontium and rare earth element (REE) fraction were isolated from the bulk matrix by ion-exchange chromatography techniques using AG 50W-X8 resin (Bio-Rad Laboratories, Inc.). The Sr.spec resin (Triskem Intl.) was used for purification of the Sr fraction (Pin et al., 2014), the Nd cut was purified using Ln resin (Triskem Intl.; Pin & Zalduegui, 1997). Isotopic analyses were performed on a Triton Plus (Thermo Fisher Scientific Inc.) thermal ionization mass spectrometer in static mode using a single Ta and double Re filament assembly for Sr and Nd, respectively. The 143Nd/144Nd ratios were corrected for mass fractionation to 146Nd/144Nd = 0.7219 and 87Sr/86Sr ratios assuming 86Sr/88Sr = 0.1194. External reproducibility and measurement accuracy was demonstrated by repeated analyses of the NBS 987 [87Sr/86Sr = 0.7102536 + 0.000018 (2r, n.54)] and JNdi-1 [143Nd/144Nd . 0.5120996 + 0.000010 (2r, n.33)] reference materials (see Jochum et al., 2005).

The Rb, Sr, Sm and Nd concentrations were obtained by inductively coupled plasma mass spectrometry (ICPMS). The decay constants applied to age-correct the isotopic ratios are from Steiger & Jäger (1977) (Sr) and Lugmair & Marti (1978) (Nd). The e values and singlestage CHUR Nd model ages were obtained using the Bulk Earth parameters of Jacobsen & Wasserburg (1980); the two-stage Depleted Mantle Nd model ages (T) were calculated after Liew & Hofmann (1988). The Sr–Nd isotopic data for the studied metagranitic migmatites are listed in Table 1.

## Analytical methods for mineral compositions

Mineral analyses and compositional maps were obtained by electron probe microanalysis (EPMA) using a JXA-8530F by JEOL with a field emission gun electron source, at the Institute of Petrology and Structural Geology, Charles University, Prague, Czech Republic. The operating conditions were accelerating voltage 15 kV and beam current 30nA in focused beam mode for spot analyses of garnet and titanite, 15 kV and 20nA with beam defocused to 5 mm for spot analyses of micas and feldspars, and 20 kV and 80 nA for element mapping. Representative analyses are given in Tables 2–7.

## Field relations and metagranite textures and Sr–Nd isotopes in the study area

### *Field relations and metagranite textures*

The focus of the study is an _2 km wide belt of metagranite around an eclogite-bearing belt in the central part of the OSD. It is dominated by narrow-banded and mylonitic rock types that can be both migmatitic and non-migmatitic, and alternate with other migmatitic types, as specified below. These metagranite rock types pass to the east and to the west into thick-banded and coarse-augen non-migmatite types. Field relations and textural types of the metagranite with the most pronounced migmatite textures were studied in a series of outcrops located to the west of the eclogite-bearing belt (Fig. 1). They involve rare augen and banded types, and metagranite with nebulitic, patchy and schlieren textures (Figs 2 and 3). We use the term banded type I where the bands of recrystallized quartz, plagioclase and/or K-feldspar have sharp boundaries and appear macroscopically monomineralic (Figs 2a and 3c); the term banded type II is used where the recrystallized K-feldspar bands appear macroscopically monomineralic, but have diffuse boundaries, and bands of quartz and plagioclase are ill defined (Figs 2b, e and 3a, b). We use the term nebulitic for nearly homogeneous distribution of phases without obvious macroscopically identified preferred mineral orientation, in places resembling a fine-grained granite (Figs 2c, e and 3a–c, e); patchy for irregular distribution of predominantly felsic domains (Fig. 3a); and schlieren for types with polymineralic or monomineralic bands elongated parallel to foliation within a nearly isotropic matrix (Figs 2d, e and 3a, b, d). In most metasedimentary migmatites (e.g. Milord et al., 2001), the schlieren are dominated by biotite, but in the metagranite migmatites studied, the schlieren are composed mostly of felsic minerals, the remnants of K-feldspar-rich bands and of quartz-rich domains (Figs 2d, e and 3b).

Two deformation fabrics are recognized in the area. The first fabric is seen as a subhorizontal foliation S1, which is folded into centimetre- to metre-scale open to isoclinal folds with north–south-trending subvertical axial planes (Fig. 2e). The second fabric is a new vertical north–south-trending foliation S2, parallel to the north– south-trending subvertical axial planes. The subhorizontal foliation S1 is defined by augen or bands of recrystallized K-feldspar, quartz and plagioclase, progressively passing through banded orthogneiss to macroscopically isotropic orthogneiss across the strain gradient. A similar textural sequence was described by Chopin et al. (2012b). In the folded banded orthogneiss, the monomineralic bands are commonly narrower in the limbs than in the hinges. This implies that the vertical fabric results from post-buckle flattening (Fig. 2e). Ultimately, the banded orthogneiss in the F2 limbs commonly passes gradually, on a centimetre-scale, from narrow-banded into patchy, schlieren and nebulitic types (Fig. 2e). The rock structure therefore shows a crenulation cleavage, where the banded orthogneiss is progressively transformed into various types of migmatite in high-strain zones parallel to the axial planes of the F2 folds. The boundaries between the migmatite types are commonly gradational, on a scale of several centimetres, parallel to the subvertical zones. The internal structure of the S2 foliation in the migmatite types is defined by orientation of polymineralic bands, schlieren and locally by quartzo-feldspathic patches. In the nebulitic orthogneiss the macroscopic foliation is illdefined, but on scale of a several centimetres it locally contains schlieren or ‘ghosts’ of banded orthogneiss layering parallel to the regional S2 fabric (Fig. 2e).

### *Sr–Nd isotopes*

To attempt to decipher the genetic link between rock types, Sr–Nd isotope values were measured in five rock samples: in the banded type II, banded type II to schlieren, nebulitic to schlieren and nebulitic (Fig. 4, Table 1). All initial isotopic ratios are recalculated to 340 Ma, the age of migmatite formation (Lange et al., 2005a). All rock types reveal relatively similar, negative e340 Nd values between −4.6 and −6.7. The two metagranite samples (banded type II and banded type II to schlieren) show most negative e340 Nd values of −6.4 and −6.2. The data correspond to fairly uniform two-stage Nd depleted-mantle model ages (TDM Nd .

1.42–1.56 Ga), which are consistent with other orthogneiss occurrences in the Bohemian Massif. Initial Sr ratios are similar for all metagranite types, ranging from 0.7236 to 0.7258. These values concur with previously published data (Kröner et al., 2001; Lange et al., 2005a)

## PETROGRAPHYAND MINERAL CHEMISTRY

Seventy thin sections were studied, from which we show petrographic and mineral chemistry data from six representative samples collected at locality S841 (50°13'56.6"N, 16°45'16.2"E; location of the thin sections is indicated in Fig. 3, and representative mineral analyses are presented in Tables 2–7). The textures and petrography are documented in Figs 5 and 6, and typical garnet, phengite and biotite compositions are shown in Figs 7 and 8. Trends in mineral composition or zoning in the text are marked with '.>', and '–' designates a range of mineral compositions; p.f.u. is per formula unit. We use white mica (Wm) where we refer to a diagram or to the rocks, because in diagrams and in rocks the composition of white mica changes continuously from phengite to muscovite. We use phengite (Ph) where the observed white mica has phengitic composition and where we refer to the observed HP mineral assemblage. Abbreviations for mineral endmembers are alm = Fe/(Ca + Fe + Mg + Mn), py = Mg/(Ca + Fe + Mg +Mn), grs = Ca/(Ca + Fe + Mg + Mn), sps = Mn/(Ca + Fe + Mg + Mn), XFe = Fe/(Fe + Mg), an = Ca/(Ca + Na + K), and ab = Na/(Ca + Na + K).

### Microstructures

The mineral assemblage in all rock types is K-feldspar, plagioclase, quartz, phengite, biotite, garnet and titanite. Accessory minerals are apatite, zircon, ilmenite and rutile. The major difference between the metagranite types is the spatial distribution of felsic phases and micas. In the banded type I rocks, K-feldspar and quartz form monomineralic layers of recrystallized grains, which alternate with plagioclase-rich bands (Fig. 5a) that have numerous interstitial white mica and some biotite at grain boundaries. White mica associated with biotite also forms bands or aggregates arranged subparallel to the foliation. These features are typical of solid-state mylonitization of granite (Schulmann et al., 2008). The microstructure of the banded type II orthogneiss is characterized by the K-feldspar bands having interstitial cuspate quartz, plagioclase droplets and myrmekite-like aggregates along the grain boundaries (Figs 5b, e–h and 6a). Similarly, interstitial plagioclase and K-feldspar with typically low dihedral angles <30° occur in the quartz bands. With increasing proportion of interstitial phases, the originally monomineralic bands appear as polycrystalline aggregates, rich in K-feldspar, whereas the plagioclase- and quartz-rich bands are ill-defined. K-feldspar grains show cuspate–lobate boundaries with respect to plagioclase and quartz, as well as myrmekite-like aggregates (Fig. 6a), the grain size of which increases from the contact of K-feldspar to the matrix. There, it appears as an equigranular aggregate of plagioclase and quartz. In the schlieren and patchy rock types, it is sometimes possible to define K-feldspar, quartz-rich and mica-rich domains within a matrix characterized by a nearly homogeneous distribution of phases (Fig. 5c). In the nebulitic rock types, the distribution of the felsic phases and micas is nearly homogeneous (Fig. 5d). Large K-feldspar, plagioclase and quartz grains (up to 0_5 cm in diameter) occur in places in some schlieren, patchy and nebulitic types, and in places contain inclusions of phengite (Figs 5d and 6b).

## Textural relations of phengite, biotite, garnet, rutile, titanite and ilmenite

In the banded types of metagranite, phengite associated with biotite tends to appear in bands or elongated aggregates parallel to the S2 foliation, whereas in the schlieren, patchy and nebulitic types, it forms randomly distributed large flakes (Fig. 6a–c). In all the types, phengite is found also at contacts with K-feldspar, quartz, plagioclase and myrmekite (Fig. 6a and b), and locally it is included in large K-feldspar, plagioclase or quartz grains (Fig. 6b). These microstructural changes are associated with destruction of monomineralic banding and with coarsening, and these microstructural changes occurred in the presence of phengite. In places, phengite crystals tend to be larger than biotite crystals (Fig. 6c). The phengite grains commonly have small to large biotite grains at their margins or along their cleavage, and in places appear in crystallographic continuity with biotite (Fig. 6c and d). These features suggest biotite formation at the expense of phengite. Garnet (<2 vol. %) is small (up to 500 lm), has irregular shapes and commonly is fragmented (Figs 6c and 7). Garnet occurs at contacts with K-feldspar, plagioclase, quartz and biotite. Some garnets are surrounded by clusters of biotite, suggesting that these biotites formed at the expense of garnet (Fig. 6c). Titanite appears as clusters or individual grains associated with, or included in, phengite and biotite, and also as individual grains in the quartzofeldspathic matrix.

Titanite cores commonly contain rutile, whereas the rims are locally surrounded by ilmenite (Fig. 6e), suggesting the crystallization succession rutile to titanite to ilmenite.

## Mineral chemistry

Garnet composition in all rock types is dominated by almandine and grossular with minor pyrope and spessartine components. Some of the core regions are chemically homogeneous. The garnet rims are zoned with increasing almandine content, decreasing grossular content and XFe, and commonly with slightly decreasing or nearly constant spessartine and slightly increasing or nearly constant pyrope contents. The most common compositions in cores are grs0.40–0.45, alm0.48–0.55, py0 .01–0.02, sps0.01–0.02, XFe 0.96–0.98 and at rims grs0.25–0.35, alm0.60–0.65, py0.03–0.08, sps0.01–0.04, XFe 0.90–0.92 (Fig. 7). Some garnet crystals show rims and/or narrow, crack-like zones or patches that interrupt the main zoning patterns. In these zones, spessartine, almandine and pyrope contents are higher and grossular content is lower compared with the composition of the host garnet, and the individual endmember values can reach grs0.18, alm0.68, py0.09, sps0.25, XFe 0.87. When garnet is fragmented, garnet zoning is cut through the core region and there is no zoning towards the fragmented edges, even at contact with biotite (Fig. 7c and e).

Phengite shows in all rock types zoning from core to rim, with decreasing Si, Ti and increasing XFe (Fig. 8a). The most phengitic mica with Si = 3.40 p.f.u. is included in large K-feldspar grains; matrix phengite core composition ranges are Si = 3.27–3.40 p.f.u., Ti = 0.02–0.10 p.f.u. and XFe = 0.47–0.55, exceptionally XFe = 0.66–0.70. The rim compositions are Si = 3.10–3.25 p.f.u., Ti = 0.01–0.08 p.f.u. and XFe = 0.50–0.70, exceptionally 0.80.

Matrix biotite has similar composition to biotite at contacts or in crystallographic continuity with phengite with XFe = 0.65–0.75, Ti= 0.10–0.25 p.f.u. and Mn = 0.00– 0.03 p.f.u. (Fig. 8b). In contact with, or in close proximity to, garnet, biotite XFe and Mn contents tend to be higher and Ti contents tends to be lower than for matrix biotite (XFe = 0.72–0.82, Ti = 0.00–0.13 p.f.u., Mn = 0.02–0.03 p.f.u.).

Plagioclase included with phengite in K-feldspar blasts is albite (an = 0.06–0.10); cores of matrix plagioclase tend to be albite (an = 0.05–0.10) whereas rims are oligoclase (an = 0.10–0.23). Anorthite content in plagioclase in myrmekite varies between 0.14 and 0.23.

K-feldspar has negligible anorthite content and cores tend to be more rich in albite (ab =

0.08–0.13) compared with rims (ab = 0.01–0.10).

## Interpretation of petrography and mineral compositions

Petrography and mineral composition are summarized and the significant features are interpreted, to give context to the modelling in the subsequent mineral equilibria section. The major difference between the metagranite types is the spatial distribution of felsic phases, which is characterized by the following features: (1) monomineralic layers of recrystallized K-feldspar, quartz and plagioclase in the banded type I orthogneiss; (2) the occurrence of few interstitial phases in the originally monomineralic layers of the banded type II orthogneiss; (3) the increasing amount of interstitial phases in the schlieren and patchy rock types; (4) the nearly homogeneous distribution of phases in the nebulitic rock type. The occurrence of interstitial grains and cuspate forms of feldspar are regarded as a result of dissolution–reprecipitation, promoted by the presence of melt and possibly also by crystallization from melt (Sawyer, 2001; Hasalová et al., 2008b; Holness & Sawyer, 2008; Závada et al., 2018). The higher amount of interstitial phases leads to the progressive disappearance of monomineralic banding from the banded type I orthogneiss, through the banded type II orthogneiss to schlieren, patchy and nebulitic migmatite, and is thus interpreted as a result of more pronounced dissolution– reprecipitation and possibly also crystallization from melt (Hasalová et al., 2008b). It is not implied that this sequence of increased textural modification relates to the amount of melt that the rocks contained at one time. The minerals that reprecipitated or crystallized from melt probably relate to the cumulative effect of the textural modification in the presence of melt.

The main characteristics of the metagranitic migmatite involve virtually the same mineral assemblage garnet+phengite+biotite+titanite+K-feldspar+plagioclase+quartz in all the studied textural types, with very similar mineral compositional trends. Phengite occurs in all the studied types, and is texturally stable with interstitial plagioclase and quartz in K-feldspar-rich bands (Fig. 6a). Phengite and albite occur as inclusions in coarse K-feldspar and quartz grains, where phengite has the highest measured Si (3.40 p.f.u.), indicating that both minerals were protected from late re-equilibration (Fig. 6b). The microstructural criteria for the presence of crystallized melt as discussed, with the presence of phengite and albite, indicate melt-assisted textural changes in migmatites at HP conditions.

Garnet element mapping shows the presence of fracture-like zones through the garnet cores that have diffuse boundaries and composition corresponding to the intermediate zone in garnet (e.g. Fig. 7c). Such areas may represent 'fractures' where garnet re-equilibrated with a melt-present assemblage, whereas core compositions may still partly reflect prograde garnet growth. This diffuse zoning contrasts with sharp zoning in areas where garnet is next to biotite, indicating that diffusion rate and therefore also temperature was lower when biotite grew at the expense of garnet compared with the temperature of the garnet diffuse zoning and fracture-like zone formation.

The retrograde textural features involve growth of biotite around phengite and along its cleavage, and local replacement of phengite in crystallographic continuity (Fig. 6d). Biotite is clustered around garnet, which also indicates replacement of garnet by biotite (Fig. 6c).

## FORWARD PHASE DIAGRAM MODELLING

A forward phase diagram modelling approach was used to calculate the mineral assemblages and mineral compositions that might have equilibrated in metagranite. We aim to find the conditions that allow for the presence of granitic melt in the field of phengite stability, as well as the conditions under which the observed migmatite formation and migmatite textural variation occurred. However, modelling of anatexis of quartzofeldspathic rocks at eclogite-facies conditions requires discussion of (1) the stability of clinopyroxene and (2) whether the granitic melt model can be used, given that it was not calibrated specifically for high-pressure conditions.

Experiments show that jadeite is stable in quartzofeldspathic rocks at (U)HP conditions, so the very rare presence of clinopyroxene in rocks with quartzofeldspathic compositions in (U)HP terranes (Young & Kylander-Clark, 2015) is enigmatic. The common absence of clinopyroxene in quartzo-feldspathic rocks led to suggestions that quartzo-feldspathic rocks do not typically equilibrate at (U)HP conditions, or equilibrate but re-equilibrate on decompression. Many studies argue for non-equilibration (Young & Kylander-Clark, 2015; Palin et al., 2017) and explain it by kinetic arguments (Koons et al., 1987), or by lack of fluid or deformation (Rubie, 1986). Samples of granitic orthogneiss studied by (Chopin et al. 2012b) occur 1 km to the NE of the study area. These samples contain phengite and Carich garnet together with dynamically recrystallized feldspar and quartz, indicating that these rocks were highly deformed and equilibrated at HP conditions. The samples from this study also lack clinopyroxene, and contain phengite and high-Ca garnet together with recrystallized feldspar and quartz. Therefore, they experienced strong deformation, inferred to have occurred at high pressure and in the presence of melt, so nonequilibration owing to the absence of fluid does not explain the absence of clinopyroxene. First, we explore phase diagrams with clinopyroxene stable, then we assume that clinopyroxene did not nucleate and phase diagrams are calculated on that basis.

The haplogranitic melt model used (White et al., 2014) is not specifically calibrated for modelling at highpressure conditions. However, it has been used successfully to explain observed mineral equilibria in quartzo-feldspathic rocks at HP_HT conditions (Štípská et al., 2008; Hopkins et al., 2010; Lexa et al., 2011; Nahodilová et al., 2014). Therefore, we also used the melt model at lower temperature conditions at high pressure and evaluated the results. The wet solidus cannot be calculated with the melt model above 1.2 GPa, therefore for melt-bearing assemblages we take into account only calculations above the experimentally determined P–T conditions of the wet solidus (e.g. 600–650°C). The results are used only below 1.8 GPa and well above the wet solidus. The H2O amount was first set to correspond to crystal-bound H2O in micas for the H2O-absent assemblage typical of a granite at MP_MT conditions. With such an amount of H2O, a P–T pseudosection was calculated to discuss stable mineral assemblages and the position of the solidus that would correspond to metamorphic assemblages and conditions of melting of a granite in a closed system. We infer that melt or H2O generated in deeper and hotter rocks can migrate upwards, therefore the effect of addition of granitic liquid on mineral equilibria was assessed. In the first approach an amount of melt is reintegrated into the whole-rock composition to discuss its influence on mineral equilibria and on the solidus position in a P–T pseudosection. It is shown that this approach is almost equivalent to just adding H2O to the bulk composition. So, in the second approach the amount of H2O, M(H2O), is varied, as a proxy for addition of hydrous granitic melt, and the influence on mineral equilibria and solidus position is discussed in P–M(H2O) and T–M(H2O) phase diagrams. Simple addition of H2O fluid as a physical process is excluded as the rocks are at a P–T well above their wet solidus. Results from each diagram are discussed in the context of the observations on the rocks studied.

## Calculation methods

Phase diagrams were calculated using THERMOCALC 3.40 with dataset 6.2 (Holland & Powell, 2011, 6 February 2012 upgrade) in the system MnO-Na2O-CaO-K2O-FeO-MgO-Al2O3-SiO2-H2O-TiO2-O (MnNCKFMASHTO). Activity models except for clinopyroxene (Green et al., 2007) are those used in White et al. (2014); albite, rutile, kyanite, sillimanite, quartz, titanite, and H2O are considered pure endmembers. For calculations, the whole-rock ICP-MS composition of the banded type II orthogneiss 2E2 was used. The Fe2O3 was calculated as a difference from a total iron oxide expressed as Fe2O3 (ICP-MS) and FeO determined by dichromate titration (performed in the Bureauveritas Laboratories). The resulting whole-rock compositions after addition of granitic liquid (first approach) or for H2O variation (second approach) used for calculation are presented as insets in each diagram, in mole per cent normalized to 100%. Mineral composition isopleths were plotted for garnet and white mica to discuss the P–T conditions of their equilibration. The diagrams calculated are used also for the discussion of the evolution of the other metagranite types, because of their closely similar bulk compositions. Phase diagrams with crystal-bound H2 O, stable and metastable with respect to omphacite The H2O content was set to 1.70 mol %, allowing stability of the H2O-free and garnet-free assemblage Wm+Bt+Ksp+Pl+Qz+Ilm+Rt at MP_MT conditions, representing the major minerals typically stable in a granite. A calculated phase diagram with omphacite stable (Fig. 9) fails to explain the main assemblage of the rocks in several ways. It shows stability of omphacite above 1.2 GPa and 600°C, but there is no evidence of omphacite in the rocks. It does not predict stability of titanite, which is present in all the rock types. The highest modelled isopleth for Si-in-phengite is Si = 3.30 p.f.u., whereas the measured phengite has up to Si = 3.43 p.f.u.; the highest modelled grossular isopleth is 40%, which only partly corresponds to up to 45% of grossular measured. A diagram excluding omphacite, in other words metastable with respect to omphacite (Fig. 10a-d), shows larger stability fields of biotite, and plagioclase and albite are stable to high pressure. The grossular and Si-in-phengite isopleths increase up pressure within the biotite stability field. However, the diagram still fails to explain the coexistence of titanite with phengite. The calculations in the next subsection, involving melt addition, show how the observed mineral assemblage and mineral compositions can be accounted for. Figure 10a–d can be used to explain part of the prograde path in the fields of Grt–Wm–Bt–Rt–Pl/Ab, with garnet cores compatible with prograde growth from 1.2 GPa and 600°C to 1.7 GPa and 700–750°C, explaining also rutile inclusions in titanite, the composition of phengite and the start of partial melting.

P–T diagram with added melt Assuming that melt generated in deeper and hotter rocks can migrate upwards, the melt composition can be taken from calculations above the Liq-in line indicated by the star in Fig. 10a and reintegrated in the whole-rock composition. The resulting P–T phase diagram with 5 mol % of integrated melt is shown in Fig. 10e–h. The diagram shows the $H_2O$-undersaturated solidus at higher pressure compared with the phase diagram with 1_70 mol % $H_2O$, a large stability field of titanite at the HP and LT side of the diagram, and the titanite–rutile transition running from 600°C, 0.7 GPa to 800°C, 2.1 GPa. The diagram shows a stability field of Grt–Wm–Bt–Ttn–Pl–Kfs–Liq between 0.8 and 1.9 GPa and 620 and 770°C, corresponding to the observed assemblage. The calculated compositional isopleths at conditions of 1.5–1.7 GPa and 690–740°C are compatible with the measured core compositions of garnet and phengite. The trends of zoning to the rims of garnet and white mica are compatible with decompression through _700_C and _1_0 GPa, with local equilibration down to the ilmenite stability field.

The P–T conditions from the migmatite are compatible with P–T estimates for other neighbouring rocks such as mylonitic orthogneiss (1.9 GPa, 700–770°C, Chopin et al., 2012b) and eclogite (2.2 GPa, 700–800°C, Štípská et al., 2012), for which a retrograde path from 0.9 GPa, 730°C to 0.5–0.6 GPa, 550–650°C was calculated (Bröcker & Klemd, 1996). This also supports the use of the haplogranitic melt model in the equilibria modelling above the wet solidus and in the range of 0.5–2.5 GPa.

## P –M (H2 O) diagram

The dependence of the assemblage and mineral composition on the amount of $H_2O$ in the whole-rock composition (taken as a proxy for the amount of $H_2O$ brought in by the hydrous granitic melt composition) was tested in a P–M($H_2O$) diagram at T = 730°C (Fig. 11). The major features of the diagram involve a Liq-in line running from M($H_2O$) = 0.85 and 0.7GPa to M($H_2O$) = 4.15 and 2.5GPa. The titanite and rutile stability are dependent on M($H_2O$) in the Liq-absent fields and are dependent on pressure in the Liq-present fields.

Titanite in the Liqpresent fields is stable at P>1.5GPa and M($H_2O$)>0.2. The compositional isopleths of Si-in-phengite and grossular and almandine in garnet are mostly dependent on the M($H_2O$) in Liq-absent fields, but are almost independent of the M($H_2O$) in the Liq-present fields, even if the equivalent amount of melt increases with higher M($H_2O$).

The diagram shows stability of the observed assemblage Grt–Wm–Bt–Ttn–Pl–Kfs–Liq for M($H_2O$)> 2.05 over the pressure interval 1.6–1.9 GPa. This is in agreement with results from the pseudosection where melt was added (Fig. 10e) and shows that using M($H_2O$) as a proxy works well. The calculated compositional isopleths in this field are in agreement with most commonly measured core compositions of phengite (Si = 3.40–3.30 p.f.u.) and garnet (grs = 0.45–0.40, alm = 0.47–0.53). The independence of the compositional variables from M($H_2O$) precludes estimation of the amount of melt in individual rock types and explains the identical assemblage and very similar mineral compositions in texturally very different types of migmatites.

In Fig. 11, at the estimated P–T of the start of melt infiltration, a horizontal arrow indicates the effect of the infiltration. The starting position at M($H_2O$) = 1.65 mol %, indicates an $H_2O$ content that allows the stability of the observed mineral assemblage with rutile and without titanite, as titanite is observed to overgrow rutile. This M($H_2O$) is reasonable for a granite, and allows stability of $H_2O$-free and garnet-free assemblage Wm-Bt-Ksp-Pl-Qz-Ilm-Rt at MP_MT conditions (see Fig. 10). From this starting position, infiltration will increase $H_2O$ in the mineral assemblage by the growth of white mica from 10 to 15 mol % (not shown). The arrow then reaches the solidus, which implies the first appearance of melt in the system. Further along the horizontal arrow, the hydrous mineral proportions do not change, with the $H_2O$ increasing in the system by the increase in melt proportion. The position of any rock along the arrow depends on the amount of melt that the rock contains at the instant the infiltration of melt proceeds; with no melt in the rock, the rock sits at the solidus. The vertical part of the arrow represents what happens during decompression if infiltration continues to occur. The system (rock plus infiltrating melt) remains in the higher M($H_2O$) part of the diagram above the solidus. In fact, given that melt migration is likely to vary as decompression occurs, the rock plus melt may lie anywhere from the solidus to the higher M($H_2O$) part of the diagram. With decompression, the hydration of the rock itself follows the solidus to lower water content. Ultimately, and not represented in this figure, the final water content of a rock depends on the amount of melt remaining in the rock that crystallizes there once infiltration ceases. This amount of melt is likely to be unrelated to the cumulative infiltration that the rock has experienced.

The dependence of the assemblage and mineral composition on the amount of H2O for Liq-absent fields and its independence for Liq-present fields is also shown in the T–M(H2O) diagram at P = 1.7 GPa (Fig. 12). The major feature of this diagram is its division by a Liq-in line running from M(H2O) = 0.86 and 800°C to M(H2O) =3.8 and 650°C. The stability of titanite, rutile and biotite in the Liq-absent fields is dependent mainly on the amount of H2O, whereas in the Liqpresent fields it depends mainly on temperature. Similarly, the mineral compositional isopleths of Si-inphengite and grossular are dependent on M(H2O) in Liq-absent fields, but independent of M(H2O) in Liqpresent fields, even if the amount of melt increases with higher M(H2O). The observed assemblage Grt–Ph–Bt–Ttn–Pl–Kfs–Liq is stable for M(H2O)>_2_1 mol % H2O, and the core mineral compositions of phengite (Si = 3.4–3.3 p.f.u.) and garnet (grs = 0.45–0.40, alm = 0.47–0.53) roughly correspond to a temperature range of 700–740°C.

## DISCUSSION AND CONCLUSIONS

Metagranitic migmatites occur in the vicinity of eclogite boudins within a continental subduction wedge exhumed in the form of a mantled gneiss dome in the Bohemian Massif. Our field study shows that the transition of metagranite to migmatite occurs in localized zones with a subvertical fabric, which formed by folding and local transposition of a precursor shallow-dipping fabric (Fig. 2). On the basis of grain-scale modification of the originally monomineralic bands into polymineralic aggregates, we argue that the cause was grainscale melt migration through the rocks (Fig. 5). The melt migration, although being pervasive at a grain scale, varied in intensity on a scale of several centimetres to several metres.

Textural variations in the migmatite appear to reflect variable intensity of melt–rock interaction (Fig. 2). These variations are similar to the migmatite transitions described in metagranite by Hasalová et al. (2008a, 2008b, 2008c), who also interpreted them to be the result of pervasive grain-scale melt migration through the rock. Whereas such melt migration has been advocated to occur in leucosome generation (Sawyer, 2001), it was first considered as a larger scale melt transport mechanism in felsic crust by Hasalová et al. (2008a, 2008b, 2008c) and Závada et al. (2018). Via forward-modelling of the mineral assemblages, migmatite formation and the formation of the observed mineral assemblage Grt+Ph+Bt+Ttn+Kfs+Pl+Qz and mineral compositions are inferred to develop only if H2O was added to the rocks (Fig. 10). Being above the P-T conditions of the wet solidus, the hydrating fluid must have been a hydrous melt. Melt infiltration occurred at about 1.5–1.7 GPa and 690–740°C and may have continued to occur down to 0.7–1.0 GPa during the retrograde history. Modelling also shows that the melt-bearing mineral assemblage and mineral compositions are almost independent of the amount of melt migrating through the system, thus accounting for virtually identical mineral assemblages and mineral compositions, independent of the migmatite type, but also hampering estimation of the amount of melt based on mineral equilibria (Figs 11 and 12).

In a larger-scale tectonic context, according to the subduction–relamination–extrusion model of Chopin et al. (2012a) and Maierová et al. (2014, 2018), the original shallow-dipping fabric is related to progressive burial. During burial, solid-state deformation and incipient melting occurred at 1.9 GPa and 700–770°C (Chopin et al., 2012b), whereas relatively low-strain domains remained melt-free (Fig. 13a and b). The steep fabric is related to crustal-scale upright folding of the orogenic crustal layers, where the lower crustal HP rocks are exhumed in cores of crustal-scale antiforms, and by this process the high-pressure rocks are juxtaposed with mid-crustal rocks that have maximum MP conditions occurring in crustal-scale synforms (Fig. 13a; see also Štípská et al., 2004, 2012; Chopin et al., 2012a). In the interior of this granulite–gneiss dome subvertical deformation zones developed, parallel to an axial planar cleavage. In these zones of high deformation, the mylonitic and migmatitic rocks were flowing considerably more rapidly towards the Earth's surface compared with surrounding rocks (Štípská et al. 2004), while the whole dome was growing en bloc and all levels were exhuming simultaneously (Štípská et al., 2012). This tectonic scenario is portrayed in the schematic illustration in Fig. 13a. The extrusion may be enhanced by delamination of the deeply subducted lithosphere (Willner et al., 2002).

In the area studied, migmatites develop along vertical deformation zones associated with upright folding of prograde shallow-dipping fabrics (Fig. 13a and c). The mineral assemblage and mineral equilibria modeling of the migmatites suggest melting at 1.5–1.7 GPa and 690–

740°C, and partial re-equilibration to 0.7–1.0 GPa, indicating that these zones were melt-bearing, with melt infiltration during exhumation from 60km to 25km (assuming lithostatic pressure and density of metagranite 2.8gcm−3) (Fig. 13a, c and e). In other words, the migmatites and the adjacent rocks not being infiltrated by melt were transported from a deeper structural level of 60km to 25km in a more or less coherent 2 km wide structural domain in the core of a crustal-scale antiform (Figs 1 and 13a; Štípská et al., 2012). Via the progression of this deformation zone, the high-pressure metagranites were juxtaposed with midcrustal rocks occurring in a crustal-scale synform 5 km to the east (Figs 1 and 13a; see also Štípská et al., 2004, 2012). Therefore, incipient melt infiltration probably already occurred at a depth of 60km in the shallow-dipping fabric. More significant melt infiltration occurred during transition to and within the subvertical fabric, starting at a depth of 60 km. Thus, melt infiltration affected a significantly larger volume of rock in which the subvertical fabric S2 was localized, compared with the volume of rock with the shallow-dipping fabric S1, occurring in the low-strain domains of the second deformation (Fig. 13a–c). Melt infiltration led to a shift of the $H_2O$-undersaturated solidus to higher pressure and allowed the localized zones infiltrated by melt to evolve as melt-bearing (Fig. 13e), whereas the zones that were not infiltrated by melt evolved as melt-free (Fig. 13d). The source of the hydrous melt is probably melt lost from similar rocks buried deeper, where they crossed their $H_2O$-undersaturated solidus (Fig. 10a). This is supported by the similarity of the Sr and Nd isotope characteristics of all the metagranite types, indicating the same protolith and that the infiltrating melt had its source in similar rocks (Fig. 4). Melt transport was not localized in leucosomes or a network of dykes, but melt migrated at grain scale. These heterogeneously developed subvertical melt-bearing zones induced a major rheological weakening (Rosenberg & Handy, 2005) that promoted further localized deformation and may have aided exhumation of deeply subducted crust (Závada et al., 2018). In the model presented, the melt is collected into dykes only at upper crustal levels, where syntectonic granite to granodiorite sills intruded the interface between high-grade dome and low-grade metasediments parallel to axial planes of large upright folds or sub-domes (Fig. 13a) (Závada et al., 2017).

## SUMMARY

- Subducted eclogite-facies granitic crust dominated by augen and banded metagranite types has subvertical localized domains with migmatite textures ranging from stromatic to schlieren, to patchy and nebulitic.
- Destruction of mineral banding of metagranites happened through grain-scale melt migration. Grain-scale melt migration may be a dominant process of melt transfer in metagranitic migmatites, which is in contrast to a localized melt transport paradigm based on observations in metasedimentary migmatites.
- The observed assemblage in all the metagranite migmatite types Grt+Ph+Bt+Ttn+Kfs+Pl+Qz is stable in the modeled phase diagrams if $H_2O$ or hydrous melt is added to the whole-rock composition. This implies that melting is not driven by dehydration melting but by addition of $H_2O$. Being above the P-T conditions of the wet solidus the hydrating fluid must have been a hydrous melt and the process cannot be called fluid-fluxed melting but is melt-fluxed melting.
- Migmatite textures result from variable degree of melt–rock interaction starting at 1.7 GPa and 730°C and ending at 0.7-1.0 GPa, implying that the granitic crust underwent melting at eclogite-facies conditions, which has, so far, rarely been reported.
- Melt infiltration along steep deformation zones facilitated exhumation of HP rocks in cores of antiforms.

## ACKNOWLEDGEMENTS

Vojtech Erban from the Czech Geological Survey is thanked for performing Sr–Nd analyses. We thank A. Willner, G. Solar, Yu Shengyao and an anonymous reviewer for constructive comments, and A. Skelton for his editorial work.

## FUNDING

This work was supported by the Czech Science Foundation (grant number 16-17457S to P.S.).

## REFERENCES

Anczkiewicz, R., Szczepa_nski, J., Mazur, S., Storey, C., Crowley, Q., Villa, I. M., Thirlwall, M. F. & Jeffries, T. E. (2007). Lu–Hf geochronology and trace element distribution in garnet: implications for uplift and exhumation of ultra-high pressure granulites in the Sudetes, SW Poland. Lithos 95, 363–380.

Auzanneau, E., Vielzeuf, D. & Schmidt, M. W. (2006). Experimental evidence of decompression melting during exhumation of subducted continental crust. Contributions to Mineralogy and Petrology 152, 125–148.

Bröcker, M. & Klemd, R. (1996). Ultrahigh-pressure metamorphism in the _Snie_znik Mountains (Sudetes, Poland): P–T constraints and geological implications. Journal of Geology 104, 417–433.

Bröcker, M., Klemd, R., Cosca, M., Brock, W., Larionov, A. N. & Rodionov, N. (2009). The timing of eclogite facies metamorphism and migmatization in the Orlica–_Snie_znik complex, Bohemian Massif: constraints from a multimethod geochronological study. Journal of Metamorphic Geology 27, 385–403.

Bröcker, M., Klemd, R., Kooijman, E., Berndt, J. & Larionov, A. (2010). Zircon geochronology and trace element characteristics of eclogites and granulites from the Orlica–Snieznik complex, Bohemian Massif. Geological Magazine 147, 339–362.

Brown, M. (2007). Crustal melting and melt extraction, ascent and emplacement in orogens: mechanisms and consequences. Journal of the Geological Society, London 164, 709–730.

Chopin, F., Schulmann, K., Skrzypek, E., Lehmann, J., Dujardin, J. R., Martelat, J. E., Lexa, O., Corsini, M., Edel, J. B., Štípská , P. & Pitra, P. (2012a). Crustal influx, indentation, ductile thinning and gravity redistribution in a continental wedge: building a Moldanubian mantled gneiss dome with underthrust

Saxothuringian material (European Variscan belt). Tectonics 31, TC1013.

Chopin, F., Schulmann, K., Štípská , P., Martelat, J. E., Pitra, P., Lexa, O. & Petri, B. (2012b). Microstructural and metamorphic evolution of a high-pressure granitic orthogneiss during continental subduction (Orlica–_Snieznik dome, Bohemian Massif). Journal of Metamorphic Geology 30, 347–376.

Clemens, J. D. & Vielzeuf, D. (1987). Constraints on melting and magma production in the crust. Earth and Planetary Science Letters 86, 287–306.

Collins, W. J., Flood, R. H., Vernon, R. H. & Shaw, S. E. (1989). The Wuluma Granite, Arunta Block, Central Australia—an example of in situ, near isochemical granite formation in a granulite-facies terrane. Lithos 23, 63–83.

Collins, W. J., Huang, H. Q. & Jiang, X. (2016). Water-fluxed crustal melting produces Cordilleran batholiths. Geology 44, 143–146.

Don, J., Dumicz, M., Wojciechowska, I. & Z_ ela_zniewicz, A. (1990). Lihology and tectonics of the Orlica–_Snie_znik Dome, Sudetes: recent state of knowledge. Neues Jahrbuch fu¨ r Geologie und Pala¨ ontologie, Abhandlungen 179, 159–188.

Ferrero, S., Wunder, B., Walczak, K., O'Brien, P. J. & Ziemann, M. A. (2015). Preserved near ultrahigh-pressure melt from continental crust subducted to mantle depths. Geology 43, 447–450.

Fischer, G. (1936). Der Bau des Glatzer Schneegebirges. Jahrbuch der Preußischen Geologischen Landesanstalt 56, 712–732.

Ganzhorn, A. C., Trap, P., Arbaret, L., Champallier, R., Fauconnier, J., Labrousse, L. & Prouteau, G. (2016). Impact of gneissic layering and localized incipient melting upon melt flow during experimental deformation of migmatites. Journal of Structural Geology 85, 68–84.

Green, E., Holland, T. & Powell, R. (2007). An order–disorder model for omphacitic pyroxenes in the system jadeite–diopside–hedenbergite–acmite, with applications to eclogitic rocks. American Mineralogist 92, 1181–1189.

Greenfield, J. E., Clarke, G. L., Bland, M. & Clark, D. J. (1996). In-situ migmatite and hybrid diatexite at Mt Stafford, central Australia. Journal of Metamorphic Geology 14, 413–426.

Hacker, B. R., Kelemen, P. B. & Behn, M. D. (2011). Differentiation of the continental crust by relamination. Earth and Planetary Science Letters 307, 501–516.

Hasalová, P., Janou_sek, V., Schulmann, K., Štípská , P. & Erban, V. (2008a). From orthogneiss to migmatite: geochemical assessment of the melt infiltration model in the Gfo¨ hl Unit (Moldanubian Zone, Bohemian Massif). Lithos 102, 508–537.

Hasalová, P., Schulmann, K., Lexa, O., Štípská , P., Hrouda, F., Ulrich, S., Haloda, J. & T_ycova´ , P. (2008b). Origin of migmatites by deformation-enhanced melt infiltration of orthogneiss: a new model based on quantitative microstructural analysis. Journal of Metamorphic Geology 26, 29–53.

Hasalová, P., Štípská , P., Powell, R., Schulmann, K., Janou_sek, V. & Lexa, O. (2008c). Transforming mylonitic metagranite by open-system interactions during melt flow. Journal of Metamorphic Geology 26, 55–80.

Hermann, J. & Rubatto, D. (2014). Subduction of continental crust to mantle depth: geochemistry of ultrahigh-pressure rocks. In: Holland, H. D. & Turekian, K. K. (eds) Treatise on Geochemistry, 2nd edn. Oxford: Elsevier–Pergamon, Vol. 4, pp. 309–340.

Hermann, J. & Spandler, C. J. (2008). Sediment melts at sub-arc depths: an experimental study. Journal of Petrology 49, 717–740.

Holland, T. J. B. & Powell, R. (2011). An improved and extended internally consistent thermodynamic dataset for phases of petrological interest, involving a new equation of state for solids. Journal of Metamorphic Geology 29, 333–383.

Holness, M. B. & Sawyer, E. W. (2008). On the pseudomorphing of melt-filled pores during the crystallization of migmatites. Journal of Petrology 49, 1343–1363.

Hopkins, M. D., Harrison, T. M. & Manning, C. E. (2010). Constraints on Hadean geodynamics from mineral inclusions in >4Ga zircons. Earth and Planetary Science Letters 298, 367–376.

Huang, W. L. & Wyllie, P. J. (1981). Phase relationships of S-type granite with H2O to 35 kbar: muscovite granite from Harney Peak, South Dakota. Journal of Geophysical Research 86, 10515–10529.

Jacobsen, S. B. & Wasserburg, G. J. (1980). Sm–Nd evolution of chondrites. Earth and Planetary Science Letters 50, 139–155.

Jastrzebski, M. (2009). A Variscan continental collision of the West Sudetes and the Brunovistulian terrane: a contribution from structural and metamorphic record of the Stronie Formation, the Orlica–_Snie_znik Dome, SW Poland. International Journal of Earth Sciences 98, 1901–1923.

Jastrzebski, M., Budzyn, B. & Stawikowski, W. (2019). Cambro-Ordovician vs Devono-Carboniferous geodynamic evolution of the Bohemian Massif: evidence from P–T–t studies in the Orlica–_Snie_znik Dome, SW Poland. Geological Magazine 156, 447–470.

Jastrzebski, M., Zelazniewicz, A., Nowak, I., Murtezi, M. &

Larionov, A. N. (2010). Protolith age and provenance of metasedimentary rocks in Variscan allochthon units: U–Pb SHRIMP zircon data from the Orlica–_Snieznik Dome, West Sudetes. Geological Magazine 147, 416–443.

Jedli_cka, R. & Faryad, S. W. (2017). Felsic granulite with layers of eclogite facies rocks in the Bohemian Massif; did they share a common metamorphic history? Lithos 286, 408–425.

Jochum, K. P., Nohl, U., Herwig, K., Lammel, E., Stoll, B. & Hofmann, A. W. (2005). GeoReM: a new geochemical database for reference materials and isotopic standards. Geostandards and Geoanalytical Research 29, 333–338.

Jung, C., Jung, S., Nebel, O., Hellebrand, E., Masberg, P. & Hoffer, E. (2009). Fluid-present melting of meta-igneous rocks and the generation of leucogranites—Constraints from garnet major- and trace element data, Lu–Hf whole rock-garnet ages and whole rock Nd–Sr–Hf–O isotope data. Lithos 111, 220–235.

Klemd, R. & Bröcker, M. (1999). Fluid influence on mineral reactions in ultrahigh-pressure granulites: a case study in the _Snie_znik Mts. (West Sudetes, Poland). Contributions to Mineralogy and Petrology 136, 358–373.

Koons, P. O., Rubie, D. C. & Fruch-Green, G. (1987). The effects of disequilibrium and deformation on the mineralogical evolution of quartz diorite during metamorphism in the eclogite facies. Journal of Petrology 28, 679–700.

Korhonen, F. J., Saito, S., Brown, M. & Siddoway, C. S. (2010). Modeling multiple melt loss events in the evolution of an active continental margin. Lithos 116, 230–248.

Košuličová , M. & Štípská , P. (2007). Variations in the transient prograde geothermal gradient from chloritoid–staurolite equilibria: a case study from the Barrovian and Buchan-type domains in the Bohemian Massif. Journal of Metamorphic Geology 25, 19–35.

Kröner, A., Jaeckel, P., Hegner, E. & Opletal, M. (2001). Single zircon ages and whole rock Nd isotopic systematics of early Palaeozoic granitoid gneisses from the Czech and Polish Sudetes (Jizerske´ hory, Krkono_se Mountains and Orlice–Sn_e_znı´k Complex). International Journal of Earth Sciences 90, 304–324.

Kröner, A., Štípská , P., Schulmann, K. & Jaeckel, P. (2000). Chronological constraints on the pre-Variscan evolution of the northeastern margin of the Bohemian Massif, Czech Republic. In: Franke, W., Haak, V., Oncken, O.& Tanner, D. (eds) Orogenic Processes; Quantification and Modelling in the Variscan Belt. Geological Society, London, Special Publications 179, 175–197.

Kryza, R., Pin, C. & Vielzeuf, D. (1996). High-pressure granulites from the Sudetes (south-west Poland): evidence of crustal subduction and collisional thickening in the Variscan Belt. Journal of Metamorphic Geology 14, 531–546.

Labrousse, L., Duretz, T. & Gerya, T. (2015). H2O-fluid-saturated melting of subducted continental crust facilitates exhumation of ultrahigh-pressure rocks in continental subduction zones. Earth and Planetary Science Letters 428, 151–161.

Lange, U., Bröcker, M., Armstrong, R., Trapp, E. & Mezger, K. (2005a). Sm–Nd and U–Pb dating of high-pressure granulites from the Zlote and Rychleby Mts (Bohemian Massif, Poland and Czech Republic). Journal of Metamorphic Geology 23, 133–145.

Lange, U., Bröcker, M., Armstrong, R., Zela_zniewicz, A., Trapp, E. & Mezger, K. (2005b). The orthogneisses of the Orlica–Snieznik complex (West Sudetes, Poland): geochemical characteristics, the importance of pre-Variscan migmatization and constraints on the cooling history. Journal of the Geological Society, London 162, 973–984.

Lange, U., Bröcker, M., Mezger, K. & Don, J. (2002). Geochemistry and Rb–Sr geochronology of a ductile shear zone in the Orlica–_Snie_znik dome (West Sudetes, Poland). International Journal of Earth Sciences 91, 1005–1016.

Lehmann, J., Schulmann, K., Edel, J. B., Jezek, J., Hrouda, F., Lexa, O. & Chopin, F. (2013). Structural and anisotropy of magnetic susceptibility records of granitoid sheets emplacement during growth of a continental gneiss dome (Central Sudetes, European Variscan Belt). Tectonics 32, 797–820.

Lexa, O., Schulmann, K., Janou_sek, V., Štípská , P., Guy, A. & Racek, M. (2011). Heat sources and trigger mechanisms of exhumation of HP granulites in Variscan orogenic root. Journal of Metamorphic Geology 29, 79–102.

Liew, T. C. & Hofmann, A. W. (1988). Precambrian crustal components, plutonic associations, plate environment of the Hercynian Fold Belt of Central Europe: indications from a Nd and Sr isotopic study. Contributions to Mineralogy and Petrology 98, 129–138.

Lugmair, G. W. & Marti, K. (1978). Lunar initial 143Nd/144Nd: differential evolution line of the lunar crust and mantle. Earth and Planetary Science Letters 39, 349–357.

Maierová , P., Lexa, O., Schulmann, K. & Štípská , P. (2014). Contrasting tectono-metamorphic evolution of orogenic lower crust in the Bohemian Massif: a numerical model. Gondwana Research 25, 509–521.

Maierová , P., Schulmann, K. & Gerya, T. (2018). Relamination styles in collisional orogens. Tectonics 37, 224–250.

Masotta, M., Laumonier, M. & McCammon, C. (2018). Transport of melt and volatiles in magmas inferred from kinetic experiments on the partial melting of granitic rocks. Lithos 318–319, 434–447.

Mazur, S. & Aleksandrowski, P. (2001). The Tepla(?)/Saxothuringian suture in the Karkonosze-Izera massif, western Sudetes, central European Variscides. International Journal of Earth Sciences 90, 341–360.

Mazur, S., Kröner, A., Szczepa_nski, J., Turniak, K., Han_zl, P., Melichar, R., Rodionov, N. V., Paderin, I. & Sergeev, S. A. (2010). Single zircon U–Pb ages and geochemistry of granitoid gneisses from SW Poland: evidence for an Avalonian affinity of the Brunian microcontinent. Geological Magazine 147, 508–526.

Mazur, S., Szczepa_nski, J., Turniak, K. & McNaughton, N. J. (2012). Location of the Rheic suture in the eastern Bohemian Massif: evidence from detrital zircon data. Terra Nova 24, 199–206.

Milord, I., Sawyer, E. W. & Brown, M. (2001). Formation of diatexite migmatite and granite magma during anatexis of semi-pelitic metasedimentary rocks: an example from St. Malo, France. Journal of Petrology 42, 487–505.

Murtezi, M. (2006). The acid metavolcanic rocks of the Orlica–_Snie_znik Dome (Sudetes): their origin and tectono-metamorphic evolution. Geologia Sudetica 38, 1–38.

Nahodilová , R., Štípská , P., Powell, R., Ko_sler, J. & Racek, M. (2014). High-Ti muscovite as a prograde relict in high pressure granulites with metamorphic Devonian zircon ages (B_estvina granulite body, Bohemian Massif): consequences for the relamination model of subducted crust. Gondwana Research 25, 630–648.

Palin, R. M., Reuber, G. S., White, R. W., Kaus, B. J. P. & Weller,

O. M. (2017). Subduction metamorphism in the Himalayan ultrahigh-pressure Tso Morari massif: an integrated geodynamic and petrological modelling approach. Earth and Planetary Science Letters 467, 108–119.

Parry, M., S_tı´pska´ , P., Schulmann, K., Hrouda, F., Jez_ek, J. & Kröner, A. (1997). Tonalite sill emplacement at an oblique plate boundary: northeastern margin of the Bohemian Massif. Tectonophysics 280, 61–81.

Pin, C., Gannoun, A. & Dupont, A. (2014). Rapid, simultaneous separation of Sr, Pb, and Nd by extraction chromatography prior to isotope ratios determination by TIMS and MC-ICP-MS. Journal of Analytical Atomic Spectrometry 29, 1858–1870.

Pin, C. & Zalduegui, J. F. S. (1997). Sequential separation of light rare-earth elements, thorium and uranium by miniaturized extraction chromatography: application to isotopic analyses of silicate rocks. Analytica Chimica Acta 339, 79–89.

Pognante, U. (1992). Migmatites and leucogranites of Tertiary age from the High Himalayan Crystallines of Zanskar (NW India): a case-history of anatexis of Palaeozoic orthogneisses. Mineralogy and Petrology 46, 291–313.

Redlin_ska-Marczyn_ska, A., Z_ laz_niewicz, A. & Fanning, C. M. (2016). An insight into a gneiss core of the Orlica–_Snie_znik Dome, NE Bohemian Massif: new structural and U–Pb zircon data. Geological Quarterly 60, 714–736.

Rosenberg, C. L. & Handy, M. R. (2005). Experimental deformation of partially melted granite revisited: implications for the continental crust. Journal of Metamorphic Geology 23, 19–28.

Rubie, D. C. (1986). The catalysis of mineral reactions by water and restrictions on the presence of aqueous fluid during metamorphism. Mineralogical Magazine 50, 399–415.

Sawyer, E. W. (1998). Formation and evolution of granite magmas during crustal reworking: the significance of diatexites. Journal of Petrology 39, 1147–1167.

Sawyer, E. W. (2001). Melt segregation in the continental crust: distribution and movement of melt in anatectic rocks. Journal of Metamorphic Geology 19, 291–309.

Sawyer, E. W. (2010). Migmatites formed by water-fluxed partial melting of a leucogranodiorite protolith: microstructures in the residual rocks and source of the fluid. Lithos 116, 273–286.

Schulmann, K. & Gayer, R. (2000). A model for a continental accretionary wedge developed by oblique collision: the NE Bohemian Massif. Journal of the Geological Society, London 157, 401–416.

Schulmann, K., Martelat, J. E., Ulrich, S., Lexa, O., Štípská , P. & Becker, J. K. (2008). Evolution of microstructure and melt Journal of Petrology, 2019, Vol. 60, No. 3 617 topology in partially molten granitic mylonite: implications for rheology of felsic middle crust. Journal of Geophysical Research 113, B10406.

Skrzypek, E., Bosse, V., Kawakami, T., Martelat, J. E. & Štípská , P. (2017). Transient allanite replacement and prograde to retrograde monazite (re)crystallization in medium-grade metasedimentary rocks from the Orlica–_Snie_znik Dome (Czech Republic/Poland): textural and geochronological arguments. Chemical Geology 449, 41–57.

Skrzypek, E., Lehmann, J., Szczepa_nski, J., Anczkiewicz, R., Štípská , P., Schulmann, K., Kröner, A. & Białek, D. (2014). Time-scale of deformation and intertectonic phases revealed by P–T–D–t relationships in the orogenic middle crust of the Orlica–_Snie_znik Dome, Polish/Czech Central Sudetes. Journal of Metamorphic Geology 32, 981–1003.

Skrzypek, E., Schulmann, K., Štípská , P., Chopin, F., Lehmann, J., Lexa, O. & Haloda, J. (2011a). Tectono-metamorphic history recorded in garnet porphyroblasts: insights from thermodynamic modelling and electron backscatter diffraction analysis of inclusion trails. Journal of Metamorphic Geology 29, 473–496.

Skrzypek, E., Štípská , P., Schulmann, K., Lexa, O. & Lexova´ , M. (2011b). Prograde and retrograde metamorphic fabrics—a key for understanding burial and exhumation in orogens (Bohemian Massif). Journal of Metamorphic Geology 29, 451–472.

Steiger, R. H. & Jäger, E. (1977). Subcommission on geochronology; convention on the use of decay constants in geo- and cosmochronology. Earth and Planetary Science Letters 36, 359–362.

Steltenpohl, M. G., Cymerman, Z., Krogh, E. J. & Kunk, M. J. (1993). Exhumation of eclogitized continental basement during Variscan lithospheric delamination and gravitational collapse, Sudety Mountains, Poland. Geology 21, 1111–1114.

Štípská , P., Chopin, F., Skrzypek, E., Schulmann, K., Pitra, P., Lexa, O., Martelat, J. E., Bollinger, C. & _Za´ _ckova´ , E. (2012). The juxtaposition of eclogite and mid-crustal rocks in the Orlica–_Snieznik Dome, Bohemian Massif. Journal of Metamorphic Geology 30, 213–234.

Štípská , P., Pitra, P. & Powell, R. (2006). Separate or shared metamorphic histories of eclogites and surrounding rocks? An example from the Bohemian Massif. Journal of

Metamorphic Geology 24, 219–240.

Schulmann, P. K. & Kröner, A. (2004). Vertical extrusion and middle crustal spreading of omphacite granulite: a model of syn-convergent exhumation (Bohemian Massif, Czech Republic). Journal of Metamorphic Geology 22, 179–198.

Štípská , P., Schulmann, K. & Powell, R. (2008). Contrastingmetamorphic histories of lenses of high-pressure rocks and host migmatites with a flat orogenic fabric (Bohemian Massif, Czech Republic): a result of tectonic mixing within horizontal crustal flow? Journal ofMetamorphic Geology 26, 623–646.

Štípská , P., Schulmann, K., Thompson, A. B., Je_zek, J. & Kröner, A. (2001). Thermo-mechanical role of a Cambro-Ordovician paleorift during the Variscan collision: the NE margin of the Bohemian Massif. Tectonophysics 332, 239–253.

Stuart, C. A., Daczko, N. R. & Piazolo, S. (2017). Local partial melting of the lower crust triggered by hydration through melt–rock interaction: an example from Fiordland, New Zealand. Journal of Metamorphic Geology 35, 213–230.

Turniak, K., Mazur, S. & Wysoczanski, R. (2000). SHRIMP zircon geochronology and geochemistry of the Orlica–_Snie_znik gneisses (Variscan belt of Central Europe) and their tectonic implications. Geodinamica Acta 13, 1–20.

Weinberg, R. F. & Hasalová, P. (2015). Water-fluxed melting of the continental crust: a review. Lithos 212–215, 158–188.

White, R. W., Pomroy, N. E. & Powell, R. (2005). An in situ metatexite–diatexite transition in upper amphibolite facies rocks from Broken Hill, Australia. Journal of Metamorphic Geology 23, 579–602.

White, R. W., Powell, R. & Holland, T. J. B. (2001). Calculation of partial melting equilibria in the system Na2O–CaO–K2O–FeO–MgO–Al2O3–SiO2–H2O (NCKFMASH).

Journal of Metamorphic Geology 19, 139–153.

White, R. W., Powell, R. & Johnson, T. E. (2014). The effect of Mn on mineral stability in metapelites revisited: new a–x relations for manganese-bearing minerals. Journal of Metamorphic Geology 32, 809–828.

Willner, A. P., Sebazungu, E., Gerya, T. V., Maresch, W. V. & Krohe, A. (2002). Numerical modelling of PT-paths related to rapid exhumation of high-pressure rocks from the crustal root in the Variscan Erzgebirge Dome (Saxony/Germany). Journal of Geodynamics 33, 281–314.

Young, D. J. & Kylander-Clark, A. R. C. (2015). Does continental crust transform during eclogite facies metamorphism? Journal of Metamorphic Geology 33, 331–357.

Závada, P., Calassou, T., Schulmann, K., Hrouda, F., Štípská , P., Hasalová, P., Míková, J., Magna, T. & Mixa, P. (2017). Magnetic fabric transposition in folded granite sills in Variscan orogenic wedge. Journal of Structural Geology 94, 166–183.

Závada, P., Schulmann, K., Racek, M., Hasalová, P., Jerábek, P., Weinberg, R. F., Štípská, P. & Roberts, A. (2018). Role of strain localization and melt flow on exhumation of deeply subducted continental crust. Lithosphere 10, 217–238.