Ambiguous isotopic and geochemical signatures resulting from limited melt interactions in a seemingly composite pluton: a case study from the Strzegom–Sobótka Massif (Sudetes, Poland)

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Abstract

The western part of the late Variscan Strzegom–Sobótka massif exemplifies an intrusion that formed in a multi-component system. The system shows ambiguous isotopic and geochemical signatures and presents a seemingly composite character [biotite granite with negligible amount of hornblende (HBG) and microgranular mafic enclaves (MMEs)]. The melts responsible for forming granitic rock and its MMEs were not derived from contrasting crustal and mantle sources but from different crustal domains. These melts exhibit hybrid characteristics due to subsequent contamination processes. The data suggest two different stages of hybridization of two geochemically distinct magmas corresponding to granite and enclaves. The processes were temporally and spatially distinct. Heterogeneous protoliths and hybridization made these magma isotopic signatures and trace element patterns ambiguous. The main mechanism involved in developing granitic rocks was fractional crystallization (FC); the mechanism responsible for MME composition was mixing, but other local processes were also involved in rock formation. Generally, the FC signature was not obscured by the hybrid mafic melt input. This felsic-mafic melt interaction was a limited one-sided process. Granitic melt was changed by the presence of small mafic magma blobs, but their influence on granite crystallization was negligible. The massif is thus an ideal example of an environment where MMEs coexist with a granitic body, although MMEs do not determine the actual composite character of the pluton. The early stage of Variscan pluton formation within the Bohemian Massif and adjacent areas was shaped by interacting mantle- and crust-derived melts. The pluton described in this paper belongs to the late stage of Variscan magmatism. It is a multi-component system shaped by the participation of melts derived from heterogeneous crustal domains. The presence of enclaves does not explicitly indicate their dominant role in the formation of composite plutons of crust-mantle origin. It is decisive to determine whether the ambiguous signatures appearing in such cases truly indicate both of these sources.

Keywords Strzegom-Sobótka massif · Granite · Microgranular mafic enclave · Magma evolution · Magma mixing and mingling · Geochemical modelling

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Introduction

Granitic rocks, which constitute a major part of the continental crust, are found in nearly all tectonic environments. The vast majority of granite intrusions are located in orogenic regions (collision zones), where the melting of crustal rocks is initiated by rising mafic melt (related to subduction) or by thickening of the crust. Melting of the crust driven by heat from the mantle and mantle melts represents a very complicated system controlled by many factors, including chaotic magma interactions.

Such interactions have been demonstrated to play significant roles in the petrogenesis of many granitic plutons worldwide. Understanding this process yields better insight

into lithosphere dynamics (e.g., Walker and Skelhorn 1966; Reid et al. 1983; Vernon 1984, 1990; Furman and Spera 1985; Didier 1987; Frost and Mahood 1987; Poli and Tomasini 1991; Didier and Barbarin 1991, 2005; Perugini and Poli 2012; Wiebe et al. 1997; Silva et al. 2000; Ratajeski et al. 2001; Janoušek et al. 2004; Liu et al. 2013; Farner et al. 2014; Kocak and Zedef 2016).

Most of the European Variscan granitic massifs may be related to crust-mantle interaction. Such granitic massifs often present calc-alkaline to high-K calc-alkaline compositions and contain mafic enclaves, features that are typically interpreted as indicators of mixed crust-mantle magma sources (Finger et al. 1997). However, this interpretation is not necessarily correct. Roberts and Clements (1993) argue that calc-alkaline magmatism need not be connected to crust-mantle magma sources and may result from crust melting only: “the chemistry of the magmas depends primarily on the nature of the protoliths rather than the processes involved in their generation”. Determining that the case of only crustal protoliths occurs is difficult because processes of magma differentiation are also important and may obscure some primary signatures. Ambiguities in trace element and isotope evidence in granite and its enclaves (low isotope contrast with their host) make solving the protolith problem difficult (Pin et al. 1990). If two melts exchange components and the process is long enough, the melts should maintain consistent Harker trends. A short exchange process may result in deviation (e.g., experiments by De Campos et al. 2008, 2011). Thus, the time and place of interaction, the volume of interacting melts and their chemical and physical features are important factors influencing homogenization and equilibration processes. For tracking these processes, whole-rock analyses and single mineral studies are very useful. Whole-rock analyses provide average information from signals (partly overlapping) of all subsequent processes and are mostly used as a base for geochemical modelling of the mechanisms of magma differentiation. Single domains of minerals may preserve evidence of different differentiation mechanisms (Słaby et al. 2007a, b, 2008, 2012; McLeod et al. 2010; Miles et al. 2013; Michel et al. 2016; Laurent et al. 2017). The combination of both sets of data may be helpful in resolving a problem regarding the real nature of a pluton that is characterized as an apparently composite pluton.

The presence of MMEs in felsic plutons is considered important evidence of melt–melt interaction (Didier 1973, 1991; Castro et al. 1990; Vernon 1984, 1990, 2010; Frost and Mahood 1987; Didier and Barbarin 1991, 2005; Hibbard 1991; Orsini et al. 1991; Wiebe et al. 1997). MMEs can also shed light on the magma source, the mode of emplacement of mafic and granitoid magmas and the dynamics of magma chambers (Vernon et al. 1988; Hrouda et al. 1999; Paterson

et al. 2004; Perugini and Poli 2000; Słaby and Martin 2008; Kumar 2010; Perugini and Poli 2012).

However, the presence of enclaves, even if they represent evidence of melt interactions, cannot be an unambiguous indicator of the main mechanism of pluton formation and testify to a crust-mantle melt origin. The Strzegom–Sobótka pluton is a good example of such a complicated environment. The apparently composite pluton shows no affinity with other Central European Variscan granitic massifs. This pluton has been dated at ~305–295 Ma (Turniak et al. 2014) and is considered an example of the most recent product of granitic plutonism in the Central European Variscides. Pilot studies have shown that the granite facies in the western part of this massif clearly exhibits the features of limited magma mixing-mingling processes due to the presence of unbalanced quantities of mafic and felsic magma (Domańska and Słaby 2004; Domańska-Siuda and Słaby 2005). Further, enclaves and granite show vagueness in their isotope signatures, which was first reported by Pin et al. (1989). Careful observations of mineral growth textures and geochemical models of magma evolution indicate that granite and enclave evolution paths were separate, which may indicate that the melts might have been emplaced into the magma chamber at least partly as hybrids. Different geochemical and isotopic characteristics of granite/enclave melts do not exclude crustal protoliths for both melts. We find this problem worth discussing because the massif is exceptional among all the Central European late Variscan granite plutons. The other plutons clearly show mantle-crust interaction without any ambiguity in isotope signatures and trace element patterns. We discuss the problem using a set of new tools aimed at identifying and quantifying the processes responsible for the magmatic evolution of the pluton. Whereas the mineral compositions and rock textures are used to define local processes, geochemical modelling is critical for the diagnosis of the whole pluton’s formation.

Geological setting and field relationships

The late stages of the development of the European Variscides were accompanied by extensive granitoid magmatism from the Late Devonian–Visean to the Late Carboniferous–Early Permian. Changes in tectono-thermal conditions within the evolving orogen resulted in the formation of a variety of granite types with distinct geochemical signatures and ages (e.g., Schaltegger 1997; Finger et al. 1997, 2009). In the Sudetes, at the north-eastern margin of the Bohemian Massif, the intermittent process of crustal melting resulted in the formation of a number of late- to post-orogenic granitic plutons that were emplaced in roughly two phases dated to ca. 340–330 Ma and ca. 320–295 Ma (Finger et al. 1997;

Mazur et al. 2007 and references therein; Finger et al. 2007; Słaby and Martin 2008; Turniak et al. 2014). Older granitoids are believed to have been generated as a result of nappe stacking that led to an increase in heat in the thickened orogenic root (e.g., Franke 2000). Younger granitoids intruded after the cessation of the Variscan convergence; the most recent granitoids are coeval with bimodal volcanism (e.g., Kryza and Awdankiewicz 2012; Awdankiewicz et al. 2014; Turniak et al. 2014). Thus, younger granitic plutonism could have resulted from decompression melting related to lithospheric thinning in the post-Variscan extensional regime (e.g., Henk 1997). Most of the granitic bodies are composite plutons that crystallized from melts derived from multiple sources (e.g., Pin et al. 1989; Finger et al. 1997, 2007; Gerdess and Wörner 2000; Žák et al. 2014).

The Strzegom–Sobótka Massif (SSM) is located in the Sudetes (SW Poland) at the north-eastern periphery of the Bohemian Massif of the Central European Variscides between the WNW–ESE-trending regional tectonic zones of the Upper Elbe and the Middle Odra (Fig. 1). The Sudetes comprise Neoproterozoic–Lower Paleozoic volcano-sedimentary sequences and magmatic suites that were metamorphosed by Cadomian and Variscan orogenic events and are covered in places by the Upper Devonian–Lower Carboniferous sediments of intramontane troughs (Żelaźniewicz 1997; Mazur et al. 2006). In the Late Carboniferous–Earliest Permian, granitic plutonism resulted in the formation of a number of granitic bodies that are widely distributed throughout the region (Fig. 1). The NE–SW-trending Sudetic Boundary Fault, a Late Variscan structure that was reactivated during the Alpine orogeny (Aleksandrowski et al. 1997; Badura et al. 2007 and ref. therein), divides the Sudetes into the elevated mountainous Sudetic Block in the SW and the strongly peneplained lowlands of the Fore-Sudetic Block in the NE.

The SSM is located on the north-eastern side of the Sudetic Boundary Fault, in the Fore-Sudetic Block, approximately 50 km southwest of Wrocław (Fig. 1). At their current level of exposure, the granites extend over 50 km WNW–ESE between Sobótka in the south-east and Jawor in the north-west. The granitic magmas intruded into a mildly metamorphosed volcano-sedimentary sequence that bears palaeontological records corresponding to ages from the Lower Paleozoic to the lower part of the Upper Carboniferous (Jerzmański 1970; Jerzmański and Teller 1971; Grocholski and Sawicki 1992). In the east, the granites are in contact with the ca. 400 Ma mafic–ultramafic suite of the Ślęza massif, which is a part of the Central Sudetic Ophiolite (Oliver et al. 1993; Dubińska et al. 2004; Kryza and Pin 2010). In the south, the SSM adjoins the Góry Sowie Block, which mainly comprises amphibolite-facies gneisses and amphibolites, partly migmatized and derived from the Neoproterozoic–Cambrian volcano-sedimentary sequence and magmatic rocks (Gunia 1999; Kryza and Fanning 2007

and references therein). The Góry Sowie Block has been interpreted as a fragment of the subducted middle lower continental crust (e.g., Mazur et al. 2006). To the west and south-west, the Sudetic Boundary Fault separates the SSM from the Kaczawa Unit, which is composed of a low-grade Neoproterozoic/Cambrian to Upper Devonian/Lower Carboniferous metasedimentary and metavolcanic succession, interpreted to comprise fragments of a Variscan accretionary prism (Baranowski et al. 1990; Kryza and Zalasiewicz 2008 and references therein).

The SSM comprises four main varieties of granite, which intruded as separate magma batches: biotite granite with negligible amount of hornblende (which is a subject of this study and is henceforth referred to as HBG), biotite granite devoid of hornblende, two-mica granite and biotite granodiorite (Majerowicz 1972; Pin et al. 1989; Puziewicz 1990) (Fig. 1). They show different paths of magma differentiation and different paths of post-magmatic alterations. The narrowing of the exposure of these granites east of Strzegom divides the SSM into an eastern part, which is built of two-mica granite and biotite granodiorite, and a western part, which is composed of biotite granite varieties. Currently, contacts between western and eastern parts is not observed in the field.

Recent low-error zircon dating (Turniak et al. 2014) has indicated that the SSM was amalgamated between ca. 305 and 295 Ma, beginning with the two-mica granite and followed by the biotite granodiorite and then by the biotite granite varieties. Similar conclusion is drawn by Pin et al. (1989) on a base of whole-rock isotope investigations. Pin et al. (1989) show very different initial Sr isotope ratios for the granites of the western and eastern parts of the SSM. They also point to significant differences between granite varieties within each of the two parts.

Sampling and analytical methods

Over 100 rock samples of the biotite granite variety HBG from western SSM block (henceforth referred to as HBG) and its enclaves (MME), xenoliths and mafic schlieren were collected (Table 1, Supplementary Material). Extreme care was taken to choose samples that were non-weathered and representative, i.e., devoid of mafic schlieren, pegmatites, aplites, and features of hydrothermal alteration.

Cathodoluminescence (CL) petrography was carried out on polished thin sections using a Nikon Eclipse E600 POL polarizing microscope equipped with a Citl-CCL 8200 cold cathode stage at the Institute of Geological Sciences, University of Wrocław, Poland. CL was excited at a gun current of 400–500 mA and an accelerating voltage of 10–15 kV, and it was digitally recorded with a Canon 590 camera at an exposure time of 1–8 s.

Fig. 1 **a** Position of the research area in the European Variscides. Variscan massifs are shadowed. **b** Position of the research area in the Bohemian Massif. Major granitoid intrusions are indicated by crosses. **c** Variscan granitoids in the Sudetic fragment of the Bohemian Massif (compilation with modifications after Aleksandrowski et al. 1997; Mazur et al. 2006; Oberc-Dziedzic et al. 2013). The Strzegom–Sobótka Massif is indicated by a black rectangle. Symbols: 1—low-grade metamorphic rocks; 2—mylonites; 3—gneisses; 4—mafites and ultramafites of the Central Sudetic Ophiolite; 5—Cadomian granitoids; 6—Variscan granitoids; 7—Moravo–Silesian metaigneous and metasedimentary rocks; 8—molasse; 9—faults; 10—thrusts; **d** Geological sketch map of the Strzegom–Sobótka massif (after Majerowicz 1972; Puziewicz 1990; modified). *ISF* Intra-Sudetic fault, *KdG* Kudowa granite, *KG* Karkonosze granite, *KZG* Kłodzko–Złoty Stok granite, *MO* Moldanubian zone, *MOF* Middle Odra fault, *MS* Moravo–Silesian zone, *NG* Niemcza granitoid, *RH* Rhenohercynian zone, *SrG* Środa Śląska granite, *SSG* Strzegom–Sobótka granite, *StG* Strzelin granite, *ST* Saxothuringian zone, *SzG* Szprotawa granite, *UEF* Upper Elbe fault, *WrG* Wrocław granite, *ZG* Žulová granite

The chemical compositions of minerals were investigated using a Cameca SX-100 electron microprobe (WDS mode) in the Electron Microprobe Laboratory at the Inter-Institute Microanalytical Complex for Minerals and Synthetic Substances, Warsaw University, Poland. The following instrumental conditions were applied: a counting time of 10–20 s; an acceleration voltage of 15 kV and a beam current of 20 nA for major elements and those of 20–30 kV and 50 nA for trace elements. The following standards were used: albite (Na); diopside (Mg, Si, Ca); wollastonite (Si, Ca); orthoclase (K, Al); haematite (Fe); rhodochrosite (Mn); apatite (P, F); phlogopite (F); barite (S, Ba); rutile (Ti); zircon (Zr); synthetic strontium titanite (Sr); YAG (Y); synthetic lanthanum hexaboride, LaB_6 (La); synthetic cerium(III) ultraphosphate, $\text{CeP}_5\text{O}_{14}$ (Ce); synthetic neodymium gallate, NdGaO_3 (Nd); synthetic chromium(III) oxide, Cr_2O_3 (Cr); synthetic NiO (Ni) and tugtupite (Cl). The typical spot size ranged between 2 and 5 μm depending on the analyzed mineral. Matrix correction was performed using the standard PAP procedure.

Feldspar and biotite formulae were recalculated on the basis of 8 and 11 oxygens, respectively. Although $\text{Fe}^{2+}/\text{Fe}^{3+}$ ratios in biotite have been estimated by Wichrowska (1974, 1977) and Puziewicz (1994, 1995), their sampling does not overlap with ours. Therefore, we decided to treat all Fe as Fe^{2+} . The amphibole formulae and estimated $\text{Fe}^{2+}/\text{Fe}^{3+}$ ratios were calculated using the procedure of Locock (2014) and following the current nomenclature of IMA (Hawthorne et al. 2012). OH contents were estimated using the equation $\text{OH} = 2\text{-Ti}$. For most analyses, the average cation contents were obtained from normalizations to 15 (Σ cations from Si to Ca = 15 apfu). In several cases, when the sum of the C-site cations was too low, the normalization to 13 (Σ cations from Si to Mg = 13 apfu) was chosen. Apatite analyses were recalculated on the basis of 8 cations. Allanite formulae were recalculated by setting the total of the T- and M-site cations

equal to 6 and adjusting the $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio to obtain a total of 25 negative charges (Ercit 2002).

The samples used for geochemical and isotopic analyses were selected based on thin section analyses. Approximately, 0.5–1.0 kg of each sample was crushed and quartered, and 100 g of each rock was milled into powder. Whole-rock major and trace element analyses were performed at the ACME Analytical Laboratories in Vancouver (Canada). Major oxide and trace element compositions were determined by ICP-ES (inductively coupled plasma emission spectrometry). Trace elements were analyzed by ICP-MS (inductively coupled plasma mass spectrometry), according to procedures described on <http://acmelab.com>. All analyses were calculated on an anhydrous basis, with iron expressed as $\text{Fe}_2\text{O}_3(t) = \text{Fe}_2\text{O}_3(t) + 1.111 \text{ FeO}$.

Rb/Sr and Sm/Nd isotopic analyses were performed on 100 mg of powdered samples at the Institute of Geological Sciences of the Polish Academy of Sciences, Warsaw (Poland) using a VG Sector 54 mass spectrometer operating in multi-collector dynamic mode. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1184$, whereas $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. The standard reference materials of NBS SRM 987 and JNdi-1 were used, which yielded a mean $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.710252 ± 18 (2σ) and a mean $^{143}\text{Nd}/^{144}\text{Nd}$ value of 0.512095 ± 13 (2σ), respectively. Time-related calculations used values of $^{143}\text{Nd}/^{144}\text{Nd} = 0.512638$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$ for the present-day depleted mantle following the radiogenic linear growth of the mantle with $\epsilon\text{Nd} = 0$ at 4.568 Ga. Epsilon values at time T were calculated using the following relation:

$$\epsilon_{\text{TNd}} = [({}^{143}\text{Nd}/{}^{144}\text{Nd}_{\text{sample}}/{}^{143}\text{Nd}/{}^{144}\text{Nd}_{\text{TCHUR}}) - 1] \times 10,000$$

where CHUR is the chondrite uniform reservoir and T is generally the time the rock was formed. Depleted mantle model ages are calculated assuming a modern upper mantle with values of $^{147}\text{Sm}/^{144}\text{Nd} = 0.2137$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.51315$.

Rock-mineral data were used in the geochemical modelling performed with the GENESIS software of Teixeira (1996). Isotopic data were processed using the ISOPLOT/EX ver. 2.4 software of Ludwig (2000).

Petrography

Hornblende-biotite granite (HBG)

The studied granite is a medium- to coarse-grained grey rock that is hypidiomorphic and equigranular to porphyritic in texture and typically exhibits non-oriented fabric. The phenocrysts are mostly K-feldspar and sometimes plagioclase. This granite is modally classified as a monzogranite and is composed of approximately 24–45% plagioclase, 21–44% K-feldspar, 17–38% quartz and 1–12% biotite, with

Table 1 Major and trace element analyses of the HBG

Sample	009	010	013	015	028	032	034	036	039	045	051	058
Locality	1	1	1	1	1	2	2	2	3	4	5	5
wt. %												
SiO ₂	72.99	72.49	71.89	72.38	73.15	73.20	72.30	70.66	72.01	73.19	72.27	70.13
TiO ₂	0.28	0.26	0.26	0.27	0.22	0.20	0.20	0.30	0.34	0.29	0.26	0.34
Al ₂ O ₃	13.30	13.68	13.94	13.78	13.30	13.33	13.57	13.99	13.89	13.09	13.56	14.14
Fe ₂ O ₃ <i>t</i>	2.85	2.57	2.63	2.80	2.27	2.09	1.48	2.72	2.54	2.83	2.71	3.10
MnO	0.05	0.05	0.05	0.06	0.05	0.05	0.03	0.05	0.05	0.05	0.06	0.07
MgO	0.30	0.27	0.28	0.29	0.21	0.22	0.29	0.38	0.54	0.33	0.27	0.41
CaO	1.42	1.39	1.31	1.40	1.21	1.26	1.41	1.39	1.94	1.40	1.39	1.54
Na ₂ O	3.70	3.68	3.52	3.64	3.41	3.20	3.18	3.46	3.28	3.46	3.61	3.80
K ₂ O	4.20	4.89	5.33	4.85	5.00	4.78	4.83	4.87	4.43	4.52	4.48	4.44
P ₂ O ₅	0.07	0.06	0.07	0.08	0.05	0.05	0.06	0.11	0.11	0.08	0.07	0.10
LOI	0.19	0.25	0.33	0.43	0.26	1.10	1.68	0.44	0.47	0.23	0.26	0.74
Total	99.46	99.70	99.76	100.10	99.24	99.48	99.15	98.51	99.71	99.58	99.08	98.94
A/CNK	1.01	0.99	1.00	1.00	1.01	1.05	1.04	1.04	1.01	1.00	1.02	1.02
Mg no.	0.17	0.17	0.17	0.17	0.15	0.17	0.28	0.22	0.30	0.19	0.16	0.21
ppm												
Ba	366.1	518.5	657.6	596.0	539.9	432.4	602.0	692.8	518.6	491.4	559.0	572.8
Cs	4.1	3.8	4.4	4.3	3.6	4.4	5.8	5.9	3.3	4.7	5.5	3.1
Rb	173.6	205.1	211.0	191.4	192.7	182.4	226.2	208.5	148.1	185.7	199.8	217.7
Sr	76.7	86.0	90.8	94.2	79.8	74.9	126.5	115.9	149.8	87.0	93.7	112.0
V	21.0	15.0	12.0	12.0	16.0	11.0	13.0	23.0	27.0	14.0	12.0	16.0
Co	5.3	2.3	2.3	2.4	2.8	1.8	2.3	4.6	3.7	3.4	3.7	3.6
Zr	294.1	246.0	231.3	257.0	189.8	179.2	162.4	223.8	200.2	231.7	279.3	220.7
Hf	7.9	7.0	6.8	6.5	5.5	5.3	5.0	5.9	5.4	6.9	7.7	5.9
Y	36.6	41.5	37.9	35.1	31.9	34.8	21.2	38.0	25.0	36.0	38.6	36.6
Nb	27.0	24.9	24.1	23.9	19.5	19.1	14.7	23.3	19.5	23.9	23.4	22.8
Ta	2.1	1.6	1.6	2.0	1.6	2.1	1.9	2.3	1.7	2.2	2.3	2.1
U	7.2	7.5	5.4	9.5	5.4	5.2	3.3	4.7	2.4	6.6	6.1	14.6
Th	28.4	27.0	21.0	22.5	15.5	16.9	18.3	13.7	20.4	23.1	23.3	18.5
La	69.5	62.8	55.0	54.6	41.5	33.5	47.0	47.9	52.4	54.0	62.7	39.7
Ce	129.9	114.5	99.8	110.0	80.2	66.9	95.1	90.3	99.7	100.8	120.0	78.6
Pr	13.3	12.8	11.2	11.9	8.4	7.7	9.4	9.4	10.5	10.8	12.6	8.7
Nd	46.6	45.2	39.5	42.0	31.9	30.7	34.1	35.7	36.3	39.0	44.7	31.8
Sm	8.4	8.0	7.0	7.2	6.5	6.1	5.7	6.6	7.1	7.4	8.7	5.8
Eu	0.9	0.9	1.0	0.9	0.9	0.8	0.9	1.2	1.0	1.0	1.0	0.9
Gd	6.9	6.5	5.9	6.7	5.4	5.1	5.0	5.9	5.8	6.4	6.7	5.9
Tb	1.2	1.2	1.0	1.0	1.0	1.0	0.8	1.1	1.0	1.2	1.2	1.0
Dy	5.2	7.0	5.7	5.9	4.9	5.8	4.0	5.6	4.5	5.3	5.7	5.9
Ho	1.2	1.3	1.2	1.2	1.0	1.2	0.8	1.2	0.9	1.2	1.3	1.1
Er	3.3	3.6	3.3	3.4	3.0	3.3	2.5	3.3	2.2	3.4	3.6	3.3
Tm	0.5	0.6	0.5	0.5	0.4	0.5	0.3	0.6	0.3	0.5	0.6	0.5
Yb	3.8	4.2	3.5	3.6	3.0	3.4	2.2	3.6	1.8	3.5	3.6	3.4
Lu	0.5	0.6	0.5	0.6	0.4	0.5	0.3	0.5	0.3	0.5	0.5	0.5
(La/Yb) _N	12.1	10.0	10.4	10.2	9.3	6.7	14.5	8.9	19.3	10.4	11.8	7.8
Eu/Eu*	0.4	0.4	0.5	0.4	0.5	0.4	0.5	0.6	0.5	0.4	0.4	0.5

Table 1 (continued)

Sample	060	065	066	067	080	082	085	087	102	107	138	140
Locality	6	7	7	7	8	8	8	8	4	4	3	3
wt. %												
SiO ₂	72.34	71.70	74.98	68.03	73.63	72.36	73.32	72.81	72.09	74.34	71.30	73.20
TiO ₂	0.25	0.25	0.14	0.59	0.16	0.21	0.23	0.23	0.29	0.17	0.36	0.20
Al ₂ O ₃	13.76	13.91	13.18	14.09	13.52	13.87	13.76	13.47	13.70	13.21	13.89	13.77
Fe ₂ O ₃ <i>t</i>	2.34	2.52	1.48	5.31	1.74	2.23	2.33	2.53	2.72	1.74	2.99	1.31
MnO	0.04	0.05	0.04	0.10	0.04	0.06	0.04	0.08	0.05	0.04	0.05	0.02
MgO	0.28	0.30	0.19	0.62	0.24	0.25	0.24	0.25	0.33	0.21	0.44	0.27
CaO	1.29	1.31	1.04	1.88	1.38	1.29	1.04	1.28	1.45	1.16	1.66	1.51
Na ₂ O	3.59	3.63	3.50	3.81	3.37	3.48	3.42	3.35	3.59	3.56	3.47	3.21
K ₂ O	4.82	4.56	4.57	3.98	4.60	5.08	5.15	4.88	4.70	4.49	4.12	4.61
P ₂ O ₅	0.06	0.10	0.04	0.15	0.04	0.05	0.06	0.06	0.08	0.05	0.12	0.06
LOI	0.32	0.33	0.32	0.26	0.39	0.28	0.39	0.24	0.28	0.36	1.10	1.30
Total	99.21	98.78	99.57	98.97	99.24	99.28	99.59	99.32	99.41	99.44	99.49	99.46
A/CNK	1.02	1.05	1.05	1.01	1.04	1.02	1.05	1.03	1.01	1.03	1.05	1.06
Mg no.	0.19	0.19	0.20	0.19	0.21	0.18	0.17	0.16	0.19	0.19	0.23	0.29
ppm												
Ba	593.7	506.4	379.1	255.4	630.6	594.9	837.0	601.1	560.3	415.0	403.0	551.8
Cs	3.4	14.8	9.4	6.0	5.8	5.0	2.8	5.5	5.2	12.5	6.6	5.2
Rb	195.9	257.0	232.2	203.0	210.3	201.4	194.7	219.1	184.8	232.4	192.4	215.2
Sr	89.3	83.4	58.6	85.3	85.7	87.1	97.2	99.1	92.3	68.6	97.8	118.1
V	11.0	11.0	0.0	22.0	0.0	9.0	11.0	9.0	14.0	< 8	18.0	11.0
Co	2.7	3.4	1.8	7.1	2.0	2.1	1.6	4.8	3.4	1.9	3.4	1.6
Zr	265.3	210.4	97.6	395.3	136.9	185.2	215.2	187.0	230.0	148.8	266.2	139.2
Hf	7.1	6.1	3.8	11.1	3.7	5.3	6.0	5.7	6.7	5.1	7.3	4.4
Y	40.4	32.3	33.7	47.0	25.8	33.6	38.1	29.8	40.2	34.3	39.2	23.8
Nb	24.5	22.3	15.9	34.5	14.1	19.9	24.7	19.7	24.8	21.5	23.5	15.1
Ta	2.2	2.6	2.8	2.6	1.0	2.1	2.7	1.6	2.2	3.9	2.7	1.6
U	6.3	10.3	11.9	6.3	7.4	6.1	9.2	4.5	7.8	14.3	3.7	3.4
Th	22.8	15.3	20.8	33.8	30.7	15.7	21.3	16.0	19.6	22.2	15.9	18.3
La	49.8	41.5	30.7	92.8	32.4	39.2	42.5	45.1	55.8	35.1	41.8	49.7
Ce	96.5	82.6	62.6	185.4	62.0	74.6	87.3	89.9	103.1	65.9	79.8	91.0
Pr	10.1	8.8	6.8	19.7	6.6	8.1	9.7	9.8	10.8	7.5	8.9	10.0
Nd	38.8	31.9	24.6	67.8	24.8	28.7	33.0	35.4	37.5	27.1	34.8	35.7
Sm	7.3	5.7	4.7	10.3	4.9	6.7	6.9	5.8	7.9	5.4	6.6	6.6
Eu	1.0	0.9	0.4	0.9	0.6	1.0	0.8	1.0	1.0	0.6	1.0	0.8
Gd	6.3	4.8	4.2	7.9	4.2	5.6	7.1	4.9	6.9	5.5	5.6	4.8
Tb	1.2	0.9	0.9	1.4	0.8	1.0	1.1	0.9	1.2	0.9	1.1	0.8
Dy	5.5	4.8	5.6	7.5	3.9	5.0	6.8	4.8	5.7	6.0	6.0	4.0
Ho	1.2	1.1	0.9	1.4	0.8	1.0	1.4	0.9	1.3	1.2	1.3	0.8
Er	3.5	2.9	3.0	4.0	2.4	2.9	4.1	2.5	3.5	3.7	3.7	2.1
Tm	0.6	0.4	0.4	0.6	0.4	0.5	0.6	0.4	0.6	0.5	0.6	0.3
Yb	3.6	3.0	3.0	4.3	2.3	3.1	4.3	2.6	3.9	3.7	3.8	2.0
Lu	0.5	0.4	0.5	0.7	0.3	0.5	0.6	0.4	0.5	0.5	0.5	0.3
(La/Yb) _N	9.1	9.4	6.8	14.6	9.3	8.4	6.6	11.8	9.6	6.4	7.3	16.5
Eu/Eu*	0.5	0.5	0.3	0.3								

LOI loss on ignition, A/CNK Al₂O₃/(CaO + NaO + K₂O) molar, Mg no. atomic Mg/(Mg + Fe), Eu/Eu* [Eu_N/√(Sm_N × Gd_N)], N chondrite normalized to values of Nakamura (1974)

Locality: 1—Strzegom, 2—Goczałków, 3—Wieśnica, 4—Kostrza, 5—Gniewków, 6—Czernica, 7—Zimnik, 8—Rogoźnica, 9—Paszowice, 10—Borów

amphibole as a minor constituent; in places, the amphibole content slightly exceeds 3% (Majerowicz 1972). There are no sharp contacts between the porphyritic and equigranular facies, and this transition appears to be gradational. Minor muscovite is present locally. Accessory minerals include apatite, zircon, allanite, titanite and rutile (Majerowicz 1972; Turniak et al. 2007). Alteration effects, such as albitization of feldspars and chloritization of biotite, are moderate, although they may be locally intensive. Aplitic and pegmatitic dikes and irregular streaks are common, and miarolitic

pegmatites of the MI-REE (miarolitic-rare-earth-element) subclass and NYF (niobium–yttrium–fluorine) affiliation, which range from centimetres to several metres in size, are locally abundant (e.g., Puziewicz 1985; Janeczek 2007; Pieczka et al. 2015).

The HBG also contains country-rock xenoliths and mafic enclaves (Majerowicz 1972; Maciejewski and Morawski 1975; Domańska 1984; Domańska and Słaby 2004; Domańska-Siuda and Słaby 2005). Detailed descriptions of all the relevant field and petrographic

Fig. 2 Field photographs of rock textures: **a** dioritic enclaves with progressive hybridization zone (indicated by dotted lines); **b** composite enclave rimmed by felsic halo; **c** hybridization zone; inside mafic body visible numerous plagioclase xenocrysts; **d** numerous mafic enclaves occurring within the pegmatite; **e** enclaves of different composition; **f** plagioclase crystal crosses the border enclave-granite and a large crystal of feldspar within the enclave. *Kfs* K-feldspar, *Plg* plagioclase

features of the HBG rocks can be found in previous works (Kural and Morawski 1968; Majerowicz 1972; Maciejewski and Morawski 1975).

Mafic magmatic enclaves (MMEs)

Mafic magmatic enclaves (MMEs) *sensu* Barbarin (1988) are relatively uniformly distributed in the granitic host and exhibit compositions ranging from monzonite, monzodiorite, granodiorite to tonalite. These MMEs range in size from a few centimetres to over a metre (Fig. 2a–f). The enclaves

with diameters up to 10 cm are generally dark grey (Fig. 2e) and fine-grained, whereas larger enclaves are lighter in colour and have relatively coarse-grained textures (Fig. 2a, b, e). MMEs show non-oriented textures; they are typically equigranular but occasionally contain 1- to 2-cm-long feldspar phenocrysts (Figs. 2a, f, 3a, d, e, 4c–e). The spherical to ellipsoidal shapes of MMEs and the common presence of fine-grained chilled margins and sharp or partly diffuse lobate to crenulated contacts suggest that they are quenched blobs of silica-poor magma that intruded the granitic host (Fig. 2a, b, e) (Didier and Barbarin 1991; Barbarin and

Fig. 3 Microphotographs showing textures of enclaves: **a** the boundary between granite and enclave; inside enclave visible large plagioclase xenocryst; **b** hybrid zone between granite and enclave; **c** automorphic crystal of hornblende inside enclave; **d** K-feldspar xenocryst

inside enclave; **e** large, zoned plagioclase xenocryst with biotite inclusion within enclave; **f** needle-like apatite morphology. All photographs taken under crossed nicols. *Ap* apatite, *Bt* biotite, *Hbl* hornblende, *Kfs* K-feldspar, *Plg* plagioclase, *Qtz* quartz

Fig. 4 CL images of: **a** miarole filled by quartz aggregates; **b** zoned plagioclase xenocryst within enclave showing different luminescence than matrix plagioclase; **c** plagioclase crystals from the contact between granite and enclave showing different luminescence in the core and identical in the rim (indicated by the yellow arrow); **d** zoned plagioclase xenocryst within enclave showing different luminescence than matrix plagioclase; **e** zoned K-feldspar xenocrysts; **f** two morphology types of apatite: prismatic and zoned, forming inclusions in plagioclase xenocrysts and acicular one. *Ap* apatite, *Bt* biotite, *Kfs* K-feldspar, *Plg* plagioclase, *Qtz* quartz

Didier 1992; Wiebe and Collins 1998). Sharp contacts are frequently observed for the smallest, darkest enclaves, which exhibit finer grained textures (Fig. 2e). However, even the contacts that appear sharp in hand specimens show interlocking textures under the microscope (Figs. 3a, 4b, c). Large feldspar phenocrysts are occasionally observed to cross the contacts between MMEs and HBG (Fig. 2f), in addition to occurring inside enclaves (Figs. 2a, f, 3a, d, e, 4b, d, e, f). MMEs are commonly surrounded by narrow felsic halos of the adjacent granite that are enriched in quartz and K-feldspar; these halos represent the so-called “bleaching” zones of Vernon (1991) (Fig. 2b). Some MMEs, particularly larger ones, are surrounded by a few centimetre-thick hybrid zones with diffuse contacts directed towards both the enclave’s core and the granitic host (Figs. 2a, c, 3b). In the hybrid zone (Fig. 2c), the grain sizes of the mineral constituents and the modal compositions are transitional between those of the MME and HBG. In comparison to the enclave-free granite, the hybrid zone contains higher contents of biotite and plagioclase and lower contents of

K-feldspar and quartz. This zone, however, often contains dispersed feldspar phenocrysts. The hybrid zone also frequently shows small mafic clots detached from the cores of the enclaves and thin schlieren-like mafic bands alternating with felsic bands (Fig. 2c). Sub-angular felsic patches with granitic textures are sometimes wrapped in thin bands of biotite. These patches most likely became incorporated into the hybrid zone from the surrounding granite as semi-rigid fragments. “Composite” enclaves are texturally distinct and consist of a mafic enclave enclosed in a hybrid zone that exhibits sharp contacts with the granite and is surrounded by a “bleaching” zone (Figs. 2b, 3b). Rare examples of MMEs included in granitic pegmatite are also observed (Fig. 2d). Generally, MMEs consist of 40–65% plagioclase, 25–30% biotite and 0–3% amphibole. Quartz and K-feldspar are subordinate, although their contents sporadically reach up to 5%. The accessory minerals include zircon (5–100 μm long), needle-like apatite (0.1–0.5 mm in length), allanite (2–50 μm), monazite (5–25 μm in diameter), and opaque phases.

Results

Mineral growth textures

The mineral growth textures of the HBG do not differ from the descriptions published by Majerowicz (1972) and Kural and Morawski (1968). Quartz in the HBG forms interstitial grains or granular aggregates with undulose to mosaic extinction. Generally anhedral, quartz locally becomes euhedral against K-feldspar. Graphic intergrowths with K-feldspar are sometimes observed. Plagioclase in the HBG occurs as subhedral to almost euhedral prisms, up to 1.5 cm long, but phenocrysts replaced by chessboard albite reach up to 3 cm. Typically, the plagioclase is a normal-zoned andesine-oligoclase with luminescence changing from green and olive-green to dull ochre from the core towards the margin. Oscillatory and patchy zoning is less common (Fig. 4d). Larger crystals show the presence of partly resorbed and altered andesine cores that are easily discernible by their bright green to bluish luminescence (Fig. 4b, c). Anhedral albitic rims with dark blue luminescence occur around plagioclase in contact with K-feldspar.

K-feldspar in the HBG is a perthitic microcline (Majerowicz 1972 and references therein). In the porphyritic facies, well-shaped phenocrysts up to 2–3 cm in size coexist with K-feldspar in the rock groundmass. K-feldspar phenocrysts usually display oscillatory zoning emphasized by the rhythmically changing intensity of blue luminescence (Fig. 4e). The boundaries between the bands are often delineated by inclusions of plagioclase and biotite, whose longer axes are approximately parallel to the former euhedral growth faces of K-feldspar. Groundmass K-feldspar is interstitial and unzoned.

Biotite in the HBG forms separate subhedral to nearly euhedral plates, usually 2–4 mm in diameter, or aggregates of randomly oriented smaller crystals. Biotite displays pale straw yellow to dark brown pleochroism. Amphibole with yellowish-green to green pleochroic colours is a minor constituent of the HBG. Subhedral to nearly euhedral prisms occur separately or in association with biotite. Biotitization and/or chloritization are common alteration features, particularly along cleavage planes, fractures and margins.

Euhedral prismatic apatite forms inclusions in biotite and amphibole. Less commonly, larger and sturdier crystals, sometimes slightly zoned as shown by varying intensity of yellow luminescence, are enclosed in feldspars and quartz. Allanite occurs separately or as inclusions in biotite and amphibole. It is euhedral and prismatic and displays complex growth zoning in scanning electron microscope SEM images, with mosaic cores overgrown by oscillatory, fine-banded margins. Zircon is mostly present as inclusions in biotite and amphibole and forms euhedral prismatic crystals

with typical igneous zoning visible in SEM images. Ilmenite is found as inclusions in biotite. Chloritized biotite and amphibole also contain secondary titanite and epidote.

Quartz in MMEs occurs as anhedral, interstitial or poikilitic grains. Rare quartz ocelli and oval to ellipsoid 5-mm quartz aggregates with euhedral plagioclase crystals projecting from the margins inwards are observed in some MMEs. The latter structures have been interpreted as voids left after the degassing of mafic magma and filled with quartz from granitic melt (e.g., Vernon 1991, 2010). The occurrence of miaroles in MMEs is evidence of the magmatic origin of the enclaves (Fig. 4a) (Vernon 1991, 2010).

Plagioclase from the MME matrix forms euhedral to subhedral laths or strongly elongated tabular crystals that are up to 1 mm in size. This plagioclase is usually normally zoned labradorite-oligoclase with very rare albitic rims. Reverse or oscillatory zoning is less common. Typically, plagioclase crystals from MMEs display blue or blue-violet bright cathodoluminescence (Fig. 4b–e). Calcic cores are strongly altered to a mixture of secondary Ca-bearing minerals. Sometimes, secondary calcite fills the microfractures and interstices adjacent to plagioclase. Some crystals (up to 2 cm) that occur close to the enclave/granite contacts have margins with olive-green to dull ochre CL, similar to the luminescence that is typical for the plagioclase in the granite (Fig. 4b–d). Therefore, these crystals are interpreted as xenocrysts that were mechanically introduced into the MMEs from the granitic host (Vernon et al. 1988; Vernon 2010). The plagioclase grains display oscillatory zoning with patchily zoned cores (Fig. 4d). Another plagioclase grain has a strongly altered core (Fig. 4b, c).

K-feldspar appears only in MMEs that contain higher than average amounts of quartz and lower contents of biotite. It represents interstitial, rarely poikilitic perthitic microcline with homogeneous blue CL. K-feldspar most likely crystallized from the late-stage hydrous granitic melts or orthomagmatic fluids that percolated from the surrounding granite. Some MMEs contain phenocrysts of K-feldspar. These phenocrysts are commonly mantled by a thin band of biotite and are identical in size, texture, luminescence properties (oscillatory zoning) and composition to those in the surrounding granite (Fig. 4e). Therefore, they are also regarded as xenocrysts (Vernon et al. 1988; Vernon 2010).

Both the plagioclase and K-feldspar xenocrysts that appear inside MMEs have sub-rounded to rounded morphologies resulting from the partial fusion of the crystal rim in higher temperature mafic magma globules (Fig. 4b, d, e).

The biotite grains in MMEs are similar to the biotite grains in the surrounding granite. In most cases, biotite forms subhedral ragged platy crystals that are up to 0.7 mm long and very uniform in size. In smaller MMEs and in the chilled margins of larger enclaves, biotite is lath-shaped. Crystals with sizes similar to those of the biotite in the

granitic host (5–6 mm) are uncommon. In contrast to the biotite from the HBG, the biotite in MMEs is poor in mineral inclusions.

Minor amphibole is usually euhedral and occurs as short prismatic crystals that are 0.5–1 mm long and often intergrown with primary biotite. Scarce skeletal amphibole crystals occur as inclusions in quartz (Fig. 3c). Biotitization and/or chloritization are common alterations.

Compared to the HBG, MMEs are relatively poor in allanite and abundant in acicular apatite (Figs. 3f, 4f). The acicular morphology of apatite is also an indication of magma mixing-mingling processes. The rapid growth of apatite in a quenched mixing system results in acicular crystal shapes, rather than stubby prismatic apatite (Wyllie et al. 1962; Hibbard 1991). Prismatic, zoned apatites are rare and occur only as inclusions in the rims of feldspar xenocrysts (Fig. 4f). This type of apatite may represent the product of an earlier crystallization episode (within the granitoid magma).

Zircon occurs as short prismatic grains typically as inclusions in biotite and amphibole, although more rarely it is also present in feldspars. The chloritization of biotite and amphibole is associated with the appearance of secondary epidote, titanite and ilmenite.

The above-described textures and microtextures may indicate that the MMEs from the HBG represent globules of more mafic magma (or hybrid magma) that mingled with a more felsic host melt.

Mineral compositions

Feldspars

The compositions of plagioclase are given in Table 2 (supplementary material) and Fig. 5. The plagioclase grains from the HBG display a wide range of compositions from andesine (max An_{46}) to albite, and their Or contents reach up to 5.3%. In MMEs, the groundmass plagioclase cores are more calcic, with An contents of 45.3–65.7%, and albitic compositions are rare and less common than in the HBG. The Or content reaches up to 8.1% (Fig. 5).

The K-feldspar phenocrysts from the HBG and associated MMEs show similar compositions (Table 3, supplementary material). In the cryptoperthitic domains, the Ab content ranges from 23.8 to 33.3%, while in the remaining regions of the crystals, where the primary distribution of Na and K has been reset by subsolidus alteration to various extents, the Ab content varies between 1.7 and 17.6% (Fig. 6). The compositional gap in the Ab content between ca. 18 and 24% may reflect the difference in temperature regime during the processes of crystallization and subsolidus alteration of alkali feldspar. Groundmass K-feldspar grains from the HBG and MMEs display very restricted Ab contents of 2–5.5%; these values correspond to the compositions of the most Na-poor

domains in the phenocrysts. The concentration of Ba in the groundmass K-feldspar reaches up to 0.62 wt.% BaO and is generally lower than that in the phenocrysts (0–1.86 wt.%).

Biotite

Generally, the composition of HBG biotite is Fe-biotite according to the nomenclature of Foster (1960); this biotite exhibits low and rather restricted $Mg/(Mg + Fe_{tot})$ ratios of 0.16–0.26 and thus corresponds to annite (Table 4, supplementary material).

The compositions of the biotite from MMEs are consistent with the compositions of the biotite from the HBG, which suggests their advanced chemical equilibration (Table 4, supplementary material) (Fig. 7).

Amphibole

Amphibole is a rare phase in both granite and enclaves. Based on the classification of Hawthorne et al. (2012), the amphibole from the HBG and associated MMEs is hastingsite (Fig. 8) with $0.91 \leq Ca/(Ca + Na) \leq 0.94$. The hastingsite from the HBG records 24.84–28.88 wt.% FeO_{tot} , 7.77–9.15 wt.% Al_2O_3 and 3.76–5.91 wt.% MgO. The $Mg/(Mg + Fe^{2+})$ values range from 0.23 to 0.36, and the calculated Fe^{3+}/Fe^{2+} values are 0.20–0.38 (Fig. 8; Table 5, supplementary material). The contents of TiO_2 and MnO do not exceed 2.26 and 1.20 wt.%, respectively.

The hastingsite from the MMEs has 26.94–29.32 wt.% FeO_{tot} and 7.38–8.56 wt.% Al_2O_3 , values similar to those observed in the amphibole from the HBG. However, the hastingsite from MMEs is slightly poorer in Mg [$MgO = 3.14–4.33$ wt.%; $Mg/(Mg + Fe^{2+}) = 0.19–0.25$] and records a lower degree of iron oxidation, with the calculated Fe^{3+}/Fe^{2+} values ranging from 0.15 to 0.23. The contents of TiO_2 and MnO are up to 2.07 and 1.17 wt.%, respectively.

Accessory minerals

The main accessory minerals in the HBG are apatite, allanite and zircon. Apatite has a nearly stoichiometric composition, with MnO contents ranging from below the detection limit to 1.06 wt.% (Table 6, supplementary material). The REE contents are below the detection limit of EPMA. Allanite is usually metamict and strongly altered. Relatively unaltered crystals display analytical totals between 99.49 and 100.34 wt.% and Si cation contents close to 3 apfu. These compositions correspond to allanite-(Ce), with cerium considerably dominant relative to the other REEs ($Ce/\Sigma REE = 0.57–0.58$) (Table 7, supplementary material). Allanite has low concentrations of MnO and TiO_2 (≤ 0.11 and 1.79–2.44 wt.%, respectively) and highly variable calculated Fe^{3+}/Fe^{2+} ratios ranging from 0.85 to 1.73. Zircon shows oscillatory zoning,

Table 2 Major and trace element analyses of MMEs

Sample	016	025	026	031	040	042	046	047	048	049	050	056
Locality	1	1	1	2	3	3	4	4	4	4	4	5
wt. %												
SiO ₂	60.32	61.28	53.30	66.26	63.46	60.44	53.09	62.54	63.99	62.98	57.81	52.75
TiO ₂	1.15	0.79	1.10	0.78	0.87	1.12	1.52	0.83	0.75	0.77	1.14	1.36
Al ₂ O ₃	15.21	15.33	16.23	15.05	16.61	15.01	16.39	15.31	14.71	14.75	15.78	17.37
Fe ₂ O ₃ <i>t</i>	9.24	9.14	13.71	6.10	5.78	10.28	15.74	8.09	8.17	8.70	11.25	13.55
MnO	0.22	0.16	0.31	0.12	0.12	0.18	0.17	0.16	0.18	0.21	0.23	0.27
MgO	1.06	0.90	1.67	0.98	1.42	1.27	1.63	0.90	1.16	1.16	1.19	2.02
CaO	3.01	2.17	3.58	2.89	3.74	3.17	2.44	2.60	2.85	2.34	2.77	3.26
Na ₂ O	4.90	5.32	4.84	4.40	4.19	4.18	4.87	4.92	3.96	4.22	5.26	3.92
K ₂ O	1.53	2.19	2.85	2.18	2.43	2.66	0.99	2.14	2.81	2.41	2.31	4.10
P ₂ O ₅	0.41	0.27	0.38	0.26	0.28	0.44	0.65	0.26	0.24	0.23	0.36	0.49
LOI	1.70	1.34	1.30	0.66	0.43	0.42	2.44	1.36	0.82	1.34	1.52	0.67
Total	98.90	99.05	99.46	99.82	99.47	99.34	99.93	99.28	99.78	99.27	99.81	100.00
A/CNK	1.00	1.02	0.92	1.01	1.02	0.97	1.21	1.01	1.00	1.07	0.97	1.03
Mg no.	0.19	0.16	0.19	0.24	0.33	0.20	0.17	0.18	0.22	0.21	0.17	0.23
ppm												
Ba	440.0	167.9	361.0	371.7	278.4	370.9	185.8	314.5	457.0	425.5	215.1	382.0
Cs	13.6	14.3	13.7	7.8	11.2	8.8	1.2	8.2	7.0	5.3	14.7	37.3
Rb	172.4	254.3	310.1	184.4	217.7	264.9	84.7	252.8	212.6	182.4	251.7	545.9
Sr	95.1	58.0	83.6	112.2	201.1	103.6	82.7	84.5	118.6	121.4	74.3	73.2
V	80.0	43.0	77.0	62.0	76.0	92.0	83.0	55.0	66.0	63.0	75.0	101.0
Co	8.4	6.7	8.9	7.6	9.9	11.7	12.0	8.2	8.0	8.2	9.3	16.2
Zr	260.0	481.2	340.9	349.3	380.1	343.5	560.9	436.3	273.8	289.5	466.4	374.0
Hf	9.5	14.9	10.6	8.6	9.6	8.8	18.2	13.7	8.7	9.7	14.0	9.8
Y	125.7	98.3	89.5	52.5	41.6	69.8	132.3	92.4	73.2	95.2	122.4	86.3
Nb	90.2	90.3	89.8	33.9	33.9	67.0	151.6	74.3	45.7	53.5	91.9	75.3
Ta	6.9	9.5	9.2	2.3	2.1	3.0	13.4	7.3	4.0	5.1	6.8	5.4
U	39.4	22.0	21.5	17.9	3.3	3.4	29.1	20.3	7.2	8.8	29.0	7.9
Th	12.1	23.4	17.0	14.6	15.4	9.7	45.2	21.2	11.6	12.1	15.8	17.2
La	27.4	18.1	23.1	55.1	64.4	44.1	28.5	55.9	35.8	34.0	26.5	30.5
Ce	71.0	36.6	54.4	101.8	114.7	95.7	57.5	112.1	80.4	74.9	62.9	67.1
Pr	10.8	5.2	7.6	11.6	13.5	11.4	7.8	12.2	10.6	9.9	9.0	9.0
Nd	50.8	24.6	33.5	45.2	49.8	50.1	34.5	50.1	47.1	44.1	43.3	41.2
Sm	15.5	8.8	10.7	9.2	9.0	11.5	12.1	11.4	11.7	11.6	13.7	11.2
Eu	1.1	0.7	0.8	1.0	1.5	1.0	0.7	0.9	1.1	1.2	0.8	0.8
Gd	18.8	11.5	12.8	8.9	7.5	12.1	16.5	11.5	12.5	13.4	16.9	12.7
Tb	3.4	2.3	2.2	1.5	1.3	2.1	3.0	2.2	2.2	2.5	3.0	2.4
Dy	21.9	13.8	14.1	8.6	6.5	11.2	19.0	13.1	13.7	15.0	19.0	14.5
Ho	4.7	2.8	3.1	1.7	1.3	2.3	3.7	2.8	2.6	3.0	3.6	2.5
Er	14.8	8.3	9.2	4.6	3.8	6.0	11.7	8.4	8.1	8.7	10.9	7.0
Tm	2.1	1.3	1.4	0.7	0.5	0.8	1.7	1.3	1.2	1.3	1.7	1.1
Yb	14.8	9.5	9.1	4.8	3.5	5.5	13.7	9.0	7.9	9.7	12.2	7.0
Lu	2.2	1.3	1.3	0.7	0.5	0.7	1.8	1.3	1.1	1.4	1.7	0.9
(La/Yb) _N	1.2	1.3	1.7	7.7	12.2	5.4	1.4	4.1	3.0	2.4	1.4	2.9
Eu/Eu*	0.2	0.2	0.2	0.3	0.6	0.3	0.2	0.2	0.3	0.3	0.2	0.2

Table 2 (continued)

Sample	061	083	092	093	097	100	103	104	105	108	117	144	145
Locality	6	8	9	9	10	10	4	4	4	4	3	4	4
wt. %													
SiO ₂	58.64	55.46	50.62	61.92	57.40	64.78	68.95	51.63	59.77	64.58	63.30	69.23	55.08
TiO ₂	1.16	1.23	1.44	0.87	1.01	0.79	0.47	1.30	0.94	0.79	0.92	0.41	0.97
Al ₂ O ₃	15.19	15.99	16.38	15.68	16.91	14.60	14.84	15.24	15.42	14.61	14.97	14.42	17.19
Fe ₂ O ₃ <i>t</i>	11.90	12.79	16.33	8.18	9.12	7.00	4.28	17.47	9.33	7.44	7.98	3.78	10.33
MnO	0.22	0.29	0.47	0.22	0.21	0.15	0.08	0.42	0.18	0.13	0.14	0.06	0.19
MgO	1.27	1.31	1.55	0.86	1.22	0.90	0.52	2.29	1.13	1.02	1.18	0.46	1.27
CaO	2.94	2.74	2.57	2.96	3.65	2.91	2.15	4.42	2.45	2.97	1.98	2.04	2.55
Na ₂ O	4.13	4.53	4.28	4.33	5.40	4.12	4.00	3.89	4.74	4.36	4.36	4.07	5.91
K ₂ O	3.21	2.94	3.40	2.63	2.16	3.02	3.84	3.30	2.93	2.16	1.41	3.25	2.06
P ₂ O ₅	0.47	0.44	0.62	0.27	0.41	0.22	0.11	0.39	0.37	0.29	0.31	0.09	0.48
LOI	0.57	1.67	1.86	0.88	1.86	0.40	0.44	0.58	1.41	0.55	2.34	1.70	3.60
Total	99.90	99.61	99.75	98.96	99.50	99.05	99.83	99.90	98.88	99.06	99.02	99.51	99.63
A/CNK	0.97	1.02	1.06	1.02	0.95	0.95	1.01	0.85	1.00	0.98	1.22	1.04	1.04
Mg no.	0.17	0.17	0.16	0.17	0.21	0.20	0.19	0.21	0.19	0.21	0.23	0.19	0.20
ppm													
Ba	449.7	289.7	409.9	335.0	177.5	383.8	587.8	368.7	354.0	279.0	351.5	356.2	361.9
Cs	18.9	12.0	34.0	18.9	11.8	8.2	5.8	32.2	10.0	6.1	1.6	6.1	14.9
Rb	367.7	354.6	488.5	260.8	222.2	207.8	237.1	451.5	334.9	168.6	88.2	187.8	239.1
Sr	89.4	70.2	74.5	121.4	114.3	121.6	93.5	68.7	101.4	115.5	124.0	95.1	89.4
V	55.0	68.0	63.0	52.0	66.0	54.0	22.0	112.0	49.0	59.0	74.0	19.0	62.0
Co	9.0	10.3	12.2	7.1	9.8	8.5	4.5	15.3	8.9	7.1	9.0	3.7	8.7
Zr	372.9	522.1	437.1	311.3	319.8	395.0	387.1	309.7	512.7	350.9	398.3	323.0	535.4
Hf	9.9	18.3	13.9	9.9	10.3	10.1	11.0	10.4	15.6	9.4	10.6	9.4	16.3
Y	77.2	105.0	102.9	59.2	149.6	67.6	46.4	198.0	110.1	50.4	74.2	45.6	97.4
Nb	72.7	111.3	114.8	65.3	81.2	49.6	30.8	92.7	86.2	39.9	55.4	27.5	77.7
Ta	7.9	13.6	12.4	4.7	6.6	2.8	1.9	8.4	7.8	2.3	5.5	2.5	7.0
U	5.5	16.5	25.6	12.5	13.1	5.6	5.6	7.2	47.0	7.4	4.6	13.3	21.9
Th	11.6	20.5	18.9	11.8	14.9	16.2	20.4	10.1	19.1	10.0	17.9	18.3	21.2
La	42.5	41.0	39.3	38.1	29.8	64.2	63.6	34.7	23.0	28.5	39.7	56.1	20.0
Ce	100.5	89.2	78.7	91.5	73.3	127.7	121.6	100.4	49.5	64.7	73.9	106.4	45.8
Pr	12.2	11.3	9.8	11.4	10.5	14.4	12.8	16.2	6.8	8.3	9.3	12.0	6.4
Nd	49.7	47.7	41.3	45.6	53.4	56.4	45.3	82.6	33.1	40.4	40.2	46.1	32.1
Sm	11.2	11.9	10.6	12.1	16.8	12.8	8.0	23.4	9.7	9.3	9.8	8.6	10.4
Eu	0.8	0.7	0.5	1.2	1.6	1.3	0.9	1.2	0.8	1.1	0.6	1.0	0.9
Gd	11.0	14.1	13.1	11.7	20.2	12.2	7.5	28.1	13.4	10.0	11.5	6.9	13.1
Tb	2.0	2.5	2.3	1.9	3.9	2.1	1.3	5.3	2.5	1.5	2.1	1.3	2.6
Dy	11.9	15.5	15.0	10.8	23.8	10.7	7.6	32.8	15.0	8.8	12.9	7.1	15.3
Ho	2.1	2.8	2.8	2.3	5.0	2.2	1.3	5.9	3.1	1.8	2.3	1.5	3.2
Er	6.1	8.2	7.9	6.6	13.9	5.9	3.8	16.6	9.4	5.5	6.2	4.2	9.2
Tm	0.9	1.3	1.2	0.9	2.0	0.9	0.6	2.6	1.6	0.8	0.9	0.7	1.5
Yb	6.2	9.3	8.2	6.0	14.0	5.8	4.0	17.2	10.5	4.8	6.1	4.4	9.6
Lu	0.9	1.4	1.2	0.9	1.8	0.8	0.6	2.4	1.6	0.7	0.8	0.6	1.4
(La/Yb) _N	4.6	2.9	3.2	4.2	1.4	7.5	10.6	1.3	1.5	4.0	4.4	8.5	1.4
Eu/Eu*	0.2	0.2	0.1	0.3	0.3	0.3	0.4	0.1	0.2	0.4	0.2	0.4	0.2

LOI loss on ignition, A/CNK Al₂O₃/(CaO + NaO + K₂O) molar; Mg no. atomic Mg/(Mg + Fe), Eu/Eu* [Eu_N/√(Sm_N × Gd_N)], N chondrite normalized to values of Nakamura (1974)

Locality: 1—Strzegom, 2—Goczałków, 3—Wieśnica, 4—Kostrza, 5—Gniewków, 6—Czernica, 7—Zimnik, 8—Rogoźnica, 9—Paszowice, 10—Borów

Fig. 5 Feldspars chemistry of HBG and MMEs. Plot of the EMPA data in the feldspar classification diagram

reflecting heterogeneous distribution of trace elements during magmatic growth (as revealed by CL and SEM analyses). This zoning pattern is most likely related to differences in the contents of elements below the detection levels of EPMA, as it shows no correlation to variations in the elements detected. The Zr/Hf ratios in the zircon vary between 80.28 and 118.09 (Table 8, supplementary material). Detailed chemical characteristics of zircon are presented by Turniak et al. (2014).

Whole-rock geochemistry

The HBG is weakly peraluminous, with molecular A/CNK ratios ranging from 0.99 to 1.06 (Fig. 9a; Table 1). The Mg number is low; the atomic Mg/(Mg + Fe) values range from 0.15 to 0.30 (Table 1). On the (Na₂O + K₂O) vs SiO₂ diagram (Fig. 9c), the HBG samples plot in the upper part of the sub-alkaline field. The rock is K₂O-rich (4.04–5.37 wt.%) with a low Na₂O/K₂O ratio (less than 1) and plots in the high-K field on the K₂O vs SiO₂ diagram (Fig. 9b). The HBG mainly exhibits calc-alkalic characteristics (Fig. 9d).

MMEs are characterized by a broad compositional range, with SiO₂ contents varying from 51 to 71 wt.% (Table 2). The Mg numbers are low and similar to those of the host granite (0.16–0.33). The Na₂O/K₂O ratios range between 0.96 and 4.92 with an average value of 1.92. Enclaves are peraluminous and metaluminous, with A/CNK ratios between 0.85 and 1.22 (Fig. 9a).

On the Harker diagrams, no visible chemical gap between the MMEs and the HBG is evident (Fig. 10). Both

rock types show well-defined trends for most of their major and trace elements. However, while most of the major elements (e.g., Ti, Al, Fe, Mg and Ca) exhibit a common line of melt evolution over a wide range of silica compositions in the range of approximately 51–75 wt.% (selected diagram shown in Fig. 10), the trends for elements such as Na, K, Ba, Rb, Zr, and La are different for the HBG and MMEs, with a marked bending at approximately 64 wt.% SiO₂. Compared to the HBG samples, MMEs are depleted in Si, K, Ba, and La and enriched in Ti, Fe, Mg, Ca, Nb, Zr, V, Y and Yb.

The chondrite-normalized (Nakamura 1974) rare-earth element patterns for the HBG and MMEs are shown in Fig. 11. The REE patterns for the HBG are homogeneous, showing moderately steep patterns for LREEs and nearly flat patterns for HREEs, with (La/Yb)_N = 6.4–19.3 (Table 1). Thus, the granite is enriched in LREEs (58–170 times more than chondrite) with respect to HREEs (11–22 times more than chondrite). As a result, the described rocks exhibit a well-defined fractionation of LREEs with (La/Sm)_N = 3.4–5.6 and a weak fractionation of HREEs with (Gd/Yb)_N = 1.1–2.5. All the samples display significant negative Eu anomalies, i.e., Eu/Eu* = 0.3–0.6.

The characteristic features of MMEs are gentle slopes of the patterns for LREEs and nearly flat slopes for HREEs with (La/Yb)_N = 1.2–12.2 (Table 2). MMEs are slightly more fractionated in LREEs relative to HREEs with (La/Sm)_N = 0.9–4.9 and (Gd/Yb)_N = 1.0–1.8. All enclaves show negative Eu anomalies, with Eu/Eu* = 0.1–0.6. Compared to the HBG, MMEs are characterized by lower concentrations of LREEs and higher concentrations of HREEs. The enrichment of HREEs in MMEs can be explained by the preferential partitioning of REEs between mafic and felsic melts (Watson 1976; Ryerson and Hess 1978).

The trace element signatures of the HBG and MMEs are presented on primitive mantle-normalized (Sun and McDonough 1989) multi-element diagrams (Fig. 12). Both the HBG and MMEs are enriched in large-ion lithophile elements (LILEs) and depleted in high field strength elements (HFSEs) compared to the primitive mantle. The HBG shows relatively homogeneous patterns with negative anomalies for Ba, Nb, Sr, P, Eu and Ti. These negative anomalies may reflect the fractionation of plagioclase, apatite, and Fe–Ti oxides during the earlier stages of magma evolution.

The trace element patterns of MMEs are much more complex than those of the HBG (Fig. 12). Most of the enclaves exhibit negative anomalies for Ba, K, Nb, La, Sr and P, although some samples show no Nb, La or P anomalies.

Table 3 Whole-rock Sr and Nd isotopic data of the HBG and MMEs

Sample	Locality	Rb ppm	Sr ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	2σ (%)	$^{87}\text{Sr}/^{86}\text{Sr}$ (298 Ma)	Sm ppm	Nd ppm	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	2σ (%)	$^{143}\text{Nd}/^{144}\text{Nd}$ (298 Ma)	ϵ Nd (298 Ma)	T_{CHUR} (Ma)	T_{DM} (Ma)	ϵ (CHUR)
HBG																	
009	1	174	77	6.567	0.73684	19	0.709440	8.4	46.6	–	–	–	–	–	–	–	–
010	1	205	86	6.919	0.73610	21	0.707235	8.0	45.2	–	–	–	–	–	–	–	–
013	1	211	91	6.742	0.73540	25	0.707276	7.0	39.5	–	–	–	–	–	–	–	–
028	1	186	70	7.681	0.73911	21	0.707064	6.5	31.9	0.1232	0.512310	15	0.512070	–3.6	681	1230	5.19
039	3	138	128	3.118	0.72164	12	0.708633	7.1	36.3	0.1182	0.512249	7	0.512018	–4.6	757	1262	5.11
058	5	208	103	5.869	0.73367	22	0.709180	5.8	31.8	0.1103	0.512325	11	0.512110	–2.8	553	1057	5.61
060	6	190	82	6.710	0.73662	18	0.708624	7.3	38.8	0.1137	0.512285	14	0.512063	–3.7	649	1153	5.37
066	7	226	56	11.718	0.75643	15	0.707534	4.7	24.6	0.1155	0.512350	16	0.512125	–2.5	542	1075	5.57
067	7	199	84	6.878	0.73791	22	0.709215	10.3	67.8	0.0918	0.512311	12	0.512132	–2.4	476	918	5.96
080	8	192	84	6.655	0.73514	15	0.707370	4.9	24.8	0.1194	0.512209	20	0.511976	–5.4	847	1342	4.92
102	4	173	86	5.834	0.73268	30	0.708340	7.9	37.5	0.1274	0.512369	21	0.512121	–2.6	592	1186	5.29
MMEs																	
025	1	254	58	12.752	0.76113	24	0.707921	8.8	24.6	0.2163	0.512526	9	0.512104	–2.9	–	–	–
031	2	184	112	4.765	0.72805	11	0.708174	9.2	45.2	0.1230	0.512303	13	0.512063	–3.7	694	1239	5.17
046	4	85	83	2.967	0.72210	31	0.709714	12.1	34.5	0.2120	0.512437	17	0.512023	–4.5	–	–	–
049	4	182	121	4.356	0.72775	21	0.709577	11.6	44.1	0.1590	0.512377	12	0.512067	–3.7	1057	1842	3.82
050	4	252	74	9.841	0.74917	15	0.708107	13.7	43.3	0.1913	0.512419	15	0.512046	–4.1	–	–	–
083	8	355	70	14.706	0.77146	21	0.710097	11.9	47.7	0.1508	0.512338	11	0.512044	–4.1	997	1695	4.13
092	9	489	75	19.125	0.79036	20	0.710566	41.3	10.6	–	–	–	–	–	–	–	–
100	10	208	122	4.955	0.73049	12	0.709817	12.8	56.4	0.1372	0.512324	12	0.512056	–3.9	806	1424	4.74
103	4	237	94	7.360	0.73956	11	0.708856	8.0	45.3	0.1068	0.512274	9	0.512066	–3.7	619	1096	5.51
104	4	452	69	19.166	0.78900	13	0.709032	23.4	82.6	0.1713	0.512425	13	0.512091	–3.2	1277	2214	3.08
105	4	335	101	9.597	0.75138	17	0.711339	9.7	33.1	0.1772	0.512472	12	0.512126	–2.5	1297	2388	2.76

(1) Localities as in Table 1

(2) Rb, Sr, Sm and Nd concentration were obtained by ICP-MS

(3) The results of isotopic measurements for Sr and Nd reference materials are: NBS SRM 987 $^{87}\text{Sr}/^{86}\text{Sr} = 0.710252 \pm 18$ (2 σ); JNdi-1 $^{143}\text{Nd}/^{144}\text{Nd} = 0.512095 \pm 13$ (2 σ)(4) T_{DM} , T_{CHUR} ages is in Ga, calculated based on values of De Paolo (1981)

(5) (–) not determined

Fig. 6 Feldspars chemistry of HBG and MMEs. Ab vs BaO diagram shows the enrichment of BaO in K-feldspars

Sr–Nd isotope data

The Sr–Nd isotope compositions of the HBG and MMEs are given in Table 3 and plotted in Fig. 13. The SHRIMP U–Pb zircon age of 298 Ma (Turniak et al. 2014) of the HBG was used to calculate its initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and $\epsilon\text{Nd}(t)$ values. The isotopic ratios of the HBG and MMEs have narrow ranges. The HBG has initial $^{87}\text{Sr}/^{86}\text{Sr}$ values ranging from 0.7070 to 0.7094 (with a mean value of 0.7082), $^{143}\text{Nd}/^{144}\text{Nd}$ values ranging from 0.51198 to 0.51213 (with a mean value of 0.51208), and $\epsilon\text{Nd}(t)$ values ranging from -5.43 to -2.39 (with a mean value of -3.47). The initial Sr ratio obtained by Pin et al. (1989) is 0.7098. The ϵNd factor given by Pin et al. (1988)

Fig. 7 Biotite chemistry of HBG and MMEs. Plot of the EMPA data in Al [apfu] vs $\text{Mg}/(\text{Mg} + \text{Fe}_{\text{tot}})$ diagram

Fig. 8 Amphibole chemistry of HBG and MMEs. Plot of the EMPA data in the calcic amphibole classification diagram (Hawthorne et al. 2012). A^* ($\text{Na} + \text{K} + 2\text{Ca}$); C^* ($\text{Al} + \text{Fe}^{3+} + \text{Cr} + 2\text{Ti}$)

is -4.2 . MMEs have similar initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7079–0.7113; mean 0.7093), $^{143}\text{Nd}/^{144}\text{Nd}$ ratios

Table 4 Results of the major element-based geochemical modelling

	Parental magma	Differentiated magma	Plagioclase	Biotite	Amphibole	Magnetite	Ilmenite	Apatite	Cumulate composition
	Co	Cl							Cs
SiO_2	69.02	75.62	62.38	35.16	41.68	0.00	0.00	0.21	48.48
TiO_2	0.60	0.14	0.00	4.28	2.26	7.53	49.89	0.00	2.02
Al_2O_3	14.30	13.29	23.58	12.78	8.65	5.74	0.00	0.00	17.34
Fe_2O_3	5.39	1.49	0.00	32.98	27.70	82.52	45.68	0.75	17.48
MnO	0.10	0.04	0.00	0.53	0.30	0.00	4.43	1.10	0.24
MgO	0.63	0.19	0.00	4.84	5.89	4.21	0.00	0.00	1.87
CaO	1.91	1.05	5.25	0.00	10.41	0.00	0.00	54.98	4.66
Na_2O	3.87	3.53	8.62	0.22	1.90	0.00	0.00	0.22	5.12
K_2O	4.04	4.61	0.17	9.21	1.22	0.00	0.00	0.00	2.45
P_2O_5	0.15	0.04	0.00	0.00	0.00	0.00	0.00	42.75	0.35
% of mineral phase in the cumulate			56.09	23.92	12.20	5.55	1.43	0.89	
Degree of FC = 24.37%									
$R^2 = 0.007$									

Mineral composition from HBG were used for mass-balance modelling

Table 5 Concentrations of particular elements in the A, B magmas and the mixture M

	C_A	C_B	C_M	$C_A - C_B$	$C_M - C_B$
Sample	092	067	061		
Locality	9	7	6		
wt. %					
SiO ₂	51.83	69.02	59.15	-17.19	-9.87
TiO ₂	1.47	0.60	1.17	0.88	0.57
Al ₂ O ₃	16.77	14.30	15.32	2.48	1.03
Fe ₂ O ₃	16.72	5.39	12.00	11.33	6.62
MnO	0.48	0.10	0.22	0.38	0.12
MgO	1.59	0.63	1.28	0.96	0.65
CaO	2.63	1.91	2.97	0.72	1.06
Na ₂ O	4.38	3.87	4.17	0.52	0.30
K ₂ O	3.48	4.04	3.24	-0.56	-0.80
P ₂ O ₅	0.63	0.15	0.47	0.48	0.32
ppm					
Co	7.10	12.20	9.00	11.70	15.30
Cs	6.00	34.00	18.90	8.80	32.20
Ga	23.80	44.40	31.50	31.20	37.10
Hf	11.10	13.90	9.90	8.80	10.40
Nb	34.50	114.80	72.70	67.00	92.70
Rb	203.00	488.50	367.70	264.90	451.50
Sr	85.30	74.50	89.40	103.60	68.70
Ta	2.60	12.40	7.90	3.00	8.40
Th	33.80	18.90	11.60	9.70	10.10
U	6.30	25.60	5.50	3.40	7.20
V	22.00	63.00	55.00	92.00	112.00
W	1.10	7.50	0.80	1.40	1.30
Y	47.00	102.90	77.20	69.80	198.00
La	92.80	39.30	42.50	44.10	34.70
Ce	185.40	78.70	100.50	95.70	100.40
Pr	19.71	9.77	12.23	11.39	16.19
Nd	67.80	41.30	49.70	50.10	82.60
Sm	10.30	10.60	11.20	11.50	23.40
Eu	0.92	0.51	0.81	1.02	1.20
Gd	7.90	13.13	11.02	12.06	28.10
Tb	1.39	2.31	1.95	2.09	5.33
Dy	7.49	15.02	11.89	11.19	32.75
Ho	1.42	2.75	2.05	2.33	5.87
Er	3.95	7.89	6.11	6.02	16.60
Tm	0.62	1.21	0.91	0.83	2.55
Yb	4.26	8.21	6.20	5.47	17.24
Lu	0.65	1.19	0.87	0.74	2.41

Localities as in Table 1

(0.51202–0.51213; mean 0.51207) and $\epsilon\text{Nd}(t)$ values (–4.51 to –2.50; mean –3.62). There are no significant isotopic differences between the HBG and MME samples (Fig. 13).

The depleted mantle model ages [T_{DM} , depleted mantle model; DePaolo 1981; DePaolo and Wasserburg (1976)] are used to estimate the average crustal residence ages of the source materials. The T_{DM} and ϵNd values at 298 Ma are shown in Table 3 and Fig. 14. The model ages (T_{DM}) range from 0.9 to 1.33 Ga for the HBG and from 1.09 to 2.38 Ga for MMEs. The ϵNd values are relatively constant for the HBG and vary from 5.11 to 5.96 (with a mean value of 5.33). The MME samples have lower ϵNd values, ranging from 2.76 to 5.51 (with a mean value of 4.08).

Discussion

Mechanism of magma evolution: geochemical modelling

As mentioned above, not all elements show consistent linear trends for the HBG and MMEs on SiO₂-dependent variation diagrams within two individual populations, separately for granite and MMEs (Fig. 10). Some trends for trace elements break down at the transition between granite and enclave. Additionally, on the (Na₂O + K₂O)/CaO vs Al₂O₃ plot (Fig. 9), the HBG and MMEs clearly indicate different trends. This result implies that at least two different processes or two different paths of granite and enclave evolution occurred. Since this difference between trends may reflect different paths of geochemical evolution, the petrogenetic modelling of these two data suites was performed separately.

Differentiation of the granitic melt

As major elements behave similarly during fractional crystallization and partial melting, some trace elements were used to recognize the differentiation-causing process. The contents of incompatible elements should remain constant during fractional crystallization or slightly increase in the melt. Decreasing contents of compatible elements result in steep trends. In contrast, partial melting results in horizontal or sub-horizontal trends due to the opposite behaviours of compatible and incompatible elements (Cocherie 1986). The logarithmic diagrams of the most compatible elements (e.g., Zr and V) versus the most incompatible elements (e.g., Rb) show sub-vertical trends for the HBG (Fig. 15). This trend supports the view that the evolution of granite was mainly controlled by fractional crystallization. Additionally, decreasing contents of Fe₂O₃, MgO, CaO TiO₂, MnO, Sr, Eu and V with increasing SiO₂ (see Harker diagrams, Fig. 10) may reflect the fractionation of biotite, amphibole, plagioclase and Ti-bearing minerals (e.g., ilmenite and titanite). The decreasing P₂O₅ contents and negative P anomalies observed on the spidergrams (Figs. 10, 12) reflect apatite fractionation.

Fig. 9 Discrimination diagrams for HBG and MMEs: **a** $\text{Al}_2\text{O}_3/(\text{Na}_2\text{O} + \text{K}_2\text{O})$ vs $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$ molar diagram (after Maniar and Piccoli 1989); **b** K_2O vs SiO_2 diagram (after Rickwood 1989); **c** $(\text{Na}_2\text{O} + \text{K}_2\text{O})$ vs SiO_2 diagram (after Irvine and Baragar 1971); **d** $(\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO})$ vs SiO_2 diagram (after Frost et al. 2001)

For this reason, fractional crystallization was modelled to reconstruct the differentiation of the high-silica (~75% wt.% SiO_2) melts derived from those with lower silica contents (~69% wt.% SiO_2). All major elements are defined by straight, linear trends, indicating the constant chemical and modal compositions of the cumulate and a small degree of differentiation. The first stage of modelling involved mass balance calculations (Stormer and Nicholls 1978) based on major element contents to determine the modal and chemical compositions of the cumulate and the degree of fractional crystallization. These calculations were executed using the assumption that the magmatic system was closed. To model fractional crystallization, the most differentiated sample with the highest silica content was taken as the composition of the most differentiated melt, whereas the most silica-poor sample was taken as the initial melt composition. The input data used for modelling, as well as the modelling results, are shown in Table 4. The best fit ($R^2 = 0.007$) was obtained for a cumulate composed of 56.09% plagioclase, 23.92% biotite, 12.20% amphibole, 5.55% magnetite, 1.43% ilmenite and 0.89% apatite. The calculated degree of fractional crystallization was equal to ca. 24%.

The obtained results were used for trace element modelling based on the equations of Rayleigh (1896) and the mineral/liquid partition coefficients ($KD_{m/l}$) that are typical for silica-rich melts (Table 9, supplementary material). The contents of the accessory phases, which are responsible for the observed behaviour of REEs, were used during this modelling step. By comparing the concentrations of particular

trace elements in the rock with the most evolved composition (Cl_{real}) with the contents obtained from the modelling (Cl_{calc}), the major element modelling data were tested. The best fit was attained by adding zircon (0.05%), monazite (0.045%) and allanite (0.005%) to the cumulate. Although slight differences between the real and calculated Cl values were observed in the HREEs (Dy and Er) and HFSEs (e.g., Nb, Ti and Y), as well as the transition metals (Co and Ni), the match between the calculated lines (Cl_{calc}) and analytical data (Cl_{real}) is satisfactory (Fig. 16).

Differentiation of the mafic melt

The Harker diagrams for the enclaves and granite do not show consistent, linear trends, which precludes a common mechanism of mafic-felsic magma evolution, i.e., neither fractional crystallization nor mixing. In turn, the textures found in MMEs (e.g., feldspar xenocrysts with disequilibrium textures inside MMEs, quartz ocelli, miaroles filled with quartz, and acicular apatite; Figs. 3, 4) point at least to local interactions of mafic blobs with felsic melt. The MMEs hosted in the HBG are generally highly dispersed, and their relative sizes are too small to effectively change the composition of the granitic melt. On the other hand, the felsic melt strongly influenced the still plastic (not fully crystallized) mafic melt. We assumed that hybridization took place at the very beginning of the granitic magma evolution. The bends in the trends on Harker diagrams occur where the enclave data and granite populations come

Fig. 10 Selected variation of major and trace element compositions vs silica (wt.%) for whole-rock samples of HBG and MMEs

Fig. 11 The chondrite-normalized (Nakamura 1974) rare-earth element patterns for HBG and MMEs

Fig. 12 Primitive mantle-normalized (Sun and McDonough 1989) trace element patterns for HBG and MMEs

Fig. 13 $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ vs $\epsilon\text{Nd}_{(t)}$ diagram for the HBG and MMEs

into contact. MMEs form consistent trends over a wide range of silica contents (from 51 to 70%), and some dispersion is observed only within the samples with the lowest silica content. For this reason, we created and assessed

Fig. 14 The depleted mantle model ages for the HBG and MMEs

the magma mixing model assuming that the granitic melt in the earliest stage of evolution (with the lowest silica content) could modify the primary enclave composition.

Since the behaviours of major and trace elements during magma mixing are identical, joint geochemical modelling was performed for both element groups. In this modelling, the equation of the mass equilibrium law was used: $C_M = X_A C_A + (1 - X_A) C_B$, where $X_A = (1 - X_B)$ (Fourcade and Allegre 1981). The least differentiated granitoid sample (69.02 wt.% SiO_2) was taken as the C_B magma composition (Table 5). This sample was treated as a felsic contaminant of the mafic melt. The MME with the lowest silica content (51.83 wt.% SiO_2) was assumed to represent the C_A magma. As the mixing product of the two magmas, C_M was assumed to indicate the MME containing 59.15 wt.% SiO_2 .

On the $(C_M - C_B) - (C_A - C_B)$ diagram, all of the points representing major and trace elements lie on the straight line crossing the centre of the coordinate axes (0; 0) (Fig. 17). The equation of this line is $y = 0.61x - 2.74$, where the value of 0.61 (line gradient) represents the amount of the contaminant C_B . Thus, the proportions of the felsic and mafic melts in the mixture were ca. 61% and 39%, respectively.

The very low sum of squared residuals testifies in favour of the model ($R^2 = 0.9676$). The rock samples chosen for the modelling fit the best-defined trend (Fig. 17). However, not all MME samples used in the modelling yield such ideal results. In particular, the rocks with the lowest differentiation index tend to deviate from the established trend. This feature may suggest that the process could have been more complex than simple two end-member magma mixing and might have occurred at a different stage of intrusion formation.

Source characteristics and origins of magmas of the HBG and MMEs

Most petrologists consider that granitoids have three possible origins: a crust-derived origin, mantle-derived origin

Fig. 15 Log(compatible element) vs log(incompatible element) plot after Cocherie (1986) for the HBG. *PM* partial melting trend, *FC* fractional crystallization trend

Fig. 16 **a** The chondrite-normalized REE patterns (Nakamura 1974) and **b** primitive mantle-normalized trace element patterns (Sun and McDonald 1989) illustrating the results of fractional crystallization modelling (bold line). The partition coefficients used in model are given in Table 9, supplementary material

Fig. 17 Major and chosen trace elements mixing test (Fourcade and Allegre 1981) for MMEs

or mixed origin that involves both crust- and mantle-derived melts. To identify the origin of the magmas contributing to pluton formation, phase and geochemical indices are used. In the latter, in the case of the HBG, ambiguity is clearly visible. The discussion of this problem and the clarification of the ambiguity is very important because the majority of Variscan plutons in the Sudetic section of the European Variscides are of mixed mantle-crust origin (Finger et al. 1997; Žák et al. 2013). Furthermore, some magma chambers within the SSM are clearly assigned to magmas originated from two sources, the mantle and the crust.

Some mineral parageneses and rock textures of the HBG from the SSM suggest that the granite probably represents a K-rich calc-alkaline granitoid (KCG), which is a subtype of calc-alkaline granitoids in the classification proposed by Barbarin (1999). The dominant mafic mineral in the HBG is biotite. The K-feldspar porphyritic texture and the composition of plagioclase (An% ranging from 15 to 40) are consistent with the KCG characteristics. However, neither the mineral association nor rock textures ultimately determine the origin of magma. For the KCG, Barbarin (1999) assumed mantle and crustal magma sources; in contrast, Roberts and Clements (1993) argued that K-rich calc-alkaline granitoids do not necessarily need such sources but can originate solely from crustal magmas. The inference of the KCG type is consistent with the geochemical features of the HBG, such as its calc-alkalic and K_2O -rich characteristics, low Mg number (0.15–0.30) and weakly peraluminous composition (Fig. 10). The obtained isotopic compositions fit with the range proposed by Barbarin (1999). Nevertheless, several indicators contradict Barbarin's (1999) hypothesis of mantle-crust sources of magma and provide some evidence in favour of Roberts and Clements's (1993) hypothesis, i.e., for the origin of magma from only a crustal source.

Nb/Ta ratios can be used to indicate the source of magma. The average Nb/Ta ratios of crust-derived magmas are

Fig. 18 The $Al_2O_3 + FeOt + MgO + TiO_2$ vs $Al_2O_3 / (FeOt + MgO + TiO_2)$ diagram for HBG and MMEs discriminates between anatectic melts from metapelites, metagreywackes and amphibolites protoliths. Domains obtained from Patiño Douce (1999), summarized by Jung et al. (2009). For comparison diorite samples from the Strzelin Massif and the Žulová Pluton (data from Pietranik and Waight 2008; Laurent et al. 2014)

approximately 12, but the values vary from approximately 8 for the lower crust to 16 for the middle crust and 13 for the upper crust (Rudnick and Gao 2003). The Nb/Ta ratios of mantle-derived magmas are 14–17 (Green 1995; Palme and O'Neill 2014). The Nb/Ta ratios of the HBG are variable over a wide range of 5.6–15.6. Thus, these values are not indicative because their variability encompasses the ranges for both crust- and mantle-derived magmas. Thus, the sources of granitic magma remain unclear.

The $Al_2O_3 + FeOt + MgO + TiO_2$ vs $Al_2O_3 / (FeOt + MgO + TiO_2)$ diagram proposed by Patiño Douce (1999) can be used to discriminate between anatectic melts from metapelite, metagreywacke and amphibolite (metagneous) protoliths. The data points for the HBG plot in the field of partial melts of metagreywackes (Fig. 18).

The CaO/Na_2O ratio may also be used to distinguish between metapelitic (plagioclase poor) and metagreywacke (plagioclase rich) sources (Sylvester 1998). Pelite-derived granite melts tend to have lower CaO/Na_2O ratios (< 0.3) than greywacke-derived granite melts (> 0.3). The HBG records CaO/Na_2O ratios between 0.30 and 0.59 with a mean value of 0.39; these values are thus more closely related to a greywacke source. It is also possible that granites with high CaO/Na_2O ratios (such as the HBG) can be derived from a gneissic rather than a greywacke source (Miller 1985). A gneiss source for the HBG was advanced by Majerowicz (1972); however, Kennan et al. (1999) excluded a gneiss

protolith based on a review of published Rb–Sr isotope data for Variscan granites in this part of the Bohemian Massif. Given isotopic composition data obtained by Pin et al. (1989), Puziewicz (1990) suggested metapelite, greywacke or acidic volcanic rock as potential sources for the HBG and proposed that the source of this melt was located in the eastern prolongation of the Kaczawa Unit rock suite, which is composed of a Cambrian to Lower Carboniferous metavolcanic–metasedimentary succession (Baranowski et al. 1990; Kryza and Muszyński 1992; Kryza and Zalasiewicz 2008). The melting of metapelitic rocks, however, would produce magma with a strongly peraluminous signature (e.g., Miller 1985), while the HBG shows only a weakly peraluminous composition. The Nb/Ta ratios for metagreywacke protoliths are variable and depend on the composition of the lithoclasts. For example, Devonian greywackes from the Rhenohercynian zone have ratios close to 10 (Floyd et al. 1991). Additionally, the temperature of the HBG magma probably exceeded 850 °C (Puziewicz 1990; Turniak et al. 2014), which is consistent with the melting of a metagreywacke protolith and could produce a sufficient amount of melt to form a granitic pluton (Johannes and Holtz 1996).

The isotope composition of the HBG shows relatively high $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ ratios from 0.7070 to 0.7094 with a mean value of 0.7082 and low $\epsilon\text{Nd}(t)$ values (-5.43 to -2.39). These data are consistent with the isotopic investigation presented by Pin et al. (1989). The authors note the high isotopic variability of the Strzegom massif, including the HBG itself. They determined the following $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ ratios: 0.7098 for hornblende-biotite granite, 0.7082 for biotite granite, 0.7058 for biotite granodiorite and 0.705 for two-mica granite. This diversity has probably been inherited from protoliths of various compositions. Thus, the isotopic data may implicate mixing of magmas from various crustal sources but does not explicitly indicate a mixed crust and mantle source for the granite. Therefore, the role of the mantle in the process of melt generation could be confined to providing heat, causing/enhancing the melting of the protolith/protoliths. However, the influence of the enriched mantle could also be related to mantle-derived fluids supporting the melting processes (Martin and De Vito 2005; Martin 2006). The influence of such fluids on the HBG may be demonstrated by the abundance of NYF miarolitic pegmatites (Pieczka et al. 2015 and references therein). The NYF signature would indicate a long-term influence of mantle-derived fluids on granite evolution. Thus, although the direct participation of mantle melt cannot be unambiguously confirmed, the indirect participation of the mantle in the formation of HBG and MMEs seems to be documented by concentrations of some trace elements as well as the geochemical characteristics of the host pegmatites.

The lack of unambiguous indicators for the participation of mantle melt opens the discussion to the possibility

of forming the HBG with the participation of only crustal material. Assuming crustal sources for granitic magma, the magma was emplaced as homogeneous molten matter into the chamber and further differentiated by fractional crystallization. The geochemical variations in the low- to high- SiO_2 samples likely resulted from the fractionation of plagioclase, biotite, and amphibole (see geochemical modelling results, Table 4). On primitive mantle-normalized spidergrams, the HBG samples are characterized by pronounced negative Ba, Nb, Sr, P, Eu and Ti anomalies but are enriched in Rb, Th, K and La. These negative anomalies may reflect the fractionation of plagioclase, apatite, and Fe–Ti oxides during some stages of magma evolution (Fig. 12). Additionally, the SiO_2 content increases with decreasing MgO , Fe_2O_3 , CaO and TiO_2 and P_2O_5 contents (Fig. 10), further suggesting the fractionation of these mineral assemblages.

MMEs in the KCG-type rocks are considered evidence of magmas from a mantle source; thus, MMEs confirm the composite nature of a pluton (Barbarin 1999, 1991, 2005, 1991; Didier 1973; Castro et al. 1990; Vernon 1984, 1990, 2010; Frost and Mahood 1987; Didier and Hibbard 1991; Orsini et al. 1991; Wiebe et al. 1997; Gerdes and Wörner 2000; Slaby and Götze 2004; Slaby and Martin 2008; Pietranik and Koepke 2009, 2014). However, this situation is not the case for the HBG because the geochemical models show that the MMEs did not influence the evolution of the granitic melt to a significant extent. In addition, although some enclave margins do exhibit local hybridization, their chemical and isotopic compositions cannot be directly related to a mantle source. A question arises as to whether the enclaves are also entirely of crustal origin or are products of the interaction of melts derived from the crust and mantle. Since their geochemical indicators are ambiguous and do not directly point to the participation of the mantle in their formation, their origin may be exclusively crustal, as suggested by the Roberts and Clements hypothesis (1993).

The existence of spheroidal dark enclaves and their petrological characteristics suggest their apparent emplacement before solidification and thus limited mafic-felsic magma interactions (Vernon 1984; Castro et al. 1990; Orsini et al. 1991; Kumar and Rino 2006; Slaby and Martin 2008; Chen et al. 2009; Kocak et al. 2011; Pietranik and Koepke 2014; Zhang et al. 2016). The investigated MMEs show some textures, such as acicular apatite, plagioclase and K-feldspar xenocrysts with resorption surfaces, miaroles filled with quartz, ocellar quartz and mafic clots, that are compatible with magma mixing-mingling with the surrounding granite. However, the precondition for running the mixing/hybridization process is the compositional gradient between the two interacting substances, regardless of their sources of origin.

MMEs contain the same minerals as the HBG but in different proportions. Enclaves are mostly composed of plagioclase and biotite with minor interstitial K-feldspar and

quartz. Microprobe analyses and CL investigations indicate that the plagioclase laths forming the groundmass of MMEs are different from those in granite. In MMEs, the groundmass plagioclase cores are more calcic, and albitic compositions are less common than in the HBG. The cores also show different CL features. In contrast, the plagioclase phenocrysts occurring in MMEs display strong similarities to those in granite. Only the rims of the plagioclase phenocrysts incorporated in the MMEs show chemical similarities to the groundmass crystals, which may reflect equilibration between mafic and felsic melts. The K-feldspar phenocrysts from the HBG and associated MMEs show similar compositions and CL features. Therefore, we interpret the large crystals of plagioclase and K-feldspar inside the MMEs as xenocrysts that were probably mechanically introduced from the granitic host. Feldspars are very sensitive indicators of magmas interaction (Waight et al. 2000; Slaby and Goetze 2004; Slaby et al. 2007a, b, 2008; Pietranik and Waight 2008; Pietranik and Koepke 2014; Ubide et al. 2014; Michel et al. 2016). In this case, however, the interaction is highly limited and it cannot be precisely determined at which stage feldspars have been incorporated into the MMEs. The compositions of the biotites from the MMEs and HBG are nearly the same, which suggests that they achieved advanced chemical equilibration. Such equilibration is also reported by Puziewicz (1995).

The fact that the minerals forming the granitoid host and MMEs have similar but not identical compositions can be explained by chemical equilibration (Fernandez and Barbarin 1991). The efficiency of this process strongly depends on the size of the enclaves; small (a few centimetres) enclaves are quenched and display more mafic compositions, whereas larger ones (up to 1 m) represent hybrid rocks with intermediate compositions. At the same time, these features exclude their having restite characteristics. We conclude that these enclaves represent globules of more mafic, generally hybrid magma that comingled with a more felsic host magma (Vernon 1983; Sparks and Marshall 1986; Castro et al. 1991; Barbarin and Didier 1992; Hawkesworth and Kemp 2006; Feeley et al. 2008; Chen et al. 2009; Kocak et al. 2011; Liu et al. 2013). The surrounding host magma had significant impacts on the chemistry of the relatively small volumes of the more mafic magma that were introduced into the granitic chamber. However, this interpretation raises the following question: what was the timing of this recharge, and from what source/place was the additional magma delivered? As mentioned above, evidence of interaction between enclaves and granite indicates only contact of molten materials with different compositions but does not provide an answer to the question about the origin of these molten materials.

The broad compositional range of MMEs, with SiO_2 contents varying from 51–71 wt.% (Table 2), may indicate the magma mixing process, but was the mixing between crustal

and mantle-derived magmas or between crustal magmas of different compositions? The Mg numbers of the MMEs are low and similar to those of the HBG (0.16–0.33). The MMEs have peraluminous and metaluminous compositions (A/CNK ratios: 0.85–1.22) (Fig. 9a). The MMEs show low Ni contents (up to 13 ppm), which are far below the Ni contents expected for a primitive basaltic magma derived from a mantle source ($\text{Ni} = 250\text{--}300$ ppm, e.g., Wilson 1989) and thus exclude a mantle source or any significant participation of mantle magma in MME formation. The low Mg numbers (less than 0.7, e.g., Wilson 1989) also indicate that even the most primitive enclave (with 51% SiO_2) had to have undergone fractionation processes or crustal contamination.

The Th/Nb ratio is a very sensitive indicator of mafic magma-crust interactions (Barth et al. 2000; Pearce 2008). On the Nb/Yb vs Th/Yb diagram (Fig. 19), almost all MMEs plot close to the upper crust field (similar to the HBG) and outside the mantle array. Enclaves approaching this trend may indicate a weak influence of the mantle source, perhaps manifested only in the presence of heat/fluids. The textural evidence that can support a fluid-rich environment may be dark reaction margins in the plagioclase grains visible on CL photos that are the result of introducing defects and OH groups into their structure (Slaby et al. 2012).

The potential impact of the mantle-derived fluids can additionally make the high trace element contents credible. The MMEs display high concentrations of large-ion

Fig. 19 Nb/Yb vs Th/Yb diagram, after Pearce (2008). Average N-MORB, E-MORB, OIB are taken from Sun and McDonough (1989); average lower crust (LC), middle crust (MC), upper crust (UC) and total continental crust (CC) are selected from Rudnick and Fountain (1995)

lithophile elements (LILEs) compared to the primitive mantle, as well as positive Rb, Th, and Sm anomalies and negative Nb and Ti anomalies (Fig. 12).

The different protoliths of the MMEs and the HBG are supported by the $\text{Al}_2\text{O}_3 + \text{FeOt} + \text{MgO} + \text{TiO}_2$ vs $\text{Al}_2\text{O}_3/(\text{FeOt} + \text{MgO} + \text{TiO}_2)$ diagram (Fig. 18); the geochemistry of the enclaves indicates that they have an amphibolitic melt source, while the geochemistry of the HBG reflects the partial melting of greywacke material. In this case, the source of both magmas might be crustal.

The lower crust and upper crust are quite heterogeneous in their isotopic compositions and are not easily characterized by a single isotopic composition. In addition, initial Sr and Nd isotope studies of the lower crust (based on xenoliths) indicated that it has ϵNd similar to that of the upper crust and that the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of both environments are variable (Downes 1993; Rudnick 1992). In the case of the HBG, Nd and Sr isotopes are not effective discriminators between potential magma sources. The MMEs have initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (with a mean value of 0.7093) and $\epsilon\text{Nd}(t)$ values (with a mean value of -3.62) (Fig. 13) similar to those of the HBG. We interpret this strong isotopic similarity as a result of complete isotopic equilibration during magma mixing at the emplacement level (Holden et al. 1987, 1991; Pin et al. 1990; Fourcade and Javoy 1991; Stephens et al. 1991; Barbarin and Didier 1992; Poli et al. 1996; Waight et al. 2001; Turnbull et al. 2010), which is more effective than geochemical equilibration (Leshner 1990). The lack of positive $\epsilon\text{Nd}(t)$ values even in the enclaves with the most primitive compositions suggests that the MMEs are hybrids that originated from the lower crust. The partial melting of amphibolitic lower crust could have been caused by the periodic influx of mantle heat/fluids and melts leading to successive contamination during their emplacement. The clear division of the HBG and MMEs into two distinct groups on the neodymium evolution diagram (Fig. 15) may indicate that their parental magmas had different sources. Therefore, the studied enclaves cannot be treated as “autoliths”, i.e., portions of cumulate generated by the fractional crystallization of the granitic melt (Dodge and Kistler 1990; Leshner 1990; Donaire et al. 2005). MMEs, which plot within the same field as the HBG, are characterized by high-silica contents. These enclaves were highly contaminated by granitoid melt during the magma mixing-mingling process. Mixing models (see geochemical modelling, Fig. 17) indicate that the enclave compositions record granitic magma contents of up to 60%.

The tectonic setting

The Sudetic granites can be divided into two groups with distinct emplacement ages of 340–330 Ma and 320–295 Ma (e.g., Mazur et al. 2007; Laurent et al. 2014). The older granites are related to the main stage of nappe stacking within the Central European Variscides (e.g., Franke 2000). The younger granites, such as the Strzegom–Sobótka Massif (as well as the Strzelin Massif and the Žulová Pluton), intruded into the cooled upper crust after the cessation of the Variscan orogeny. Turniak et al. (2014) suggested that the melt generation of the SSM was initiated by crustal heating due to the influx of basaltic magmas rising from the lithospheric mantle. The same mechanism was proposed by Pietranik and Waight (2008) for the Strzelin Massif (~306–291 Ma) and by Laurent et al. (2014) for the Žulová Pluton (~298–291 Ma) in the Sudetic domain of the Central European Variscides (Fig. 1). In contrast to the previously presented models, detailed research on the HBG does not point to the widespread direct participation of mantle melts in granitic magma generation. However, these models have defined the mantle role in a qualitative manner.

The neodymium model ages (T_{DM}) obtained for the HBG (0.9–1.33 Ga) from the Strzegom–Sobótka Massif show a crustal residence age similar to that of the Strzelin Massif (1.09–1.46 Ga for bt-ms granite and tonalite samples; Oberc-Dziedzic et al. 2016; 1.3 Ga for the Gęsinięc quartz diorite; Pietranik and Waight 2008) and the Žulová Pluton (1.18–1.30 Ga for granite and quartz monzodiorite samples; Laurent et al. 2014), which may confirm similar times but partly different mechanisms of generation for their parental melts.

The chemical compositions of the MMEs from the HBG, diorite samples from the Strzelin Massif and quartz diorite samples from the Žulová Pluton (data from Pietranik and Waight 2008; Laurent et al. 2014) appear to be compatible with the partial melting of intermediate metaigneous material, i.e., the crustal source. This interpretation would be in accord with our results (Fig. 18).

Conclusion

The HBG composing the western part of the Strzegom–Sobótka Massif is an example of an intrusion that formed in a multi-component system. The system shows ambiguous isotopic and geochemical signatures. The pluton appears to have a composite character. However, the melts responsible for the granitic rock formation and MMEs were derived from different crustal domains and afterwards were contaminated. These melts exhibit hybrid characteristics. The origins of both the granitic and MME hybrids cannot

be explained by a two end-member system. This system was supplemented by other components, which influenced its isotopic signature. The main mechanism involved in developing granitic rocks was fractional crystallization (FC), but other local processes were involved in rock formation. Generally, the FC signature was not obscured by the input of the hybrid mafic melt. This felsic-mafic interaction was limited and can be viewed as one-sided. The influence of small-volume mafic magma blobs on granite crystallization was negligible. The massif is thus an ideal example of an environment where MMEs coexist with a granitic body; however, the MMEs do not determine the actual composite characteristics of the pluton. MMEs are not cognate with the granitoid magma. The data point to two different stages of the hybridization of two different magmas: one stage led to granitic magma formation, and the other stage led to the formation of hybrid MME magma. The two processes were temporally and spatially distinct.

Either the granitic melt was derived from a metagreywacke crustal source and might have been contaminated during emplacement, or it was originally derived from a compositionally heterogeneous crustal protolith where some of the domains might have been metagreywacke and emplaced into the final chamber.

The parental melts of MMEs were derived from the partial melting of mafic (amphibolitic) lower crust influenced by underlying enriched lithospheric mantle. The hybridization process tended to modify the entire composition of the mafic magma by possibly interacting with fluids. Finally, the mafic magma reached the silicic magma chamber as small blobs and cooled in the nearly crystallized host magma. During this stage, the mingling-mixing textural features developed, and the original hybrid composition of the MMEs was changed due to isotopic and partial geochemical equilibration with the volumetrically prevailing granitic melt. The last process was one-sided as the influence of mafic melt on granitic melt was negligible due to the insignificant volume of the former.

On the basis of field observations, as well as geochemical and isotopic analyses supported by geochemical modelling, we conclude that the main process responsible for the differentiation of the parental melt of the HBG was fractional crystallization. In contrast to the granitic melt, the enclaves have chemical compositions that were mainly obtained by magma mixing-mingling, which likely occurred prior to or during emplacement, as well as in their current locations.

Composite plutons mainly show strong imprints of two end-member systems. Many of these plutons demonstrate that their entire range of granitic facies could have been produced by the mixing of crust- and mantle-derived magmas. Such a system dominated during the early stage of Variscan pluton formation. The pluton described in this paper belongs to the late stage of Variscan magmatism. It presents

a multi-component system shaped by the participation of melts derived from heterogeneous crust domains. This pluton mimics the KCG system (granite with MMEs) but is only seemingly similar to those originating from a two-component system. The proper role of MMEs and their influence on pluton formation can be defined as one-sided reactions with granite magma with only a very limited local scope of mutual interactions. Thus, the presence of enclaves does not explicitly indicate their dominant role in the formation of composite plutons of crust-mantle origin. It is decisive to determine whether the ambiguous signatures appearing in such cases truly indicate both of these sources or merely one source with heterogeneous composition.

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