Syn-deformational melt percolation through a high-pressure orthogneiss and the exhumation of a subducted continental wedge (Orlica-Śnieżnik Dome, NE Bohemian Massif)

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ABSTRACT

High-pressure granitic orthogneiss of the south-eastern Orlica–Śnieżnik Dome (NE Bohemian Massif) shows relics of a shallow-dipping foliation, reworked by upright folds and a mostly pervasive N-S trending subvertical axial planar foliation. Based on macroscopic observations, a gradual transition from banded to schlieren and nebulitic orthogneiss was distinguished. All rock types comprise plagioclase, K-feldspar, quartz, white mica, biotite and garnet. The transition is characterized by increasing presence of interstitial phases along like-like grain boundaries and by progressive replacement of recrystallized K-feldspar grains by fine-grained myrmekite. These textural changes are characteristic for syn-deformational grain-scale melt

percolation, which is in line with the observed enrichment of the rocks in incompatible elements such as REEs, Ba, Sr, and K, suggesting open-system behaviour with melt passing through the rocks. The *P*–*T* path deduced from the thermodynamic modelling indicates decompression from ~15−16 kbar and ~650–740 ºC to ~6 kbar and ~640 ºC. Melt was already present at the *P*–*T* peak conditions as indicated by the albitic composition of plagioclase in films, interstitial grains and in myrmekite. The variably re-equilibrated garnet suggests that melt content may have varied along the decompression path, involving successively both melt gain and loss. The 6–8 km wide zone of vertical foliation and migmatite textural gradients is interpreted as vertical crustal-scale channel where the grain-scale melt percolation was associated with horizontal shortening and vertical flow of partially molten crustal wedge en masse.

Keywords: H*P*-granitic orthogneiss; petrological modelling; melt percolation; exhumation

INTRODUCTION

Recent petrological and microstructural studies show increased role of interplay between grain-scale melt transfer and deformation in various tectonic settings ranging from continental subduction (Závada et al. 2018), extrusion of subducted crust in continental wedges (Štípská et al. 2019) and in Cordilleran magmatic arc (Stuart et al. 2018). In all these settings melt passes through deforming crust exploiting grain boundaries mostly of felsic granitic protoliths without segregation into veins and dykes typical for metasedimentary migmatites (Collins and Sawyer, 1996; Brown and Solar, 1998; Weinberg, 1999; Vanderhaeghe, 2001). For all the above mentioned studies the typical feature is connection of this pervasive grain-scale flow with ductile shear zones, in particular with domains characterized by almost isotropic granite-like texture (Hasalová et al. 2008a). Závada et al. (2018) also suggested that prerequisite of pervasive porous flow of granitic melt is granular flow or cohesion-less grain boundary sliding of relictual parental grains enabling dynamic dilatancy of grain boundaries, typical for ultramylonitic cores of shear zones (Závada et al. 2007; Schulmann et al. 2008; Oliot et al. 2014).

Petrological studies characterizing role of melt for vertical transfer of rocks through crust are scarce, but the existing studies indicate that chemistry of crystallized interstitial melt and compositional variations of minerals can record exhumation of rocks of the order of ~10–20 kilometers (Hasalová et al. 2008b; Štípská et al. 2019; Nahodilová et al. 2020). In addition, these studies indicate that dynamically moving melt plays a major role in exhuming large portions of continental crust. Quantitative petrological studies of simultaneously migmatized and deformed granitoids thus provide unique insight into mechanisms of exhumation of deeply buried crust as proposed already by pioneering works of Hollister (1993) and Brown and Solar (1998).

The Orlica-Śnieżnik Dome (OSD) located at the NE extremity of the Bohemian Massif represents an ideal site where simultaneous deformation and melt transfer can be studied (see Fig. 1). The OSD is part of a continental wedge that shows structural and petrological records of continental subduction followed by vertical extrusion of partially molten deep crust in form of a giant gneiss dome (Chopin et al. 2012a). The extrusion process is characterized by elevation of two antiforms. The western Międzygórze antiform is marked by good preservation of shallow-dipping H*P* subduction fabrics reworked by vertical and localized melt-bearing zones, associated with heterogeneous exhumation of buried rocks (Štípská et al. 2019). In contrast, the eastern and larger Králíky-Śnieżnik antiform shows widespread melting of the crust associated with almost complete reworking of previously subducted rocks in up to 10-kilometer-wide zone of migmatites and granitoids (Chopin et al. 2012a; Lehmann et al. 2013).

Our goal is to examine a key outcrop section of the Králíky-Śnieżnik antiform where the continental subduction fabrics are almost completely transposed during simultaneous horizontal shortening and vertical melt transfer. On this key section, rock types range from augen to banded and migmatitic orthogneiss. These rocks are studied using microstructural qualitative analysis, whole-rock geochemistry and quantitative petrological modelling to illustrate open-system behaviour where melt percolates through granitic protolith during homogeneous crustal-scale deformation. By means of thermodynamic modelling, it will be shown that homogeneous pervasive melt percolation associated with grain-scale deformation contributes to the exhumation of partially molten crust en masse. We discuss homogeneous deformation of crust, mineral transformations of solid phases and the mutual interactions of melt with the rocks during syn-deformational melt percolation. Furthermore, we argue that pervasive melt percolation is the principal mechanism controlling exhumation of deep continental crust in hot collisional orogens of which the Bohemian Massif is a world example.

GEOLOGICAL SETTING

The OSD is considered to be a part of the high-grade Moldanubian–Lugian zone representing deeply eroded Variscan orogenic root (Fig. 1a; e.g. Franke and Zelaźniewicz, 2000; Aleksandrowski and Mazur, 2002). The structure of the OSD corresponds to a mantled gneiss dome (e.g. Eskola, 1948) with a core formed by medium- to high-grade metamorphic rocks surrounded by low-grade slates and metavolcanic rocks (Fig. 1b; e.g. Don et al. 1990; Mazur and Aleksandrowski, 2001). The central part of the dome is made up by antiforms cored by felsic orthogneiss with eclogite and H*P* granulite lenses (Kozłowski, 1961; Smulikowski, 1967; Pouba et al. 1985), alternating with N-S trending synforms formed mainly by metasedimentary sequences (Fig. 1b). The felsic orthogneiss types are traditionally divided into coarse-grained augen to banded Śnieżnik and fine-grained mylonitic to migmatitic Gierałtów orthogneiss (e.g. Fischer, 1936; Lange et al. 2002; Żelaźniewicz et al. 2002; Chopin et al. 2012b). The orthogneiss protolith corresponds to a granite emplaced at *c*. 510–490 Ma within a host metasedimentary sequence (e.g. Oliver et al. 1993; Borkowska and Dörr, 1998; Kröner et al. 2000; 2001; Lange et al. 2005; Bröcker et al. 2009; Mazur et al. 2010). The protolith age of the (U)H*P* granulite facies rocks remains still controversial but it is probably older than *c*. 470 Ma (Štípská et al. 2004; Lange et al. 2005; Anczkiewicz et al. 2007; Bröcker et al. 2009; Bröcker et al. 2010). The metasedimentary sequences (Młynowiec and Stronie formations) are dominated by Neoproterozoic to Cambro-Ordovician paragneiss and micaschist sporadically interbedded with layers of varied lithologies, such as metavolcanic rocks and few lenses of quartzite, graphite schist and marble (Don et al. 1990, 2003).

The metasedimentary rocks and orthogneiss together with (U)H*P* rocks were affected by an Early Carboniferous metamorphism (*c*. 340 Ma; e.g. Turniak et al. 2000; Štípská et al. 2004; Lange et al. 2005; Anczkiewicz et al. 2007; Bröcker et al. 2009; Jastrzębski, 2009; Bröcker et al. 2010; Walczak et al. 2017). The orthogneiss shows amphibolite- to eclogite-facies conditions with peak estimates ranging from ~20 kbar/700 ºC to ~27 kbar/700–800 ºC (e.g. Bakun-Czubarow, 1992; Bröcker and Klemd, 1996; Klemd and Bröcker, 1999; Štípská et al. 2004; Chopin et al. 2012b), whereas the metapelites preserve an amphibolite-facies metamorphism reaching *P*–*T* conditions of ~5–9 kbar and ~460–650 ºC (Murtezi, 2006; Jastrzębski, 2009; Jastrzębski et al. 2010, 2016; Skrzypek et al. 2011a,b, 2014; Štípská et al. 2012). UH*P* metamorphism (reaching ~30 kbar) was proposed for eclogite and H*P*-granulite (Bakun-Czubarow, 1991, 1992; Kryza et al. 1996; Klemd and Bröcker, 1999; Ferrero et al. 2015; Jedlička et al. 2017; Walczak et al. 2017; Majka et al. 2019), while other authors suggest lower *P*–*T* conditions of ~19–22 kbar at 700–800 ºC for the eclogite (Bröcker and

Klemd, 1996; Štípská et al. 2012) and ~18–22 kbar at 800–1000 ºC for the granulite (Pouba et al. 1985; Steltenpohl et al. 1993; Štípská et al. 2004; Bröcker et al. 2010; Budzyń et al. 2015). Retrograde metamorphism has been estimated at ~4–9 kbar and 550–700 ºC for the eclogite (Klemd et al. 1995; Bröcker and Klemd, 1996; Štípská et al. 2012), and at ~8–12 kbar and 560–800 ºC for the granulite (Steltenpohl et al. 1993; Štípská et al. 2004). Numerous cooling ages at *c*. 350–330 Ma were obtained using Rb–Sr and ^{40}Ar–^{39}Ar phengite and biotite dating (e.g. Steltenpohl et al. 1993; Maluski et al. 1995; Białek and Werner, 2004; Lange et al. 2005; Schneider et al. 2006; Anczkiewicz et al. 2007; Bröcker et al. 2009; Chopin et al. 2012a).

Available geological, geochronological and geophysical data allowed to divide the tectonic evolution of the OSD into three main stages: (1) H*P* metamorphism of Late Devonian – Early Carboniferous age associated with continental underthrusting of the Saxothuringian crust beneath the autochthonous Teplá-Barrandian type Neoproterozoic crust (Fig. 1a,b; Chopin et al. 2012b; Mazur et al. 2012; Majka et al. 2019). (2) Extrusion of a Saxothuringian-derived high-grade metamorphic core through the upper crust of the upper plate in front of the Brunia microcontinent. This event is responsible for exhumation of H*P* rocks and their re-equilibration at mid-crustal conditions (Štípská et al. 2004, 2012; Majka et al. 2019), while metasedimentary rocks were simultaneously buried in marginal synforms (Skrzypek et al. 2011a,b; Štípská et al. 2012; Skrzypek et al. 2014). (3) Ductile thinning event associated with formation of detachments and unroofing of the apical part of the dome (Pressler et al. 2007; Chopin et al. 2012a; Lehmann et al. 2013).

The south-eastern part of the OSD

The south-eastern part of the OSD represents a type section through the extruded Saxothuringian-derived crustal portion, where structural and petrological evolution of rocks affected by vertical ductile flow can be studied (see Fig. 1b). This section consists of the gneissic Międzygórze antiform in the west and the Králíky-Śnieżnik antiform in the east, separated from each other by a metasedimentary synform cropping out mainly in the Morava valley (Figs. 1c and 2a; Don, 1964, 1982; Chopin et al. 2012a; Štípská et al. 2012).

The previously well studied Międzygórze antiform is formed by an orthogneiss cored by a N-S trending belt of eclogite lenses (Figs. 1c and 2a; Chopin et al. 2012a,b; Štípská et al. 2012). Here, low-strain augen orthogneiss was progressively transformed into a banded to migmatitic orthogneiss (Chopin et al. 2012b). The banded and migmatitic orthogneiss are the most abundant in a ~1.5 km wide zone around the eclogite belt, whereas the augen orthogneiss is developed east of the high-grade orthogneiss-eclogite core (see Fig. 2a). The distinct textural varieties of the three orthogneiss types were interpreted as a result of deformation and metamorphism of the same Cambro-Ordovician granite protolith, reflecting mostly strain gradient and migmatization (Turniak et al. 2000; Lange et al. 2005; Chopin et al. 2012b), but also variable degree of melt-rock interaction (Štípská et al. 2019).

In the Międzygórze antiform, the augen orthogneiss displays a sub-horizontal S1 foliation (Fig. 2a), which is deformed by asymmetrical m- to km-scale open upright F2 folds associated with progressive transposition into a N-S striking subvertical S2 metamorphic foliation in the banded to migmatitic orthogneiss (Chopin et al. 2012a,b; Štípská et al. 2012; 2019). Finally, the S2 migmatitic foliation was heterogeneously reworked by a weakly developed F3 recumbent folds and rare sub-horizontal S3 axial plane cleavage (Fig. 2a). The banded to migmatitic orthogneiss adjacent to the eclogite experienced prograde evolution along a H*P* gradient reaching eclogite-facies conditions similar to that of the adjacent eclogite. This implies that both lithologies shared the same process of burial to ~19–20 kbar and >700 ºC (Chopin et al. 2012b; Štípská et al. 2012). This contrasts with the conditions from augen orthogneiss in direct continuity with paragneiss in the synforms, which show maximum burial conditions of ~7.5–8.0 kbar and ~600–620 ºC (Jastrzebski et al. 2017). Recent mineral equilibria modelling of Štípská et al. (2019) reveals that the migmatitic orthogneiss types were equilibrated at ~15–17 kbar and ~690–740 ºC during infiltration of a granitic melt. Retrograde equilibration down to ~7−10 kbar was largely restricted to retrograde zoning in phengite, garnet and plagioclase, and crystallization of biotite around phengite and garnet. Here, Early Carboniferous metamorphic ages of *c.* 340–360 Ma were determined on zircon rims in migmatitic orthogneiss (Turniak et al. 2000; Lange et al. 2005) and on zircon in leucosomes and leucocratic veins (Bröcker et al. 2009).

In the Morava valley metasedimentary synform, microstructural and petrological evidence suggests that the S1 fabric is associated with prograde metamorphism along a M*P*–M*T* gradient reaching garnet- to staurolite-grade conditions at ~6 kbar/580 ºC (Štípská et al. 2012). A continuous growth of both garnet and staurolite parallel to S2 indicates prograde metamorphism up to ~7.5 kbar/630 ºC in the sub-vertical fabric. However, chlorite parallel to the S2 foliation suggests that retrograde metamorphism and exhumation to ~5 kbar/500 ºC also occurred during this deformation event. The *P*–*T* path related to D3 was a prolonged retrograde evolution towards temperature lower than 550 ºC (Štípská et al. 2012).

In the eastern N-S trending Králíky-Śnieżnik antiform, the augen to banded mylonitic orthogneiss is rare and the bulk of the antiform is formed by a migmatitic orthogneiss. Eclogite lenses occur in places in the migmatitic orthogneiss and in the north the core of the antiform is formed by intermediate to felsic H*P* granulites (Fig. 1c; Pouba et al. 1985; Štípská et al. 2004). The whole Králíky-Śnieżnik antiform is characterized by significantly higher degree of migmatization compared to the westerly Międzygórze antiform (Figs. 1c and 2b). Recent petrological studies from the granulite belt indicate UH*P* conditions of ~20–25 kbar and ~550–950 ºC (Budzyń et al. 2015; Jedlička et al. 2017; Walczak et al. 2017) which is consistent with melt inclusion study from felsic granulites that show trapping conditions for melt at ~27 kbar and ~875 ºC, suggesting near UH*P* conditions of melting (Ferrero et al. 2015). The age of metamorphism is constrained by U–Pb dating of zircon to *c*. 340 Ma (Štípská et al. 2004; Walczak et al. 2017).

In the Králíky-Śnieżnik antiform, the rare relics of the earliest S1 fabric are systematically related to various types of migmatitic orthogneiss. These structures are mostly transposed by the dominant N-S sub-vertical S2 foliation (Fig. 2b). The whole domain is characterized by decimetre-scale transitions from remnants of partially molten porphyritic 'augen' to 'banded' orthogneiss and finally, to 'fine-grained' migmatitic orthogneiss (Figs. 1c and 2b; Don, 1982; Chopin et al. 2012a). This type of transition is classically referred to as the Śnieżnik augen gneiss – Gierałtów gneiss transition in the whole OSD (Lange et al. 2002 and references therein). However, this distinction is descriptive and only rarely took into account the processes leading to the gradual development of these rock types (Chopin et al. 2012a,b; Štípská et al. 2012).

FIELD RELATIONS AND STRUCTURAL EVOLUTION

Field relations and textural types of the orthogneiss were examined along a representative outcrop in the central part of the Králíky-Śnieżnik antiform (UTM WGS84 coordinates 50° 5' 37.54"N and 16° 51' 12.90"E; Figs. 1c, 2b and 3a). Here, the orthogneiss shows macroscopically a range of textures with continuous transitions among them (see Fig. 3b–f). In description of different rock types, we follow Štípská et al. (2019): we use the term augen or banded type I where the augen/bands of recrystallized quartz, plagioclase and/or K-feldspar have sharp boundaries and appear macroscopically monomineralic; the term augen or banded type II where the recrystallized K-feldspar augen/bands appear macroscopically monomineralic, but have diffuse boundaries, and bands of quartz and plagioclase are ill defined; the term schlieren for rock types with polymineralic and/or monomineralic, felsic and/or mafic bands elongated parallel to foliation within a nearly isotropic matrix; and nebulitic for rock types with nearly homogeneous distribution of phases. The studied outcrop is dominated by banded type II, schlieren and nebulitic orthogneiss (see Fig. 3b–f). Because of continuous transitions among the rock types, it is sometimes difficult to assign each sample unequivocally to a single rock type, and also, looking at rocks at different scale may lead to assigning them to different types (e.g. schlieren and nebulitic in Fig. 3e,f).

The S1 foliation is tightly folded by cm- to dm-scale upright close to isoclinal F2 folds with N-S trending subhorizontal axes and subvertical N-S trending axial planes (Fig. 3a,c,d). The intersection lineation L2 is parallel to the F2 fold hinges (Fig. 2b). This upright folding locally results in formation of N-S striking subvertical S2 axial planar cleavage (Fig. 3d). Here, the foliation in limbs of the isoclinal folds appear in subvertical position, subparallel to the S2 cleavage, and therefore the foliation in these limbs represents a composite S1-2 foliation (Fig. 3c,d). Rock types in the S1 foliation occurring mainly in the F2 limbs are represented by banded type II orthogneiss (e.g. Fig. 3b,c,d). Rock types in the cleavage are represented mainly by schlieren and nebulitic orthogneiss (e.g. Fig. 3e). In areas where only the subvertical fabric is present, the rock transitions from banded type II to schlieren and nebulitic orthogneiss are most typical (Fig. 3f).

In order to put valorisation to the field observations, we make here some basic interpretation of the observed textural transitions (for similar interpretation see Hasalová et al. 2008a,b,c; Schulmann et al. 2008; Chopin et al. 2012b; Štípská et al. 2019). From the gradual transitions of the rock types in the field, it is concluded that the banded type II, schlieren and nebulitic orthogneiss types resulted from progressive disappearance of the originally monomineral augen/banded texture typical of the augen/banded type I orthogneiss. From the observed textural gradients occurring parallel to the S2 foliation and ranging from the banded type II to schlieren and nebulitic orthogneiss types, it is concluded that these transformations occurred mainly during the D2 deformation. As the schlieren and nebulitic textures are commonly assigned to migmatites (Mehnert, 1971), it is supposed that the observed textural transitions are a result of transformation under the presence of melt (for similar interpretation see also Hasalová et al. 2008a,b,c; Schulmann et al. 2008; Chopin et al. 2012b; Závada et al. 2018; Štípská et al. 2019).

MICROSTRUCTURAL AND PETROGRAPHIC FEATURES

Microstructural and petrological studies were carried out on 10 samples collected along the outcrop section (see Fig. 3a for location of samples). Characterization of the different orthogneiss types was carried out by combining optical microscopy, back-scattered electron imaging (BSE) and optical cathodoluminescence (CL) on XZ oriented thin sections (Figs. 4 and 5). BSE images were acquired by a Tescan VEGA\\XMU electron microscope at the Charles University in Prague (Czech Republic). The CL mosaics were compiled from around 90–250 images that were captured using the CITL Mk5–2 CL microscope at 700 mA with a capture time of 1.5 s at the Czech Geological Survey in Prague.

In all the rock types, the mineral assemblage is plagioclase, K-feldspar, quartz, white mica, biotite and garnet. Apatite and zircon are present as accessory minerals in the matrix, ilmenite is only enclosed in biotite. Mineral abbreviations used here are: ksp, K-feldspar; pl, plagioclase; q, quartz; bi, biotite; g, garnet; ap, apatite; ilm, ilmenite; myr, myrmekite (mainly after Holland and Powell, 1998). White mica (wm) is used in diagrams because the composition ranges from phengite (ph) to muscovite (mu). Phengite is used where the observed white mica has phengitic composition.

In the following microstructural descriptions, we characterize the spatial distribution of K-feldspar, plagioclase, quartz and micas in individual rock types (Fig. 4a–d).

Banded type II orthogneiss

At microscopic scale, this rock type is characterized by alternation of K-feldspar, quartz and plagioclase-rich layers or domains. The boundaries between individual layers range from sharp (e.g. white arrows in Fig. 3b and Fig. 4a) to diffuse (e.g. Fig. 4b). The K-feldspar-rich layers have an interconnected network of K-feldspar grains separated by ~5% (e.g. Fig. 4a) up to ~20% (e.g. Fig. 4b) of interstitial quartz, plagioclase and myrmekite. The quartz-rich layers are composed of interconnected grains of quartz with ~5% up to 30% of interstitial feldspar, white mica, biotite, garnet and accessory minerals (e.g. Fig. 4a,b). The plagioclase-rich layers are thinner and more irregular compared to the K-feldspar and quartz-rich layers and have up to 30% of interstitial quartz, K-feldspar, white mica, biotite, garnet and accessory minerals in some samples (e.g. Fig. 4a), while in other samples only ill-defined plagioclase-rich domains with up to 50% of other phases are identified (Fig. 4b).

K-feldspar grains in K-feldspar-rich layers are slightly elongated, with irregular shape and 50–1000 µm in size (Fig. 5a). Large grains of K-feldspar show a well-developed shape preferred orientation parallel to the macroscopic foliation, whereas preferred orientation of smaller grains is less developed (Fig. 5a). The boundaries between elongated K-feldspar grains are serrated and lined by irregular <10–30 µm wide films of plagioclase, irregular plagioclase grains 10−300 µm in size and by rounded quartz grains 10−500 µm in size. K-feldspar grains show cuspate-lobate boundaries with respect to interstitial plagioclase and quartz (Fig. 5b). Small myrmekite-like aggregates 10−200 µm in size are also present along the K-feldspar grains within the K-feldspar-rich layers (Fig. 5b). Myrmekite is composed mostly of fine-grained plagioclase and quartz, with small and rare grains of K-feldspar and white mica.

Plagioclase aggregates have granoblastic microstructure and an average grain size of 400 µm. The plagioclase grains do not display a visible shape preferred orientation (Fig. 5c). The rounded and isolated quartz (10−100 µm) and cuspate small K-feldspar (10−50 µm in size) occur as interstitial grains at triple junctions of the plagioclase grains. Boundaries between plagioclase grains are mostly irregular. In addition, layers rich in polygonal plagioclase (1500−2500 μm) with antiperthite core are locally observed parallel to the S2 foliation (Fig. 5d).

Quartz forms recrystallized aggregates composed of large and inequigranular grains 30−2000 μm in size with amoeboid to highly lobate boundaries (Fig. 4a,b). They show a weak shape preferred orientation. Feldspar crystals penetrate heterogeneously into quartz aggregates along mutually lobate boundaries (Fig. 5c).

Schlieren orthogneiss

At microscopic scale, the schlieren are composed mostly by relict K-feldspar-rich domains with up to 40% of interstitial phases. The K-feldspar-rich domains have very irregular boundaries with the surrounding matrix (see white-dashed lines in Fig. 4c). The internal textures within these K-feldspar-rich domains are identical to the textures described for the banded type II orthogneiss and are mainly characterized by cuspate-lobate boundaries of K-feldspar grains with respect to interstitial plagioclase, quartz, and myrmekite (see above). The matrix has nearly homogeneous distribution of all the phases typical of nebulitic texture, and is described below.

Nebulitic orthogneiss

This rock type is characterized by nearly homogeneous distribution of all the phases. K-feldspar, plagioclase and quartz grains are highly irregular and show large variation in grain size from <1 μm up to 1.5 mm (Figs. 4c,d and 5e). The larger grains of K-feldspar are surrounded by abundant myrmekite. Locally, K-feldspar has at its rim quartz-white mica symplectite when it occurs at contact with white mica (Fig. 5f).

Textural relations of white mica, biotite and garnet

In the banded type II orthogneiss, both white mica and biotite tend to form aggregates or also occur as individual grains within the layers dominated by plagioclase and quartz, and are strongly to weakly oriented parallel to the S1-2 foliation (Fig. 5c). In the schlieren and nebulitic orthogneiss, micas tend to appear as individual grains that are strongly oriented parallel to the S2 foliation (Fig. 5e). The proportion of white mica is higher compared to biotite (around 70% and 30% of all micas respectively). However, the amount of biotite is higher in the banded type II, and lower in the schlieren and nebulitic orthogneiss (reaching ~40%, ~30% and 20% of the total amount of the micas, respectively). White mica forms commonly large laths (250–1000 μm in size) with small or large biotite at its margins, along its cleavage or in its crystallographic continuity (Fig. 5g). These features indicate that this biotite grew at the expense of white mica. Biotite is partially or completely chloritized. Garnet (<2 vol. %) occurs as small grains (around 50–200 μm) inside plagioclase or quartz, or as small (up to 500 μm) grains at contact with plagioclase and quartz, but in some places it is also in contact with white mica and biotite. Garnet has irregular shapes and is commonly fragmented (Fig. 5e).

From the observation at microscale, we conclude that the progressively higher amount of interstitial phases observed along like-like grain boundaries from the banded type II, to schlieren and to nebulitic orthogneiss is responsible for progressive disintegration of originally monomineral layers typical of weakly migmatized banded type II orthogneiss (Fig. 4a–d). The nucleation of interstitial phases is typically explained by crystallization from melt (for identical interpretation of similar rock sequence see also Hasalová et al. 2008a,b,c; Schulmann et al. 2008; Chopin et al. 2012b; Štípská et al. 2019).

MINERAL CHEMISTRY

Samples corresponding to different orthogneiss types were selected for mineral chemical analysis. Minerals were analysed using the *Electron Probe Micro-Analyser* (EPMA) JEOL

8200 at the University of Lausanne (Switzerland) and Jeol FEG-EPMA JXA-8530F at the Charles University in Prague (Czech Republic). The analyses were made in point beam mode at 15-kV acceleration voltage and 20-nA beam current, with a spot diameter of 5 μm and a counting time of 20–30 s. Representative analyses of feldspar, micas and garnet are presented in Tables 1, 2 and 3 and shown in Figures 6–9. Abbreviations used for mineral end-members in molar proportions are an = Ca/(Ca+Na+K); ab = (Na/(Ca+Na+K); or = K/(Ca+Na+K); X_{Fe} (ph, bi, g) = $Fe_{total}/(Fe_{total} + Mg)$; alm = $Fe^{+2}/(Fe^{+2} + Mg + Ca + Mn)$; sps = $Mn/(Fe^{+2} + Mg + Ca + Mn)$; grs = $Ca/(Fe^{+2} + Mg + Ca + Mn)$ and prp = $Mg/(Fe^{+2} + Mg + Ca + Mn)$. The sign "→" indicates a trend in mineral composition or zoning, the sign "–" depicts a range of mineral compositions and p.f.u. is per formula unit.

Feldspar

Recrystallized K-feldspar has subtle zoning within all the orthogneiss types (Fig. 6a–f, Table 1), with cores more albite-rich (or = 0.86–0.90; ab = 0.10–0.14) compared to rims (or = 0.90–0.94; ab = 0.06–0.10). K-feldspar in small grains in myrmekite (Fig. 6a–c), and K-feldspar forming thin films or small interstitial grains in the layers dominated by plagioclase and quartz has a lower content of albite (or = 0.92–0.96; ab = 0.04–0.08) (Fig. 6a, b).

Two distinct plagioclase populations have been measured according to their microstructural position (Fig. 6a–f, Table 1). Plagioclase aggregates exhibit homogeneous oligoclase composition (an = 0.12–0.17; ab = 0.83–0.88), whereas more sodic composition (an = 0.02–0.05; ab = 0.95–0.98) is found in small interstitial plagioclase grains and thin films coating the K-feldspar aggregates of the banded type II and schlieren orthogneiss (Fig. 6a). Oligoclase (an = 0.12–0.22; ab = 0.78–0.88) and albite (an = 0.02–0.09; ab = 0.91–0.98) compositions are also found in myrmekite-like aggregates within all the rock types (Fig. 6e, f).

Micas

White mica shows in all rock types zoning from core to rim, with decreasing Si from ~3.20−3.40 p.f.u. to ~3.10−3.20 p.f.u., Ti from ~0.03−0.08 p.f.u. to ~0.01−0.03 p.f.u. and with increasing $X_{Fe(tot)}$ from ~0.50−0.55 to > 0.55 (Fig. 7a–d, Table 2). F content varies from 0.04 to 0.14 p.f.u. with higher values in the core (Fig. 7c, Table 2).

Biotite has similar compositional ranges in all the rock types: in banded type II orthogneiss, Ti = 0.05–0.20 p.f.u., Al = 1.65–1.95 p.f.u., X_{Fe} = 0.74–0.78 and F = 0.10–0.25 p.f.u.; in schlieren orthogneiss, Ti = 0.05–0.12 p.f.u., Al = 1.75–1.90 p.f.u., X_{Fe} = 0.75–0.78 and F = 0.10–0.20 p.f.u.; and in nebulitic orthogneiss, Ti = 0.12–0.20 p.f.u., Al = 1.65–1.75 p.f.u., X_{Fe} = 0.76–0.78 and F = 0.10–0.20 p.f.u. (Fig. 8, Table 2).

Garnet

Garnet exhibits a flat or significant compositional zoning in samples of banded type II to schlieren orthogneiss (Fig. 9a–c, Table 3), whereas in samples of the nebulitic orthogneiss garnet displays a strong zoning (Fig. 9d, Table 3). In banded type II to schlieren orthogneiss, some garnet shows flat profiles with high X_{Fe} (0.95–0.98) and almandine ($alm_{0.65–0.72}$), low pyrope ($prp_{0.02–0.04}$), grossular ($grs_{0.07–0.09}$) and spessartine ($sps_{0.10–0.19}$) (Fig. 9a–c, Table 3). In some garnets of the banded type II to schlieren orthogneiss, grossular decreases ($grs_{0.18\rightarrow0.09}$) and spessartine increases ($sps_{0.09\rightarrow0.19}$) from core to rim (Fig. 9a–c, Table 3). In the nebulitic orthogneiss, garnet has higher grossular and lower spessartine in the core compared to banded type II and schlieren orthogneiss, and garnet zoning is from core to rim marked by decrease in grossular ($grs_{0.34\rightarrow0.09}$) and increase in almandine ($alm_{0.59\rightarrow0.68}$) and spessartine ($sps_{0.04\rightarrow0.21}$), and accompanied by a flat trend in high X_{Fe} (0.95–0.98) and low pyrope ($prp_{0.02–0.04}$) (Fig. 9d, Table 3).

WHOLE-ROCK CHEMISTRY

Whole-rock major- and trace-element analyses were performed by *Inductively Coupled Plasma-Atomic Emission Spectroscopy* (ICP-AES) and -*Mass Spectrometry* (ICP-MS) at the Acme laboratories of Canada for each rock type. Analyses are summarized in Table 4 and presented in a series of isocon diagrams (Fig. 10a,b; Grant, 1986) and spider plot normalized to Chondrite (Evensen et al. 1978) to show geochemical variations for different rock types (Fig. 10c).

The isocon diagrams point to weak variations of major elements for all the schlieren orthogneiss (Fig. 10a) and a nebulitic orthogneiss (Fig. 10b), as were typically reported from deformed and migmatized granitic rocks (Chopin et al. 2012b; Závada et al. 2018). The trace-element concentrations of individual rock types are compared to different rock groups described by Chopin et al. (2012b) in Figure 10c. According to Chopin et al. (2012b), the LREE (La, Ce, Pr and Nd) and MREE (Sm, Eu and Gd) contents drop slightly in the sequence from augen to banded and mylonitic/migmatitic orthogneiss, whereas the HREE distribution is homogeneous (see shaded fields in Fig. 10c). For studied rocks, the character of distribution patterns for all studied samples is fairly homogeneous, showing similar REE contents (~ΣREE = 55–89 ppm., Table 4) and subparallel distribution patterns except in the nebulitic orthogneiss (sample FC076C, Fig. 10c). The nebulitic orthogneiss is slightly depleted in LREE and MREE and enriched in HREE compared to the banded type II and schlieren orthogneiss (see red diamonds in Fig. 10c). In general, the REE compositional ranges overlap the mylonitic/migmatitic group described by Chopin et al. (2012b), characterized by pronounced negative Eu anomalies (Fig. 10c). This fact indicates that the studied samples can be explained by increasing degree of percolating melt fractionation as was described by Hasalová et al. (2008b) and Chopin et al. (2012b).

In Figure 10d, a compositional variation between the different rock types is presented in the spider plot normalized by a weakly migmatized banded type II orthogneiss (sample

FC076A). Here, the character of distribution patterns for all rock types shows only slight differences in chemical composition. According to Lange et al. (2002) and Chopin et al. (2012b), slight differences in major and trace elements between individual rock types may be likely due to original heterogeneity of the protolith. On the other hand, Marquer and Burkhard (1992) and Hasalová et al. (2008b) attributed these variations to the presence of external fluids or melt in similar felsic rocks.

To identify the mass transfers related to melt and/or fluid flux through the studied rock types, mass balance calculations of incompatible elements have been performed using the normalized Potdevin diagram for the comparison of weakly and strongly migmatized orthogneiss (see Fig. 11a–c and Table 5). In this diagram, relative element transfers between the selected rock types – typically a protolith and the altered rock – is conveniently expressed using the relative abundance of specific element (i) and F_v volume factor (Potdevin and Marquer, 1987; López-Moro, 2012). While the F_v is given by the volume ratio between the transformed rock and the initial one, the difference in relative abundance of specific element (i) is expressed by:

$$\Delta m_i = F_v \cdot (\rho_a/\rho_0)\, C_a^{\,i} - C_0^{\,i},$$

where Δm_i is the relative gain or loss of mass, and $C_0^{\,i}$ and $C_a^{\,i}$ are the initial and final concentrations and ρ_0 and ρ_a the densities of these rocks, respectively.

Resulting calculations show a slight gain in a range of incompatible elements, such as REEs, Ba, Sr, and K in banded type II (sample FC076E) and schlieren orthogneiss (sample FC076J) (Fig. 11a–b) with respect to weakly migmatized banded type II orthogneiss (sample FC076A). This trend is coupled with a slight loss in HFSE's, such as U, Zr, Nb, Hf and Ta. Enrichment in REEs, Ba, Sr, and loss of HFSE's is less significant in the nebulitic orthogneiss, which is marked mainly by depletion in Th, Cs, Pr, La, U and Ta (Fig. 11c).

FORWARD MODELLING OF MIGMATITIC ORTHOGNEISS

In the modelling of anatexis of granitic rocks at eclogite-facies conditions we follow the approach discussed in Štípská et al. (2019). The assemblages are modelled metastable with respect to the stability of clinopyroxene, as clinopyroxene is commonly absent in quartzo-feldspathic rocks at (U)H*P* conditions (e.g. Young and Kylander-Clark, 2015). The haplogranitic melt model is used as it can explain well mineral equilibria in quartzo-feldspathic rocks at H*P*−H*T* conditions, even if it was not calibrated explicitly for these conditions (Štípská et al. 2008; Hopkins et al. 2010; Lexa et al. 2011; Nahodilová et al. 2014).

Pseudosections were calculated using THERMOCALC 3.33 (Powell et al. 1998; version 2009) and dataset 5.5 (Holland and Powell, 1998; January 2006 upgrade), in the system MnO-Na_2O-CaO-K_2O-FeO-MgO-Al_2O_3-SiO_2-H_2O-TiO_2-O (MnNCKFMASHTO). The activity-composition relationships used are as follows: for silicate melt (liq), from White et al. (2007); for garnet (g), biotite (bi) and ilmenite (ilm), from White et al. (2005); for feldspar (pl, ksp), from Holland and Powell (2003); for white mica (wm), from Coggon and Holland (2002); and, for cordierite (cd), from Holland and Powell (1998). The calculations are done for the whole-rock composition of a nebulitic orthogneiss (sample FC076C; Table 4). Because of closely similar whole-rock compositions of the other samples (Fig. 10b) the calculated diagrams are also used for interpretation of their metamorphic evolution. The amount of H_2O in the whole-rock composition was deduced from $T−M(H_2O)$ pseudosection (see description below, Fig. 12). Mineral composition isopleths of garnet, white mica, plagioclase and molar proportion of liquid were plotted to discuss P–T conditions of mineral equilibration. The isopleth notation used is: m(sps) = Mn/(Ca+Mg+Fe+Mn)*100; x(alm) = Fe/(Ca+Mg+Fe+Mn)*100; z(grs) = Ca/(Ca+Mg+Fe+Mn)*100; Si(wm) p.f.u.; ca(pl) = Ca/(Ca+Na+K) and liq (mol. %).

$T−M(H_2O)$ pseudosection at 7 kbar

In order to estimate the conditions of last equilibration in migmatites, it is assumed that the assemblage tends to equilibrate until it becomes melt poor or melt absent near or at the solidus (e.g. Štípská and Powell, 2005; Hasalová et al. 2008c). Therefore, a $T-M(H_2O)$ pseudosection at a pressure of 7 kbar was calculated first, simulating conditions of last equilibration with melt near the solidus (Fig. 12). The observed assemblage of q-pl-ksp-g-bi-wm-ilm-liq is stable at $T-M(H_2O)$ = ~0.30−0.53 and at ~630−710 ºC and the observed garnet rim composition (~alm_{68}; grs_{10}; sps_{20}) fits the calculated isopleths at conditions of $T-M(H_2O)$ = ~0.52 and ~630 ºC (Fig. 12). Therefore, it is assumed that the rocks crossed the solidus with minimum H_2O content corresponding to 2.07 mol. % of H_2O in the whole-rock composition.

Closed-system *P–T* pseudosection

The *P–T* diagram (Fig. 13) was calculated for an amount of H_2O = 2.07 mol. % deduced from the $T-M(H_2O)$ diagram (see Fig. 12 and related discussion). This H_2O amount allows the stability of a typical mineral assemblage of a granite at M*P*–M*T* conditions, which is H_2O-free and garnet-free (~5.2−6.7 kbar and <550−620 ºC), and therefore may illustrate rock evolution in a closed system, even for H_2O. The major features and topology of the diagram involve a H_2O-saturated solidus up to ~7.8 kbar followed by a steeply inclined H_2O-undersaturated solidus from ~7.8 kbar to higher pressure at progressively higher temperature. The biotite-out and garnet-out lines are temperature sensitive at suprasolidus and subsolidus conditions, respectively and pressure sensitive at subsolidus conditions (Fig. 13).

The resulting *P–T* phase diagram shows a stability field of q-pl-ksp-g-bi-wm-ilm-liq between ~4–12 kbar and ~630–740 ºC, corresponding to the observed assemblage in all the rock types. The nebulitic orthogneiss preserves high grossular content of garnet in the core ($alm_{60\text{-}65}$; $grs_{30\text{–}34}$; sps_4; Fig. 9d), pointing to a *P–T* peak in the melt-free stability field of q-pl-ksp-g-bi-wm-ru at ~14–16 kbar and ~600–740 ºC (ellipse 1 in Fig. 13a−b). The high Si content in white mica in all the rock types (Si = 3.35–3.40 p.f.u.; Fig. 7a) supports its crystallization at high pressure conditions at ~16 kbar (Fig. 13c). The plagioclase composition measured in films and interstitial grains is albite ($an_{0.05–0.02}$; $ab_{0.95–0.98}$), compatible with plagioclase crystallization at H*P* conditions at ~16 kbar (Fig. 13d). The presence of interstitial plagioclase, quartz and K-feldspar is interpreted as crystallized from melt at grain boundaries (e.g. Hasalová et al. 2008a). We documented also albite and oligoclase composition in myrmekite-like aggregates. According to Barker (1970), the myrmekite is typically associated with higher anorthite content in plagioclase then observed in this study or other similar studies (Štípská et al. 2019). The albite compositions measured in myrmekite may be result of the presence of melt at H*P* conditions, and partial reequilibration of plagioclase to oligoclase on decompression. The oligoclase compositions measured in cores of matrix plagioclase we interpret as composition attained during recrystallization of the original aggregates during burial (see also Chopin et al. 2012), part of these compositions, mainly at rims may reflect also the last reequilibration of rocks on decompression. Combination of the textural argument for melt presence with the very low anorthite content of plagioclase suggests crossing of the H_2O-undersaturated solidus and beginning of partial melting at ~16 kbar. The calculated isopleths of melt mode suggest a very small melt production around 1 mol. % (Fig. 13e). This calculated volume of melt suggests that the melt was isolated in melt films, pools or pockets, compatible with the observed microstructure (see Fig. 5b). Even if along this $P-T$ path the calculated molar proportion of melt increases for the whole rock, in individual textural positions, as are the like-like grain boundaries, the phases may crystallize from melt in order to lower the surface energy of the monomineral aggregates (e.g. Lexa et al. 2006; Hasalová et al. 2008a). Likely mechanism for crystallization of albite and other phases from melt, while the melt content of the rock increases, is also a process of dissolution-precipitation of these phases promoted by the presence of melt and possibly also by crystallization from melt (Sawyer, 2001; Hasalová et al. 2008b; Holness and Sawyer, 2008; Závada et al. 2018, Štípská et al. 2019). The observed mineral assemblage q-pl-ksp-g-bi-wm-ilm with supposed former melt in all rock types together with rim compositions of garnet ($alm_{65–70}$; $grs_{5–10}$; $sps_{20–25}$; Fig. 9) and white mica (Si = 3.10 p.f.u.; Fig. 7a) point to last partial equilibration in the middle-*P* part of the q-pl-ksp-g-bi-wm-ilm-liq stability field, at ~6 kbar and ~640 ºC (ellipse 2 in Fig. 13a). Therefore, the absence of rutile and the core-to-rim zoning trends of garnet and white mica in the nebulitic orthogneiss suggest a *P*−*T* path from ~14−16 kbar and ~600–740 ºC to ~6 kbar and ~640 ºC, with local equilibration down to the ilmenite stability field and close to the solidus.

Additionally, the core composition of garnet in samples of banded type II and schlieren orthogneiss is also partially to completely re-equilibrated at ~6 kbar and ~640 ºC (see Fig. 9a–c), suggesting that the mineral assemblage of these rocks were melt-bearing close to the solidus.

P–$X_{added\ melt}$ pseudosection for discussion of open-system melt infiltration

Previously calculated diagrams and experiments for granitic rocks in the OSD and other parts of the Bohemian Massif show that depending on the exact *P*−*T* path, the rocks may produce up to 7−20% of melt on decompression from peak *P*−*T* conditions of ~25 kbar (for OSD: Walczak et al. 2017, for other occurrences in the Bohemian Massif: Hasalová et al. 2008c; Lexa et al. 2011; Nahodilová et al. 2014). At such melt proportions, melt can be lost from the rocks undergoing melting deeper in the crust, and can be added to the rocks above (Štípská et al. 2019). Therefore, the effect of a possible melt gain in the studied rocks was explored in P–$X_{added\ melt}$ diagram (Fig. 14). In the calculations, the composition of the melt is taken at 16 kbar and 730 ºC (see black star in Fig. 13a). The P–$X_{added\ melt}$ diagram was calculated for the whole-rock composition of the nebulitic orthogneiss at conditions of the estimated peak metamorphism (T = 700 ºC), and for a range of compositions between the H_2O-undersaturated whole-rock composition ($X_{added\ melt}$ = 0) and the composition with 20 mol. % of melt added

($X_{\mathrm{added\ melt}}$ = 1; Fig. 14). The main characteristics of the diagram show a liquid-in line heading from ~12 kbar for $X_{\mathrm{added\ melt}}$ = 0 to ~20 kbar for $X_{\mathrm{added\ melt}}$ = 0.25. The stability of biotite, rutile and ilmenite depends on pressure in the melt-present fields and on both pressure and $X_{\mathrm{added\ melt}}$ in the melt-absent fields. The compositional isopleths of garnet and Si-in-phengite are pressure sensitive in the melt-present fields, and do not depend on the amount of melt added (Fig. 14). The horizontal arrow illustrates the effect of addition of melt to a rock with original H_2O content inferred for the protolith (Fig. 13a). The starting position at $X_{\mathrm{added\ melt}}$ = 0.05, allows the stability of the observed mineral assemblage q-pl-ksp-g-bi-wm, calculated is also stability of infinitesimal amount (around 0.13–0.15 mol. %) of rutile which is not observed in the rocks. The reason for this difference in the calculated and observed assemblage with respect to rutile may be reequilibration on decompression to the ilmenite stability or also the fact that Ti is not included in the model of white mica.

At $X_{\mathrm{added\ melt}}$ = 0.09 where the arrow crosses the solidus, the first melt appears and from this point along the horizontal arrow melt proportion increases while the high grossular content in garnet and high Si content in micas do not change. The position of any rock along the horizontal arrow depends on the amount of melt added. The vertical part of the arrows labelled A, B and C represents examples of what happens during decompression. Along the arrows A, B and C the modelled mineral assemblage evolves from q-pl-ksp-g-bi-wm-ru-liq, through q-pl-ksp-g-bi-wm-ru-ilm-liq to q-pl-ksp-g-bi-wm-ilm-liq until ~6.5 kbar where sillimanite starts to be stable. Along the paths A, B and C the major difference is the amount of melt present on decompression being at 7 kbar for A ~2 mol. % melt, B ~8 mol. % melt and C ~ 22 mol. % melt. As the compositional isopleths are not dependent on the amount of melt added, all the rocks evolve with identical mineral composition of garnet, white mica and biotite. As the melt migration occurs, the amount of melt may increase or decrease depending whether the melt is added or lost at any particular $P-T$ conditions on decompression. The rock plus melt on decompression may therefore lie anywhere from the solidus to the higher values of $X_{\text{added melt}}$ of the diagram.

For the studied rock types, the observed assemblage q-pl-ksp-g-bi-wm-ilm, absence of rutile and the core-to-rim zoning trends of garnet ($alm_{60\rightarrow70}$; $grs_{34\rightarrow5}$, $sps_{4\rightarrow22}$; Fig. 9d) and white mica (Si = 3.40–3.10 p.f.u.; Fig. 7a) are compatible with a decompression path from ~16 kbar to ilmenite-bearing stability field at ~8−13 kbar. The garnet and white mica compositional isopleths are not sensitive to the amount of melt added; therefore, the melt amount that percolated in individual rock types is unknown. However, consequence of different melt amount in different samples may be the difference in re-equilibration along the decompression path. We suggest that for the nebulitic orthogneiss that preserves H*P* garnet core the amount of melt was low at lower pressure, thus precluding complete garnet reequilibration on decompression. Re-equilibration of garnet core close to the solidus observed in the banded type II and schlieren orthogneiss may be caused by higher proportion of melt at lower pressure compared to the nebulitic orthogneiss. Therefore, a new *P–T* diagram with ~3.40 mol. % of re-integrated melt in the whole-composition was calculated for explaining better the observed mineral assemblage and mineral compositions of the banded type II and schlieren orthogneiss (Fig. 15).

P–T pseudosection with added melt

The resulting whole-rock composition after the re-integration of 3.40 mol. % melt is presented in mole percent normalized to 100% (Fig. 15a). The major features and topology of the diagram involve a displacement of the H_2O-undersaturated solidus to higher pressure, whereas the other features and topology are similar to the H_2O-undersaturated diagram (see Fig. 13).

The resulting *P–T* diagram shows a stability field of q-pl-ksp-g-bi-wm-ilm-liq between ~4–13 kbar and ~610–740 ºC, corresponding to the observed assemblage. The *P–T* peak conditions are defined by the high grossular content measured in garnet cores ($alm_{60–65}$; $grs_{30–34}$; sps_4; Figs. 9d and 15b) preserved in nebulitic orthogneiss and the high Si content in phengite cores measured in all the rock types (Si = 3.35–3.40 p.f.u.; Figs. 7a and 15c). The measured albite content of plagioclase films and interstitial grains is consistent with the calculated isopleth of albite at these *P–T* peak conditions ($an_{0.05-0.02}$; Fig. 15d). The last equilibration at ~630–640 ºC and ~6 kbar is constrained by the re-equilibrated compositions of garnet ($alm_{65–70}$; $grs_{5–10}$; $sps_{20–25}$; Fig. 15b) and white mica (Si = 3.10 p.f.u.; Fig. 15c). The calculated isopleths of melt mode suggest a melt production during retrograde path of the order of 1 mol. % resulting in up to 4 mol. % for 3.40 mol. % melt added (Fig. 15e). Therefore, the mineral chemistry together with the absence of rutile record a decompression path from ~15−16 kbar and ~650–740 ºC to ~6 kbar and ~640 ºC, with local equilibration down to the ilmenite stability field and close to the solidus with the presence of melt. This decompression path is compatible with the one previously described in the H_2O-undersaturated diagram (Fig. 13), and differs only in the melt amount present during the peak and retrograde evolution, suggesting that higher melt amount may result in more profound re-equilibration of the assemblage close to the solidus. The melt might have been added to rocks, explaining the textures, e.g. disintegration of the monomineral banding or more profound re-equilibration of garnet composition, but the rocks may also undergo variable degree of melt loss on decompression before last cooling through the solidus.

DISCUSSION AND CONCLUSIONS

The core of the Orlica-Śnieżnik Dome (Fig. 1b) consists of two antiforms affected by various degrees of migmatization related to their distance from the site of presumed continental subduction further west (see Fig. 13 in Chopin et al. 2012a). The proximal and small Międzygórze antiform (~1 km across and ~4 km long; Fig. 1c) is characterized by heterogeneously developed zones of partial melting surrounding blocks of well-preserved H*P* rocks. The distal large-scale Králíky-Śnieżnik antiform (~6–8 km across and ~20 km long; Fig. 1c) was affected by widespread melting and almost complete re-equilibration of H*P* mineral assemblages. In this work the orthogneiss of the Králíky-Śnieżnik antiform are compared with those of the Międzygórze antiform in order to understand the role of melt-deformation interplays during the exhumation of large portions of continental crust in Variscan continental collision zone.

Microstructural and geochemical arguments for grain-scale melt percolation during D2 deformation

In the study area, relics of shallow-dipping S1 foliation are folded by upright F2 folds and transposed by almost pervasive N-S trending subvertical S2 foliation (Fig. 3). This structural succession is the same as in the Międzygórze antiform but the degree of fabric transposition to subvertical fabric is significantly broader. It reaches the width of ~6–8 km and length of ~20 km, thereby attesting to a large-scale orogenic process.

A well-preserved and continuous transition from banded type II to schlieren and nebulitic orthogneiss (Fig. 4) is documented on a representative outcrop section in the central part of the Králíky-Śnieżnik antiform (Fig. 1c). The continuous transition from banded II to schlieren and nebulitic orthogneiss is commonly gradational and perpendicular to subvertical S2 transposition (Fig. 3). Therefore, by analogy to the Międzygórze antiform the studied sequence of rocks can be interpreted to reflect the higher intensity of D2 deformation.

Furthermore, the q-pl-ksp-g-bi-wm-ilm assemblage observed in all orthogneiss types is the same despite variations in meso- and micro-scale structural and textural features (see Figs. 3 and 4). The presence of cuspate K-feldspar in plagioclase layers, quartz and albite-rich plagioclase intergrowths in K-feldspar aggregates, amoeboid grains of K-feldspar in quartz layers and diffuse boundaries between different felsic layers are interpreted as a result of grain-scale melt percolation through the solid felsic rock (Figs. 4 and 5). This interpretation is in agreement with previously reported examples of Hasalová et al. (2008b), Závada et al. (2007, 2018) and Štípská et al. (2019) where grain boundaries were open at the micron scale to fluid/melt circulation (e.g. Oliot et al. 2014). Such interpretation is also supported by slight differences in chemical composition. For instance, mass balance calculations of incompatible elements show the depletion of U, Zr and Hf compatible with partial dissolution of zircon in the melt, implying that some melt must have been lost or must have percolated through the banded type II and schlieren orthogneiss, and the gain in Ba, Sr, Eu, K and Rb corresponding to a heterogeneous nucleation of interstitial feldspar from percolated melt (Fig. 11a–b). On the other hand, in the nebulitic orthogneiss, mass balance calculations show the depletion of Th, Cs, Pr, La, U and Ta compatible with partial dissolution of monazite in the melt, implying that some melt must have been lost (Fig. 11c). Therefore, the observed textural trend from banded type II to nebulitic orthogneiss can be considered not only as a result of a deformation gradient but also as a result of different degree of melt infiltration/percolation of granitic sources under open-system conditions (e.g. Hasalová et al. 2008; Goncalves et al. 2012; Závada et al. 2018).

***P*–*T* evolution and melt transfer**

The *P*–*T* path deduced from the forward-modelling indicates that the orthogneiss underwent a decompression from *P*–*T* peak conditions of ~15–16 kbar and ~650–740 ºC to ~6 kbar and ~640 ºC under the presence of melt (Figs. 13 and 15). The estimated *P*–*T* peak conditions are recorded by the presence of Ca-rich garnet cores in the nebulitic orthogneiss and composition of phengite in all the rock types. Subsequent decompression to ~6 kbar is recorded by the core-to-rim zoning trends of garnet and white mica and absence of rutile. Similar *P*–*T* evolutions have been repeatedly reported from other rock types such as eclogite or H*P* granulite occurring in the orthogneiss (~18–30 kbar and ~700–1000 ºC; e.g. Pouba et al. 1985; Bakun-Czubarow, 1991, 1992; Steltenpohl et al. 1993; Bröcker and Klemd, 1996;

Kryza et al. 1996; Klemd and Bröcker, 1999; Bröcker et al. 2010; Štípská et al. 2012; Ferrero et al. 2015; Budzyń et al. 2015; Jedlička et al. 2017; Walczak et al. 2017; Majka et al. 2019) with retrograde paths ranging from ~9 kbar/~730 ºC to ~6 kbar/~560 ºC (Steltenpohl et al. 1993; Bröcker and Klemd, 1996; Štípská et al. 2004, 2012). Recently, prograde H*P* metamorphism was attributed to the S1 fabric in the granitic orthogneiss close to the eclogite in the Międzygórze antiform (~13−18 kbar/~700–800 ºC; e.g. Chopin et al. 2012b). It was also shown that migmatization of orthogneiss surrounding the Międzygórze eclogite started in eclogite-facies conditions at ~15−17 kbar in the S1 fabric (Štípská et al. 2019). These authors also suggested that decompression along the S2 fabric from ~15−17 kbar to ~7−10 kbar/~690–740 ºC was locally associated with anatexis. In analogy, we argue that the $P-T-d$ path for the S2 vertical fabric is comparable in both antiforms (Fig. 16).

The modelling showed that the orthogneiss protolith from the Králíky-Śnieżnik antiform is able to produce only ~1 mol. % of melt along the $P-T$ path. However, the macro- and microstructural features are typical for advanced migmatization and attest to melt presence along grain boundaries (Figs. 4 and 5), suggesting that a higher melt proportion was present. This is possible to achieve only if H_2O is added to the rocks, but being above the conditions of the wet solidus, the hydrating fluid must have been external melt (Štípská et al. 2019). Such melt is supposed to be released by similar rocks buried deeper and this is simulated in the modelling by adding granitic melt to the whole-rock composition. The modelling did not allow estimation of the melt proportion in the rocks based on the mineral chemical composition, as the mineral chemical composition for the observed assemblage is independent on the amount of melt added (Fig. 14). However, it is supposed, that rocks that contain garnet with high grossular content characteristic for H*P* conditions contained on decompression less melt compared with rocks that show garnet with re-equilibrated grossular content (see Figs. 13−15).

The melt proportion remains unknown, and may have varied along the $P-T$ path. Melt percolation started already at ~15–16 kbar and ~650–740 ºC as indicated by albitic composition of plagioclase in myrmekite and in interstitial films, then melt equilibrated with the minerals (at least with their rim compositions) during the retrograde history to ~6 kbar and ~640 ºC. However, the amount of melt percolated is not likely to be sufficient to produce a melt-supported structure required for diatexite formation (Brown, 2007; Hasalová et al. 2008a,b,c), but may be sufficient to allow melt-assisted granular flow (Rosenberg and Handy, 2005; Závada et al. 2007; Schulmann et al. 2008). This is supported by the results of Štípská et al. (2019) from the Międzygórze antiform, where it was shown that different migmatite textures originated from variable degree of melt-rock interaction starting at ~17 kbar and ~730 ºC and ending at ~7–10 kbar.

Back-stop extrusion of partially molten crust

Grain-scale melt percolation started locally in the S1 structure as demonstrated in the Międzygórze antiform (Chopin et al. 2012b), and continued in the subvertical S2 foliation where the grain-scale percolation of melt occurred along heterogeneously developed subvertical narrow zones of intense D2 deformation (Fig. 16a; Štípská et al. 2019). Within these subvertical zones occur low-strain domains, where the rocks preserve the shallow-dipping S1 fabric and high pressure conditions of 17–20 kbar. These zones of S1 foliation preserve a strain gradient from banded to narrow zones of fine-grained mylonitic/migmatitic orthogneiss. Across this deformation gradient the rocks show slight depletion of REE and HFSE (Chopin et al. 2012b), that was explained by presence of melt in the mylonitic/migmatitic orthogneiss. Because the proportion of fine-grained S1 mylonites is subordinate, the proportion of melt-bearing rocks during the D1 is also subordinate. There is not a geochemical study of REE and HFSE across a strain gradient related to the S2 fabric in the Międzygórze antiform. However, as the proportion of schlieren and nebulitic orthogneiss types is higher in the S2 compared to S1 we extrapolate the depletion of REE and HFSE also to the S2 migmatitic fabric (Fig. 16a).

In the Králíky-Śnieżnik antiform the D2 deformation affects almost homogenously the whole volume of the felsic orthogneiss with only rare relics of low-strain S1 domains (Fig. 16b). The degree of transposition of S1 by S2, together with high proportion of schlieren and nebulitic orthogneiss types in S2 fabric implies also that melt percolation affected significantly larger proportion of crust during D2 compared to the Międzygórze antiform. The orthogneiss samples from this study, from a large zone of almost homogeneous D2 deformation show similar degree of REE and HFSE depletion (Fig. 10c) as the mylonitic type of Chopin et al. (2012b) from the Międzygórze antiform, which was interpreted as a result of presence of melt during deformation. Because the zone of D2 deformation and proportion of the migmatitic types is significantly higher in the Králíky-Śnieżnik antiform, the total depletion REE and HFSE in this portion of crust is much more important compared to the Międzygórze antiform (Fig. 16b). Another important feature is the last reequilibration of rocks marked by mineral rim chemistry, under the presence of melt, along the decompression $P-T$ path. The rocks from the Międzygórze antiform show last reequilibration under the presence of melt at ~10 kbar while the rocks from the Králíky-Śnieżnik antiform show reequilibration under the presence of melt at ~ 5 kbar. This implies that the difference in total exhumation between two antiforms under the presence of melt within the D2 zone is significantly higher in the case of the Králíky-Śnieżnik antiform compared to the Międzygórze antiform.

Based on our petrological data it may be concluded that melt percolation along vertical D2 deformation zone facilitated exhumation of H*P* rocks in core of the two antiforms from ~60 km to 35 km (Międzygórze antiform, Fig, 16a) and ~50 km to >20 km (Králíky-Śnieżnik antiform, Fig. 16b), allowing the juxtaposition of originally H*P* orthogneiss with M*P* rocks in crustal-scale synforms (see also Štípská et al. 2004; Chopin et al. 2012a; Štípská et al. 2012, 2019). Another point that reinforces the role of the presence of melt for facilitating extrusion of these deep rocks is a fact that the presence even of small melt volumes plays a major role in decreasing the strength of the rocks (Rosenberg and Handy, 2005).

The above described deformation was related to horizontal shortening of collisional wedge that contributed to the vertical extrusion of partially molten crust en masse in the eastern part of the OSD. Such an extrusion of weak material is pronounced in particular close to the Brunia back-stop where massive portions of rheologically weak and partially molten rocks flowed upwards under horizontal stress from the root area of the orogenic wedge (Fig. 16). The extrusion model proposed by Thompson et al. (1997a,b) is suitable to well explain the exceptionally high rate of exhumation suggested already by Steltenpohl et al. (1993) and the shape of *P*–*T* path depicted by this study. Majka et al. (2019) suggested that first part of exhumation occurred in channel flow region of the subducted continental margin along the rigid buttress of the Brunovistulian microcontinent and that folding of the entire crustal sequence occurred in the shallower part of the accretionary wedge. The minimum depth for the F2 folding from this study and from Štípská et al. (2019) is ~16−17 kbar. Recently, the numerical modelling of Maierová et al. (2014) tested various parameters controlling the exhumation rates of hot gneiss domes and corresponding *P*–*T*–*t* paths such as rate of convergence, heat production and erosion. These authors concluded that in the case of the Orlica-Śnieżnik Dome the gravitational instability contribution was minor compared to laterally forced folding leading to gneiss dome formation and exhumation of hot felsic lower crust. All the above mentioned models tacitly suppose extreme weakness of thermally softened hot felsic lower crust allowing homogeneous vertical flow. However, in detail, both meso- and micro-scale mechanism allowing the extreme drop of strength of the felsic crust remain enigmatic. Only recently, natural observations from hot gneiss domes and their analogue modelling suggest that partial melting can trigger detachment folding and vertical flow of migmatites and granitoids (Lehmann et al. 2017). Based on our study we argue that the grain-scale melt percolation (called also reactive porous flow, melt infiltration) represents such a principal weakening mechanism allowing homogeneous flow of crust typical for extrusion. It is probably also a principal mechanism controlling exhumation of deep partially molten crust in hot collisional orogens such as the European Variscan belt.

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FIGURE AND TABLE CAPTIONS

Fig. 01. (a) Lithotectonic map of North European Variscan belt (modified after Aleksandrowski et al. 1997; Don et al. 2003, Żelaźniewicz et al. 2006). The position of Orlica-Śnieżnik Dome is outlined. (b) Geological sketch map of the Orlica-Śnieżnik Dome

(modified after Aleksandrowski et al. 1997; Don et al. 2003; Żelaźniewicz et al. 2006). (c) Geological and structural map of Międzygórze and Králíky-Śnieżnik antiforms (lithologies after Don et al. 2003; Chopin et al. 2012a,b; Štípská et al. 2012). The location of the orthogneiss used in this study and the position of the structural profiles are indicated.

Fig. 02. Interpretative cross-sections showing the main structural relationships between both antiforms (a–b) and the different orthogneiss types. See Figure 1c for locations of profiles.

Fig. 03. (a) Outcrop photograph illustrating a lithological sketch (see legend in the Figures 1 and 2) and the locations of studied samples with white stars. (b–f) Field photographs showing the different textural rocks displayed, ranging from banded type II to schlieren and nebulitic orthogneiss.

Fig. 04. Representative CL image mosaics showing textural features of the different rock types: (a) banded type II orthogneiss (sample FC076B): alternation of individual layers of recrystallized K-feldspar, quartz and plagioclase distinctly defined, with less diffuse boundaries between individual layers. (b) Banded type II orthogneiss (samples FC076E): alternation of almost individual layers of recrystallized K-feldspar, quartz and plagioclase with highly diffuse boundary between feldspar layers. (c) Schlieren orthogneiss (sample FC076J) characterised by relics of K-felsdpar-rich layers in otherwise isotropic matrix. (d) Nebulitic orthogneiss (samples FC076C) with no relics of original banding.

Fig. 05. Detailed back scattered electron (BSE) and cathodoluminescence (CL) images of different rock types. (a) Diffuse boundary between a layers of mixed aggregates and a K-feldspar-rich layer with numerous interstitial plagioclase, quartz and myrmekite. (b) Detail of cuspate plagioclase, rounded quartz and myrmekite as interstitial phases in K-feldspar layer. (c) Quartz- and plagioclase-rich layers with interstitial grains of feldspar and quartz. Phengite is surrounded by biotite and biotite occurs also along cleavage of phengite. (d) Large crystal of plagioclase with and antiperthitic core rich in exsolutions of K-feldspar. (e) Garnet in a

matrix of quartz, plagioclase, K-feldspar, phengite, biotite and myrmekite. (f) Symplectite of quartz and white mica at the boundary of K-feldspar and phengite. (g) Large phengite laths with biotite at margins and along the cleavage.

Fig. 06. Composition of feldspar (an = (Ca/(Ca+Na+K))*100 in blue color and or = (K/(Ca+Na+K))*100 in yellow color) for the different rock types localized in different microstructural positions (a−f). Representative analyses are listed in Table 1.

Fig. 07. Composition of white mica: (a) Si (p.f.u.) content of white micas *v.* samples. Note the decrease in Si (p.f.u.) content from cores to rim. (b−c) Compositional ranges from cores to rims of white micas. Grey and yellow arrows indicate zoning from core to rim. Representative analyses are listed in Table 2.

Fig. 08. Composition of biotite for the different rock types (a, b). Representative analyses are listed in Table 2.

Fig. 09. Garnet compositional trends of the different rock types: (a) weakly migmatized banded type II orthogneiss (sample FC076B); (b) banded type II orthogneiss (sample FC076E); (c) schlieren orthogneiss (sample FC076J): and (d) nebulitic orthogneiss (FC076C). Representative analyses are listed in Table 3.

Fig. 10. Isocon diagrams after Grant (1986) comparing: (a) the average whole-rock composition of schlieren orthogneiss and (b) the whole-rock composition of the nebulitic orthogneiss (vertical axis) with respect to the composition of the weakly migmatized banded type II orthogneiss (sample FC076A). The diagrams (a) and (b) show loss and/or gain of major elements with respect to composition of reference rock plotted as reference line. (c) Spider plot normalized to chondrite (Evensen et al. 1978). Shaded field corresponds to the orthogneiss types described by Chopin et al. (2012b) in the Międzygórze antiform. (d) Spider plot normalized to weakly migmatized banded type II orthogneiss (sample FC076A).

Fig. 11. Potdevin diagrams (Potdevin and Marquer, 1987; Lopez-Moro, 2012) illustrating the loss-gain relationships for a range of incompatible elements for banded type II, schlieren and nebulitic orthogneiss compared to weakly migmatized banded type II orthogneiss (sample FC076A). Relative gain or loss of mass (Δm_i) is normalized by the initial mass (m_i^0) in the volume-composition diagrams. The volume factor (F_v) is given by the volume ratio between the transformed rocks and initial one. Shaded area refers to the magmatic fluctuation ranges (immobile elements) and solid lines to mobile elements. LILE = Large Ion Lithophile Elements, HFSE = High Field Strength Elements, REE = Rare Earth Elements. (see Table 5).

Fig. 12. P-M(H_2O) pseudosection calculated at 7 kbar for a nebulitic orthogneiss (sample FC076C) and contoured for the calculated spessartine (m(sps)), almandine (x(alm)) and grossular (z(grs)) contents of garnet and for the molar proportion of liquid (liq (mol. %)). The solidus is emphasized by a dark-dashed line. Quartz, plagioclase and K-feldspar are present in all fields.

Fig. 13. (a) P–T pseudosection calculated for the analysed whole-rock composition of a nebulitic orthogneiss (sample FC076C). (b−e) Simplified pseudosections with compositional isopleths of spessartine (m(sps)), almandine (x(alm)) and grossular (z(grs)) in garnet; Si content of white mica (Si(wm) p.f.u.); anorthite content of plagioclase (ca(pl)); and molar proportion of melt (liq (mol. %)). The ellipses indicate the P–T ranges compatible with the observed assemblage and core and rim compositions of garnet and white mica. The star indicates P–T conditions from which the melt composition was taken to be reintegrated into whole-rock composition shown in Figures 14 and 15. The solidus is underlined by a thick black dashed line. See text for discussion of the P–T path.

Fig. 14. T–$X_{added\ melt}$ pseudosection calculated at 700 ºC for a range of compositions representing mixtures of H_2O-undersaturated whole-rock composition of the nebulitic orthogneiss (x = 0) and the composition of melt calculated at 680 ºC and 16 kbar (x = 1, black

star in Figure 13). $X_{added\ melt}$ is the proportion of melt added in the migmatitic orthogneiss at 16 kbar. x = 1 corresponds to 20 mol. % of melt added. The solidus is underlined by a thick black dashed line. The calculated isopleths show molar proportion of garnet (g mol. %), almandine (x(alm)) and grossular (z(grs)) content of garnet, Si content of white micas (Si(wm) p.f.u.) and molar proportion of melt (liq mol. %). The ellipses indicate the *P–T* ranges compatible with the observed assemblage and core and rim compositions of garnet and white mica. Evolution along three decompression paths at different $X_{added\ melt}$ is discussed in the text.

Fig. 15. (a) *P–T* pseudosection calculated with 3.40 mol. % melt added to the whole-rock composition of the nebulitic orthogneiss. (b−e) Simplified pseudosections with compositional isopleths of spessartine (m(sps)), almandine (x(alm)) and grossular (z(grs)) in garnet; Si content of muscovite (Si(wm) p.f.u.); anorthite content of plagioclase (ca(pl)); and molar proportion of melt (liq (mol. %)). The ellipses indicate the *P–T* ranges compatible with the observed assemblage and core and rim compositions of garnet and white mica. The solidus is underlined by a thick black dashed line. See text for discussion of the *P–T* path.

Fig. 16. Sketches summarizing the variations of different textural parameters and melt loss-gain relationship as indicated by geochemical signatures (REE – Rare Earth Elements; HFSE – High Field Strength Element) across both antiforms. (a) Międzygórze antiform: melt-absent orthogneiss with local migmatite formation. Shallow-dipping S1 fabric preserved in the low-strain D2 domains. Grain-scale melt percolation is mainly localized in narrow zones of vertical S2 fabric (modified after Štípská et al. 2019). Compilation of *P–T* paths for orthogneiss reported by Chopin et al. (2012b) (1), and Štípská et al. (2019) (2). (b) Králíky-Śnieżnik antiform: melt-present migmatitic orthogneiss. High-strain D2 domains with subvertical S2 foliation are connected with crustal-scale shear zones. The wide zone of subvertical S2 fabric is to a large extent percolated by melt. Simplified *P–T* diagram with the *P–T* path obtained in this study. *P–T* paths are mostly within the melt-present assemblages.

(c) Tectonic sketch of the Orlica-Śnieżnik Dome as a part of the Sudetes showing shallow-dipping S1 fabrics related to subduction of the continental crust up to eclogite- and (U)H*P* granulite-facies conditions, and subvertical S2 fabrics related to its vertical exhumation to the middle and upper crust (modified after Chopin et al. 2012a). The positions in subduction wedge of two antiforms are indicated as a proximal part (a) and a more distal part (b) with respect to the subduction.

Table 01. Representative analyses of feldspar.

Table 02. Representative analyses of micas.

Table 03. Representative analyses of garnet.

Table 04. Major and trace-element compositions (ICP-AES and -MS). LOI: Loss On Ignition.

Table 05. Gain/loss of incompatible elements of the studied rocks relative to weakly migmatized banded type II orthogneiss (sample FC076A).