

Tracking the origin of post-Variscan granites through the geochemical diversity of mafic microgranular enclaves: Strzegom-Sobótka Massif, Poland

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Domańska-Siuda, J., Szuszkiewicz, A., Turniak, K., 2026. Tracking the origin of post-Variscan granites through the geochemical diversity of mafic microgranular enclaves: Strzegom-Sobótka Massif, Poland. *Geological Quarterly*, 70, 21; <https://doi.org/10.7306/gq.1866>

Associate Editor: Stanisław Mikulski

The Strzegom-Sobótka Massif, a composite post-Variscan granitoid intrusion in SW Poland, contains abundant mafic microgranular enclaves that provide crucial insights into crust-mantle interactions during magma evolution. We present new geochemical and Sr-Nd isotopic data from enclaves in its eastern part. These are mostly tonalitic and show narrow isotopic ranges (initial $^{87}\text{Sr}/^{86}\text{Sr}$: 0.7054–0.7060; $\epsilon\text{Nd}(300)$: –2.9 to –1.8) with relatively young depleted mantle model ages (TDM) (1.24–1.65 Gyr). In contrast, western enclaves, previously described in the literature, range from monzodiorite to tonalite and show more variable, more crustal isotopic compositions (initial $^{87}\text{Sr}/^{86}\text{Sr}$: 0.7079–0.7113; $\epsilon\text{Nd}(298)$: –4.5 to –2.5) and older TDM ages (1.10–2.40 Gyr). The western enclaves record extensive chemical and isotopic equilibration with the host hornblende-biotite granite during emplacement, while the isotopic similarity between the eastern enclaves and the host biotite granodiorite reflects derivation from a common juvenile hybrid source. This source incorporated ancient crustal and juvenile mantle-derived components, whereas the western suite was derived from evolved continental crust with amphibolite- and greywacke-like affinities. These contrasting signatures indicate assembly of the pluton from isotopically heterogeneous source domains during post-collisional lithospheric thinning. Geochemical similarities to Permian intermediate volcanic rocks of the North Sudetic Basin further suggest a common melting history within metasomatized subcontinental lithosphere reactivated during post-orogenic extension.

Key words: mafic microgranular enclaves, felsic-mafic magma interaction, Sr–Nd isotopes, Strzegom-Sobótka Massif, post-collisional granites.

INTRODUCTION

Granitic rocks are a significant constituent of the continental crust. Their formation is of fundamental significance in interpreting the generation, growth and reworking of the continents. Granitoids have been identified in almost all tectonic environments, predominantly in collision zones, where the melting of crustal rocks is initiated by the rise of mafic melt (related to subduction) or by the thickening of the crust. Furthermore, granitoids have been observed to form in post-orogenic environments, suggesting a transition from convergence to extension tectonic regimes, and the onset of lithospheric thinning. Due to their geological significance, the petrogenesis

of granitoids has been a long-standing topic of debate, as unraveling this process provides deeper insight into lithospheric dynamics.

Recent research on the genesis of granitic rocks has focused on models suggesting that felsitic magmas are formed by crystal fractionation of mantle-derived melts, by intracrustal melting, or as a mix of mantle- and crustal-derived magmas (see Janoušek et al., 2020; Moya et al., 2021; Jacob et al., 2021; Pichavant et al., 2024 and references therein). Mafic microgranular enclaves are commonly found in metaluminous, calc-alkaline granitic plutons, but are also found in other types of granitic intrusive rocks. These enclaves characteristically have a darker colour than their host rocks and are generally of mafic to intermediate composition. Although the origin of enclaves remains a subject of debate, they are recognized as providing a unique opportunity to constrain the petrogenesis and source of granitoid magmas (e.g., Barbarin and Didier, 1991; Perugini et al., 2003; Barbarin, 2005). The presence of enclaves in felsic plutons is also considered as important evidence of crust–mantle interaction and provides evidence of

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Received: May 4, 2026; accepted: August 20, 2026; first published online: October 13, 2026

the role of mafic magmas in the initiation and evolution of felsic magmas (e.g., [Castro et al., 1990](#); [Vernon, 1990, 2010](#); [Didier and Barbarin, 1991](#); [Hibbard, 1991](#); [Orsini et al., 1991](#); [Wiebe et al., 1997](#); [Barbarin, 2005](#); [Kumar and Rino, 2006](#); [Słaby and Martin, 2008](#); [Clemens et al., 2016](#); [Kumar et al., 2017](#)). Consequently, their origin is of crucial importance in interpreting the history of granitoid intrusions.

Different models have been proposed for the origin of mafic microgranular enclaves, of which the three most important are:

1. The 'restite model', which interprets the enclaves as unmelted remnants of the granitoid source region (e.g., [Chappell et al., 1987](#); [Chappell and White, 1991](#) and references therein);
2. The 'autolith model' (also named the 'cognate' or 'cumulate' model), in which the enclaves have a cognate origin with their felsic host and represent disrupted cumulates or the fine-grained, early precipitating chilled margin of the magma chamber (e.g., [Fershtater and Borodina, 1977, 1991](#); [Dorais et al., 1997](#); [Donaire et al., 2005](#); [Shellnutt et al., 2010](#); [Flood and Shaw, 2014](#));
3. The 'magma mixing' model, in which the enclaves may be regarded as mafic magma 'globules' or 'blobs' that have mingled or partly mixed with felsic magmas (e.g., [Didier, 1973](#); [Vernon, 1984, 1991](#); [Castro et al., 1990](#); [Barbarin and Didier, 1991](#); [Elburg, 1996](#); [Collins et al., 2000](#); [Kumar and Rino, 2006](#); [Słaby and Martin, 2008](#); [Perugini and Poli, 2012](#); [Chen et al., 2015](#); [Kumar, 2020](#) and others).

The Strzegom-Sobótka Massif (SSM) is one of the largest granitic plutons in the Sudetes, located in the northeasternmost part of the Bohemian Massif within the European Variscides ([Fig. 1](#)). It is a composite intrusion formed through the successive emplacement of magma batches with distinct chemical and modal compositions over a period of 10–15 Myr, between ~305 and 295 Ma ([Majerowicz, 1972](#); [Pin et al., 1989](#); [Puziewicz, 1990](#); [Turniak et al., 2014](#)). According to [Turniak et al. \(2014\)](#), the SSM intruded shortly after the cessation of compression in the Variscan orogenides of SW Poland, contemporaneous with the rapid uplift of the Variscides and widespread bimodal volcanism in nearby intramontane troughs ([Awdankiewicz et al., 2014, 2025](#)). This suggests that the gradual heating of the crust was driven by thermal influx into the lower crust, likely related to the ascent of basaltic magmas. These basaltic magmas are thought to have formed through decompressional melting of the lithospheric mantle, implying that the emplacement of the SSM, together with the associated bimodal volcanism, marked the onset of lithospheric thinning beneath Central Europe.

The central segment of the Sudetes, where the SSM is located, consists of a mosaic of fault-bounded tectono-stratigraphic units, each with distinct geochemical characteristics and different metamorphic records. This assemblage has been interpreted as a Variscan subduction-accretionary complex (e.g., [Chopin et al., 2012](#); [Mazur et al., 2015](#); [Tabaud et al., 2020](#) for details and further references). The heterogeneous nature of the bedrock is reflected in the compositional diversity of the SSM magma sources, which likely range from basaltic rocks and their pelitic weathering products to rock series of crustal origin ([Pin et al., 1989](#); [Puziewicz, 1990](#)). This study investigates the extent to which the mafic component of the SSM magmatic system, represented by microgranular mafic enclaves, reflects this source heterogeneity. We focus on enclaves from the western and eastern parts of the massif, hosted by two different granitoids (hornblende-biotite granite and biotite granodiorite, respectively), where such enclaves are particularly common ([Domańska and Słaby, 2004](#); [Birski et](#)

[al., 2018](#); [Domańska-Siuda et al., 2019](#); [Domańska-Siuda and Bagiński, 2019](#)). We provide a comprehensive study of the mineralogy, and whole-rock and Sr–Nd geochemistry, of enclaves hosted in the biotite granodiorite and compare the results with previously published data for enclaves from the hornblende-biotite granite. Our aim is to determine the origin of these enclaves and assess their implications for the evolution of the SSM.

GEOLOGICAL SETTING

The SSM is located in the Sudetes (SW Poland), at the north-eastern margin of the Bohemian Massif within the Central European Variscides ([Fig. 1](#)). The Sudetes comprise Neoproterozoic to lower Paleozoic volcano-sedimentary sequences and magmatic suites metamorphosed during the Cadomian and Variscan orogenies and, in some areas, overlain by the Upper Devonian–Lower Carboniferous deposits of intramontane troughs (e.g., [Żelaźniewicz, 1997](#); [Franke and Żelaźniewicz, 2000](#); [Mazur et al., 2006](#); for review). During the Late Carboniferous–earliest Permian, granitic magmatism led to the emplacement of several plutons across the region ([Fig. 1](#)). The NE–SW-trending Sudetic Boundary Fault, a late Variscan structure reactivated during the Alpine orogeny ([Aleksandrowski et al., 1997](#); [Badura et al., 2007](#)), divides the Sudetes into the elevated Sudetic Block in the SW and the replenished lowlands of the Fore-Sudetic Block in the NE.

The SSM is located in the central sector of the Sudetes (Central Sudetes), within the Fore-Sudetic Block, ~50 km SW of Wrocław ([Fig. 1](#)). At the current level of exposure, the SSM extends over 50 km WNW–ESE, between the towns of Sobótka in the south-east and Jawor in the north-west. The Central Sudetes represent a collage of crustal- and oceanic-related tectono-metamorphic units, collectively interpreted as an accretionary prism formed at the junction of the peri-Gondwanan terranes during the Late Devonian to Early Carboniferous collision (e.g., [Mazur et al., 2015](#); [Jastrzębski et al., 2021](#); [Tabaud et al., 2021](#); [Szczepański et al., 2022](#); for review).

To the east of the SSM, the granites are in intrusive contact with the ~400 Ma mafic-ultramafic Śląska Ophiolite suite ([Oliver et al., 1993](#); [Dubińska et al., 2004](#); [Kryza and Pin, 2010](#); [Awdankiewicz et al., 2021](#)). To the south, the SSM adjoins the high- to medium-grade metamorphic domain of the Góry Sowie Massif, which is mainly composed of partially migmatized paragneisses with minor orthogneisses and amphibolites. These rocks were derived from a volcano-sedimentary succession with Early Cambrian maximum depositional ages and associated Upper Cambrian magmatic rocks (e.g., [Kryza and Fanning, 2007](#); [Tabaud et al., 2021](#); [Jastrzębski et al., 2021, 2025](#) and references therein). The Góry Sowie suite records a polymetamorphic evolution, including a high-pressure–high-temperature event at ~400–390 Ma, amphibolite-facies retrogression at ~380–370 Ma and rapid cooling at ~370–360 Ma ([O'Brien et al., 1997](#); [Gordon et al., 2005](#); [Kryza and Fanning, 2007](#); [Budzyń et al., 2018](#); [Jastrzębski et al., 2021](#); [Tabaud et al., 2021](#)). To the west and SW, 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 meta-volcanic succession. This sequence is interpreted to comprise fragments of a Variscan accretionary prism ([Baranowski et al., 1990](#); [Kryza and Zalasiewicz, 2008](#) and references therein). The Kaczawa Unit extends across the Sudetic Boundary Fault

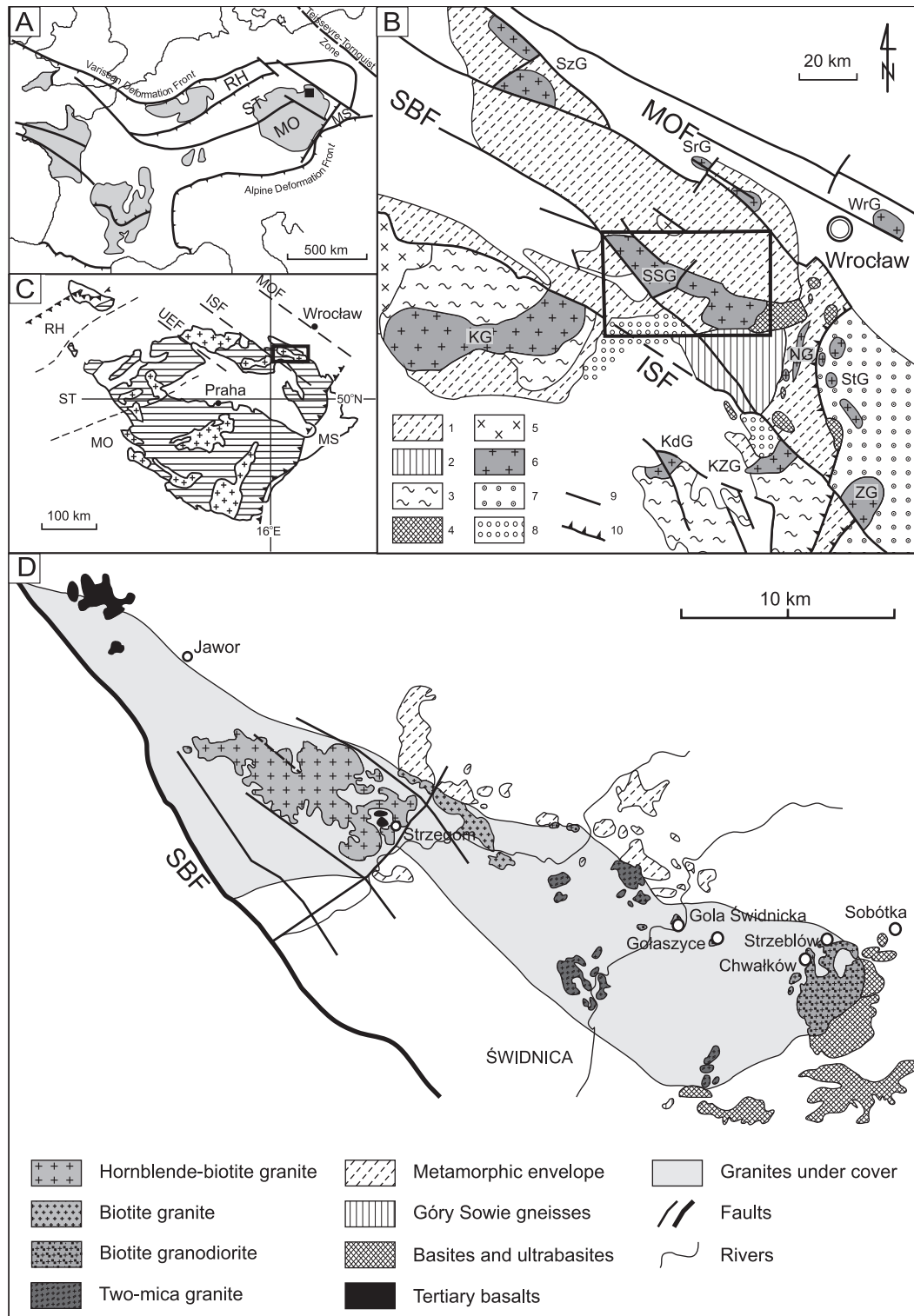


Fig. 1A – location of the study area in the European Variscides (black square). Variscan massifs are shaded; **B** – position of the study area in the Bohemian Massif. Major granitoid intrusions are indicated by crosses; **C** – Variscan granitoids in the Sudetic part of the Bohemian Massif (compilation 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, SBF – Sudetic Boundary Fault, 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 – Żulová Granite

onto the Fore-Sudetic Block and likely formed the metamorphic cover of the SSM, currently mostly eroded or cropping out as isolated patches to the north and NE of the granite exposures.

The SSM consists of four main granitic rock types, emplaced as separate magma batches: (1) hornblende-biotite granite, (2) biotite granite, (3) biotite granodiorite, and (4) two-mica granite (Majerowicz, 1972; Pin et al., 1989; Puziewicz, 1990). The narrowing of their exposures east of Strzegom divides the SSM into an eastern part, comprising biotite granodiorite and two-mica granite, and a western part, composed of hornblende-biotite granite and subordinate biotite granite (Fig. 1). Contacts between these granite varieties are currently not observed in the field. Detailed petrographic descriptions of the SSM were given by Kural and Morawski (1968), Majerowicz (1972), Maciejewski and Morawski (1975) and Puziewicz (1990). The most relevant aspects are summarized below.

The hornblende-biotite granite and the biotite granite display strong petrographic similarities. Both rocks are medium- to coarse-grained, grey, hypidiomorphic and equigranular to porphyritic in texture, typically with a non-oriented fabric. The phenocrysts are mostly K-feldspar, with occasional plagioclase. These granites are modally classified as a monzogranite, composed of ~24–45% plagioclase, 21–44% K-feldspar, 17–38% quartz and 1–12% biotite, with amphibole as an accessory to minor constituent (up to 3%; Majerowicz, 1972). Accessory minerals include apatite, zircon, allanite, titanite, rutile, xenotime and monazite (Majerowicz, 1972; Turniak et al., 2007). Locally, miarolitic pegmatites of the NYF (Niobium–Yttrium–Fluorine) geochemical affiliation, ranging from centimetres to several metres in size, are abundant in the hornblende-biotite granite (e.g., Janeczke, 2007; Pieczka et al., 2015). Furthermore, this granite contains numerous mafic microgranular enclaves (Majerowicz, 1972; Maciejewski and Morawski, 1975; Domańska and Słaby, 2004, 2005; Domańska-Siuda et al., 2019).

The biotite granodiorite occurs in the eastern and central parts of the SSM (Fig. 1). This rock is light-grey, medium-grained, and equigranular, in places showing a slightly directional fabric. It is composed of plagioclase (36–48%), K-feldspar (20–25%), quartz (23–35%) and biotite (3–6%), with accessory zircon, apatite, allanite, monazite, xenotime and opaque minerals (Majerowicz, 1972). Pegmatites and aplites are rare, typically occurring as veins with a massive texture and simple mineral composition (e.g., Puziewicz, 1985; Janeczke, 2007; Pieczka et al., 2015). This facies is also rich in mafic microgranular enclaves (Majerowicz, 1963; Domańska-Siuda and Bagiński, 2019).

The two-mica granite is found in the eastern part of the SSM (Fig. 1). This is a light-grey to yellowish, fine- to medium-grained rock, with a well-defined directional fabric. Locally, it has a porphyritic texture due to the presence of relatively large alkali feldspar crystals. The two-mica granite contains plagioclase (32–34%), perthitic K-feldspar (24–28%), quartz (32–36%) and biotite (2–4%) and muscovite (2–5%) with accessory zircon, apatite, monazite, xenotime and occasionally garnet (Majerowicz, 1972). This variety also contains rare-element vein pegmatites of LCT (Lithium–Cesium–Tantalum) affinity (Janeczke, 1985, 2007; Janeczke and Sachanbiński, 1989; Pieczka et al., 2015) and surmicaceous enclaves with a restitic character (Majerowicz, 1972; Puziewicz, 1987, 1990).

According to Turniak et al. (2014), the amalgamation of the composite SSM during the late Variscan post-collisional extension began with the intrusion of two-mica granite at ~305 Ma, followed by biotite granodiorite dated at ~302–298 Myr. The biotite and hornblende-biotite granites yielded overlapping ages of ~301–298 Myr. Pin et al. (1989) reached a comparable

conclusion based on whole-rock isotope investigations. Zircon saturation temperature calculations (Turniak et al., 2014) revealed a gradual yet significant increase in magma temperatures with decreasing intrusion age. The temperatures rose from ~720°C in the two-mica granite, to 740–780°C in the biotite granodiorite, and finally to 800–860°C in the biotite–hornblende granites. This gradual temperature increase correlates well with the observation that the subsequent magma pulses became progressively enriched in a mantle component, as indicated by higher proportions of mafic microgranular enclaves. This trend is likely associated with thermal thinning of the lithosphere and the subsequent rise of basaltic magmas.

ANALYTICAL METHODS

Twenty-five samples of mafic enclaves and nineteen samples from the host biotite granodiorite from the eastern part of the SSM were selected for geochemical analysis. Sample selection was based on thin-section examinations. Each sample was crushed, quartered, and milled into a fine powder (100 g per sample). Whole-rock major and trace element analyses were conducted at Bureau Veritas Minerals (BVM) Vancouver, Canada. Major oxide and trace element compositions were determined using inductively coupled plasma emission spectrometry (ICP-ES), while trace elements were analysed by inductively coupled plasma mass spectrometry (ICP-MS), following the procedures outlined on BVM's website (<https://www.bureauveritas.com.au>). All results were recalculated on an anhydrous basis, with iron expressed as $\text{Fe}_2\text{O}_{3(\text{total})} = \text{Fe}_2\text{O}_3 + 1.111 \text{ FeO}$.

The chemical compositions of minerals were investigated using the Cameca SX-100 electron microprobe (WDS mode) at the Electron Microprobe Laboratory, part of 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 10 nA; beam-diameter range from 1–5 µm depending on the mineral analysed, and 20 s and 10 s counting times for peak and background positions, respectively. The spectral interferences of the REE analytical lines were automatically corrected in the quantitative result files by the SX100 software. During the quantification process, interference intensity from overlapping lines was calculated based on measurements on the standard. The $\Phi(\rho Z)$ correction model (X-PHI in the electron microprobe software) developed by Merlet (1994) was used to quantify the chemical composition of the minerals investigated. The following standards were used for calibration: albite (Na); diopside (Mg, Si, Ca); wollastonite (Si, Ca); orthoclase (K, Al); hematite (Fe); rhodochrosite (Mn); apatite (P, F); phlogopite (F); barite (S, Ba); rutile (Ti); zircon (Zr); synthetic strontium titanate (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).

Feldspar and biotite formulae were recalculated on the basis of 8 and 10 oxygens, respectively. Although $\text{Fe}^{2+}/\text{Fe}^{3+}$ ratios in biotite were previously estimated by Wichrowska (1974, 1977) and Puziewicz (1994, 1995), these studies did not overlap with our sampling. Therefore, we treated all Fe as Fe^{2+} .

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 in multi-collector

dynamic mode. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1184$, and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. Standard reference materials NBS SRM 987 and JNdi-1 were used, yielding mean values of $^{87}\text{Sr}/^{86}\text{Sr} = 0.710252 \pm 18$ (2σ) and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512095 \pm 13$ (2σ), respectively. Time-related calculations were based on $^{143}\text{Nd}/^{144}\text{Nd} = 0.512638$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$ for the present-day depleted mantle, assuming radiogenic linear growth with $\epsilon_{\text{Nd}} = 0$ at 4.568 Ga. Epsilon values at time T were calculated using the relation: $\epsilon_{\text{Nd}} = [(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}} / (^{143}\text{Nd}/^{144}\text{Nd})_{\text{TCHUR}} - 1] \times 10000$, where CHUR is the chondrite uniform reservoir and T denotes the time the rock was formed. Depleted mantle model ages were calculated using the single-stage model of DePaolo (1981), assuming a modern upper mantle with values of $^{147}\text{Sm}/^{144}\text{Nd} = 0.2137$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.51315$. Isotopic ratios of $^{143}\text{Nd}/^{144}\text{Nd}$ were measured with the MC ICPMS Neptune spectrometer by ThermoFinnigan at the Environmental Geochemistry of Isotopes Laboratory of ING PAN (Kraków, Poland). The $^{147}\text{Sm}/^{144}\text{Nd}$ ratios were calculated from the data obtained, along with Sm and Nd concentrations.

Zircon saturation temperatures T_{Zr} were calculated following the calibration of Watson and Harrison (1983). The composition-dependent factor M [where: $M = (\text{Na} + \text{K} + 2\text{Ca})/(\text{Al} + \text{Si})$ as cation ratios] was verified for all rocks analysed to ensure the applicability of the model. For the biotite granodiorite and enclaves hosted within, the M factor ranges from 0.84 to 1.85, strictly satisfying the experimental calibration limits of 0.90–1.90. However, for four highly mafic enclaves from the hornblende-biotite granite (with $\text{SiO}_2 \leq 53$ wt.%), the M factor exceeds 1.90 (reaching 2.20–2.50) due to elevated Ca and lower Si contents. Because the model is inapplicable to these specific compositions, these samples were excluded from the temperature estimations to avoid unreliable mathematical extrapolations.

RESULTS

PETROGRAPHY

Mafic microgranular enclaves occur mainly in the hornblende-biotite granite and biotite granodiorite in the western and eastern parts of the SSM, where they are referred to as enclaves from the western part (EWP) and enclaves from the eastern part (EEP), respectively. EWP are distributed relatively uniformly within the granitic host. These enclaves have spherical to ellipsoidal shapes, ranging in size from a few centimetres to over a metre (Fig. 2A–C). Smaller enclaves are generally dark grey and fine-grained, whereas larger ones are lighter in colour and have relatively coarse-grained textures (Fig. 2A, B). Most enclaves show non-oriented textures; they are typically equigranular, though some occasionally contain feldspar phenocrysts that measure 1–2 cm in length (Fig. 2A). Large feldspar phenocrysts are sometimes observed to cross the contacts between the enclaves and their granitic host (Fig. 2A) or to be present within the enclaves themselves (Fig. 2A). Contacts between EWP and the host granite can vary: sharp contacts are frequently observed in smaller, finer-grained and darker examples (Fig. 2B), whereas partly diffuse lobate to crenulated contacts are typical of larger EWP (Fig. 2). Some EWP are surrounded by narrow felsic halos enriched in quartz and K-feldspar, while others (especially larger ones) form composite enclaves (Fig. 2C).

EWP were previously studied by Domańska and Ślaby (2004, 2005) and Domańska-Siuda et al. (2019); the published mineralogical and geochemical data are used here for comparison. According to these studies, the compositions of EWP range from monzonite and monzodiorite to granodiorite and tonalite. Generally, EWP consist of 40–65% plagioclase, 25–30% biotite and 0–3% amphibole, with quartz and K-feldspar as subordinate minerals (up to 5%). Typical accessory minerals include zircon (5–100 μm long), needle-like apatite (0.1–0.5 mm), allanite and opaque phases, while monazite (5–25 μm) is rare.

In this contribution, we focus on EEP, which are particularly abundant in the Strzeblów quarry (Fig. 1). Our field, macro- and microscopic investigations reveal that they occur as uniformly distributed discrete individual bodies or, sometimes, form swarms or narrow steeply inclined trains (Fig. 2D–F). Compared to EWP, dark and fine-grained EEP are smaller, ranging from 3 to 50 cm in size, but typically measure between 5 and 15 cm in diameter. Most EEP are oval or elliptical in shape, although some are sub-angular, especially those forming trains (Fig. 2D–F). Contacts between EEP and the host rock vary from sharp but unchilled, to lobate or diffuse over distances of a few centimetres (Fig. 2D). Most EEP are equigranular but may occasionally contain phenocrysts (Fig. 2E). Feldspar phenocrysts also extend across the boundary between these enclaves and their granitic host, indicating dynamic magma flow and comingling. Hybrid or composite enclaves, though common in EWP, are relatively rare among EEP (Fig. 2G). In general, our observations suggest that EEP are petrographically more homogeneous than EWP.

Although the petrography of EEP was initially provided by Domańska-Siuda and Bagiński (2019), our new sample set allows for a more comprehensive characterization. Most EEP are modally classified as tonalites, consisting mainly of plagioclase (50–65%) and biotite (30–35%), with subordinate quartz (5–15%) and K-feldspar (0–5%). The enclaves contain needle-like apatite, up to 0.5 mm in size, although finer crystals (<0.1 mm) are more common. Zones particularly rich in apatite are found close to the boundary between the enclaves and their host rock. Another abundant accessory phase in these contact zones is monazite, with the largest crystals (up to 500 μm) typically occurring near the boundaries of the enclaves, where adjacent enclaves are closely spaced. Zircon (5–80 μm long) and opaque minerals are also common in EEP, while allanite is relatively rare compared to EWP.

MINERAL COMPOSITIONS

Results from this and previous mineralogical studies (Domańska-Siuda et al., 2019) are compiled in Figure 3. Mineral compositions determined for EEP in the present study are given in Appendices 1–3.

Variations in feldspar compositions from EWP and EEP are illustrated in an An–Ab–Or ternary plot (Fig. 3A). In both cases, small euhedral laths of feldspar form the groundmass of enclaves. However, a wider range of plagioclase compositions is observed in EWP compared to EEP. Most of the EWP plagioclase shows distinct normal zoning, with a calcic core (45–66 mol% An) and a sodic rim (1–33 mol% An). The Or content can reach up to 8%. In contrast, EEP plagioclase shows clear normal or oscillatory zoning, with a more calcic core (27–45 mol% An) and a less calcic rim (13–25 mol% An). Reverse zoning (12–22 mol% An) is less common, and the Or content can reach up to 4%. Large plagioclase phenocrysts

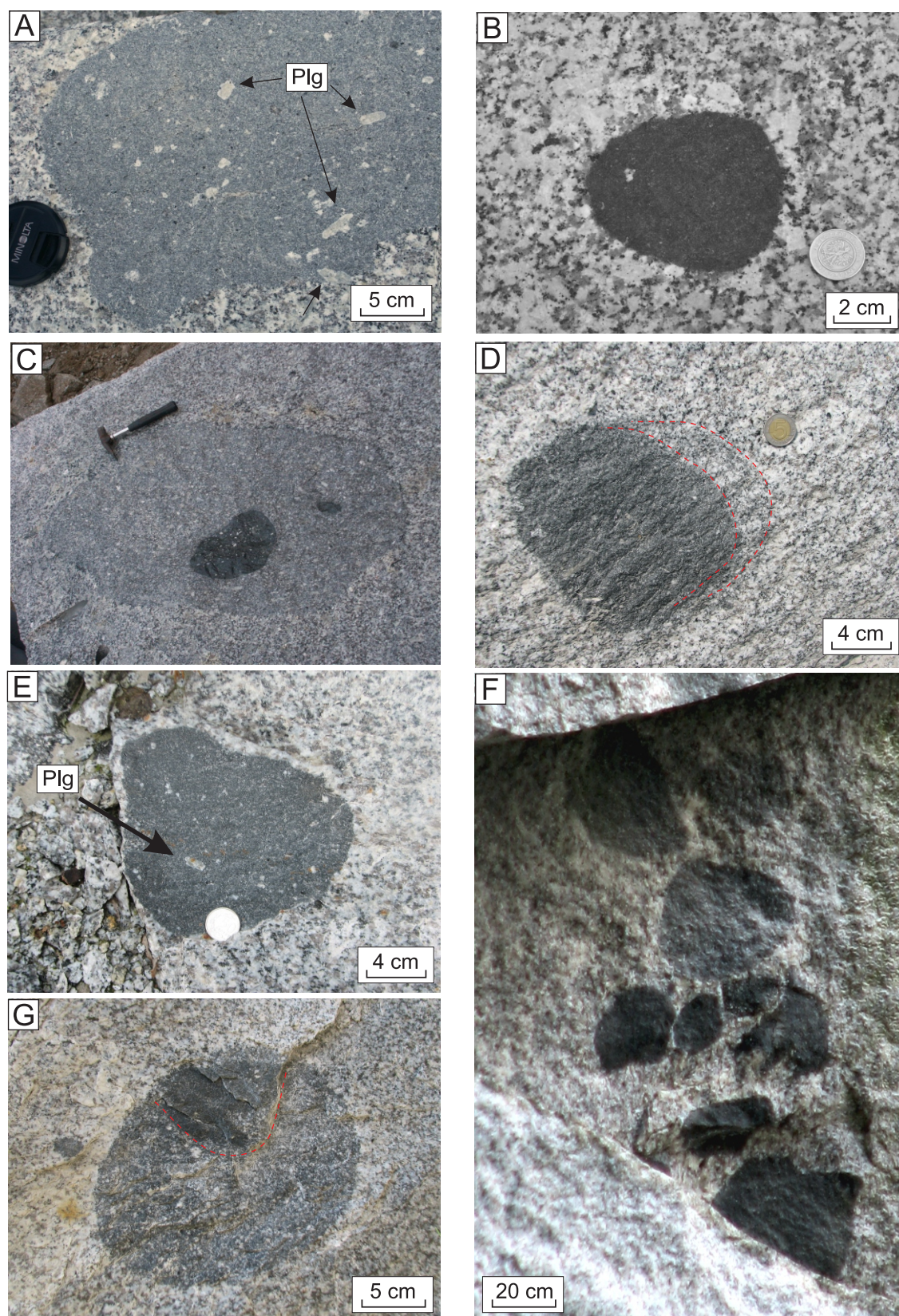


Fig. 2. Field photographs of the enclave-granitoid associations from the Strzegom-Sobótka Massif

A – plagioclase crystal crossing the enclave-granite boundary with with a large feldspar phenocrysts present within the enclave (EWP); **B** – dark, fine-grained enclave (EWP); **C** – composite enclave rimmed by thin felsic halo (EWP); **D** – dioritic enclaves (EWP) with progressive hybridization zone (indicated by dotted lines); **E** – enclave rimmed by felsic halo with a crystal of feldspar within (EEP); **F** – enclave swarm (EEP); **G** – composite enclave (EEP)

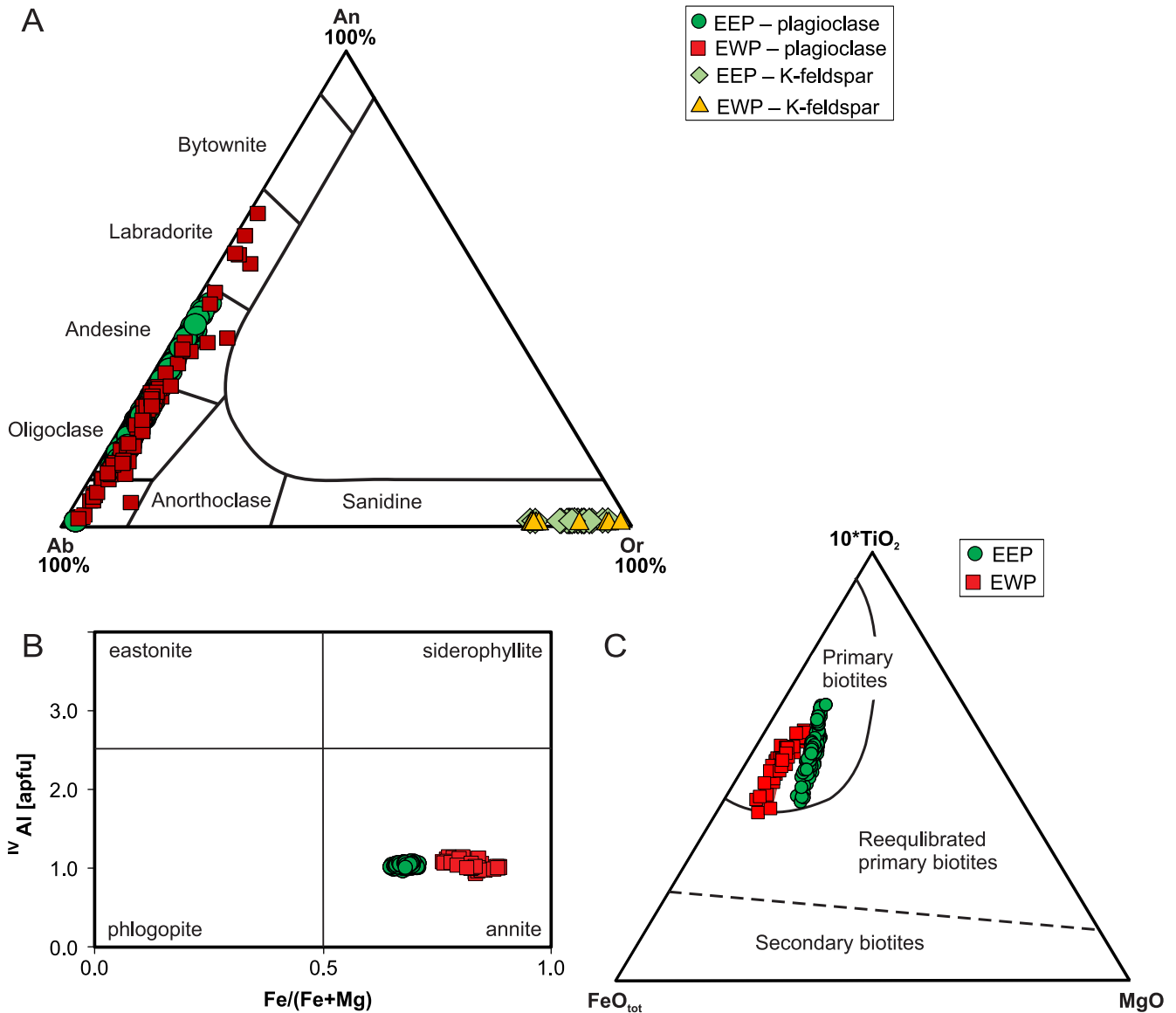


Fig. 3. Mineral chemistry of biotite and feldspar from EEP (this study) and EWP (data from Domańska-Siuda et al., 2019)

A – plot of the EMPA data in the feldspar classification diagram; **B** – plot of the EMPA data from biotite in a ^{IV}Al [apfu] vs. $Fe/(Fe+Mg)$ diagram; **C** – $10*(TiO_2)-FeO_{tot}+MgO$ ternary diagram (Nachit et al., 2005) plotted for EWP and EEP biotites suggesting their magmatic nature

that occur close to the contacts between the enclaves and the granitic host, and less frequently within the interior of the enclaves, are characteristic of EWP. They commonly display oscillatory zoning, with patchily zoned or strongly altered cores. Their compositions range from andesine (up to 45 mol% An) to albite. Similar plagioclase crystals from EEP are generally rounded due to resorption, with poikilitic rims containing biotite inclusions. Their compositions range from labradorite (up to 53 mol% An) to albite. The Ab contents of groundmass K-feldspar grains from EWP range from 2 to 5.5 mol%, with Ba concentrations reaching up to 0.62 wt.% BaO. Groundmass K-feldspar from EEP shows a broader range of Ab contents, from 2 to 17%, with up to 1.2 wt.% BaO.

Biotite is the most common mafic mineral found in the enclaves. In EWP, it forms subhedral, ragged platy crystals up to 0.7 mm long, all very uniform in size. In smaller and darker enclaves, biotite is lath-shaped. In EEP, biotite occurs only as elongated, lath-shaped grains or as inclusions in plagioclase. Biotite in both EWP and EEP is classified as annite (Fig. 3B).

The EWP biotite has high $Fe^{2+}/(Fe^{2+}+Mg)$ ratios ranging from 0.71 to 0.84, with a mean TiO_2 content of 3.38 wt.% (Domańska-Siuda et al., 2019). EEP biotite has lower $Fe^{2+}/(Fe^{2+}+Mg)$ ratios of 0.60–0.66, though with similarly high Ti contents, with a mean of 3.67 wt.% TiO_2 . According to the $10*(TiO_2)-FeO_{tot}+MgO$ ternary components (Nachit et al., 2005), both biotites belong to primary biotites (Fig. 3C).

Amphibole occurs only in EWP, where it is a rare mineral phase. It has high FeO contents (26.94 to 29.32 wt.%), low MgO contents (3.14 to 4.33 wt.%), and TiO_2 contents between 0.78 and 2.07 wt.%. The $[Mg/(Mg+Fe^{2+})]$ ranges from 0.19 to 0.25. The amphibole is classified as hastingsite, with $0.91 \leq Ca/(Ca + Na) \leq 0.94$ (Domańska-Siuda et al., 2019). Its magmatic origin is indicated by Si contents (mean = 6.60), which are below 7.50 a.p.f.u. limit for igneous amphiboles (Leake, 1971; Fleet and Barnett, 1978). It crystallized under intermediate oxygen fugacity conditions, as indicated by an average $Fe/(Fe+Mg)$ ratio of 0.78 (Anderson and Smith, 1995).

Table 1

Major and trace element analyses of EEP and BG host

Sample	022*	122	123*	124*	125-A*	125-B*	152	153	154	156-A	156-B	157	158	160	161
Locality	1	1	1	1	1	1	1	1	1	1	1	1	1	1	1
Lithology	e	e	e	e	e	e	e	e	e	e	e	e	e	e	e
	wt. %														
SiO ₂	62.76	62.30	64.70	65.55	63.06	67.16	62.30	64.44	62.15	65.69	64.32	66.20	64.67	63.05	60.27
TiO ₂	0.87	0.89	0.72	0.69	0.82	0.59	0.80	0.71	0.85	0.66	0.75	0.61	0.74	0.84	0.94
Al ₂ O ₃	17.48	17.16	16.55	16.21	17.10	16.39	17.33	16.87	17.25	16.58	16.94	16.66	16.78	17.35	17.71
Fe ₂ O ₃ T	5.68	5.96	4.88	4.76	5.61	3.97	5.71	4.80	6.06	4.63	5.22	4.37	5.07	5.42	6.98
MnO	0.16	0.18	0.15	0.15	0.16	0.12	0.17	0.13	0.17	0.13	0.15	0.13	0.14	0.14	0.22
MgO	1.68	1.68	1.38	1.34	1.64	1.13	1.67	1.42	1.74	1.35	1.49	1.25	1.47	1.64	2.02
CaO	3.61	3.23	2.97	2.97	3.35	2.96	3.02	3.22	2.98	3.02	3.38	3.14	3.21	3.34	2.58
Na ₂ O	4.64	4.80	5.05	4.80	4.79	5.25	5.14	4.95	4.93	5.11	4.73	4.95	4.93	4.97	4.98
K ₂ O	2.13	2.30	1.93	1.84	2.11	1.57	2.25	1.85	2.34	1.80	1.95	1.67	1.90	2.08	3.26
P ₂ O ₅	0.38	0.33	0.28	0.24	0.36	0.23	0.29	0.31	0.34	0.28	0.38	0.29	0.30	0.34	0.37
Total	99.80	99.52	99.42	99.59	99.80	99.97	99.80	99.81	99.82	99.97	99.92	99.98	99.93	99.79	99.93
A/CNK	1.06	1.06	1.05	1.06	1.05	1.04	1.06	1.05	1.07	1.05	1.06	1.05	1.06	1.05	1.08
Mg no.	0.37	0.36	0.36	0.36	0.37	0.36	0.37	0.37	0.36	0.37	0.36	0.36	0.36	0.37	0.36
	ppm														
Ba	506.6	446.1	331.6	360.4	462.6	351.9	511.1	422.1	382.5	255.7	283.1	427.5	450.7	473.4	630.3
Cs	8.9	12.5	10.3	10.2	10.6	6.8	16.4	9.1	15.9	13.2	11.9	9.2	13.3	11.1	25.0
Ga	22.3	27.0	26.1	25.6	26.1	23.1	28.4	23.8	26.1	23.4	23.1	23.1	26.9	23.2	36.1
Hf	4.7	6.3	6.3	6.1	4.8	5.1	6.6	5.5	6.1	4.9	4.9	5.3	5.3	4.8	10.9
Nb	27.3	42.4	35.1	36.4	30.8	26.2	49.5	26.8	39.0	29.2	25.9	28.0	34.4	26.2	68.7
Rb	142.6	189.2	157.0	150.3	165.4	119.0	187.9	136.0	190.8	151.6	139.4	142.7	170.7	152.1	264.9
Sr	355.2	330.3	256.2	297.3	355.2	287.5	282.8	346.0	297.2	261.9	267.7	335.3	389.9	277.1	226.5
Ta	1.5	3.4	2.6	2.9	2.1	2.0	5.0	1.9	3.8	1.9	1.7	2.2	2.2	1.9	8.1
Th	4.6	7.0	8.0	9.2	5.3	7.6	8.4	7.2	6.1	7.4	6.6	7.3	7.2	5.6	6.7
U	8.9	12.1	3.8	3.9	17.7	3.8	3.7	21.2	33.5	4.0	3.6	27.1	6.0	2.5	17.7
Co	7.3	8.5	6.4	6.0	8.4	5.2	8.3	6.9	8.7	7.1	6.2	7.8	7.7	8.7	9.9
V	55.0	56.0	42.0	43.0	58.0	37.0	54.0	51.0	56.0	42.0	40.0	52.0	50.0	58.0	70.0
Zr	154.7	189.2	209.7	180.5	169.6	172.5	216.9	185.5	187.5	175.1	168.1	190.4	202.0	174.3	288.7
Y	42.5	53.9	45.0	41.7	46.8	30.4	44.5	40.2	54.0	36.2	32.9	43.1	44.4	32.1	57.1
La	26.1	25.4	34.4	27.7	27.6	26.6	26.3	32.2	27.6	30.5	25.9	27.2	29.8	29.4	15.3
Ce	52.3	54.5	71.3	56.4	55.2	53.1	53.7	64.0	57.9	60.6	52.7	55.2	62.3	60.2	34.8
Pr	6.6	6.4	7.6	6.4	6.8	6.1	6.4	7.3	7.0	7.0	6.1	6.6	7.2	7.1	4.4
Nd	25.4	25.6	28.7	24.4	26.6	22.9	25.4	27.8	26.8	26.0	22.8	25.3	26.6	26.8	18.4
Sm	6.5	6.7	6.6	5.7	7.0	5.8	7.1	7.1	7.7	6.7	5.9	6.8	7.3	6.4	6.6
Eu	0.8	0.7	0.7	0.6	0.8	0.7	0.7	0.8	0.8	0.6	0.6	0.8	0.9	0.7	0.6
Gd	7.2	7.8	7.0	6.1	7.9	5.7	7.7	7.4	8.5	6.8	6.4	7.5	8.1	6.4	7.9
Tb	1.4	1.6	1.4	1.2	1.5	1.1	1.6	1.4	1.7	1.3	1.2	1.5	1.5	1.2	1.8
Dy	7.5	7.9	7.2	6.4	8.0	5.5	8.4	7.3	8.8	6.8	6.2	7.7	8.1	6.2	10.3
Ho	1.3	1.7	1.4	1.2	1.5	1.0	1.5	1.3	1.7	1.2	1.1	1.5	1.4	1.1	1.8
Er	3.6	4.6	3.7	3.6	4.1	2.7	4.0	3.5	4.7	3.2	2.9	3.8	4.0	2.8	5.0
Tm	0.5	0.7	0.6	0.5	0.6	0.4	0.6	0.5	0.7	0.5	0.4	0.6	0.6	0.4	0.9
Yb	3.3	3.9	3.3	3.3	3.7	2.3	3.6	3.0	4.1	2.8	2.5	3.2	3.3	2.4	5.7
Lu	0.5	0.6	0.5	0.5	0.6	0.3	0.6	0.4	0.6	0.4	0.4	0.5	0.5	0.4	0.8
(La/Yb) _N	5.2	4.3	7.0	5.7	5.0	7.7	4.9	7.2	4.5	7.4	6.8	5.7	6.0	8.2	1.8
Eu/Eu*	0.3	0.3	0.3	0.3	0.3	0.4	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.4	0.2
TZr °C	768	784	797	787	776	783	796	786	785	782	781	788	794	778	820

GEOCHEMISTRY

Geochemical data from EEP and the host biotite granodiorite are provided in Table 1 [together with data for 8 samples published by Domańska-Siuda and Bagiński (2019), added for comparison]. Major and trace element data for EWP and their granitic host were published by Domańska-Siuda et al., (2019). EWP have a high alkali sum (Na₂O+K₂O) ranging from 5.8 to 8.0 wt.% and a broad compositional range, with SiO₂ contents between 50.6 and 69.2 wt.% (Table 1 and Fig. 4A). According to the total alkali-silica (TAS) diagram of Middelmost (1994), EWP are classified mainly as monzodiorite, monzonite and granodiorite. The enclaves are peraluminous to metaluminous, with the Aluminous Saturation Index [A/CNK = molar ratio of Al₂O₃/(CaO+Na₂O+K₂O)] ranging from 0.85 to 1.22 (Fig. 4B). The Mg# (Mg/(Mg + Fe²⁺)) is low, ranging from 0.16 to 0.33. Following the classification of Frost et al. (2001), EWP are ferroan and alkalic to calc-alkalic (Fig. 4C, D).

EEP display a narrower range of SiO₂ contents (60.3–68.8 wt.%), similar alkali concentrations (Na₂O+K₂O = 6.5–8.2 wt.%), and mainly plot in the granodiorite field on the TAS diagram (Fig. 4A). The Mg# is in the range of 0.36–0.40, dis-

tinctly higher than in EWP. EEP are magnesian (Fig. 4D) and peraluminous, with an A/NCK ratio ranging between 1.04 and 1.10 (Fig. 4B), and plot mostly in the calc-alkalic field (Fig. 4C).

For both EWP and EEP, most elements (e.g., TiO₂, Al₂O₃, Fe₂O₃, MnO, MgO, CaO, P₂O₅, Rb, Zr, Hf, Nb, Y, V, Ni) show negative correlations with SiO₂ (selected diagrams in Fig. 5). For, EWP, K₂O and Na₂O do not show straightforward correlations with SiO₂; instead, they form curved trends. In EEP, K₂O displays a negative trend, while Na₂O does not correlate with the differentiation index. The greatest differences in major element contents are observed for Fe, Mg and Ca. EWP are enriched in Fe₂O₃ relative to EEP (3.78–17.47 wt.% vs. 3.21–7.84 wt.%, respectively) but have lower contents of MgO (0.46–2.29 wt.% vs. 0.98–2.34 wt.%, respectively) and CaO (1.98–4.42 wt.% vs. 2.58–3.61 wt.%, respectively). Smaller differences were also observed in the trace element contents, including Sr, Nd, Sm, Y, Nb, Zr (Fig. 5E–J).

Trace element signatures are shown in primitive mantle-normalized (Sun and McDonough, 1989) multi-element diagrams (Fig. 6B, D). All the enclaves are enriched in large-ion lithophile elements (LILE) and high-field strength elements (HFSE) relative to the primitive mantle. EWP display patterns

Tabl. 1 cont.

Sample	163	164	SS48A	SS156	SS159	SS160	SS229	SS235	SS240	SS241	003	007	008	021*	023*
Locality	1	1	1	1	1	1	2	3	3	3	2	4	4	1	1
	e	e	e	e	e	e	e	e	e	e	g	g	g	g	g
wt. %															
SiO ₂	68.24	65.61	62.98	68.76	64.80	65.85	66.76	62.70	65.33	67.04	69.81	69.23	69.65	74.08	72.57
TiO ₂	0.49	0.68	0.80	0.50	0.70	0.66	0.60	0.87	0.71	0.60	0.45	0.47	0.45	0.19	0.25
Al ₂ O ₃	16.58	16.69	16.90	15.77	16.68	16.52	16.50	16.97	16.62	16.20	15.13	15.37	15.39	13.97	14.47
Fe ₂ O ₃ T	3.41	4.61	5.46	3.35	4.61	4.55	3.21	5.55	4.57	3.85	2.61	2.67	2.57	1.50	1.85
MnO	0.10	0.13	0.16	0.10	0.13	0.13	0.06	0.12	0.14	0.11	0.05	0.05	0.05	0.04	0.06
MgO	0.98	1.32	1.63	0.99	1.38	1.34	1.10	1.86	1.37	1.13	0.85	0.87	0.83	0.37	0.49
CaO	2.83	3.20	3.19	2.81	3.02	2.92	2.87	2.63	3.01	2.91	2.49	2.47	2.52	1.59	1.89
Na ₂ O	5.08	4.97	4.78	5.03	5.02	5.03	4.68	4.72	5.19	4.98	3.84	4.05	4.13	3.78	4.16
K ₂ O	1.43	1.79	2.11	1.69	1.81	1.75	2.74	3.35	1.72	1.76	3.63	3.75	3.51	3.87	3.42
P ₂ O ₅	0.17	0.30	0.34	0.16	0.25	0.24	0.19	0.21	0.27	0.22	0.13	0.13	0.13	0.08	0.06
Total	99.83	99.80	99.77	99.16	98.40	98.99	98.71	99.75	99.75	99.75	98.99	99.06	99.23	99.47	99.22
A/CNK	1.10	1.05	1.06	1.04	1.06	1.07	1.04	1.05	1.05	1.05	1.02	1.01	1.01	1.05	1.03
Mg no.	0.36	0.36	0.37	0.37	0.37	0.37	0.40	0.40	0.37	0.37	0.39	0.39	0.39	0.33	0.34
ppm															
Ba	354.5	418.1	458.0	334.0	371.0	352.0	655.0	826.0	356.0	366.0	731.0	683.5	673.5	583.7	590.3
Cs	11.7	11.4	12.5	7.5	8.0	9.1	9.0	10.1	4.4	3.8	2.3	3.1	2.6	3.9	4.4
Ga	23.1	26.2	26.2	22.0	24.9	24.4	20.2	26.6	23.9	21.6	18.3	19.3	20.5	16.0	18.5
Hf	6.0	5.4	4.9	5.1	6.0	6.8	5.6	8.4	5.6	4.9	4.0	4.5	4.4	3.4	3.4
Nb	22.8	29.6	36.2	24.3	32.4	35.4	18.5	32.1	30.8	25.4	13.2	13.2	14.7	9.6	14.0
Rb	112.4	144.6	181.2	111.3	144.0	141.7	134.4	142.5	124.2	110.3	107.4	125.8	112.8	110.3	114.7
Sr	291.5	330.6	327.2	283.4	292.8	295.1	371.9	325.6	290.7	273.4	386.9	372.3	385.1	229.8	268.3
Ta	2.0	2.1	2.7	2.2	2.5	3.2	2.0	2.1	2.5	2.4	1.2	1.1	1.4	1.2	1.3
Th	7.1	6.8	7.9	8.5	8.0	8.2	13.3	23.0	7.9	8.9	16.2	17.3	13.3	9.8	11.2
U	3.7	3.7	8.0	11.6	4.2	4.4	6.3	9.3	6.4	6.2	3.6	3.9	4.5	6.2	4.5
Co	5.0	7.1	63.7	101.2	69.8	71.6	72.4	62.9	99.9	64.0	5.4	5.8	5.4	3.6	2.9
V	34.0	46.0	55.0	34.0	46.0	44.0	45.0	60.0	46.0	48.0	40.0	39.0	36.0	13.0	15.0
Zr	183.9	197.6	176.9	160.3	213.3	223.4	181.4	253.1	175.6	162.4	139.7	145.8	150.7	83.6	99.5
Y	27.0	37.1	42.3	27.9	37.2	35.1	24.3	34.9	34.5	28.1	18.7	17.8	19.8	19.2	21.8
La	23.0	30.0	27.5	25.4	32.2	30.6	35.2	30.8	29.0	32.2	30.7	26.4	31.1	13.9	17.1
Ce	46.4	62.8	55.5	50.5	63.8	59.6	74.0	66.1	58.5	63.4	66.8	55.8	68.0	27.6	36.0
Pr	5.1	7.2	6.6	5.6	7.3	6.9	8.0	7.5	6.4	6.9	7.1	6.0	7.1	3.4	3.9
Nd	18.8	27.0	24.4	21.3	27.0	27.3	30.0	30.4	24.4	27.4	25.8	21.7	25.1	12.2	15.4
Sm	4.5	6.8	7.2	5.1	6.9	6.7	5.6	6.8	5.9	5.5	4.4	4.0	4.3	2.9	3.0
Eu	0.7	0.8	0.8	0.7	0.8	0.7	1.0	0.7	0.7	0.7	0.9	0.7	0.9	0.5	0.5
Gd	4.5	7.0	8.3	5.5	7.4	6.8	4.9	6.7	6.6	5.9	3.5	3.1	3.6	2.8	3.0
Tb	0.9	1.3	1.5	1.0	1.3	1.2	0.8	1.2	1.2	1.0	0.6	0.5	0.6	0.6	0.6
Dy	4.9	6.9	8.4	5.4	7.1	6.8	4.5	6.2	6.2	5.1	3.0	2.7	3.0	3.2	3.1
Ho	0.9	1.2	1.5	1.0	1.2	1.2	0.8	1.2	1.2	1.0	0.6	0.6	0.6	0.6	0.7
Er	2.5	3.2	4.2	2.7	3.4	3.4	2.4	3.3	3.3	2.6	1.7	1.7	1.9	1.8	2.0
Tm	0.4	0.5	0.6	0.4	0.5	0.5	0.4	0.5	0.5	0.4	0.3	0.3	0.3	0.3	0.3
Yb	2.3	2.8	3.6	2.7	3.2	3.2	2.3	3.1	2.8	2.3	1.7	1.7	1.8	1.8	2.0
Lu	0.4	0.4	0.4	0.4	0.4	0.4	0.3	0.3	0.5	0.5	0.3	0.3	0.3	0.3	0.3
(La/Yb) _N	6.7	7.3	5.2	6.3	6.7	6.4	10.3	6.7	7.0	9.3	12.0	10.4	11.9	5.1	5.9
Eu/Eu*	0.5	0.3	0.3	0.4	0.3	0.3	0.6	0.3	0.4	0.4	0.7	0.7	0.7	0.6	0.6
TZr °C	796	793	781	779	801	807	787	810	782	779	768	769	773	734	745

with pronounced negative anomalies for Ba, Sr, P and Ti, while EEP show negative anomalies for Ba, Th, Nb, Ti, and weaker negative anomalies for P. These negative anomalies probably reflect the fractionation of plagioclase, apatite (and other phosphates, such as monazite), and Fe-Ti oxides during earlier stages of magma evolution.

Calculated zircon saturation temperatures T_{Zr} , based on the calibration of [Watson and Harrison \(1983\)](#), for the rock suites studied range from 702°C to 787°C for the biotite granodiorite and from 768°C to 820°C for EEP, with all individual values listed in [Table 1](#).

Sr-Nd ISOTOPE DATA

The Sr-Nd isotopic compositions of EEP and their host biotite granodiorite are listed in [Table 2](#). For comparison, data for EWP and the hornblende-biotite granite host were sourced from [Domańska-Siuda et al. \(2019\)](#). The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and ϵNd_t values for EEP were calculated using the SHRIMP U-Pb zircon age of ~300 Myr obtained for the biotite granodiorite by [Turniak et al. \(2014\)](#).

EEP are characterized by Rb contents ranging from 110 to 265 ppm and Sr concentrations between 227 and 390 ppm ([Table 2](#)), yielding low and restricted $^{87}\text{Rb}/^{86}\text{Sr}$ ratios (0.92–1.77). Their initial $^{87}\text{Sr}/^{86}\text{Sr}$ values range from 0.7054 to 0.7060. The host biotite granodiorite shows similarly low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7054–0.7059) and $^{87}\text{Rb}/^{86}\text{Sr}$ ratios between 1.24 and 1.52. In terms of the Sm–Nd system, EEP contain 18–30 ppm Nd and 5–8 ppm Sm ([Table 2](#)). The initial $\epsilon(\text{Nd})_t$ values for EEP display a narrow range from –1.8 to –2.9. A close isotopic relationship is observed between the enclaves and their host biotite granodiorite, with the $\epsilon(\text{Nd})_t$ values of the host ranging from –1.8 to –3.1 ([Table 2](#)). The corresponding depleted mantle model ages (T_{DM}), calculated using the procedure of [DePaolo \(1981\)](#), range from 1.24 to 1.65 Gyr for EEP ([Table 2](#)). The chondritic uniform reservoir model ages (T_{CHUR}) for these enclaves fall between 0.66 and 0.76 Gyr ([Table 2](#)). EEP data cluster within a narrow field defined by these $\epsilon(\text{Nd})_t$ values and T_{DM} ages, indicating a relatively uniform isotopic evolution ([Fig. 7D](#)).

Tabl. 1 cont.

Sample	024*	120	128	SS27a	SS30a	SS42a	SS127a	SS146a	SS155a	SS161a	SS162a	SS224	SS230	SS239a
Locality	1	2	3	2	2	3	3	4	1	1	1	4	2	3
	g	g	g	g	g	g	g	g	g	g	g	g	g	g
wt. %														
SiO ₂	73.07	67.53	72.57	70.05	69.19	73.17	71.77	68.12	71.78	72.05	71.77	69.20	69.34	72.22
TiO ₂	0.27	0.54	0.26	0.45	0.43	0.20	0.24	0.51	0.24	0.20	0.23	15.28	15.52	14.82
Al ₂ O ₃	14.56	16.11	14.04	15.06	15.51	14.43	15.07	15.99	14.96	14.92	15.14	2.69	2.57	1.43
Fe ₂ O ₃ T	1.97	3.02	1.84	2.55	2.53	1.33	1.70	2.82	1.74	1.44	1.62	0.91	0.91	0.43
MnO	0.07	0.06	0.06	0.05	0.04	0.03	0.05	0.05	0.05	0.04	0.05	2.55	2.57	1.85
MgO	0.53	1.01	0.51	0.91	0.85	0.39	0.50	0.99	0.51	0.41	0.48	3.82	4.00	4.19
CaO	1.83	2.81	1.82	2.47	2.56	1.71	1.87	2.82	1.92	1.83	1.94	3.70	3.39	3.83
Na ₂ O	4.14	4.23	4.18	3.81	4.04	3.87	4.19	3.85	4.25	3.99	4.23	0.47	0.47	0.22
K ₂ O	3.51	3.50	2.99	3.58	3.50	3.83	3.72	3.51	3.55	4.15	3.57	0.15	0.15	0.06
P ₂ O ₅	0.06	0.15	0.08	0.12	0.14	0.05	0.07	0.15	0.07	0.06	0.06	0.05	0.05	0.05
Total	100.01	98.96	98.35	99.05	98.79	99.01	99.18	98.81	99.07	99.09	99.09	98.82	98.97	99.10
A/CNK	1.04	1.02	1.05	1.03	1.03	1.06	1.05	1.05	1.04	1.04	1.06	1.02	1.04	1.03
Mg no.	0.35	0.40	0.35	0.41	0.40	0.37	0.37	0.41	0.37	0.36	0.37	0.40	0.41	0.37
ppm														
Ba	547.0	710.1	471.9	734.0	729.0	689.0	655.0	880.0	556.0	846.0	621.0	765.0	702.0	597.0
Cs	6.3	2.4	2.3	2.6	3.0	3.5	3.8	1.9	5.1	5.1	3.7	2.7	5.5	3.3
Ga	18.5	18.9	16.6	17.7	18.4	15.1	18.0	18.7	18.6	17.1	17.8	16.5	17.3	17.8
Hf	4.2	5.0	3.6	3.8	4.5	3.1	3.5	4.2	3.0	2.7	3.2	4.4	4.4	3.5
Nb	17.2	14.1	15.1	11.7	14.4	9.3	12.9	13.9	13.0	10.9	12.3	12.9	15.0	12.0
Rb	138.0	106.3	107.3	112.0	105.9	98.0	119.4	99.4	118.1	125.4	114.1	104.6	114.1	110.5
Sr	262.9	388.2	245.7	367.1	379.0	263.0	265.1	425.1	266.3	273.3	287.8	369.3	354.8	251.5
Ta	2.9	1.2	1.9	1.1	1.5	1.0	1.9	1.3	1.6	1.3	1.5	1.2	1.6	1.7
Th	10.7	13.6	9.8	10.5	10.4	10.8	11.6	9.9	11.9	9.2	10.3	12.9	11.8	11.2
U	5.3	5.3	7.0	3.6	3.1	2.2	4.5	3.3	5.6	3.6	4.0	3.3	5.8	5.7
Co	3.2	6.3	2.6	117.5	90.7	127.3	101.0	179.5	114.1	142.0	173.1	97.7	95.0	99.9
V	16.0	41.0	19.0	39.0	36.0	14.0	19.0	41.0	17.0	15.0	16.0	48.0	49.0	12.0
Zr	110.7	165.5	106.6	136.9	142.9	83.1	95.8	158.6	86.8	89.6	94.1	140.7	150.0	91.5
Y	23.0	19.6	40.7	14.1	20.3	19.0	19.1	19.4	21.7	16.0	16.9	16.5	21.5	16.7
La	18.5	36.7	18.7	22.5	27.8	19.3	19.2	32.3	20.2	14.4	17.4	27.7	28.9	17.3
Ce	38.0	75.4	35.5	45.3	57.6	39.1	37.7	65.7	39.3	29.4	34.5	56.2	60.7	36.2
Pr	4.3	7.7	4.0	5.1	6.5	4.4	4.3	7.4	4.4	3.3	3.9	6.1	6.6	3.7
Nd	15.7	26.5	15.3	19.4	23.4	17.1	15.8	26.7	16.8	11.7	14.4	23.3	25.7	14.4
Sm	3.4	4.5	3.3	3.2	4.5	3.3	3.3	5.0	3.7	2.5	3.1	4.0	4.8	3.1
Eu	0.6	0.9	0.5	0.8	0.9	0.6	0.6	1.0	0.6	0.6	0.6	0.8	0.9	0.5
Gd	3.2	3.4	3.9	2.7	3.7	3.0	3.2	4.1	3.7	2.7	2.9	3.5	4.3	2.9
Tb	0.6	0.6	0.8	0.4	0.6	0.5	0.6	0.6	0.6	0.5	0.5	0.5	0.7	0.5
Dy	3.3	3.0	4.9	2.1	3.4	3.0	3.2	3.4	3.5	2.4	2.7	2.8	3.7	3.0
Ho	0.7	0.6	1.1	0.5	0.7	0.6	0.6	0.6	0.7	0.5	0.6	0.6	0.7	0.6
Er	2.0	1.8	3.6	1.3	1.9	1.9	2.0	1.9	2.1	1.6	1.7	1.7	2.1	1.7
Tm	0.3	0.2	0.6	0.2	0.3	0.3	0.3	0.3	0.3	0.2	0.3	0.3	0.3	0.3
Yb	2.1	1.6	3.2	1.3	1.9	1.7	1.8	1.8	2.0	1.5	1.5	1.7	2.0	1.8
Lu	0.3	0.3	0.5	0.2	0.3	0.3	0.3	0.3	0.3	0.2	0.3	0.3	0.3	0.3
(La/Yb) _N	5.9	15.0	3.9	11.2	10.0	7.5	7.1	11.9	6.9	6.5	7.7	11.0	9.8	6.4
Eu/Eu*	0.5	0.7	0.4	0.8	0.6	0.5	0.6	0.7	0.5	0.7	0.6	0.7	0.6	0.5
TZr °C	754	778	752	767	769	734	742	779	734	736	741	768	775	737

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); TZr °C – zircon saturation temperatures (Watson and Harrison, 1983); lithology: e – enclave, g –granodiorite; locality: 1 – Strzeblów; 2 – Gola Świdnicka; 3 – Chwałków; 4 – Gołaszycze; * – the data for the samples marked in italics were previously published in Domańska-Siuda and Bagiński (2019)

DISCUSSION

The SSM is a granitic intrusion emplaced during the Carboniferous to Permian transition into a subduction-accretionary complex composed of various tectono-stratigraphic units, each with a distinct origin and geochemical characteristics. This complex formed at the junction of peri-Gondwana terranes during the Late Devonian to Early Carboniferous collision (e.g., Mazur et al., 2015; Jastrzębski et al., 2021; Tabaud et al., 2021; Szczepański et al., 2022 for details). As discussed by Turniak et al. (2014), the SSM is a product of latest Carboniferous–early Permian extensional magmatism within the Central European Variscides, likely driven by progressive thermal input into the crust from ascending basaltic magmas during post-orogenic lithosphere thinning. The studied granitoid massif is contemporaneous with the extensive silicic magmatism of the

Intra-Sudetic Basin and the North-Sudetic Basin (Awdankiewicz et al., 2014, 2025). According to Awdankiewicz et al. (2025), this magmatism records a transition from predominantly crust-derived to increasingly mantle-influenced sources, reflecting progressive lithospheric thinning and enhanced mantle heat flux. This regional geodynamic framework provides a broader context for the emplacement of the SSM, which may represent the deeper plutonic expression of these processes.

The petrography and geochemistry of the granitic facies within the pluton reflect diverse source regions of the SSM melts (Majerowicz, 1972; Maciejewski and Morawski, 1974; Pin et al., 1989; Puziewicz, 1990; Turniak et al., 2014). However, it remains unclear whether the mafic component of the SSM magmatic system, represented by microgranular mafic enclaves, also records this source heterogeneity.

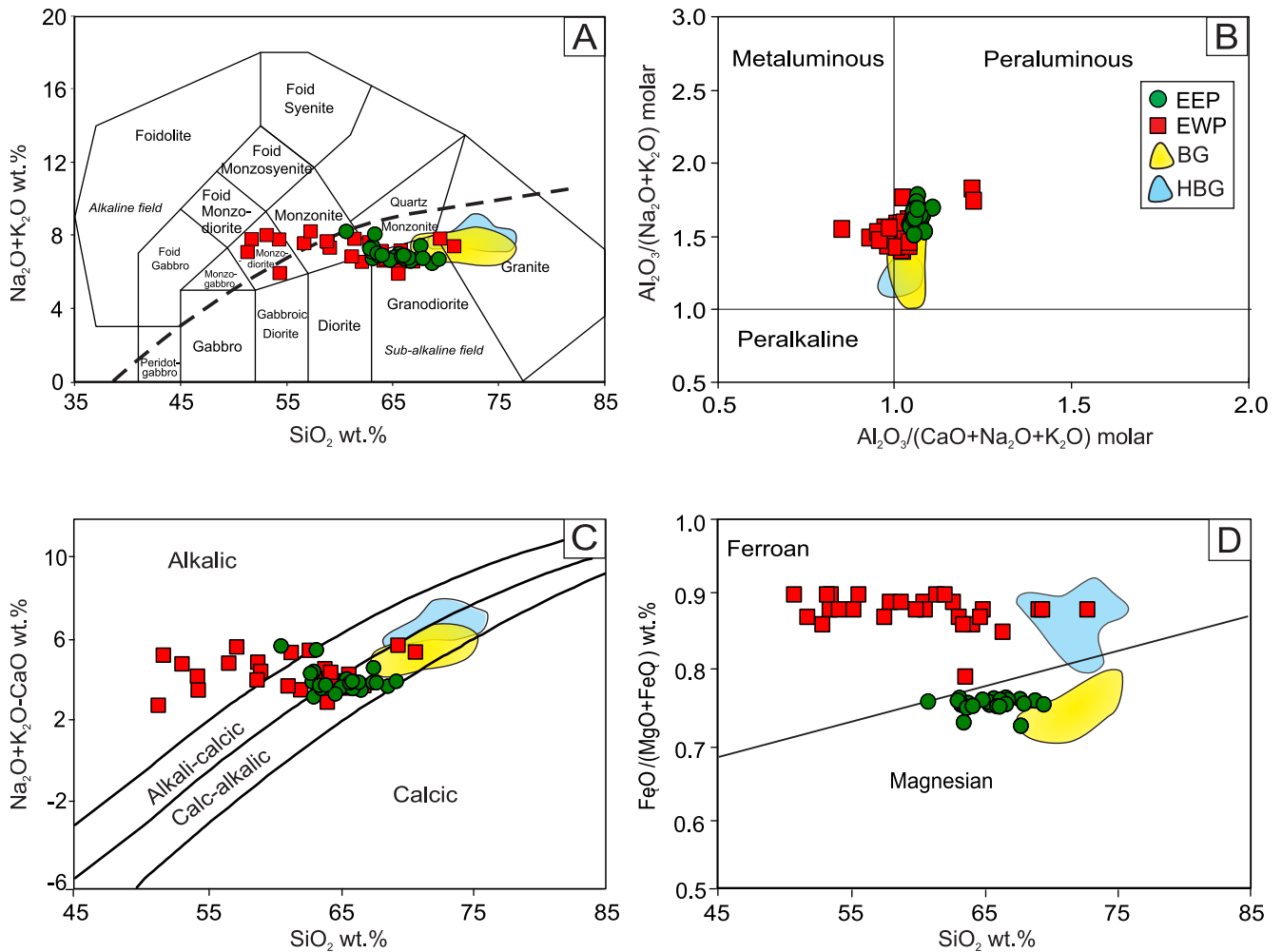


Fig. 4. Discrimination diagrams for EEP (this study) and EWP (data from Domańska-Siuda et al., 2019) compared to their host: BG (this study) and HBG (data from Domańska-Siuda et al., 2019)

A – $(\text{Na}_2\text{O} + \text{K}_2\text{O})$ vs. SiO_2 diagram (Irvine and Baragar, 1971); **B** – $\text{Al}_2\text{O}_3 / (\text{Na}_2\text{O} + \text{K}_2\text{O})$ vs. $\text{Al}_2\text{O}_3 / (\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$ molar diagram (Shand, 1943); **C** – $(\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO})$ vs. SiO_2 diagram (Frost et al., 2001); **D** – $\text{FeO} / (\text{MgO} + \text{FeO})$ vs. SiO_2 diagram (Frost et al., 2001)

Previous studies (Domańska and Słaby, 2003, 2005; Birski et al., 2018; Domańska-Siuda et al., 2019) suggested that the hornblende-biotite granite in the western part of the SSM formed by fractional crystallization, with only limited magma mixing. The parental melt was largely crustal in origin but shows evidence of a minor mantle contribution. The felsic melt was likely derived from a meta-greywacke source and may have undergone contamination during emplacement. The influence of the lithospheric mantle appears to have been mostly thermal, supplying heat to the lower crust and triggering partial melting. The parental melts of the mafic enclaves were probably generated by partial melting of amphibolitic lower crust that had previously been modified by an enriched lithospheric mantle. Hybridization likely modified the composition of the mafic magmas through interaction with fluids. The enclaves, preserved as discrete blobs within the granite, are interpreted to have formed by mingling and undercooling of mafic magma within a cooler, partially crystalline felsic host. During this process, mingling and mixing textures developed, and the enclaves underwent partial geochemical and isotopic equilibration with the more voluminous granitic melt. This equilibration was likely one-sided, as the relatively small volume of mafic melt had little influence on the composition of the host granite (Birski et al., 2018; Domańska-Siuda et al., 2019).

In contrast, Domańska-Siuda and Bagiński (2019) described mixing–mingling textures in enclaves from the eastern part of the SSM, indicating interaction between compositionally contrasting melts. Their petrographic observations suggest that the microgranular magmatic enclaves represent globules of hybrid magma formed through mingling with a more felsic host melt. However, that study did not address the origin of the enclave-forming melts.

Textural evidence indicates that the host magma in the western part of the SSM was in a highly fluid, low-crystallinity state during the injection of the mafic melt, in contrast to the eastern part, where the host granodiorite had already crystallized to near-solidus conditions. Within the hornblende-biotite granite, the mafic magma readily disintegrated into discrete blobs that became widely dispersed within the felsic crystal mush, forming enclaves. Efficient mingling increased the contact area between the two melt components, facilitating chemical transfer between the enclaves and the granitoid host. The wide compositional variability of the enclaves is consistent with such a prolonged and dynamic mingling process (Domańska-Siuda et al., 2019). In contrast, the granodioritic magma in the eastern part of the SSM had reached a more advanced stage of crystallization. Consequently, the incoming mafic melt was channelled into pre-existing fractures within the rigid,

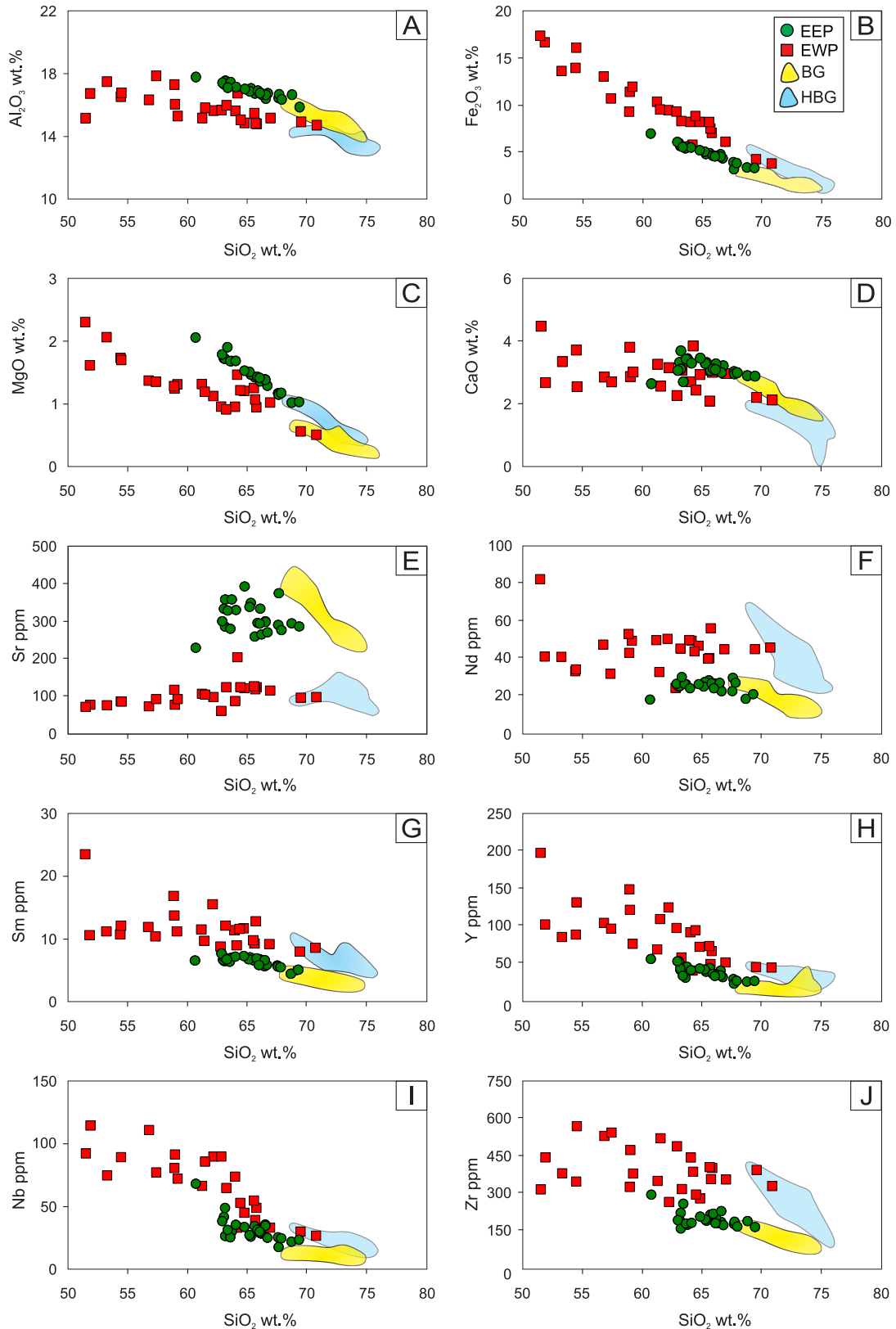


Fig. 5. Selected variations of major and trace element compositions vs. silica (wt.%) for whole-rock samples of EEP (this study) and EWP (data from Domańska-Siuda et al., 2019) compared to their hosts: BG (this study) and HBG (data from Domańska-Siuda et al., 2019)

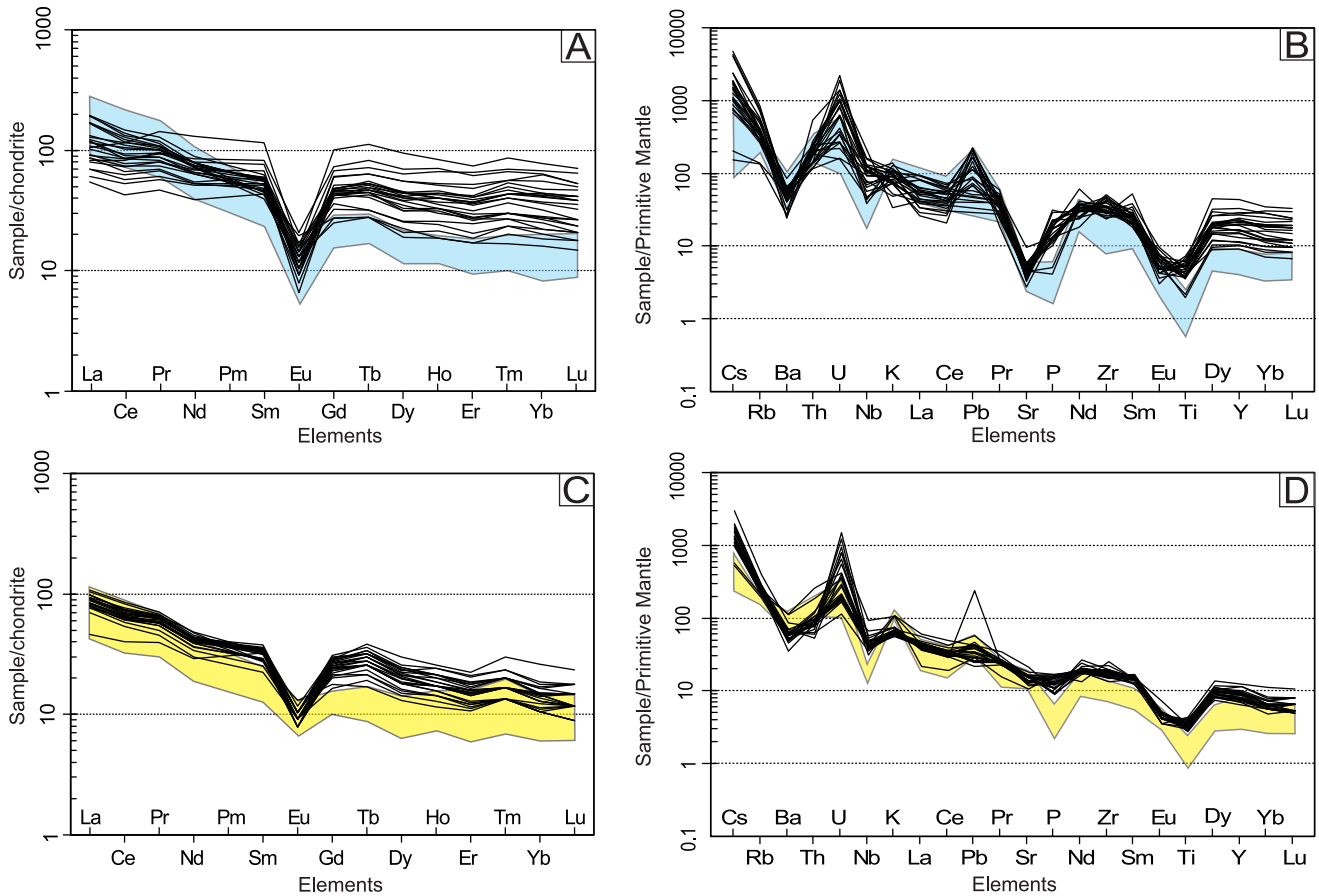


Fig. 6. Chondrite-normalized REE distribution diagrams and primitive mantle-normalized trace element spider diagrams for (A, B) EEP (this study) and (C, D) EWP (data from Domańska-Siuda et al., 2019) compared to their hosts BG (this study) and HBG (data from Domańska-Siuda et al., 2019), respectively

Chondrite and primitive mantle data were referenced from Nakamura (1974) and Sun and McDonough (1989), respectively (Fig. 6). Chondrite-normalized REE distribution diagrams and primitive mantle-normalized trace element spider diagrams for (A, B) EEP (this study) and (C, D) EWP (data from Domańska-Siuda et al., 2019) compared to their host BG (this study) and HBG (data from Domańska-Siuda et al., 2019), respectively. Chondrite and primitive mantle data were referenced from Nakamura (1974) and Sun and McDonough (1989), respectively

nearly solidified granitoid framework and interacted only locally with the remaining interstitial melt, resulting in the formation of composite or fragmented dikes. Lower temperatures in the eastern part of the SSM are independently supported by zircon saturation thermometry. Turniak et al. (2014) estimated temperatures $\sim 120^\circ\text{C}$ lower than those in the western part. Our calculated zircon saturation temperatures (T_{Zr}) for the eastern part of the massif (Table 1) and, for comparison, for the western part using data from Domańska-Siuda et al. (2019), show higher temperatures for the hornblende-biotite granite (748–860°C) than for the biotite granodiorite (702–787°C). This is consistent with the gradual heating trend of the magma source proposed by Turniak et al. (2014). Furthermore, the enclaves also yield higher T_{Zr} temperatures (EEP: 768–820°C; EWP: 800–878°C) than their respective hosts, supporting the advective heat input mechanism and mantle-derived injections suggested by Turniak et al. (2014). To compare and reconstruct the sources of the mafic melts in the western and eastern parts of the SSM, we used trace element ratio diagrams. The Th/Yb vs. Nb/Yb diagram (Pearce, 2008) is commonly used to distinguish between crustal- and mantle-derived signatures in magmas (Fig. 8). Magmas that have interacted with continental crust during ascent or possess a subduction-related component typically have elevated Th/Yb values. Nearly all enclave

samples plot outside the mantle array. EWP plot between the lower- and upper-crust fields and overlap the field occupied by the host hornblende-biotite granite. EEP, characterized by slightly higher Nb/Yb ratios, are shifted towards the ocean island basalt field but still fall between the upper- and lower-crust fields and overlap the field of their granodioritic host. This distribution is consistent with magma mixing processes that modified the primary compositions of the enclaves (Birski et al., 2018; Domańska-Siuda et al., 2019; Domańska-Siuda and Bagiński, 2019). During ascent, the mafic melts may also have become contaminated by crustal material, further modifying their primary composition. Consequently, the present compositions of the enclaves preserve only a weak mantle signature, possibly limited to the thermal or fluid contribution of mantle-derived melts (Domańska-Siuda et al., 2019).

Biotite chemistry provides further insight into the crystallization environment. The $\text{FeO}/[\text{FeO}+\text{MgO}]$ vs. MgO plot for biotites from both EWP and EEP (Fig. 9) indicates crystallization from crustally derived host melts (Zhou, 1986). However, biotite from EEP are shifted towards compositions indicative of crystallization from a relatively more mafic melt than that from EWP.

Table 2

Whole rock Sr and Nd isotopic data of EEP and BG host

Sample	Locality	Rb ppm	Sr ppm	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr	2σ (%)	⁸⁷ Sr/ ⁸⁶ Sr (300 Ma)	Sm ppm	Nd ppm	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd	2σ (%)	¹⁴³ Nd/ ¹⁴⁴ Nd (300 Ma)	εNd (300 Ma)	T _{CHUR1} (Ga)	T _{DM1} (Ga)	T _{DM2} (Ga)
biotite granodiorite																	
021	1	110.3	229.8	1.3892	0.71129	17	0.70536	2.9	12.2	0.14374	0.512374	11	0.512092	-3.1	0.76	1.45	1.27
023	1	114.7	268.3	1.2373	0.71117	20	0.70589	3.0	15.4	0.11780	0.512385	16	0.512154	-1.9	0.49	1.05	1.17
024	1	138.0	262.9	1.5193	0.71199	15	0.70550	3.4	15.7	0.13096	0.512385	12	0.512128	-2.4	0.59	1.21	1.21
120	2	—	—	—	—	—	—	4.5	26.5	0.10269	0.512358	14	0.512156	-1.9	0.46	0.95	1.17
128	3	107.3	245.7	1.2640	0.71122	18	0.70582	3.3	15.3	0.11369	0.512372	11	0.512149	-2.0	0.49	1.03	1.18
SS146	4	—	—	—	—	—	—	4.95	26.7	0.11211	0.512382	9	0.512161	-1.8	0.46	1.00	1.16
enclaves																	
122	1	189	330	1.6581	0.712558	20	0.70548	6.7	25.6	0.1583	0.512472	12	0.512161	-1.8	0.66	1.56	1.16
123	1	157	256	1.7739	0.712945	11	0.70537	6.6	28.7	0.1391	0.512433	10	0.512160	-1.8	0.54	1.24	1.16
124	1	150	297	1.4633	0.712074	21	0.70583	5.7	24.4	0.1413	0.512433	15	0.512156	-1.9	0.57	1.28	1.17
125-A	1	165	355	1.3478	0.711543	19	0.70579	7.0	26.6	0.1591	0.512451	12	0.512138	-2.2	0.76	1.65	1.19
125-B	1	119	288	1.1980	0.711104	15	0.70599	5.8	22.9	0.1532	0.512428	9	0.512127	-2.4	0.74	1.54	1.21
161	1	—	—	—	—	—	—	6.6	18.4	0.2169	0.512550	10	0.512124	-2.5	—	—	1.20
SS156	1	—	—	—	—	—	—	5.1	21.3	0.1459	0.512390	11	0.512104	-2.9	0.75	1.46	1.25

The isotopic data (Fig. 7 and Table 2) provide additional insights into the evolution of the enclaves. EWP show negative εNd₍₂₉₈₎ values ranging from -4.5 to -2.5, and highly variable initial ⁸⁷Sr/⁸⁶Sr ratios between 0.7079 and 0.7113. These values broadly overlap those of the host hornblende-biotite granite (⁸⁷Sr/⁸⁶Sr₍₂₉₈₎ ratios 0.7070–0.7094; εNd₍₂₉₈₎ values: -5.4 to -2.4) (Domańska-Siuda et al., 2019). The pronounced Sr-isotopic heterogeneity of the enclaves may reflect contamination by ⁸⁷Sr-rich components during magma ascent, or alternatively, derivation from source regions characterized by variable time-integrated Rb/Sr ratios. In contrast, EEP display εNd₍₃₀₀₎ values between -1.8 and -2.9, and lower initial ⁸⁷Sr/⁸⁶Sr ratios (0.7054–0.7060), closely matching those of their biotite granodiorite host (⁸⁷Sr/⁸⁶Sr₍₃₀₀₎: 0.7054–0.7059; εNd₍₃₀₀₎: -1.8 to -3.1). These isotopic characteristics suggest that the biotite granodiorite and its enclaves evolved from a distinct magma batch derived from a source with lower time-integrated Rb/Sr and Nd/Sm ratios, or alternatively experienced less crustal contamination than the hornblende-biotite granite.

Overall, the isotopic compositions of the enclaves closely parallel those of their granitic hosts. Granitoids with the least radiogenic Sr and highest εNd contain enclaves with similarly low ⁸⁷Sr/⁸⁶Sr and more radiogenic Nd, and vice versa (Fig. 7). Such relationships suggest that isotopic equilibration occurred during magma mingling and mixing at the level of emplacement (e.g., Pin et al., 1990; Fourcade and Javoy, 1991; Barbarin, 2005), effectively overprinting the original isotopic signatures of the enclave-forming magmas. Similar processes involving melt interaction, diffusion-controlled isotopic equilibration and chemical exchange between compositionally contrasting magmas have been documented in other Variscan granitoid systems in the Sudetes (Pietranik and Waight, 2008), supporting the interpretation that prolonged mingling at high temperatures can lead to near-complete isotopic homogenization of mafic enclaves within felsic hosts. Field observations confirm that such mingling was particularly effective in the western part of the SSM. There, mafic magma was injected into a largely molten felsic melt, where it disintegrated into blobs that formed enclaves. This produced an extensive interface between the mafic and felsic magmas, facilitating prolonged chemical and isotopic exchange. The pronounced compositional variability of EWP supports such long-lived interaction (Domańska-Siuda et al., 2019), while the elevated magmatic temperatures in the western SSM (~120°C higher than in the eastern part; Turniak et al., 2014 and this study) would have further enhanced diffusion and isotopic equilibration. In contrast, the mafic magma forming EEP was emplaced into the eastern SSM after the host granodioritic magma had reached a much more advanced stage of crystallization. It was channelled into fractures within a nearly solidified granitoid framework and interacted only locally with the remaining residual melt. Under such conditions, large-scale isotopic homogenization would have been limited. Therefore, the isotopic similarity between EEP and their host granodiorite is more likely to reflect derivation from a common juvenile, isotopically homogeneous source than secondary equilibration during magma mingling.

Despite similarities between the enclaves and their respective granitic hosts, EWP and EEP differ from one another isotopically and define two distinct geochemical groups (Fig. 7). In both the Sm–Nd and Rb–Sr isotope systems, EWP display greater geochemical variability, with highly variable Rb (88–546 ppm) and Nd (25–83 ppm) concentrations and relatively low Sr contents (58–201 ppm). In contrast, EEP are characterized by a narrower range of Rb (110–265 ppm) and Nd (18–30 ppm) concentrations and consistently higher Sr contents (227–390 ppm; Table 2 and Domańska-Siuda et al., 2019). Likewise,

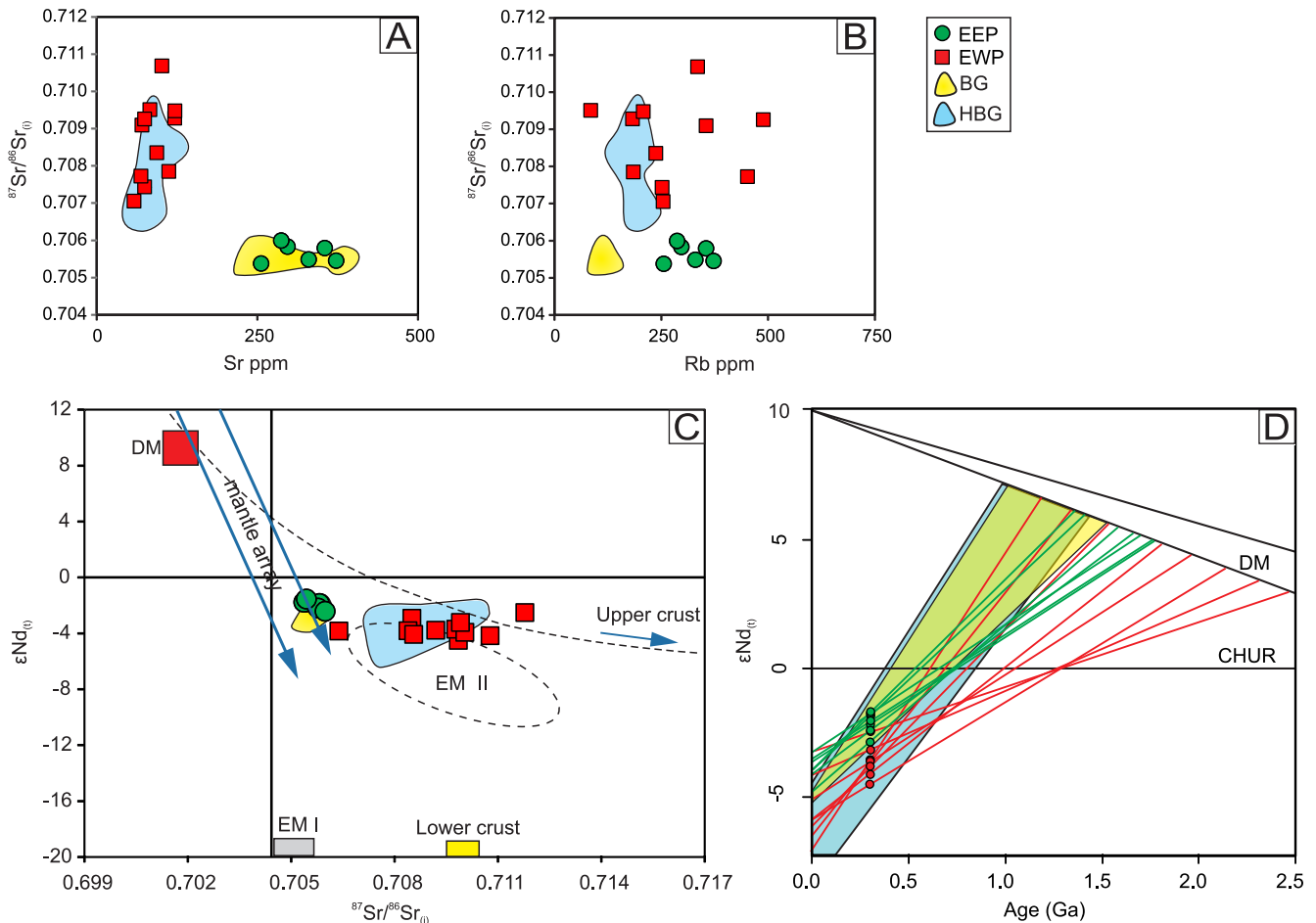


Fig. 7. Diagrams (A) $^{87}\text{Sr}/^{86}\text{Sr}_0$ vs. Sr (ppm), (B) $^{87}\text{Sr}/^{86}\text{Sr}_0$ vs. Rb (ppm) and (C) $^{87}\text{Sr}/^{86}\text{Sr}_0$ vs. ϵNd_0 and (D) ϵNd_0 vs. TDM model age (Ga) showing evolution lines for EEP (this study) and EWP (data from Domańska-Siuda et al., 2019)

CHUR – Chondritic Uniform Reservoir; DM – Depleted Mantle (DePaolo and Wasserburg, 1979; DePaolo, 1981). Fields for Enriched Mantle components (EMI, EMII) are plotted according to Zindler and Hart (1986), and the Lower Crust values are according to Rudnick and Gao (2003)

EWP display a wide range of $^{87}\text{Rb}/^{86}\text{Sr}$ ratios (2.97–21.76), while EEP are characterized by lower and more uniform values (0.92–1.77; Fig. 7A, B). If isotopic equilibration with the host granitoids occurred during emplacement, the remaining isotopic differences between EWP and EEP most likely reflect differences in the source regions of their felsic host melts. This interpretation is consistent with the conclusions of Pin et al. (1988, 1989), who documented pronounced isotopic heterogeneity within the SSM and attributed it to inheritance from crustal protoliths of contrasting compositions. They proposed that the biotite granodiorite, with low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and less negative ϵNd values, was emplaced into a crustal block containing abundant remnants of oceanic lithosphere, such as fragments of the Śląska ophiolite. The relatively low model ages (T_{DM} : 1.24–1.65 Gyr) obtained for the biotite granodiorite and EEP (Fig. 7D) are typical of many Variscan granitoids in Europe (Liew and Hofmann, 1988) and likely represent mixed model ages rather than the age of a single source reservoir. Although these values could formally indicate a contribution from remelting of Mesoproterozoic juvenile mafic crust, they are more plausibly interpreted as reflecting a mixed source comprising ancient crustal material and a juvenile mantle-derived component (Liew and Hofmann, 1988; Gawęda et al., 2019). Subsequent contamination by upper crustal material

during magma ascent or emplacement could account for the negative ϵNd values, while interaction with enriched mantle-derived fluids may have further modified the isotopic signatures.

In contrast, the hornblende-biotite granite and EWP are characterized by higher T_{DM} ages (up to 2.40 Gyr), elevated initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, and strongly negative ϵNd values (Domańska-Siuda et al., 2019). These features indicate derivation from a source region dominated by evolved continental crust. Based on the isotopic data of Pin et al. (1989), Puziewicz (1990) suggested that the parental melt was derived from metapelitic, greywacke, or silicic volcanic protoliths. He further proposed that the source region was located within the eastern prolongation of the Kaczawa Unit, which comprises Cambrian to Lower Carboniferous metavolcanic–metasedimentary successions (Baranowski et al., 1990; Kryza and Muszyński, 1992; Kryza and Zalasiewicz, 2008). Domańska-Siuda et al. (2019) similarly concluded that crustal material with greywacke-like chemical affinities represented the most probable source of the granitoid magmas. These interpretations show notable similarities to the inferred source characteristics of the Permian intermediate volcanic rocks in the North-Sudetic Basin. According to Awdankiewicz et al. (2006), these volcanic rocks represent variably evolved high-K calc-alkaline magmas emplaced in a post-collisional extensional setting. Their geoche-

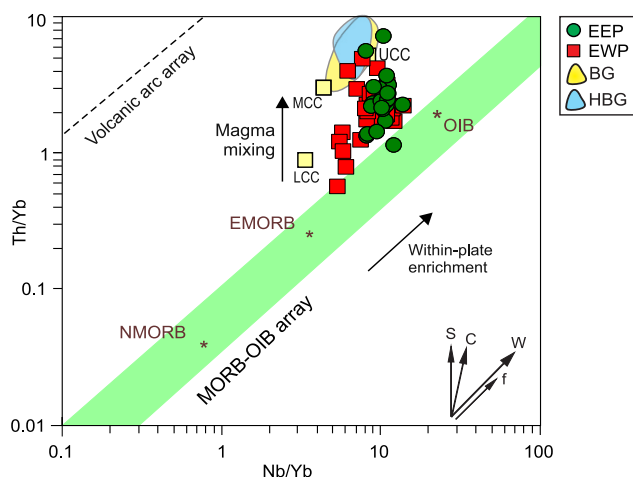


Fig. 8. Th/Yb vs. Nb/Yb diagram (Pearce, 2008) for EEP (this study) and EWP (data from Domańska-Siuda et al., 2019)

Reference values for average N-MORB (normal mid-ocean ridge basalt), E-MORB (enriched mid-ocean ridge basalt), and OIB (ocean island basalt) are plotted after Sun and McDonough (1989). Compositions of LCC, MCC, and UCC (lower, middle, and upper continental crust, respectively) are according to Rudnick and Gao (2003). The bold arrows in the bottom right are: S – subduction component, C – crustal contaminant component, W – within plate, and f – fractional crystallisation vectors

mical evolution involved fractional crystallization, repeated mafic replenishment, and assimilation of crustal material with greywacke-like compositions. The trace-element characteristics (e.g., high Th/Nb and La/Nb ratios) suggest derivation from an enriched lithospheric mantle source carrying a crustal signature, possibly inherited through subduction-related metasomatism during the Variscan Orogeny.

The geochemical similarities, together with the broadly synchronous timing of magmatism (Awdankiewicz et al., 2014, 2025), suggest that the SSM and the Permian volcanics formed within a common tectono-magmatic framework. Although the granitoids crystallized at deeper crustal levels, the comparable source characteristics of both suites are consistent with melting of a heterogeneous lithospheric domain that had been metasomatized during the Variscan Orogeny and was subsequently reactivated during post-orogenic extension.

CONCLUSIONS

Geochemical and isotopic data indicate that the granitoids and associated mafic microgranular enclaves in the western and eastern parts of the SSM originated from two distinct magma source domains that record contrasting crustal evolution.

1. The hornblende-biotite granite and EWP are characterized by negative $\epsilon\text{Nd}_{(298)}$ values, relatively high and variable initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, and variable depleted mantle model ages (T_{DM}), indicating derivation from evolved continental crust.

2. The isotopic characteristics of EWP closely match those of their host granite, consistent with extensive chemical and isotopic equilibration during prolonged magma mingling and mixing at the level of emplacement.

3. The isotopic characteristics of the hornblende-biotite granite and EWP indicate derivation from distinct crustal sources with greywacke- and amphibolite-like affinities, respectively.

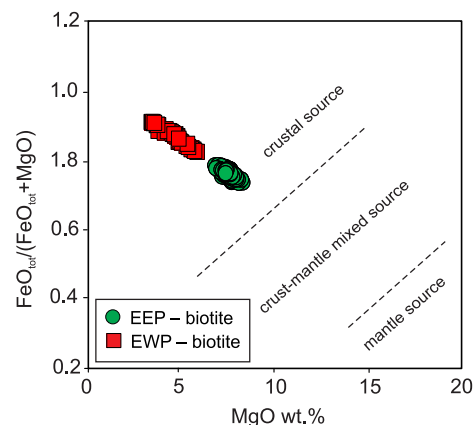


Fig. 9. $\text{FeO}_{\text{tot}}/(\text{FeO}_{\text{tot}}+\text{MgO})$ vs. MgO diagram (Zhou, 1986) plotted for biotite from EEP (this study) and EWP (data from Domańska-Siuda et al., 2019) biotites which depicts the source for mafic magma

4. The biotite granodiorite and EEP are characterized by lower and more uniform initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, less negative $\epsilon\text{Nd}_{(300)}$ values, and younger, more homogeneous T_{DM} ages than their western counterparts.

5. The isotopic similarity between EEP and the host biotite granodiorite likely reflects derivation from a common juvenile hybrid source rather than extensive isotopic equilibration during magma mingling.

6. The geochemical and isotopic characteristics of the biotite granodiorite and EEP indicate melting of a hybrid source containing both ancient crustal material and juvenile mantle-derived components, followed by limited upper-crustal contamination during magma ascent or emplacement. A contribution from enriched mantle-derived fluids is also possible.

7. Collectively, these results demonstrate that the SSM was derived from isotopically heterogeneous crustal source regions, reflecting a complex tectono-magmatic evolution during the late stages of the Variscan orogeny and post-Variscan lithospheric thinning.

8. The geochemical similarities between the SSM and Permian intermediate volcanic rocks of the North-Sudetic Basin indicate that both magmatic suites formed within a common post-collisional tectono-magmatic framework related to melting of a heterogeneous, sub-continental lithosphere that was metasomatized during the Variscan orogeny and subsequently reactivated during post-orogenic extension.

Acknowledgements. The authors would like to express their sincere gratitude to the Editor and the reviewers (Aleksandra Gawęda, Ewa Krzemińska and Ewa Słaby) for their constructive and insightful comments, which significantly improved the quality of this manuscript. This study was supported by a research grant no. PSP 501-D113-20-0004316 (BOB-661-1240/2025), under the framework of the project Excellence Initiative – Research University (IDUB), at the University of Warsaw to JDS and a research grant no. NN307 112735, from the Ministry of Science and Higher Education of Poland (MNiSW) to KT.

Declarations

Conflict of interest

All Authors declare their consent to publish the results and are not aware of any financial or non-financial interests that are directly or indirectly related to the work submitted for publication.

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APPENDIX 1

Representative wavelength dispersive of plagioclase; based on 8 oxygen

EEP

[illegible]

APPENDIX 2

Representative wavelength dispersive of K-feldspar; based on 8 oxygen

EEP

Sample	157	157	157	157	157	161	161	161	161	161	161	161	161	161	161	161	161	161	161	161
Profile	1/1	1/2	1/3	1/4	1/5	1/1	1/2	1/3	1/4	2/1	2/2	2/3	3/1	3/2	3/3	3/4	4/1	4/2	4/3	4/4
wt. %																				
SiO ₂	64.79	64.76	65.16	65.03	64.86	64.38	64.54	65.98	64.21	64.89	64.44	64.50	64.29	64.59	64.73	64.79	64.53	64.29	64.17	64.55
Al ₂ O ₃	18.41	18.41	18.58	18.57	18.51	18.59	18.39	18.88	18.35	18.56	18.41	18.30	18.15	18.33	18.63	18.51	18.52	18.57	18.46	18.54
CaO	0.06	0.05	0.04	0.03	0.08	0.26	0.10	0.02	0.04	0.07	0.02	0.00	0.04	0.08	0.11	0.00	0.00	0.00	0.02	0.00
FeO	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.01	0.01	0.02	0.00	0.00	0.00	0.00
Na ₂ O	0.75	0.60	1.00	0.80	0.88	0.81	0.78	1.85	0.66	1.82	1.11	0.79	0.39	1.00	1.12	0.27	1.22	1.27	0.68	1.03
K ₂ O	15.88	16.05	15.30	15.68	15.21	15.69	15.92	13.96	16.20	14.70	15.58	16.13	16.74	16.22	15.24	16.38	15.05	15.15	16.27	15.66
total	99.89	99.87	100.09	100.12	99.55	99.72	99.73	100.69	99.47	100.03	99.55	99.71	99.62	100.22	99.84	99.97	99.33	99.28	99.59	99.77
Formula																				
Si ⁴⁺	3.00	3.00	3.00	3.00	3.00	2.98	2.99	3.00	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.98	2.98	2.98
Al ³⁺	1.00	1.00	1.01	1.01	1.01	1.01	1.00	1.01	1.01	1.01	1.01	1.00	1.00	1.00	1.01	1.01	1.01	1.01	1.01	1.01
Ca ²⁺	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Fe ²⁺	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na ⁺	0.07	0.05	0.09	0.07	0.08	0.07	0.07	0.16	0.06	0.16	0.10	0.07	0.04	0.09	0.10	0.02	0.11	0.11	0.06	0.09
K ⁺	0.94	0.95	0.90	0.92	0.90	0.93	0.94	0.81	0.96	0.86	0.92	0.96	0.99	0.96	0.90	0.97	0.89	0.90	0.96	0.92
total	5.01	5.00	4.99	5.00	4.98	5.00	5.01	4.98	5.02	5.02	5.02	5.02	5.02	5.04	5.00	4.99	5.00	5.00	5.02	5.01
An	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.10	0.10	0.00	0.10	0.00	0.00	0.00	0.00
Ab	6.77	5.39	9.02	7.24	8.09	7.31	6.92	16.77	5.87	15.80	9.78	6.92	3.40	8.59	10.03	2.42	11.01	11.29	5.95	9.06
Or	93.23	94.61	90.98	92.76	91.91	92.69	93.08	83.23	94.13	84.20	90.22	93.08	96.50	91.32	89.97	97.47	88.99	88.71	94.05	90.94

APPENDIX 3

Representative wavelength dispersive of biotite; based on 10 oxygen

EEP

n.a. - not analyzed