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

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16
17 **ABSTRACT**

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

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

37

38 **Keywords:** *HP*-granitic orthogneiss; petrological modelling; melt percolation; exhumation

39

40 INTRODUCTION

41 Recent petrological and microstructural studies show increased role of interplay between
42 grain-scale melt transfer and deformation in various tectonic settings ranging from continental
43 subduction (Závada et al. 2018), extrusion of subducted crust in continental wedges (Štípská
44 et al. 2019) and in Cordilleran magmatic arc (Stuart et al. 2018). In all these settings melt
45 passes through deforming crust exploiting grain boundaries mostly of felsic granitic protoliths
46 without segregation into veins and dykes typical for metasedimentary migmatites (Collins and
47 Sawyer, 1996; Brown and Solar, 1998; Weinberg, 1999; Vanderhaeghe, 2001). For all the
48 above mentioned studies the typical feature is connection of this pervasive grain-scale flow
49 with ductile shear zones, in particular with domains characterized by almost isotropic granite-

50 like texture (Hasalová et al. 2008a). Závada et al. (2018) also suggested that prerequisite of
51 pervasive porous flow of granitic melt is granular flow or cohesion-less grain boundary
52 sliding of relictual parental grains enabling dynamic dilatancy of grain boundaries, typical for
53 ultramylonitic cores of shear zones (Závada et al. 2007; Schulmann et al. 2008; Oliot et al.
54 2014).

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

64 The Orlica-Śnieżnik Dome (OSD) located at the NE extremity of the Bohemian Massif
65 represents an ideal site where simultaneous deformation and melt transfer can be studied (see
66 Fig. 1). The OSD is part of a continental wedge that shows structural and petrological records
67 of continental subduction followed by vertical extrusion of partially molten deep crust in form
68 of a giant gneiss dome (Chopin et al. 2012a). The extrusion process is characterized by
69 elevation of two antiforms. The western Międzygórze antiform is marked by good
70 preservation of shallow-dipping HP subduction fabrics reworked by vertical and localized
71 melt-bearing zones, associated with heterogeneous exhumation of buried rocks (Štípská et al.
72 2019). In contrast, the eastern and larger Králíky-Śnieżnik antiform shows widespread
73 melting of the crust associated with almost complete reworking of previously subducted rocks

74 in up to 10-kilometer-wide zone of migmatites and granitoids ([Chopin et al. 2012a](#); [Lehmann](#)
75 [et al. 2013](#)).

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

89 **GEOLOGICAL SETTING**

90 The OSD is considered to be a part of the high-grade Moldanubian–Lugian zone representing
91 deeply eroded Variscan orogenic root ([Fig. 1a](#); e.g. [Franke and Zelaźniewicz, 2000](#);
92 [Aleksandrowski and Mazur, 2002](#)). The structure of the OSD corresponds to a mantled gneiss
93 dome (e.g. [Eskola, 1948](#)) with a core formed by medium- to high-grade metamorphic rocks
94 surrounded by low-grade slates and metavolcanic rocks ([Fig. 1b](#); e.g. [Don et al. 1990](#); [Mazur](#)
95 [and Aleksandrowski, 2001](#)). The central part of the dome is made up by antiforms cored by
96 felsic orthogneiss with eclogite and *HP* granulite lenses ([Kozłowski, 1961](#); [Smulikowski,](#)
97 [1967](#); [Pouba et al. 1985](#)), alternating with N-S trending synforms formed mainly by
98 metasedimentary sequences ([Fig. 1b](#)). The felsic orthogneiss types are traditionally divided

99 into coarse-grained augen to banded Śnieżnik and fine-grained mylonitic to migmatitic
100 Gierałtów orthogneiss (e.g. Fischer, 1936; Lange et al. 2002; Żelaźniewicz et al. 2002;
101 Chopin et al. 2012b). The orthogneiss protolith corresponds to a granite emplaced at *c.* 510–
102 490 Ma within a host metasedimentary sequence (e.g. Oliver et al. 1993; Borkowska and
103 Dörr, 1998; Kröner et al. 2000; 2001; Lange et al. 2005; Bröcker et al. 2009; Mazur et al.
104 2010). The protolith age of the (U)HP granulite facies rocks remains still controversial but it
105 is probably older than *c.* 470 Ma (Štípská et al. 2004; Lange et al. 2005; Anczkiewicz et al.
106 2007; Bröcker et al. 2009; Bröcker et al. 2010). The metasedimentary sequences (Młynowiec
107 and Stronie formations) are dominated by Neoproterozoic to Cambro-Ordovician paragneiss
108 and micaschist sporadically interbedded with layers of varied lithologies, such as
109 metavolcanic rocks and few lenses of quartzite, graphite schist and marble (Don et al. 1990,
110 2003).

111 The metasedimentary rocks and orthogneiss together with (U)HP rocks were affected by
112 an Early Carboniferous metamorphism (*c.* 340 Ma; e.g. Turniak et al. 2000; Štípská et al.
113 2004; Lange et al. 2005; Anczkiewicz et al. 2007; Bröcker et al. 2009; Jastrzębski, 2009;
114 Bröcker et al. 2010; Walczak et al. 2017). The orthogneiss shows amphibolite- to eclogite-
115 facies conditions with peak estimates ranging from ~20 kbar/700 °C to ~27 kbar/700–800 °C
116 (e.g. Bakun-Czubarow, 1992; Bröcker and Klemd, 1996; Klemd and Bröcker, 1999; Štípská
117 et al. 2004; Chopin et al. 2012b), whereas the metapelites preserve an amphibolite-facies
118 metamorphism reaching *P–T* conditions of ~5–9 kbar and ~460–650 °C (Murtezi, 2006;
119 Jastrzębski, 2009; Jastrzębski et al. 2010, 2016; Skrzypek et al. 2011a,b, 2014; Štípská et al.
120 2012). UHP metamorphism (reaching ~30 kbar) was proposed for eclogite and HP-granulite
121 (Bakun-Czubarow, 1991, 1992; Kryza et al. 1996; Klemd and Bröcker, 1999; Ferrero et al.
122 2015; Jedlička et al. 2017; Walczak et al. 2017; Majka et al. 2019), while other authors
123 suggest lower *P–T* conditions of ~19–22 kbar at 700–800 °C for the eclogite (Bröcker and

124 [Klemd, 1996; Štípská et al. 2012](#)) and ~18–22 kbar at 800–1000 °C for the granulite ([Pouba et](#)
125 [al. 1985; Steltenpohl et al. 1993; Štípská et al. 2004; Bröcker et al. 2010; Budzyń et al. 2015](#)).
126 Retrograde metamorphism has been estimated at ~4–9 kbar and 550–700 °C for the eclogite
127 ([Klemd et al. 1995; Bröcker and Klemd, 1996; Štípská et al. 2012](#)), and at ~8–12 kbar and
128 560–800 °C for the granulite ([Steltenpohl et al. 1993; Štípská et al. 2004](#)). Numerous cooling
129 ages at *c.* 350–330 Ma were obtained using Rb–Sr and ⁴⁰Ar–³⁹Ar phengite and biotite dating
130 (e.g. [Steltenpohl et al. 1993; Maluski et al. 1995; Białek and Werner, 2004; Lange et al. 2005;](#)
131 [Schneider et al. 2006; Anczkiewicz et al. 2007; Bröcker et al. 2009; Chopin et al. 2012a](#)).

132 Available geological, geochronological and geophysical data allowed to divide the
133 tectonic evolution of the OSD into three main stages: (1) *HP* metamorphism of Late Devonian
134 – Early Carboniferous age associated with continental underthrusting of the Saxothuringian
135 crust beneath the autochthonous Teplá-Barrandian type Neoproterozoic crust ([Fig. 1a,b;](#)
136 [Chopin et al. 2012b; Mazur et al. 2012; Majka et al. 2019](#)). (2) Extrusion of a Saxothuringian-
137 derived high-grade metamorphic core through the upper crust of the upper plate in front of the
138 Brunia microcontinent. This event is responsible for exhumation of *HP* rocks and their re-
139 equilibration at mid-crustal conditions ([Štípská et al. 2004, 2012; Majka et al. 2019](#)), while
140 metasedimentary rocks were simultaneously buried in marginal synforms ([Skrzypek et al.](#)
141 [2011a,b; Štípská et al. 2012; Skrzypek et al. 2014](#)). (3) Ductile thinning event associated with
142 formation of detachments and unroofing of the apical part of the dome ([Pressler et al. 2007;](#)
143 [Chopin et al. 2012a; Lehmann et al. 2013](#)).

144 **The south-eastern part of the OSD**

145 The south-eastern part of the OSD represents a type section through the extruded
146 Saxothuringian-derived crustal portion, where structural and petrological evolution of rocks
147 affected by vertical ductile flow can be studied (see [Fig. 1b](#)). This section consists of the

148 gneissic Międzygórze antiform in the west and the Králíky-Śnieżnik antiform in the east,
149 separated from each other by a metasedimentary synform cropping out mainly in the Morava
150 valley (Figs. 1c and 2a; Don, 1964, 1982; Chopin et al. 2012a; Štípská et al. 2012).

151 The previously well studied Międzygórze antiform is formed by an orthogneiss cored by a
152 N-S trending belt of eclogite lenses (Figs. 1c and 2a; Chopin et al. 2012a,b; Štípská et al.
153 2012). Here, low-strain augen orthogneiss was progressively transformed into a banded to
154 migmatitic orthogneiss (Chopin et al. 2012b). The banded and migmatitic orthogneiss are the
155 most abundant in a ~1.5 km wide zone around the eclogite belt, whereas the augen
156 orthogneiss is developed east of the high-grade orthogneiss-eclogite core (see Fig. 2a). The
157 distinct textural varieties of the three orthogneiss types were interpreted as a result of
158 deformation and metamorphism of the same Cambro-Ordovician granite protolith, reflecting
159 mostly strain gradient and migmatization (Turniak et al. 2000; Lange et al. 2005; Chopin et al.
160 2012b), but also variable degree of melt-rock interaction (Štípská et al. 2019).

161 In the Międzygórze antiform, the augen orthogneiss displays a sub-horizontal S1 foliation
162 (Fig. 2a), which is deformed by asymmetrical m- to km-scale open upright F2 folds associated
163 with progressive transposition into a N-S striking subvertical S2 metamorphic foliation in the
164 banded to migmatitic orthogneiss (Chopin et al. 2012a,b; Štípská et al. 2012; 2019). Finally,
165 the S2 migmatitic foliation was heterogeneously reworked by a weakly developed F3
166 recumbent folds and rare sub-horizontal S3 axial plane cleavage (Fig. 2a). The banded to
167 migmatitic orthogneiss adjacent to the eclogite experienced prograde evolution along a *HP*
168 gradient reaching eclogite-facies conditions similar to that of the adjacent eclogite. This
169 implies that both lithologies shared the same process of burial to ~19–20 kbar and >700 °C
170 (Chopin et al. 2012b; Štípská et al. 2012). This contrasts with the conditions from augen
171 orthogneiss in direct continuity with paragneiss in the synforms, which show maximum burial

172 conditions of ~7.5–8.0 kbar and ~600–620 °C (Jastrzebski et al. 2017). Recent mineral
173 equilibria modelling of Štípská et al. (2019) reveals that the migmatitic orthogneiss types
174 were equilibrated at ~15–17 kbar and ~690–740 °C during infiltration of a granitic melt.
175 Retrograde equilibration down to ~7–10 kbar was largely restricted to retrograde zoning in
176 phengite, garnet and plagioclase, and crystallization of biotite around phengite and garnet.
177 Here, Early Carboniferous metamorphic ages of *c.* 340–360 Ma were determined on zircon
178 rims in migmatitic orthogneiss (Turniak et al. 2000; Lange et al. 2005) and on zircon in
179 leucosomes and leucocratic veins (Bröcker et al. 2009).

180 In the Morava valley metasedimentary synform, microstructural and petrological evidence
181 suggests that the S1 fabric is associated with prograde metamorphism along a *MP–MT*
182 gradient reaching garnet- to staurolite-grade conditions at ~6 kbar/580 °C (Štípská et al.
183 2012). A continuous growth of both garnet and staurolite parallel to S2 indicates prograde
184 metamorphism up to ~7.5 kbar/630 °C in the sub-vertical fabric. However, chlorite parallel to
185 the S2 foliation suggests that retrograde metamorphism and exhumation to ~5 kbar/500 °C
186 also occurred during this deformation event. The *P–T* path related to D3 was a prolonged
187 retrograde evolution towards temperature lower than 550 °C (Štípská et al. 2012).

188 In the eastern N-S trending Králíky-Śnieżnik antiform, the augen to banded mylonitic
189 orthogneiss is rare and the bulk of the antiform is formed by a migmatitic orthogneiss.
190 Eclogite lenses occur in places in the migmatitic orthogneiss and in the north the core of the
191 antiform is formed by intermediate to felsic *HP* granulites (Fig. 1c; Pouba et al. 1985; Štípská
192 et al. 2004). The whole Králíky-Śnieżnik antiform is characterized by significantly higher
193 degree of migmatization compared to the westerly Międzygórze antiform (Figs. 1c and 2b).
194 Recent petrological studies from the granulite belt indicate UHP conditions of ~20–25 kbar
195 and ~550–950 °C (Budzyń et al. 2015; Jedlička et al. 2017; Walczak et al. 2017) which is
196 consistent with melt inclusion study from felsic granulites that show trapping conditions for

197 melt at ~27 kbar and ~875 °C, suggesting near UHP conditions of melting (Ferrero et al.
198 2015). The age of metamorphism is constrained by U–Pb dating of zircon to *c.* 340 Ma
199 (Štípská et al. 2004; Walczak et al. 2017).

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

210 **FIELD RELATIONS AND STRUCTURAL EVOLUTION**

211 Field relations and textural types of the orthogneiss were examined along a representative
212 outcrop in the central part of the Králíky-Šniežnik antiform (UTM WGS84 coordinates 50° 5'
213 37.54"N and 16° 51' 12.90"E; Figs. 1c, 2b and 3a). Here, the orthogneiss shows
214 macroscopically a range of textures with continuous transitions among them (see Fig. 3b–f).
215 In description of different rock types, we follow Štípská et al. (2019): we use the term augen
216 or banded type I where the augen/bands of recrystallized quartz, plagioclase and/or K-feldspar
217 have sharp boundaries and appear macroscopically monomineralic; the term augen or banded
218 type II where the recrystallized K-feldspar augen/bands appear macroscopically
219 monomineralic, but have diffuse boundaries, and bands of quartz and plagioclase are ill
220 defined; the term schlieren for rock types with polymineralic and/or monomineralic, felsic

221 and/or mafic bands elongated parallel to foliation within a nearly isotropic matrix; and
222 nebulitic for rock types with nearly homogeneous distribution of phases. The studied outcrop
223 is dominated by banded type II, schlieren and nebulitic orthogneiss (see [Fig. 3b–f](#)). Because
224 of continuous transitions among the rock types, it is sometimes difficult to assign each sample
225 unequivocally to a single rock type, and also, looking at rocks at different scale may lead to
226 assigning them to different types (e.g. schlieren and nebulitic in [Fig. 3e,f](#)).

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

238 In order to put valorisation to the field observations, we make here some basic
239 interpretation of the observed textural transitions (for similar interpretation see [Hasalová et al.](#)
240 [2008a,b,c](#); [Schulmann et al. 2008](#); [Chopin et al. 2012b](#); [Štípská et al. 2019](#)). From the gradual
241 transitions of the rock types in the field, it is concluded that the banded type II, schlieren and
242 nebulitic orthogneiss types resulted from progressive disappearance of the originally
243 monomineral augen/banded texture typical of the augen/banded type I orthogneiss. From the
244 observed textural gradients occurring parallel to the S2 foliation and ranging from the banded
245 type II to schlieren and nebulitic orthogneiss types, it is concluded that these transformations

246 occurred mainly during the D2 deformation. As the schlieren and nebulitic textures are
247 commonly assigned to migmatites (Mehnert, 1971), it is supposed that the observed textural
248 transitions are a result of transformation under the presence of melt (for similar interpretation
249 see also Hasalová et al. 2008a,b,c; Schulmann et al. 2008; Chopin et al. 2012b; Závada et al.
250 2018; Štípská et al. 2019).

251 **MICROSTRUCTURAL AND PETROGRAPHIC FEATURES**

252 Microstructural and petrological studies were carried out on 10 samples collected along the
253 outcrop section (see Fig. 3a for location of samples). Characterization of the different
254 orthogneiss types was carried out by combining optical microscopy, back-scattered electron
255 imaging (BSE) and optical cathodoluminescence (CL) on XZ oriented thin sections (Figs. 4
256 and 5). BSE images were acquired by a Tescan VEGA\\XMU electron microscope at the
257 Charles University in Prague (Czech Republic). The CL mosaics were compiled from around
258 90–250 images that were captured using the CITL Mk5–2 CL microscope at 700 mA with a
259 capture time of 1.5 s at the Czech Geological Survey in Prague.

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

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

269 **Banded type II orthogneiss**

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

281 K-feldspar grains in K-feldspar-rich layers are slightly elongated, with irregular shape and
282 50–1000 μm in size (Fig. 5a). Large grains of K-feldspar show a well-developed shape
283 preferred orientation parallel to the macroscopic foliation, whereas preferred orientation of
284 smaller grains is less developed (Fig. 5a). The boundaries between elongated K-feldspar
285 grains are serrated and lined by irregular <10–30 μm wide films of plagioclase, irregular
286 plagioclase grains 10–300 μm in size and by rounded quartz grains 10–500 μm in size. K-
287 feldspar grains show cusped-lobate boundaries with respect to interstitial plagioclase and
288 quartz (Fig. 5b). Small myrmekite-like aggregates 10–200 μm in size are also present along
289 the K-feldspar grains within the K-feldspar-rich layers (Fig. 5b). Myrmekite is composed
290 mostly of fine-grained plagioclase and quartz, with small and rare grains of K-feldspar and
291 white mica.

292 Plagioclase aggregates have granoblastic microstructure and an average grain size of 400
293 μm . The plagioclase grains do not display a visible shape preferred orientation (Fig. 5c). The
294 rounded and isolated quartz (10–100 μm) and cusped small K-feldspar (10–50 μm in size)

295 occur as interstitial grains at triple junctions of the plagioclase grains. Boundaries between
296 plagioclase grains are mostly irregular. In addition, layers rich in polygonal plagioclase
297 (1500–2500 μm) with antiperthite core are locally observed parallel to the S2 foliation (Fig.
298 5d).

299 Quartz forms recrystallized aggregates composed of large and inequigranular grains
300 30–2000 μm in size with amoeboid to highly lobate boundaries (Fig. 4a,b). They show a
301 weak shape preferred orientation. Feldspar crystals penetrate heterogeneously into quartz
302 aggregates along mutually lobate boundaries (Fig. 5c).

303 **Schlieren orthogneiss**

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

312 **Nebulitic orthogneiss**

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

318 **Textural relations of white mica, biotite and garnet**

319 In the banded type II orthogneiss, both white mica and biotite tend to form aggregates or also
320 occur as individual grains within the layers dominated by plagioclase and quartz, and are
321 strongly to weakly oriented parallel to the S1-2 foliation (Fig. 5c). In the schlieren and
322 nebulitic orthogneiss, micas tend to appear as individual grains that are strongly oriented
323 parallel to the S2 foliation (Fig. 5e). The proportion of white mica is higher compared to
324 biotite (around 70% and 30% of all micas respectively). However, the amount of biotite is
325 higher in the banded type II, and lower in the schlieren and nebulitic orthogneiss (reaching
326 ~40%, ~30% and 20% of the total amount of the micas, respectively). White mica forms
327 commonly large laths (250–1000 μm in size) with small or large biotite at its margins, along
328 its cleavage or in its crystallographic continuity (Fig. 5g). These features indicate that this
329 biotite grew at the expense of white mica. Biotite is partially or completely chloritized. Garnet
330 (<2 vol. %) occurs as small grains (around 50–200 μm) inside plagioclase or quartz, or as
331 small (up to 500 μm) grains at contact with plagioclase and quartz, but in some places it is
332 also in contact with white mica and biotite. Garnet has irregular shapes and is commonly
333 fragmented (Fig. 5e).

334 From the observation at microscale, we conclude that the progressively higher amount of
335 interstitial phases observed along like-like grain boundaries from the banded type II, to
336 schlieren and to nebulitic orthogneiss is responsible for progressive disintegration of
337 originally monomineral layers typical of weakly migmatized banded type II orthogneiss (Fig.
338 4a–d). The nucleation of interstitial phases is typically explained by crystallization from melt
339 (for identical interpretation of similar rock sequence see also Hasalová et al. 2008a,b,c;
340 Schulmann et al. 2008; Chopin et al. 2012b; Štípská et al. 2019).

341 **MINERAL CHEMISTRY**

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

344 8200 at the University of Lausanne (Switzerland) and Jeol FEG-EPMA JXA-8530F at the
345 Charles University in Prague (Czech Republic). The analyses were made in point beam mode
346 at 15-kV acceleration voltage and 20-nA beam current, with a spot diameter of 5 μm and a
347 counting time of 20–30 s. Representative analyses of feldspar, micas and garnet are presented
348 in [Tables 1, 2 and 3](#) and shown in [Figures 6–9](#). Abbreviations used for mineral end-members
349 in molar proportions are an = $\text{Ca}/(\text{Ca}+\text{Na}+\text{K})$; ab = $(\text{Na}/(\text{Ca}+\text{Na}+\text{K}))$; or = $\text{K}/(\text{Ca}+\text{Na}+\text{K})$; X_{Fe}
350 (ph, bi, g) = $\text{Fe}_{\text{total}}/(\text{Fe}_{\text{total}} + \text{Mg})$; alm = $\text{Fe}^{+2}/(\text{Fe}^{+2} + \text{Mg} + \text{Ca} + \text{Mn})$; sps = $\text{Mn}/(\text{Fe}^{+2} + \text{Mg}$
351 $+ \text{Ca} + \text{Mn})$; grs = $\text{Ca}/(\text{Fe}^{+2} + \text{Mg} + \text{Ca} + \text{Mn})$ and prp = $\text{Mg}/(\text{Fe}^{+2} + \text{Mg} + \text{Ca} + \text{Mn})$. The sign
352 “ \rightarrow ” indicates a trend in mineral composition or zoning, the sign “–” depicts a range of
353 mineral compositions and p.f.u. is per formula unit.

354 **Feldspar**

355 Recrystallized K-feldspar has subtle zoning within all the orthogneiss types ([Fig. 6a–f](#), [Table](#)
356 [1](#)), with cores more albite-rich (or = 0.86–0.90; ab = 0.10–0.14) compared to rims (or = 0.90–
357 0.94; ab = 0.06–0.10). K-feldspar in small grains in myrmekite ([Fig. 6a–c](#)), and K-feldspar
358 forming thin films or small interstitial grains in the layers dominated by plagioclase and
359 quartz has a lower content of albite (or = 0.92–0.96; ab = 0.04–0.08) ([Fig. 6a, b](#)).

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

368 **Micas**

369 White mica shows in all rock types zoning from core to rim, with decreasing Si from
370 ~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
371 with increasing $X_{\text{Fe}(\text{tot})}$ from ~0.50–0.55 to > 0.55 (Fig. 7a–d, Table 2). F content varies from
372 0.04 to 0.14 p.f.u. with higher values in the core (Fig. 7c, Table 2).

373 Biotite has similar compositional ranges in all the rock types: in banded type II
374 orthogneiss, Ti = 0.05–0.20 p.f.u., Al = 1.65–1.95 p.f.u., $X_{\text{Fe}} = 0.74–0.78$ and F = 0.10–0.25
375 p.f.u.; in schlieren orthogneiss, Ti = 0.05–0.12 p.f.u., Al = 1.75–1.90 p.f.u., $X_{\text{Fe}} = 0.75–0.78$
376 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
377 p.f.u., $X_{\text{Fe}} = 0.76–0.78$ and F = 0.10–0.20 p.f.u. (Fig. 8, Table 2).

378 **Garnet**

379 Garnet exhibits a flat or significant compositional zoning in samples of banded type II to
380 schlieren orthogneiss (Fig. 9a–c, Table 3), whereas in samples of the nebulitic orthogneiss
381 garnet displays a strong zoning (Fig. 9d, Table 3). In banded type II to schlieren orthogneiss,
382 some garnet shows flat profiles with high X_{Fe} (0.95–0.98) and almandine ($\text{alm}_{0.65–0.72}$), low
383 pyrope ($\text{prp}_{0.02–0.04}$), grossular ($\text{grs}_{0.07–0.09}$) and spessartine ($\text{sps}_{0.10–0.19}$) (Fig. 9a–c, Table 3). In
384 some garnets of the banded type II to schlieren orthogneiss, grossular decreases ($\text{grs}_{0.18→0.09}$)
385 and spessartine increases ($\text{sps}_{0.09→0.19}$) from core to rim (Fig. 9a–c, Table 3). In the nebulitic
386 orthogneiss, garnet has higher grossular and lower spessartine in the core compared to banded
387 type II and schlieren orthogneiss, and garnet zoning is from core to rim marked by decrease in
388 grossular ($\text{grs}_{0.34→0.09}$) and increase in almandine ($\text{alm}_{0.59→0.68}$) and spessartine ($\text{sps}_{0.04→0.21}$),
389 and accompanied by a flat trend in high X_{Fe} (0.95–0.98) and low pyrope ($\text{prp}_{0.02–0.04}$) (Fig. 9d,
390 Table 3).

391 **WHOLE-ROCK CHEMISTRY**

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

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

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

417 FC076A). Here, the character of distribution patterns for all rock types shows only slight
418 differences in chemical composition. According to [Lange et al. \(2002\)](#) and [Chopin et al.](#)
419 [\(2012b\)](#), slight differences in major and trace elements between individual rock types may be
420 likely due to original heterogeneity of the protolith. On the other hand, [Marquer and Burkhard](#)
421 [\(1992\)](#) and [Hasalová et al. \(2008b\)](#) attributed these variations to the presence of external
422 fluids or melt in similar felsic rocks.

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

$$432 \quad \Delta m_i = F_v \cdot (\rho_a / \rho_0) C_a^i - C_0^i,$$

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

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

441 **FORWARD MODELLING OF MIGMATITIC ORTHOGNEISS**

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

449 Pseudosections were calculated using THERMOCALC 3.33 (Powell et al. 1998; version
450 2009) and dataset 5.5 (Holland and Powell, 1998; January 2006 upgrade), in the system MnO-
451 Na₂O-CaO-K₂O-FeO-MgO-Al₂O₃-SiO₂-H₂O-TiO₂-O (MnNCKFMASHTO). The activity-
452 composition relationships used are as follows: for silicate melt (liq), from White et al. (2007);
453 for garnet (g), biotite (bi) and ilmenite (ilm), from White et al. (2005); for feldspar (pl, ksp),
454 from Holland and Powell (2003); for white mica (wm), from Coggon and Holland (2002);
455 and, for cordierite (cd), from Holland and Powell (1998). The calculations are done for the
456 whole-rock composition of a nebulitic orthogneiss (sample FC076C; Table 4). Because of
457 closely similar whole-rock compositions of the other samples (Fig. 10b) the calculated
458 diagrams are also used for interpretation of their metamorphic evolution. The amount of H₂O
459 in the whole-rock composition was deduced from *T–M*(H₂O) pseudosection (see description
460 below, Fig. 12). Mineral composition isopleths of garnet, white mica, plagioclase and molar
461 proportion of liquid were plotted to discuss *P–T* conditions of mineral equilibration. The
462 isopleth notation used is: $m(\text{sps}) = \text{Mn}/(\text{Ca}+\text{Mg}+\text{Fe}+\text{Mn})\cdot 100$; $x(\text{alm}) =$
463 $\text{Fe}/(\text{Ca}+\text{Mg}+\text{Fe}+\text{Mn})\cdot 100$; $z(\text{grs}) = \text{Ca}/(\text{Ca}+\text{Mg}+\text{Fe}+\text{Mn})\cdot 100$; Si(wm) p.f.u.; $ca(\text{pl}) =$
464 $\text{Ca}/(\text{Ca}+\text{Na}+\text{K})$ and liq (mol. %).

465 ***T–M*(H₂O) pseudosection at 7 kbar**

466 In order to estimate the conditions of last equilibration in migmatites, it is assumed that the
467 assemblage tends to equilibrate until it becomes melt poor or melt absent near or at the solidus
468 (e.g. Štípská and Powell, 2005; Hasalová et al. 2008c). Therefore, a T - $M(\text{H}_2\text{O})$ pseudosection
469 at a pressure of 7 kbar was calculated first, simulating conditions of last equilibration with
470 melt near the solidus (Fig. 12). The observed assemblage of q-pl-ksp-g-bi-wm-ilm-liq is
471 stable at T - $M(\text{H}_2\text{O}) = \sim 0.30$ – 0.53 and at ~ 630 – 710 °C and the observed garnet rim
472 composition ($\sim \text{alm}_{68}$; grs_{10} ; sps_{20}) fits the calculated isopleths at conditions of T - $M(\text{H}_2\text{O}) =$
473 ~ 0.52 and ~ 630 °C (Fig. 12). Therefore, it is assumed that the rocks crossed the solidus with
474 minimum H_2O content corresponding to 2.07 mol. % of H_2O in the whole-rock composition.

475 **Closed-system P - T pseudosection**

476 The P - T diagram (Fig. 13) was calculated for an amount of $\text{H}_2\text{O} = 2.07$ mol. % deduced from
477 the T - $M(\text{H}_2\text{O})$ diagram (see Fig. 12 and related discussion). This H_2O amount allows the
478 stability of a typical mineral assemblage of a granite at MP - MT conditions, which is H_2O -
479 free and garnet-free (~ 5.2 – 6.7 kbar and < 550 – 620 °C), and therefore may illustrate rock
480 evolution in a closed system, even for H_2O . The major features and topology of the diagram
481 involve a H_2O -saturated solidus up to ~ 7.8 kbar followed by a steeply inclined H_2O -
482 undersaturated solidus from ~ 7.8 kbar to higher pressure at progressively higher temperature.
483 The biotite-out and garnet-out lines are temperature sensitive at suprasolidus and subsolidus
484 conditions, respectively and pressure sensitive at subsolidus conditions (Fig. 13).

485 The resulting P - T phase diagram shows a stability field of q-pl-ksp-g-bi-wm-ilm-liq
486 between ~ 4 – 12 kbar and ~ 630 – 740 °C, corresponding to the observed assemblage in all the
487 rock types. The nebulitic orthogneiss preserves high grossular content of garnet in the core
488 (alm_{60-65} ; grs_{30-34} ; sps_4 ; Fig. 9d), pointing to a P - T peak in the melt-free stability field of q-pl-
489 ksp-g-bi-wm-ru at ~ 14 – 16 kbar and ~ 600 – 740 °C (ellipse 1 in Fig. 13a–b). The high Si
490 content in white mica in all the rock types (Si = 3.35–3.40 p.f.u.; Fig. 7a) supports its

491 crystallization at high pressure conditions at ~16 kbar (Fig. 13c). The plagioclase composition
492 measured in films and interstitial grains is albite ($an_{0.05-0.02}$; $ab_{0.95-0.98}$), compatible with
493 plagioclase crystallization at HP conditions at ~16 kbar (Fig. 13d). The presence of interstitial
494 plagioclase, quartz and K-feldspar is interpreted as crystallized from melt at grain boundaries
495 (e.g. Hasalová et al. 2008a). We documented also albite and oligoclase composition in
496 myrmekite-like aggregates. According to Barker (1970), the myrmekite is typically associated
497 with higher anorthite content in plagioclase than observed in this study or other similar studies
498 (Štípská et al. 2019). The albite compositions measured in myrmekite may be result of the
499 presence of melt at HP conditions, and partial reequilibration of plagioclase to oligoclase on
500 decompression. The oligoclase compositions measured in cores of matrix plagioclase we
501 interpret as composition attained during recrystallization of the original aggregates during
502 burial (see also Chopin et al. 2012), part of these compositions, mainly at rims may reflect
503 also the last reequilibration of rocks on decompression. Combination of the textural argument
504 for melt presence with the very low anorthite content of plagioclase suggests crossing of the
505 H₂O-undersaturated solidus and beginning of partial melting at ~16 kbar. The calculated
506 isopleths of melt mode suggest a very small melt production around 1 mol. % (Fig. 13e). This
507 calculated volume of melt suggests that the melt was isolated in melt films, pools or pockets,
508 compatible with the observed microstructure (see Fig. 5b). Even if along this *P–T* path the
509 calculated molar proportion of melt increases for the whole rock, in individual textural
510 positions, as are the like-like grain boundaries, the phases may crystallize from melt in order
511 to lower the surface energy of the monomineral aggregates (e.g. Lexa et al. 2006; Hasalová et
512 al. 2008a). Likely mechanism for crystallization of albite and other phases from melt, while
513 the melt content of the rock increases, is also a process of dissolution-precipitation of these
514 phases promoted by the presence of melt and possibly also by crystallization from melt
515 (Sawyer, 2001; Hasalová et al. 2008b; Holness and Sawyer, 2008; Závada et al. 2018, Štípská

516 [et al. 2019](#)). The observed mineral assemblage q-pl-ksp-g-bi-wm-ilm with supposed former
517 melt in all rock types together with rim compositions of garnet (alm_{65-70} ; grs_{5-10} ; sps_{20-25} ; [Fig.](#)
518 [9](#)) and white mica ($\text{Si} = 3.10$ p.f.u.; [Fig. 7a](#)) point to last partial equilibration in the middle- P
519 part of the q-pl-ksp-g-bi-wm-ilm-liq stability field, at ~ 6 kbar and ~ 640 °C (ellipse 2 in [Fig.](#)
520 [13a](#)). Therefore, the absence of rutile and the core-to-rim zoning trends of garnet and white
521 mica in the nebulitic orthogneiss suggest a P - T path from ~ 14 - 16 kbar and ~ 600 - 740 °C to
522 ~ 6 kbar and ~ 640 °C, with local equilibration down to the ilmenite stability field and close to
523 the solidus.

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

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

529 Previously calculated diagrams and experiments for granitic rocks in the OSD and other parts
530 of the Bohemian Massif show that depending on the exact P - T path, the rocks may produce
531 up to 7-20% of melt on decompression from peak P - T conditions of ~ 25 kbar (for OSD:
532 [Walczak et al. 2017](#), for other occurrences in the Bohemian Massif: [Hasalová et al. 2008c](#);
533 [Lexa et al. 2011](#); [Nahodilová et al. 2014](#)). At such melt proportions, melt can be lost from the
534 rocks undergoing melting deeper in the crust, and can be added to the rocks above ([Štípská et](#)
535 [al. 2019](#)). Therefore, the effect of a possible melt gain in the studied rocks was explored in P -
536 $X_{\text{added melt}}$ diagram ([Fig. 14](#)). In the calculations, the composition of the melt is taken at 16 kbar
537 and 730 °C (see black star in [Fig. 13a](#)). The P - $X_{\text{added melt}}$ diagram was calculated for the
538 whole-rock composition of the nebulitic orthogneiss at conditions of the estimated peak
539 metamorphism ($T = 700$ °C), and for a range of compositions between the H_2O -undersaturated
540 whole-rock composition ($X_{\text{added melt}} = 0$) and the composition with 20 mol. % of melt added

541 ($X_{\text{added melt}} = 1$; Fig. 14). The main characteristics of the diagram show a liquid-in line heading
542 from ~12 kbar for $X_{\text{added melt}} = 0$ to ~20 kbar for $X_{\text{added melt}} = 0.25$. The stability of biotite, rutile
543 and ilmenite depends on pressure in the melt-present fields and on both pressure and $X_{\text{added melt}}$
544 in the melt-absent fields. The compositional isopleths of garnet and Si-in-phengite are
545 pressure sensitive in the melt-present fields, and do not depend on the amount of melt added
546 (Fig. 14). The horizontal arrow illustrates the effect of addition of melt to a rock with original
547 H₂O content inferred for the protolith (Fig. 13a). The starting position at $X_{\text{added melt}} = 0.05$,
548 allows the stability of the observed mineral assemblage q-pl-ksp-g-bi-wm, calculated is also
549 stability of infinitesimal amount (around 0.13–0.15 mol. %) of rutile which is not observed in
550 the rocks. The reason for this difference in the calculated and observed assemblage with
551 respect to rutile may be reequilibration on decompression to the ilmenite stability or also the
552 fact that Ti is not included in the model of white mica.

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

566 rock plus melt on decompression may therefore lie anywhere from the solidus to the higher
567 values of $X_{\text{added melt}}$ of the diagram.

568 For the studied rock types, the observed assemblage q-pl-ksp-g-bi-wm-ilm, absence of
569 rutile and the core-to-rim zoning trends of garnet ($\text{alm}_{60 \rightarrow 70}$; $\text{grs}_{34 \rightarrow 5}$, $\text{sps}_{4 \rightarrow 22}$; Fig. 9d) and
570 white mica ($\text{Si} = 3.40\text{--}3.10$ p.f.u.; Fig. 7a) are compatible with a decompression path from
571 ~ 16 kbar to ilmenite-bearing stability field at $\sim 8\text{--}13$ kbar. The garnet and white mica
572 compositional isopleths are not sensitive to the amount of melt added; therefore, the melt
573 amount that percolated in individual rock types is unknown. However, consequence of
574 different melt amount in different samples may be the difference in re-equilibration along the
575 decompression path. We suggest that for the nebulitic orthogneiss that preserves HP garnet
576 core the amount of melt was low at lower pressure, thus precluding complete garnet
577 reequilibration on decompression. Re-equilibration of garnet core close to the solidus
578 observed in the banded type II and schlieren orthogneiss may be caused by higher proportion
579 of melt at lower pressure compared to the nebulitic orthogneiss. Therefore, a new $P\text{--}T$
580 diagram with ~ 3.40 mol. % of re-integrated melt in the whole-composition was calculated for
581 explaining better the observed mineral assemblage and mineral compositions of the banded
582 type II and schlieren orthogneiss (Fig. 15).

583 **$P\text{--}T$ pseudosection with added melt**

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

589 The resulting $P\text{--}T$ diagram shows a stability field of q-pl-ksp-g-bi-wm-ilm-liq between
590 $\sim 4\text{--}13$ kbar and $\sim 610\text{--}740$ °C, corresponding to the observed assemblage. The $P\text{--}T$ peak

591 conditions are defined by the high grossular content measured in garnet cores (alm_{60-65} ; grs_{30-}
592 $_{34}$; sps_4 ; Figs. 9d and 15b) preserved in nebulitic orthogneiss and the high Si content in
593 phengite cores measured in all the rock types ($\text{Si} = 3.35-3.40$ p.f.u.; Figs. 7a and 15c). The
594 measured albite content of plagioclase films and interstitial grains is consistent with the
595 calculated isopleth of albite at these $P-T$ peak conditions ($\text{an}_{0.05-0.02}$; Fig. 15d). The last
596 equilibration at $\sim 630-640$ °C and ~ 6 kbar is constrained by the re-equilibrated compositions
597 of garnet (alm_{65-70} ; grs_{5-10} ; sps_{20-25} ; Fig. 15b) and white mica ($\text{Si} = 3.10$ p.f.u.; Fig. 15c). The
598 calculated isopleths of melt mode suggest a melt production during retrograde path of the
599 order of 1 mol. % resulting in up to 4 mol. % for 3.40 mol. % melt added (Fig. 15e).
600 Therefore, the mineral chemistry together with the absence of rutile record a decompression
601 path from $\sim 15-16$ kbar and $\sim 650-740$ °C to ~ 6 kbar and ~ 640 °C, with local equilibration
602 down to the ilmenite stability field and close to the solidus with the presence of melt. This
603 decompression path is compatible with the one previously described in the H_2O -
604 undersaturated diagram (Fig. 13), and differs only in the melt amount present during the peak
605 and retrograde evolution, suggesting that higher melt amount may result in more profound re-
606 equilibration of the assemblage close to the solidus. The melt might have been added to rocks,
607 explaining the textures, e.g. disintegration of the monomineral banding or more profound re-
608 equilibration of garnet composition, but the rocks may also undergo variable degree of melt
609 loss on decompression before last cooling through the solidus.

610 **DISCUSSION AND CONCLUSIONS**

611 The core of the Orlica-Śnieżnik Dome (Fig. 1b) consists of two antiforms affected by various
612 degrees of migmatization related to their distance from the site of presumed continental
613 subduction further west (see Fig. 13 in Chopin et al. 2012a). The proximal and small
614 Międzygórze antiform (~ 1 km across and ~ 4 km long; Fig. 1c) is characterized by
615 heterogeneously developed zones of partial melting surrounding blocks of well-preserved HP

616 rocks. The distal large-scale Králiky-Śnieżnik antiform (~6–8 km across and ~20 km long;
617 [Fig. 1c](#)) was affected by widespread melting and almost complete re-equilibration of *HP*
618 mineral assemblages. In this work the orthogneiss of the Králiky-Śnieżnik antiform are
619 compared with those of the Międzygórze antiform in order to understand the role of melt-
620 deformation interplays during the exhumation of large portions of continental crust in
621 Variscan continental collision zone.

622 **Microstructural and geochemical arguments for grain-scale melt percolation during D2** 623 **deformation**

624 In the study area, relics of shallow-dipping S1 foliation are folded by upright F2 folds and
625 transposed by almost pervasive N-S trending subvertical S2 foliation ([Fig. 3](#)). This structural
626 succession is the same as in the Międzygórze antiform but the degree of fabric transposition
627 to subvertical fabric is significantly broader. It reaches the width of ~6–8 km and length of
628 ~20 km, thereby attesting to a large-scale orogenic process.

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

635 Furthermore, the q-pl-ksp-g-bi-wm-ilm assemblage observed in all orthogneiss types is
636 the same despite variations in meso- and micro-scale structural and textural features (see [Figs.](#)
637 [3](#) and [4](#)). The presence of cusped K-feldspar in plagioclase layers, quartz and albite-rich
638 plagioclase intergrowths in K-feldspar aggregates, amoeboid grains of K-feldspar in quartz
639 layers and diffuse boundaries between different felsic layers are interpreted as a result of
640 grain-scale melt percolation through the solid felsic rock ([Figs. 4 and 5](#)). This interpretation is

641 in agreement with previously reported examples of [Hasalová et al. \(2008b\)](#), [Závada et al.](#)
642 [\(2007, 2018\)](#) and [Štípská et al. \(2019\)](#) where grain boundaries were open at the micron scale
643 to fluid/melt circulation (e.g. [Oliot et al. 2014](#)). Such interpretation is also supported by slight
644 differences in chemical composition. For instance, mass balance calculations of incompatible
645 elements show the depletion of U, Zr and Hf compatible with partial dissolution of zircon in
646 the melt, implying that some melt must have been lost or must have percolated through the
647 banded type II and schlieren orthogneiss, and the gain in Ba, Sr, Eu, K and Rb corresponding
648 to a heterogeneous nucleation of interstitial feldspar from percolated melt ([Fig. 11a–b](#)). On
649 the other hand, in the nebulitic orthogneiss, mass balance calculations show the depletion of
650 Th, Cs, Pr, La, U and Ta compatible with partial dissolution of monazite in the melt, implying
651 that some melt must have been lost ([Fig. 11c](#)). Therefore, the observed textural trend from
652 banded type II to nebulitic orthogneiss can be considered not only as a result of a deformation
653 gradient but also as a result of different degree of melt infiltration/percolation of granitic
654 sources under open-system conditions (e.g. [Hasalová et al. 2008](#); [Goncalves et al. 2012](#);
655 [Závada et al. 2018](#)).

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

657 The *P–T* path deduced from the forward-modelling indicates that the orthogneiss underwent a
658 decompression from *P–T* peak conditions of ~15–16 kbar and ~650–740 °C to ~6 kbar and
659 ~640 °C under the presence of melt ([Figs. 13 and 15](#)). The estimated *P–T* peak conditions are
660 recorded by the presence of Ca-rich garnet cores in the nebulitic orthogneiss and composition
661 of phengite in all the rock types. Subsequent decompression to ~6 kbar is recorded by the
662 core-to-rim zoning trends of garnet and white mica and absence of rutile. Similar *P–T*
663 evolutions have been repeatedly reported from other rock types such as eclogite or *HP*
664 granulite occurring in the orthogneiss (~18–30 kbar and ~700–1000 °C; e.g. [Pouba et al.](#)
665 [1985](#); [Bakun-Czubarow, 1991, 1992](#); [Steltenpohl et al. 1993](#); [Bröcker and Klemd, 1996](#);

666 Kryza et al. 1996; Klemd and Bröcker, 1999; Bröcker et al. 2010; Štípská et al. 2012; Ferrero
667 et al. 2015; Budzyń et al. 2015; Jedlička et al. 2017; Walczak et al. 2017; Majka et al. 2019)
668 with retrograde paths ranging from ~9 kbar/~730 °C to ~6 kbar/~560 °C (Steltenpohl et al.
669 1993; Bröcker and Klemd, 1996; Štípská et al. 2004, 2012). Recently, prograde *HP*
670 metamorphism was attributed to the S1 fabric in the granitic orthogneiss close to the eclogite
671 in the Międzygórze antiform (~13–18 kbar/~700–800 °C; e.g. Chopin et al. 2012b). It was
672 also shown that migmatization of orthogneiss surrounding the Międzygórze eclogite started in
673 eclogite-facies conditions at ~15–17 kbar in the S1 fabric (Štípská et al. 2019). These authors
674 also suggested that decompression along the S2 fabric from ~15–17 kbar to ~7–10
675 kbar/~690–740 °C was locally associated with anatexis. In analogy, we argue that the *P–T–d*
676 path for the S2 vertical fabric is comparable in both antiforms (Fig. 16).

677 The modelling showed that the orthogneiss protolith from the Králíky-Śnieżnik antiform
678 is able to produce only ~1 mol. % of melt along the *P–T* path. However, the macro- and
679 microstructural features are typical for advanced migmatization and attest to melt presence
680 along grain boundaries (Figs. 4 and 5), suggesting that a higher melt proportion was present.
681 This is possible to achieve only if H₂O is added to the rocks, but being above the conditions
682 of the wet solidus, the hydrating fluid must have been external melt (Štípská et al. 2019).
683 Such melt is supposed to be released by similar rocks buried deeper and this is simulated in
684 the modelling by adding granitic melt to the whole-rock composition. The modelling did not
685 allow estimation of the melt proportion in the rocks based on the mineral chemical
686 composition, as the mineral chemical composition for the observed assemblage is independent
687 on the amount of melt added (Fig. 14). However, it is supposed, that rocks that contain garnet
688 with high grossular content characteristic for *HP* conditions contained on decompression less
689 melt compared with rocks that show garnet with re-equilibrated grossular content (see Figs.
690 13–15).

691 The melt proportion remains unknown, and may have varied along the P – T path. Melt
692 percolation started already at ~15–16 kbar and ~650–740 °C as indicated by albitic
693 composition of plagioclase in myrmekite and in interstitial films, then melt equilibrated with
694 the minerals (at least with their rim compositions) during the retrograde history to ~6 kbar and
695 ~640 °C. However, the amount of melt percolated is not likely to be sufficient to produce a
696 melt-supported structure required for diatexite formation (Brown, 2007; Hasalová et al.
697 2008a,b,c), but may be sufficient to allow melt-assisted granular flow (Rosenberg and Handy,
698 2005; Závada et al. 2007; Schulmann et al. 2008). This is supported by the results of Štípská
699 et al. (2019) from the Międzygórze antiform, where it was shown that different migmatite
700 textures originated from variable degree of melt-rock interaction starting at ~17 kbar and
701 ~730 °C and ending at ~7–10 kbar.

702 **Back-stop extrusion of partially molten crust**

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

716 types is higher in the S2 compared to S1 we extrapolate the depletion of REE and HFSE also
717 to the S2 migmatitic fabric (Fig. 16a).

718 In the Králíky-Śnieżnik antiform the D2 deformation affects almost homogeneously the
719 whole volume of the felsic orthogneiss with only rare relics of low-strain S1 domains (Fig.
720 16b). The degree of transposition of S1 by S2, together with high proportion of schlieren and
721 nebulitic orthogneiss types in S2 fabric implies also that melt percolation affected
722 significantly larger proportion of crust during D2 compared to the Międzygórze antiform. The
723 orthogneiss samples from this study, from a large zone of almost homogeneous D2
724 deformation show similar degree of REE and HFSE depletion (Fig. 10c) as the mylonitic type
725 of Chopin et al. (2012b) from the Międzygórze antiform, which was interpreted as a result of
726 presence of melt during deformation. Because the zone of D2 deformation and proportion of
727 the migmatitic types is significantly higher in the Králíky-Śnieżnik antiform, the total
728 depletion REE and HFSE in this portion of crust is much more important compared to the
729 Międzygórze antiform (Fig. 16b). Another important feature is the last reequilibration of
730 rocks marked by mineral rim chemistry, under the presence of melt, along the decompression
731 $P-T$ path. The rocks from the Międzygórze antiform show last reequilibration under the
732 presence of melt at ~10 kbar while the rocks from the Králíky-Śnieżnik antiform show
733 reequilibration under the presence of melt at ~ 5 kbar. This implies that the difference in total
734 exhumation between two antiforms under the presence of melt within the D2 zone is
735 significantly higher in the case of the Králíky-Śnieżnik antiform compared to the
736 Międzygórze antiform.

737 Based on our petrological data it may be concluded that melt percolation along vertical D2
738 deformation zone facilitated exhumation of *HP* rocks in core of the two antiforms from ~60
739 km to 35 km (Międzygórze antiform, Fig. 16a) and ~50 km to >20 km (Králíky-Śnieżnik
740 antiform, Fig. 16b), allowing the juxtaposition of originally *HP* orthogneiss with *MP* rocks in

741 crustal-scale synforms (see also Štípská et al. 2004; Chopin et al. 2012a; Štípská et al. 2012,
742 2019). Another point that reinforces the role of the presence of melt for facilitating extrusion
743 of these deep rocks is a fact that the presence even of small melt volumes plays a major role
744 in decreasing the strength of the rocks (Rosenberg and Handy, 2005).

745 The above described deformation was related to horizontal shortening of collisional
746 wedge that contributed to the vertical extrusion of partially molten crust en masse in the
747 eastern part of the OSD. Such an extrusion of weak material is pronounced in particular close
748 to the Brunia back-stop where massive portions of rheologically weak and partially molten
749 rocks flowed upwards under horizontal stress from the root area of the orogenic wedge (Fig.
750 16). The extrusion model proposed by Thompson et al. (1997a,b) is suitable to well explain
751 the exceptionally high rate of exhumation suggested already by Steltenpohl et al. (1993) and
752 the shape of P – T path depicted by this study. Majka et al. (2019) suggested that first part of
753 exhumation occurred in channel flow region of the subducted continental margin along the
754 rigid buttress of the Brunovistulian microcontinent and that folding of the entire crustal
755 sequence occurred in the shallower part of the accretionary wedge. The minimum depth for
756 the F2 folding from this study and from Štípská et al. (2019) is ~16–17 kbar. Recently, the
757 numerical modelling of Maierová et al. (2014) tested various parameters controlling the
758 exhumation rates of hot gneiss domes and corresponding P – T – t paths such as rate of
759 convergence, heat production and erosion. These authors concluded that in the case of the
760 Orlica-Śnieżnik Dome the gravitational instability contribution was minor compared to
761 laterally forced folding leading to gneiss dome formation and exhumation of hot felsic lower
762 crust. All the above mentioned models tacitly suppose extreme weakness of thermally
763 softened hot felsic lower crust allowing homogeneous vertical flow. However, in detail, both
764 meso- and micro-scale mechanism allowing the extreme drop of strength of the felsic crust
765 remain enigmatic. Only recently, natural observations from hot gneiss domes and their

766 analogue modelling suggest that partial melting can trigger detachment folding and vertical
767 flow of migmatites and granitoids (Lehmann et al. 2017). Based on our study we argue that
768 the grain-scale melt percolation (called also reactive porous flow, melt infiltration) represents
769 such a principal weakening mechanism allowing homogeneous flow of crust typical for
770 extrusion. It is probably also a principal mechanism controlling exhumation of deep partially
771 molten crust in hot collisional orogens such as the European Variscan belt.

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1082 **FIGURE AND TABLE CAPTIONS**

1083 **Fig. 01.** (a) Lithotectonic map of North European Variscan belt (modified after
1084 [Aleksandrowski et al. 1997](#); [Don et al. 2003](#), [Żelaźniewicz et al. 2006](#)). The position of
1085 Orlica-Śnieżnik Dome is outlined. (b) Geological sketch map of the Orlica-Śnieżnik Dome

1086 (modified after Aleksandrowski et al. 1997; Don et al. 2003; Żelaźniewicz et al. 2006). (c)
1087 Geological and structural map of Międzygórze and Králíky-Śnieżnik antiforms (lithologies
1088 after Don et al. 2003; Chopin et al. 2012a,b; Štípská et al. 2012). The location of the
1089 orthogneiss used in this study and the position of the structural profiles are indicated.

1090 **Fig. 02.** Interpretative cross-sections showing the main structural relationships between both
1091 antiforms (a–b) and the different orthogneiss types. See Figure 1c for locations of profiles.

1092 **Fig. 03.** (a) Outcrop photograph illustrating a lithological sketch (see legend in the Figures 1
1093 and 2) and the locations of studied samples with white stars. (b–f) Field photographs showing
1094 the different textural rocks displayed, ranging from banded type II to schlieren and nebulitic
1095 orthogneiss.

1096 **Fig. 04.** Representative CL image mosaics showing textural features of the different rock
1097 types: (a) banded type II orthogneiss (sample FC076B): alternation of individual layers of
1098 recrystallized K-feldspar, quartz and plagioclase distinctly defined, with less diffuse
1099 boundaries between individual layers. (b) Banded type II orthogneiss (samples FC076E):
1100 alternation of almost individual layers of recrystallized K-feldspar, quartz and plagioclase
1101 with highly diffuse boundary between feldspar layers. (c) Schlieren orthogneiss (sample
1102 FC076J) characterised by relics of K-feldspar-rich layers in otherwise isotropic matrix. (d)
1103 Nebulitic orthogneiss (samples FC076C) with no relics of original banding.

1104 **Fig. 05.** Detailed back scattered electron (BSE) and cathodoluminescence (CL) images of
1105 different rock types. (a) Diffuse boundary between a layers of mixed aggregates and a K-
1106 feldspar-rich layer with numerous interstitial plagioclase, quartz and myrmekite. (b) Detail of
1107 cusped plagioclase, rounded quartz and myrmekite as interstitial phases in K-feldspar layer.
1108 (c) Quartz- and plagioclase-rich layers with interstitial grains of feldspar and quartz. Phengite
1109 is surrounded by biotite and biotite occurs also along cleavage of phengite. (d) Large crystal
1110 of plagioclase with and antiperthitic core rich in exsolutions of K-feldspar. (e) Garnet in a

1111 matrix of quartz, plagioclase, K-feldspar, phengite, biotite and myrmekite. (f) Symplectite of
1112 quartz and white mica at the boundary of K-feldspar and phengite. (g) Large phengite laths
1113 with biotite at margins and along the cleavage.

1114 **Fig. 06.** Composition of feldspar ($an = (Ca/(Ca+Na+K))*100$ in blue color and $or =$
1115 $(K/(Ca+Na+K))*100$ in yellow color) for the different rock types localized in different
1116 microstructural positions (a–f). Representative analyses are listed in [Table 1](#).

1117 **Fig. 07.** Composition of white mica: (a) Si (p.f.u.) content of white micas *v.* samples. Note the
1118 decrease in Si (p.f.u.) content from cores to rim. (b–c) Compositional ranges from cores to
1119 rims of white micas. Grey and yellow arrows indicate zoning from core to rim. Representative
1120 analyses are listed in [Table 2](#).

1121 **Fig. 08.** Composition of biotite for the different rock types (a, b). Representative analyses are
1122 listed in [Table 2](#).

1123 **Fig. 09.** Garnet compositional trends of the different rock types: (a) weakly migmatized
1124 banded type II orthogneiss (sample FC076B); (b) banded type II orthogneiss (sample
1125 FC076E); (c) schlieren orthogneiss (sample FC076J); and (d) nebulitic orthogneiss (FC076C).
1126 Representative analyses are listed in [Table 3](#).

1127 **Fig. 10.** Isocon diagrams after [Grant \(1986\)](#) comparing: (a) the average whole-rock
1128 composition of schlieren orthogneiss and (b) the whole-rock composition of the nebulitic
1129 orthogneiss (vertical axis) with respect to the composition of the weakly migmatized banded
1130 type II orthogneiss (sample FC076A). The diagrams (a) and (b) show loss and/or gain of
1131 major elements with respect to composition of reference rock plotted as reference line. (c)
1132 Spider plot normalized to chondrite ([Evensen et al. 1978](#)). Shaded field corresponds to the
1133 orthogneiss types described by [Chopin et al. \(2012b\)](#) in the Międzygórze antiform. (d) Spider
1134 plot normalized to weakly migmatized banded type II orthogneiss (sample FC076A).

1135 **Fig. 11.** Potdevin diagrams (Potdevin and Marquer, 1987; Lopez-Moro, 2012) illustrating the
1136 loss-gain relationships for a range of incompatible elements for banded type II, schlieren and
1137 nebulitic orthogneiss compared to weakly migmatized banded type II orthogneiss (sample
1138 FC076A). Relative gain or loss of mass (Δm_i) is normalized by the initial mass (m_i^0) in the
1139 volume-composition diagrams. The volume factor (F_v) is given by the volume ratio between
1140 the transformed rocks and initial one. Shaded area refers to the magmatic fluctuation ranges
1141 (immobile elements) and solid lines to mobile elements. LILE = Large Ion Lithophile
1142 Elements, HFSE = High Field Strength Elements, REE = Rare Earth Elements. (see Table 5).

1143 **Fig. 12.** P - $M(\text{H}_2\text{O})$ pseudosection calculated at 7 kbar for a nebulitic orthogneiss (sample
1144 FC076C) and contoured for the calculated spessartine ($m(\text{sps})$), almandine ($x(\text{alm})$) and
1145 grossular ($z(\text{grs})$) contents of garnet and for the molar proportion of liquid (liq (mol. %)). The
1146 solidus is emphasized by a dark-dashed line. Quartz, plagioclase and K-feldspar are present in
1147 all fields.

1148 **Fig. 13.** (a) P - T pseudosection calculated for the analysed whole-rock composition of a
1149 nebulitic orthogneiss (sample FC076C). (b-e) Simplified pseudosections with compositional
1150 isopleths of spessartine ($m(\text{sps})$), almandine ($x(\text{alm})$) and grossular ($z(\text{grs})$) in garnet; Si
1151 content of white mica (Si(wm) p.f.u.); anorthite content of plagioclase (ca(pl)); and molar
1152 proportion of melt (liq (mol. %)). The ellipses indicate the P - T ranges compatible with the
1153 observed assemblage and core and rim compositions of garnet and white mica. The star
1154 indicates P - T conditions from which the melt composition was taken to be reintegrated into
1155 whole-rock composition shown in Figures 14 and 15. The solidus is underlined by a thick
1156 black dashed line. See text for discussion of the P - T path.

1157 **Fig. 14.** T - $X_{\text{added melt}}$ pseudosection calculated at 700 °C for a range of compositions
1158 representing mixtures of H_2O -undersaturated whole-rock composition of the nebulitic
1159 orthogneiss ($x = 0$) and the composition of melt calculated at 680 °C and 16 kbar ($x = 1$, black

1160 star in [Figure 13](#)). $X_{added\ melt}$ is the proportion of melt added in the migmatitic orthogneiss at 16
1161 kbar. $x = 1$ corresponds to 20 mol. % of melt added. The solidus is underlined by a thick black
1162 dashed line. The calculated isopleths show molar proportion of garnet (g mol. %), almandine
1163 ($x(alm)$) and grossular ($z(grs)$) content of garnet, Si content of white micas (Si(wm) p.f.u.)
1164 and molar proportion of melt (liq mol. %). The ellipses indicate the P – T ranges compatible
1165 with the observed assemblage and core and rim compositions of garnet and white mica.
1166 Evolution along three decompression paths at different $X_{added\ melt}$ is discussed in the text.

1167 **Fig. 15.** (a) P – T pseudosection calculated with 3.40 mol. % melt added to the whole-rock
1168 composition of the nebulitic orthogneiss. (b–e) Simplified pseudosections with compositional
1169 isopleths of spessartine ($m(sps)$), almandine ($x(alm)$) and grossular ($z(grs)$) in garnet; Si
1170 content of muscovite (Si(wm) p.f.u.); anorthite content of plagioclase ($ca(pl)$); and molar
1171 proportion of melt (liq (mol. %)). The ellipses indicate the P – T ranges compatible with the
1172 observed assemblage and core and rim compositions of garnet and white mica. The solidus is
1173 underlined by a thick black dashed line. See text for discussion of the P – T path.

1174 **Fig. 16.** Sketches summarizing the variations of different textural parameters and melt loss-
1175 gain relationship as indicated by geochemical signatures (REE – Rare Earth Elements; HFSE
1176 – High Field Strength Element) across both antiforms. (a) Międzygórze antiform: melt-absent
1177 orthogneiss with local migmatite formation. Shallow-dipping S1 fabric preserved in the low-
1178 strain D2 domains. Grain-scale melt percolation is mainly localized in narrow zones of
1179 vertical S2 fabric (modified after [Štípská et al. 2019](#)). Compilation of P – T paths for
1180 orthogneiss reported by [Chopin et al. \(2012b\)](#) (1), and [Štípská et al. \(2019\)](#) (2). (b) Králíky-
1181 Śnieżnik antiform: melt-present migmatitic orthogneiss. High-strain D2 domains with
1182 subvertical S2 foliation are connected with crustal-scale shear zones. The wide zone of
1183 subvertical S2 fabric is to a large extent percolated by melt. Simplified P – T diagram with the
1184 P – T path obtained in this study. P – T paths are mostly within the melt-present assemblages.

1185 (c) Tectonic sketch of the Orlica-Śnieżnik Dome as a part of the Sudetes showing shallow-
1186 dipping S1 fabrics related to subduction of the continental crust up to eclogite- and (U)HP
1187 granulite-facies conditions, and subvertical S2 fabrics related to its vertical exhumation to the
1188 middle and upper crust (modified after [Chopin et al. 2012a](#)). The positions in subduction
1189 wedge of two antiforms are indicated as a proximal part (a) and a more distal part (b) with
1190 respect to the subduction.

1191 **Table 01.** Representative analyses of feldspar.

1192 **Table 02.** Representative analyses of micas.

1193 **Table 03.** Representative analyses of garnet.

1194 **Table 04.** Major and trace-element compositions (ICP-AES and -MS). LOI: Loss On Ignition.

1195 **Table 05.** Gain/loss of incompatible elements of the studied rocks relative to weakly
1196 migmatized banded type II orthogneiss (sample FC076A).