

Indications of HP events in the volcanosedimentary succession of the Orlica–Śnieżnik Dome, NE Bohemian Massif: data from a marble-amphibolite interface

Maksymilian TWYRDY¹, * and Andrzej ŻELAŻNIEWICZ¹

¹ Institute of Geological Sciences, Polish Academy of Sciences, Research Center in Wrocław, Podwale 75, 50-449 Wrocław, Poland



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A volcanosedimentary succession of the Młynowiec–Stronie Group (MSG) in the Orlica–Śnieżnik Dome (OSD), the Sudetes, NE Bohemian Massif underwent multiple folding and shearing during the Variscan Orogeny. In the sheared domains, there are less deformed pods in which rocks preserve better records of metamorphic events prior to the regional temperature peak. In one such pod, near Gniewosów on the western limb of the dome, marbles enclosed by massive amphibolites occur. In these rocks, zoned plagioclase with actinolite and epidote inclusions and zoned amphibole grains allowed recognition of three mineral assemblages and three P-T stages at: (1) 310°C/3–4 kbar, (2) 480–500°C/10.5 kbar, (3) 500–530°C/6–6.5 kbar, based on isopleth intersections and checked against conventional thermobarometry. These define a steep clockwise P-T path and a geothermal gradient of 17°C/km before peak conditions were attained, which suggests subduction of the metavolcano-sedimentary rocks (Stronie Formation of the MSG) on the western limb of the OSD, with a transient yet discrete higher pressure episode. Mineral relicts capable of demonstrating a higher pressure event are scarce in the supracrustal rocks of the dome, mainly because they became more thoroughly equilibrated and obliterated during the temperature peak at mid-amphibolite facies conditions and the subsequent ubiquitous greenschist facies overprint.

Key words: high-pressure episode, P-T path, marble-amphibolite association, Sudetes.

INTRODUCTION

The Orlica–Śnieżnik Dome (OSD) is a Variscan tectonostratigraphic unit of the NE Bohemian Massif (Fig. 1). It either occurs in the hanging wall of a terrane suture that welded Saxothuringia and Brunovistulia which was in the footwall (Żelaźniewicz et al., 2009; Chopin et al., 2012; Jastrzębski et al., 2015b), or is assigned to the Moldanubian Terrane (Skácel, 1989; Matte et al., 1990; Schulmann and Gayer, 2000). In the eastern part of the OSD, close to the suture that coincides with the Stáre Město Belt, there are eclogite and granulite bodies tectonically or diapirically inserted in gneisses which occupy the core of the dome (Brueckner et al., 1991; Kryza et al., 1996; Bröcker et al., 2009; Štípská et al., 2012). Analysed using several isotopic methods, the HP rocks yielded almost exclusively ages of ~350–330 Ma, consistent, within error, with the timing of the peak of the MP-MT metamorphic and deformational event recognized in the host core gneisses as well as in felsic meta-volcanic rocks and metapelitic schists of the Młynowiec–Stronie

Group (MSG) that mantle the core of the dome (review in Żelaźniewicz et al., 2014). The age data apparently assign the (U)HP rocks to the main tectonometamorphic episode. Such an assignment would suggest that all rocks in the OSD were metamorphosed in a short time span yet at dramatically different crustal/lithosphere depths. This observation requires that an intense tectonism brought rocks with various P-T signatures to their present attitudes in the same short time span, hence, in tectonic terms, almost instantaneously. Although, in the western part of the dome, no (U)HP rock was known, Faryad and Kachlík (2013) were able to find there mica schists in which muscovite flakes contain phengitic cores indicative of pressure between 11 kbar and 20 kbar. Such high pressure must have affected the schists prior to the main regional metamorphic episode under medium amphibolite facies conditions. The time gap between the two episodes remains undetermined in case of the schists and is at least unclear in the case of the eclogites and granulites (Anczkiewicz et al., 2007; Bröcker et al., 2009). The discovery made by Faryad and Kachlík (2013) and its tectonic consequences need confirmation because so far no estimate of P-T conditions made for various rocks of the Młynowiec–Stronie Group throughout the OSD has revealed pressures in excess of 10–11 kbar (Murtezi, 2006; Jastrzębski, 2009; Szczepański, 2010; Skrzypek et al., 2011a, b, 2014; Ilnicki et al., 2013; Jastrzębski et al., 2015a). However, this might be due to selective obliteration of (U)HP mineral assem-

* Corresponding author, e-mail: maksymilian.twyrdy@twarda.pan.pl

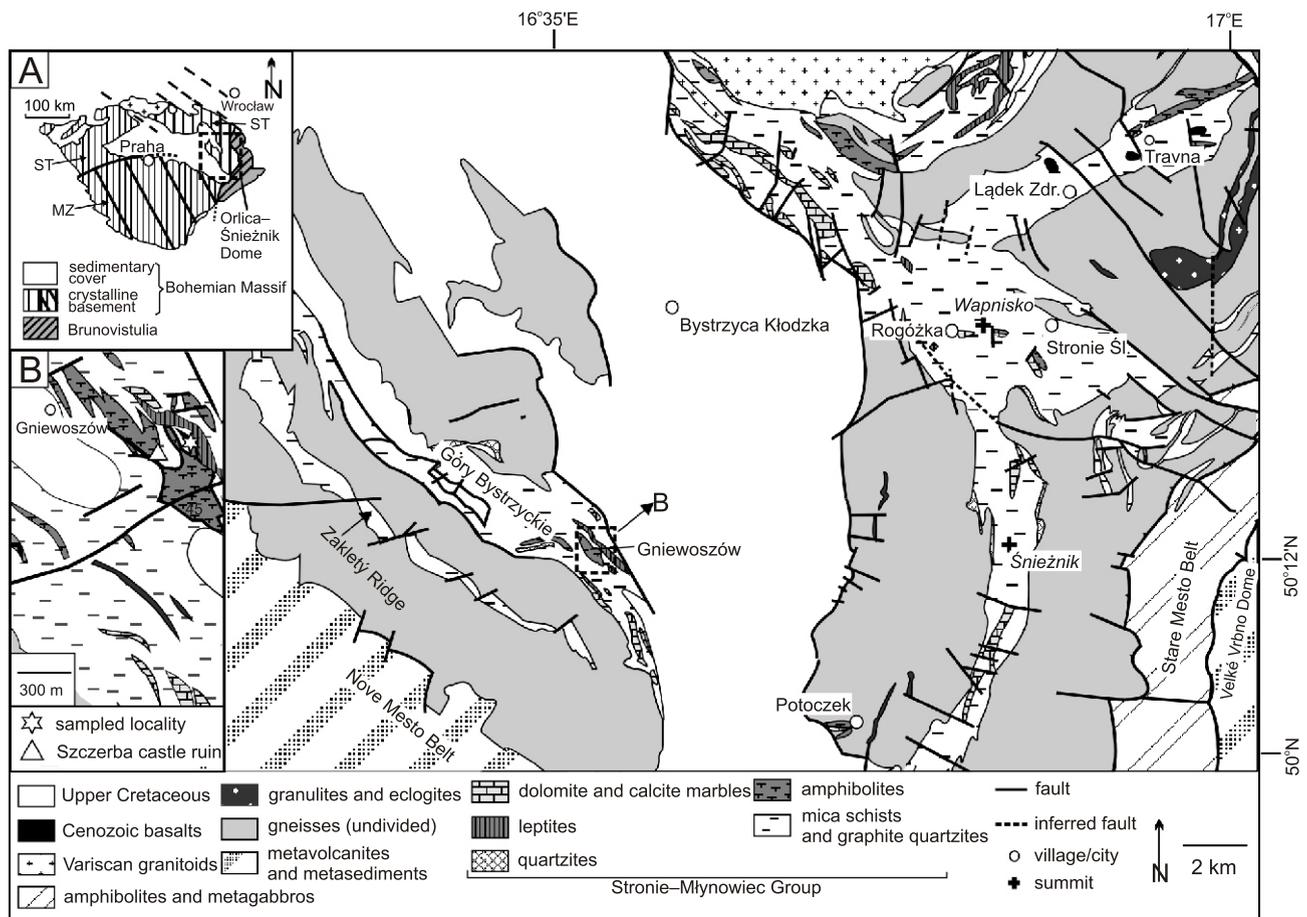


Fig. 1. Geological sketch of the Orlica-Śnieżnik Dome after Żelaźniewicz et al. (2014, simplified)

A – location in the Bohemian Massif, MZ – Moldanubian Zone, ST – Saxothuringian Zone;
 B – details of the study area after Kozdrój (1990, modified)

blage(s) during later metamorphic processes under an amphibolite and greenschist facies overprint. Such a possibility is to be tested in supracrustal rocks (MSG) throughout the OSD. In this study, we decided to perform this test on amphibolites that show a bluish shade hinting at the presence of Na-amphibole, which were found by us on the western limb of the dome as intercalations in mica schists. The amphibolites enclose marble lenses and the interface between the two rocks is layered, the layers being mafic impurities in the marbles. We here seek relic records of possible earlier HP event(s), determine the P-T conditions under which the rocks were metamorphosed, discuss implications for the evolution of the OSD, and decipher the origin of the mafic-carbonate associations.

GEOLOGICAL SETTING

In the OSD, supracrustal rocks – mica schists with subordinate bimodal metavolcanic rocks, marble and quartzites – are distinguished as the Młynowiec-Stronie Group of Ediacaran/Cambrian to Ordovician age (review in Żelaźniewicz et al., 2014). Mica schists defined as the Stronie Formation (SF) in this group contain both acid and basic metavolcanic rocks and lenses of marbles tens of centimetres up to a few tens of metres thick (Fig. 1). The marbles were studied by Koszela (1997) who showed that the prevailing pelitic sedimentation in the SF locally gave way to carbonate sedimentation accompanied by episodic

underwater lava extrusions and pyroclastic ash deposition. A generalized lithostratigraphic profile of the SF reveals that the marbles usually overlie amphibolites (Koszela, 1997). Indeed such a position suggests that carbonate sedimentation followed volcanic eruptions and cyclically active volcanoes were likely overgrown by carbonate reefs.

On the western limb of the OSD, in the Bystrzyckie Mts. (Fig. 1), near the village of Gnievoszów, mica schists contain intercalations of metavolcanogenic rocks. They appear as lenticular bodies of schistose to massive quartz-alkali feldspar-mica metarhyolites several kilometers long, dated at ca. 500 Ma (Murtezi, 2006; Mazur et al., 2015), and hornblende-plagioclase-epidote metabasalts which often occur close to calcite marbles (Fig. 2). The metabasalts locally preserve identifiable pillow structures typical of underwater extrusions (Ilnicki et al., 2013). They are represented by medium- to fine-blastic, moderately foliated metabasalts.

ANALYTICAL METHODS

The amphibolites and marble lenses were studied in thin section under a polarizing microscope and using a JEOL JSM/JXM 840A machine at the Institute of Geological Sciences, PAS, equipped with an EDS Thermo-NORAN phase identifier (acc. voltage 15 kV during data acquisition). Structural formulas of minerals were calculated using CALCMIN software

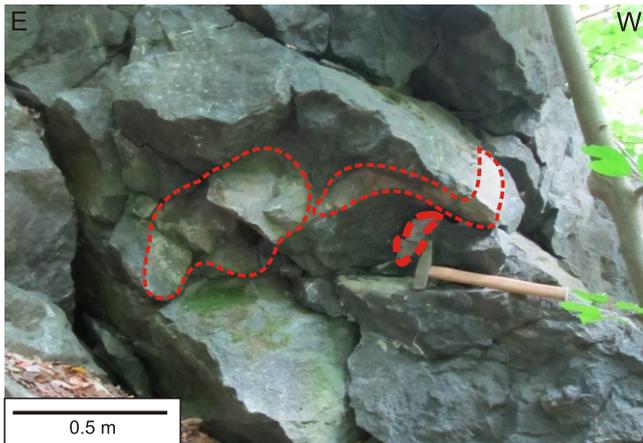


Fig. 2. Marble lenses (dotted outlines) within massive amphibolite

Abandoned quarry near Szczerba castle, Gniewosów

(Brandelik, 2009). Mineral abbreviations follow those of Whitney and Evans (2010). Amphibole microprobe analyses were recalculated to 23 O according to the method recommended by IMA97 with calibration after Esawi (2004) most accurate for low-Ti amphiboles (e.g., Schumacher, 2007). Whole-rock analysis of massive amphibolite was performed at the Acme Analytical Laboratories, Vancouver, according to the standard procedure LF600 of the lab. Conventional thermobarometric calculations were done by methods and calibrations proposed for amphibole-plagioclase pairs by Plyusnina (1982), Holland and Blundy (1994), Anderson and Smith (1995), Bhadra and Bhattacharya (2007), and using XFe and XAl in amphiboles (Ridolfi et al., 2010), or XTr, XTsch, XParg in amphiboles (Gerya et al., 1997; Zenk and Schulz, 2004). In addition, the thermodynamic modelling was performed using *Theriak Domino* software (Capitani and Petrakakis, 2010) with an internal database after Berman (1988; Jun92). The applied calibration used an oxygen fugacity buffer (fO_2) and was based on estimation of the dehydrogenisation rate which can be determined by comparison of calculated sum of cations with that of the bulk rock composition [$fO_{2act} = X_{calc}(pfu) - 23 O$].

DESCRIPTION OF THE ROCKS STUDIED

In an abandoned quarry near Szczerba castle (50°19'58"N, 16°61'96"E), structurally 50 m above the pillow metabasites, massive amphibolites enclose several irregular lenses of white or grey marble which are a few cm to 1 m long and 15–30 cm thick (Fig. 2). Within such marble lenses, there are several 1–3 mm thick dark bands arranged parallel to the lenses' longer edges. The carbonate rocks were likely incorporated into original basalts. A contact between amphibolites, mica schists and schistose metarhyolites has a subvertical attitude, being oriented at a high angle to the dominant foliation in these rocks, which, along with small-scale structures, points to the presence of large-scale overturned folds in the area. Pillows that are discernible in medium- to fine-blastic, moderately foliated metabasalts are flattened parallel to the SW-wards dipping foliation and elongated in the WNW–ESE direction, consistent with those folds.

¹ Ps ratio = mole % of the pistacite molecule ($Ca_2(Fe,Al)_3(SiO_4)_3(OH)$) in epidote; the composition of epidote is expressed in terms of the theoretical pistacite end-member, $Ps = 100 \frac{Fe^{3+}}{(Fe^{3+} + Al)}$

AMPHIBOLITES

Next to the layered margin of marble lenses, the massive amphibolite is a granoblastic to porphyroblastic, poorly foliated rock. It is composed of (1) subhedral or anhedral needles of Mg-hornblende and (2) mainly anhedral albite blasts which carry actinolite and rutile inclusions. Many plagioclase blasts lack inclusions in their outer parts, which imparts a zonal appearance to them (Fig. 3A, B). Microprobe analyses confirm the compositional zonation of such blasts and reveal that they have albite cores ($An_{1-1.5}$) but oligoclase (An_{25-27}) rims. In other grains, the rims remain albitic yet with slightly higher An content ($An_{1.5-2.5}$). There are also accessory calcite, epidote, apatite, ilmenite, magnetite, chlorite and titanite blasts, the latter containing single inclusions of rutile (Fig. 3C). These minerals are often concentrated immediately next to Al-rich Mg-hornblende blasts (Fig. 3D–F) characterized by bluish-green pleochroism and compositional zonation due to increasing Al content towards the rims of the blasts.

In contrast, amphibolites derived from pillow lava contain zoned Ca-amphibole blasts with actinolitic cores but tschermakitic rims. Plagioclase blasts have albitic cores and up to andesine rims (An_{23-33}). In the albitic cores, actinolite and ilmenite inclusions are in evidence (Ilnicki et al., 2013; Szczepański and Ilnicki, 2014).

MARBLES

The marbles are composed mainly of medium- to fine-blastic calcite (~70 vol.%, estimation made using alizarin-S pigment) and subhedral, medium-blastic albite, often with polysynthetic twinning. The albite forms aggregates with quartz (ca. 20 vol.%), 1–2 mm across. Epidote, titanite, actinolite, Mg-hornblende and opaque minerals appear as minor phases (10 vol.%). Calcite blasts with well-developed twinning contain randomly distributed inclusions of actinolite, quartz, albite and rarely magnetite and ilmenite. In contrast, albite blasts contain abundant inclusions of epidote, actinolite and calcite (Fig. 4), mostly randomly distributed, or, much more rarely, arranged to form a kind of foliation parallel to the longer edges of the marble lenses. Interestingly, dark arrays at these edges are composed of bimineralic actinolite-epidote aggregates, 1–2 mm long and roughly oval in shape.

LAYERED MARGINS OF MARBLE LENSES

The marble lenses possess more or less regular layers, up to a few cm thick, close to the boundary with the surrounding massive amphibolites, which makes the lens margins layered or striped (Fig. 5). Such layers differ in mineral composition both among themselves and from the marble host.

In the layered margin, the innermost layer 1 is built mainly of calcite and scattered albite and quartz. Layer 2 is amphibolite which consists of tschermakite, zoned plagioclase (cores An_{2-5} , rims An_{20-23}) and epidote/clinozoisite (Ps_{10-30})¹. Between layers 1 and 2, locally an additional layer 1A occurs, which is composed of calcite with scattered actinolite and epidote (Ps_{12-25}) blasts.

Layer 3 is discontinuous (Fig. 5), has unique composition and porphyroblastic texture. In the fine-blastic Mg-chlorite – clinozoisite (Ps_{10-40}) matrix, porphyroblasts of subhedral amphibole and epidote (Ps_{35-40}) are randomly dispersed. The amphibole blasts of layer 3 uniquely display bluish-green

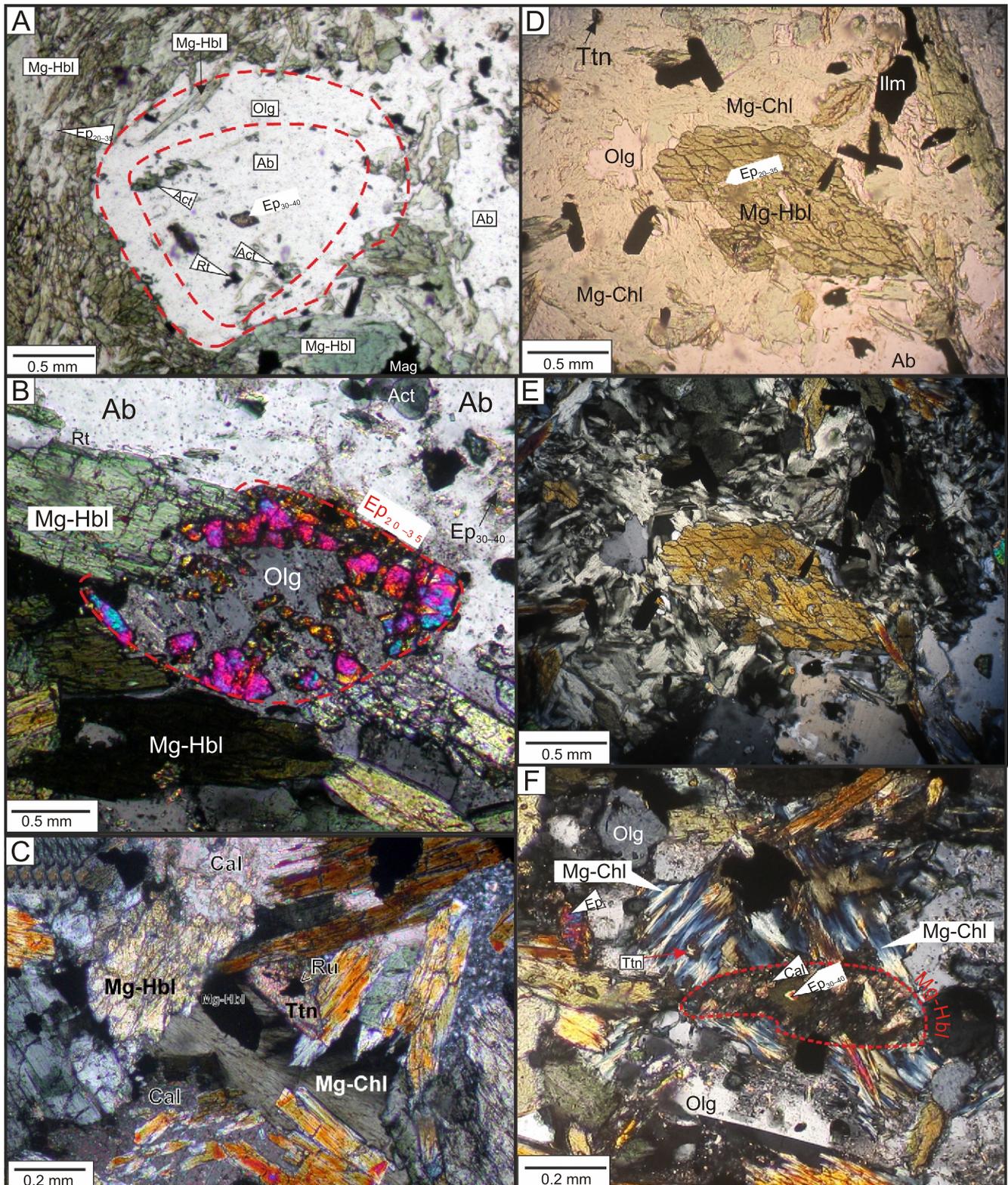


Fig. 3. Massive amphibolites

A – zoned plagioclase, note Act+Ep_{P_s30–40}+Rt inclusions of assemblage IA in an albite core, albite coexists with Mg-Hbl, crossed polars; **B** – details of albite-oligoclase interface, note Olg intergrowth with lower-Fe epidote Ep_{P_s20–35} and Mg-Hbl of assemblage IIA, crossed polars; **C** – assemblage IIIA of Mg-Chl and Ttn with Rt relics, crossed polars; **D** – Ilm and Ttn in Mg-Chl+Olg+An assemblage (IIA), plane polarized light; **E** – Ilm and Ttn in Mg-Chl+Olg+An matrix, crossed polars; **F** – Mg-Hbl (dotted lines) with Ep_{P_s30–40}+Cal inclusions replaced by Mg-Chl, crossed polars; Ab – albite, Act – actinolite, Cal – calcite, Ep – epidote, Ilm – ilmenite, Mag – magnetite, Mg-Chl – magnesio-chlorite, Mg-Hbl – magnesio-hornblende, Olg – oligoclase, Rt(Ru) – rutile, Ttn – titanite

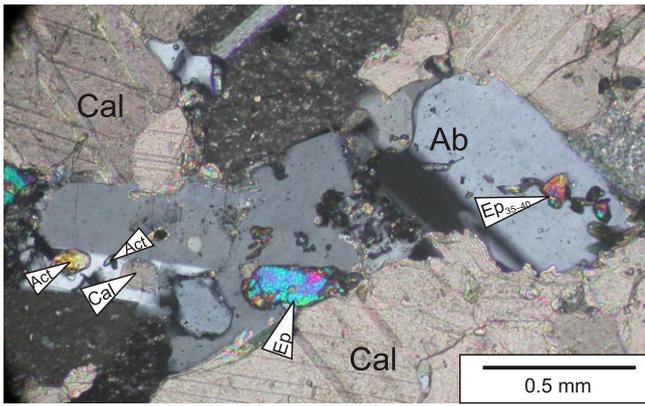


Fig. 4. Marble with dispersed albite which contains Act+Ep₃₅₋₄₀ inclusions

Ep₃₅₋₄₀ – epidote(Ps₃₅₋₄₀), other explanations as in Figure 3

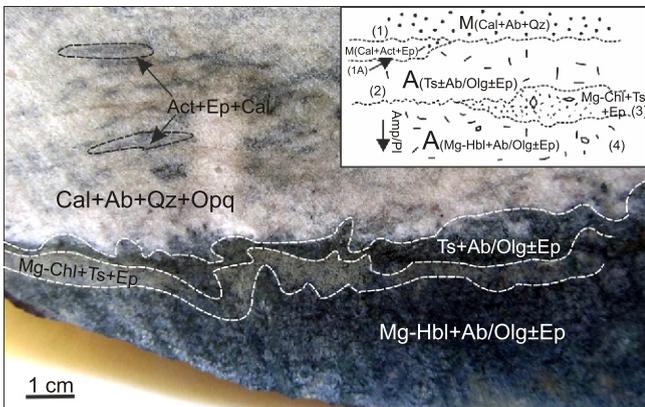


Fig. 5. Details of the layered margin of a marble lens, a transition to surrounding massive amphibolite

Inset: generalized sketch profile showing position of all layers distinguished within the marble (M) and contact with the amphibolite (A); Amp – amphibole, Chl – chlorite, Pl – plagioclase, Qz – quartz, Ts – tschermakite, other explanations as in Figure 3

pleochroism and show compositional zoning from tschermakite in the cores through winchite to barrosite in the rims (Fig. 6A–C).

Layer 3 occurs directly at the border with massive amphibolite composed of Mg-hornblende + oligoclase ± epidote (Ps₃₅₋₄₀) (Fig. 5). The massive amphibolite differs from layer 2 by the absence of tschermakite. It also differs from the (pillowed) metabasites that consist of zoned Ca-amphiboles (actinolite cores to tschermakite rims) and zoned plagioclase (Szczepański and Ilnicki, 2014).

MINERAL ASSEMBLAGES AND TEXTURAL OBSERVATIONS

Based on the observed textural relationships between minerals both in the massive amphibolites and in the amphibolite of layer 2, three mineral assemblages can be distinguished in these rocks (Table 1):

1. Assemblage IA represented by inclusions in the albite cores of plagioclase blasts, Act/Tr+Ep(Ps₃₀₋₄₀)+Rt (Fig. 3A, B);

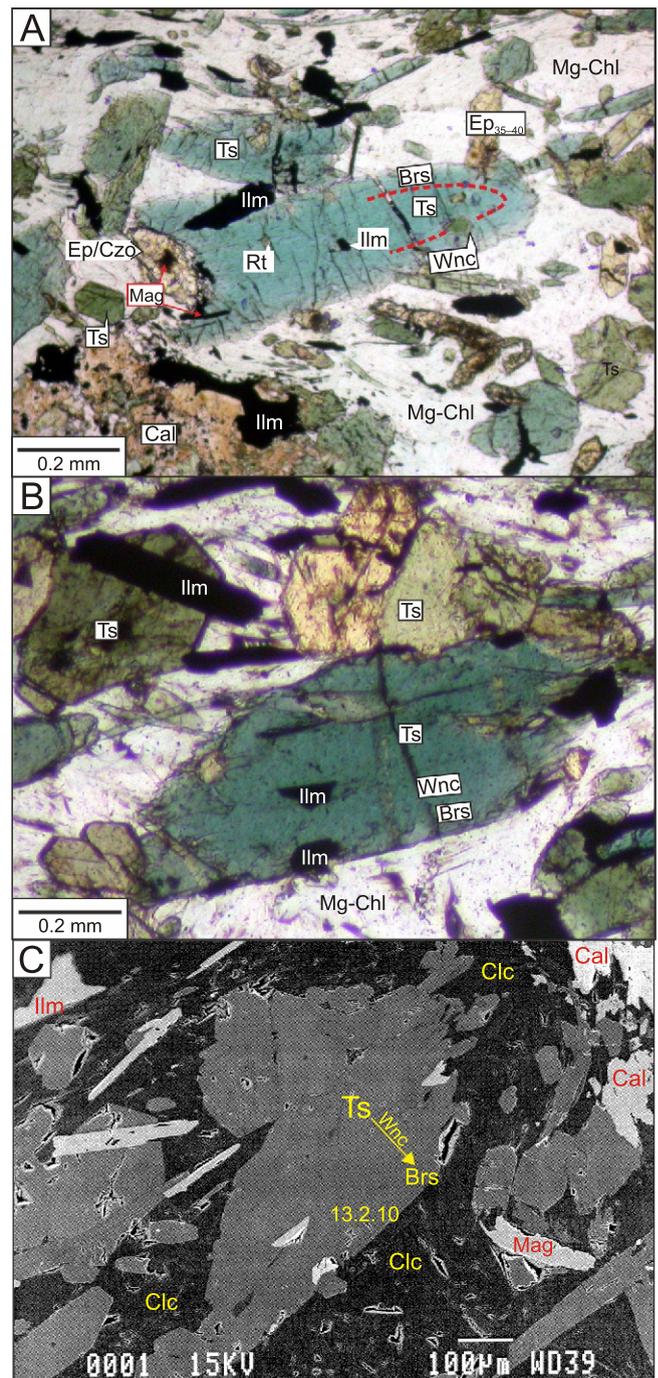


Fig. 6. Mafic layer 3 in calcite marble

A – Ts amphibole with Brs+Wnc rim within Mg-Chl-Cal matrix with Ep₃₅₋₄₀, Ilm and Mag, plane polarized light; **B** – Ts largely replaced by Wnc+Brs and Ilm set in Mg-Chl, plane polarized light; **C** – BSE image of amphibole, the Ts core which changes rimwards to Wnc and Brs; Brs – barrosite, Czo – clinozoisite, Wnc – winchite; other explanations as in Figures 3–5

2. Assemblage IIA is the major rock component, composed of Mg-Hbl+Ab+Olg+Ep(Ps₂₀₋₃₅)+Opq(Mag+Ilm) ± Cal ± Ttn (Fig. 3A, B);
3. Assemblage IIIA consists of Mg-Chl+Ttn with Rt inclusions; chlorite either surrounds Mg-hornblende or appears as single flakes between the minerals of the IIA assemblage (Fig. 3C–F).

Table 1

Compositional characteristics of the mineral assemblages distinguished

Mineral	ASSEMBLAGE				
	IA	IIA	IIIA	IM	layer 3
Amphibole	actinolite inclusions in albite	low- to high-Al Mg-hornblende	–	actinolite inclusions in albite	tschermakitic cores with barroisitic or winchitic rims
Epidote	epidote (Ps _{30–40}) with constant Ps ratios	epidote cores (Ps _{20–35}) with rimward decreasing Ps ratio	–	epidote (Ps _{35–40}) with constant Ps ratios	Porphyroblasts epidote (Ps _{35–40}) with rimward increasing Ps ratio
					Matrix clinozoisite cores (Ps ₁₀) with epidote (Ps ₄₀) rims
Plagioclase	albite (An _{<2%mol})	albite cores (An _{<2.5}) with oligoclase rims (An _{18–28}) and increasing calcium content	–	–	–
Chlorite	–	–	chamosite or nimitite surrounding Mg-hornblende or dispersed	–	sheridanite, nimitite (leptochlorite) or clinocllore
Rutile/Titanite	rutile inclusions within albite	dispersed titanite in matrix	rutile inclusions in titanite	–	rutile inclusions within tschermakite
Other	–	magnetite with epidote (Ps ₃₅) and clinozoisite (Ps ₂₀)	–	–	1. ilmenite inclusions in tschermakite or in the groundmass 2. magnetite inclusions within epidote (Ps ₄₀) or clinozoisite (Ps ₁₀)

Oligoclase outgrowths on zoned plagioclase grains are in direct contact with Mg-hornblende. The Olg-Hbl pair overgrew the albite blasts which contain inclusions of older epidote (Ps_{30–40}) (Fig. 3A, B).

In the marbles, two mineral assemblages can be distinguished:

1. Assemblage IM consisting of Ep_{35–40}+Cal+Act that are inclusions in albite blasts (Fig. 4),
2. Assemblage IIM, the main rock component, in which, between the dominant calcite grains, Ab+Qz (Fig. 4) and opaque minerals accompanied by ±Bt and ±Czo are dispersed.

In layer 3, Mg-Chl + fine-grained Ep (Ps_{10–40}) form a matrix in which porphyroblasts of Ts+Ep_{35–40}±Wnc±Brs+Mag+Ilm are embedded (Fig. 3A–C).

In assemblage IIM, metamorphic transformations were not capable of equilibrating plagioclase which remained zonal with albite cores and oligoclase rims.

In the marbles, actinolite and epidote of IM composition do not occur merely as inclusions in albite but also form independent aggregates characteristically concentrated close to the boundaries of the marble lenses (Fig. 7). Such observations allow us to assume that Ep_{35–40}+Cal+Act of assemblage IM were also in equilibrium with the minerals of the calcite-dominated assemblage IIM.

In terms of structural observations, minerals IA included in the albite blasts have some preferred orientation parallel to the weak external foliation displayed by the amphiboles of assemblage IIA. However, in perpendicular sections only a random distribution of the included minerals is in evidence. These observations suggest a constrictional strain geometry owing to low L>S type deformation at the time of albite blastesis and stronger S>L type deformation during the formation of the dominant assemblage IIA. Evidence for strain in the marble lenses is

even poorer. In the albite blasts, actinolite and epidote inclusions of assemblage IM are randomly distributed. Minerals IIM do not show any preferred orientation either.

In contrast, in the matrix of unique layer 3, fine-grained Mg-chlorite and epidote (Ps_{10–40}) along with tschermakite porphyroblasts are linearly arranged parallel to the marble-amphibolite boundaries (Fig. 5). In perpendicular sections, the minerals reveal L>S fabric and constrictional strain, which was characteristic of only the early mineral assemblage IA in the amphibolites. Moreover, in layer 3, a few epidote porphyroblasts are oriented perpendicular to these arrays

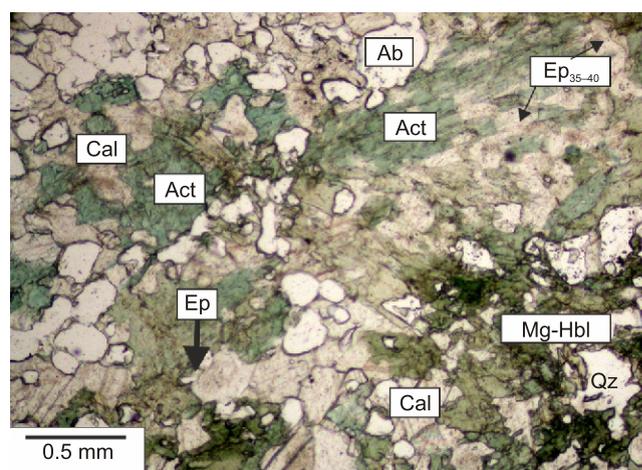


Fig. 7. Marble, actinolite-epidote aggregates within calcite background with sparse quartz and Mg-hornblende (plane polarized light)

Explanation as in Figures 3–5

whereas some chlorite blasts are random. Such observations suggest that the unique assemblage of layer 3 developed concurrently with the assemblages IA and IM, thus relatively early in the history of the parent rocks.

In general, the marbles and the host massive amphibolites are not intensely deformed. This stays in contrast with the strongly schistose adjacent metarhyolite and mica schist surroundings. Therefore, the former appear as a structural pod which is significantly less strained than is its neighbourhood.

MINERAL CHEMISTRY

Chemical characteristics of the minerals in the rocks studied and the distinguished mineral assemblages are summarized in Table 1. All amphiboles in the massive amphibolites and marginal mafic layers in the marble lenses represent the Ca-amphibole group with minor sodic-calcic amphiboles (classification by Leake et al., 2004). They belong to the ^W(OH, F, Cl)-amphibole supergroup (classification after Hawthorne et al., 2012) and are characterized by relatively high magnesian ratios XMg > 0.5 (Fig. 8A and Appendix 1*).

ASSEMBLAGES IA, IM AND IIIA

In assemblages IA and IM, actinolite inclusions within the albite blasts have similar composition [$Si^{4+} > 7.0$, $Na^B = 0.05–0.15$, $(Ca+Na)^B > 1.3$ pfu]. Fe^{2+} contents markedly decrease in inclusions located in the outer parts of the host albite blasts (Fig. 8A). Poikiloblastic albite does not differ between IA and IM, being of almost identical composition ($An_{1.3–2.5}$). Iron-rich epidote inclusions in the albite blasts are characterized by $Ps_{30–40}$ in IA and $Ps_{35–40}$ in IM.

Amphibole and plagioclase occur in assemblage IIA but are missing in assemblage IIM. In IIA, amphiboles are represented by low- to high-Al Mg-hornblende (Fig. 8A). High-Al amphiboles are characterized by increasing core-to-rim Al_{tot} values (cores: $Al_{tot} = 2.8$; rims: $Al_{tot} = 3.2$ pfu) and a constant silica content, whereas low-Al ones have constant Al_{tot} (ca. 2.2 pfu) with somewhat increasing silica (+0.7 pfu). Mg^B increases towards the rims (+1.0 pfu) proportionally to decreasing Fe^B . Zoned plagioclases have albite cores similar in composition to albite IA. The oligoclase rims ($An_{18–28}$) further reveal zonality marked by the slightly increasing calcium content (An_{18} – An_{28}) towards the blast boundaries. In IIA, epidote blasts also reveal zonality, which is marked by a decrease in the pistacite molecule toward their rims ($Ps_{30–35}$ – $Ps_{20–30}$).

As observed in the massive amphibolites and in layer 2, assemblage IIIA contains Mg-chlorite which mostly developed by replacement of older Mg-hornblende blasts. Its chemical composition is rather uniform (Appendix 2).

LAYER 3

In contrast, subhedral amphiboles of layers 3 are characterized by tschermakitic cores with the XMg ratio = 0.6–0.7, $Si^{4+} = 5.5–6.5$ and the $Al^{IV}-[Na+K]^A$ ratio = 1.5–2.3 (Fig. 8). Their rims are winchitic or barroisitic ($Na^{M4} = 0.65–0.7$ pfu, $Si^{4+} = 7–7.5$ pfu, XMg > 0.5, $Al^{IV}-[Na+K]^A < 0.5$ pfu) and Na^{M4} increases proportionally to the decreasing Ca^B value. Plagioclase is absent.

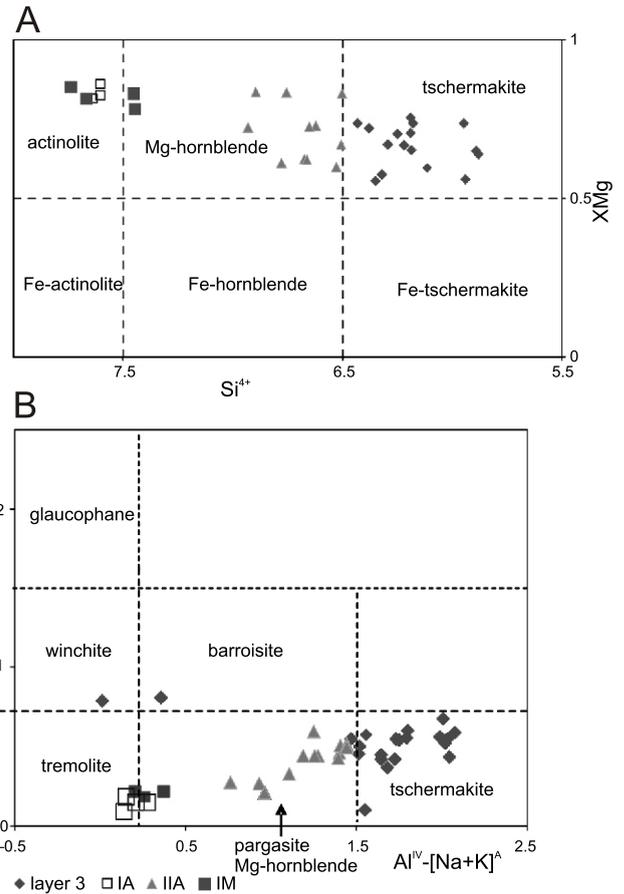


Fig. 8. Classification plots of amphiboles from the Gniezów metabasites

A – Si^{4+} –XMg classification diagram for the calcic amphiboles analysed; $Ca^B/(Ca^B + Na^B) = 0.75$ and $Ca^B = 1.5$ pfu, XMg = $Mg/(Fe^{2+}+Mg)$ (diagram template after Leake et al., 2004); **B** – the Na^{M4} vs. $Al^{IV}-[Na+K]^A$ plot with glaucophane/tschermakite vectors correlation, respectively (diagram template after Schumacher, 2007; modified); IA, IIA, IM – mineral assemblages (see Table 1)

Zoned epidote porphyroblasts have Fe-rich cores and Fe contents slightly increase towards their rims: $Ps_{35–38}$ – $Ps_{38–40}$. However, much stronger zonality is shown by fine epidote blasts which coexist with chlorite in the chlorite-epidote matrix of layer 3. Although they have similar Fe-rich rims ($Ps_{35–40}$), their cores are distinctly Fe-poor ($Ps_{10–15}$). XRD analyses helped to reveal, in the matrix chlorite of layer 3, orthochlorite with a sheridanite composition ($Si^{4+} = 3.8–4.2$ pfu, XFe = 0.2), which coexists with nickel-bearing lepto-chlorite (nimite: $Ni+Mg+Al^{IV} = 6$ pfu) or chamosite ($Mg > Fe$ and $Mg + Fe = 5$ pfu) and tschermakitic amphibole (Table 1 and Appendix 2). The chlorite of layer 3 differs significantly from the Mg-chlorite of assemblage IIIA.

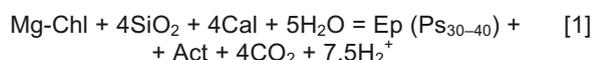
METAMORPHIC REACTIONS

The textural relationships and chemical compositions of the mineral assemblages described above were used to determine

* Supplementary data associated with this article can be found, in the online version, at doi: 10.7306/gq.1345

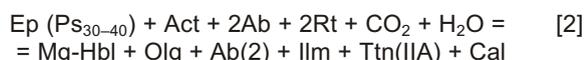
metamorphic reactions that may have brought about the mineral transformations in the rocks studied. The number of realistic reactions that may have actually occurred is constrained by the lack of garnet, K-feldspar and biotite. The observed coexistence of Mg-hornblende-actinolite-albite with accessory titanite and an opaque phase points to the lower amphibolite facies, which concurs with the data obtained for the foliated metabasalts with relict pillow structure (Szczepański and Ilnicki, 2014). However, it is unclear what transformations occurred to the originally underwater extrusions of pillow basalts that during the protolith stage were likely composed mainly of pyroxene and plagioclase. Did this pair undergo early hydration and metamorphism under low P-T and higher pressure conditions during the subsequent subduction episode? The above-described mineral assemblages and their textural relationships suggest that the metamorphic transformations likely occurred in the following order:

The first identifiable assemblage IA (Fig. 3A) included in albite may have developed due to a reaction similar to that described by Bucher and Grapes (2011):



Early Mg-chlorite was likely formed via hydration of pyroxene and albite blastesis presumably continued at the expense of more Ca-rich original plagioclase.

Under ongoing progressive conditions, assemblage IIA (Fig. 3B) could develop owing to the reaction between actinolite, epidote and rutile of assemblage IA with albite (Apted and Liou, 1983):



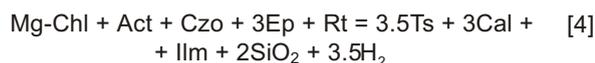
This reaction explains the coexistence of Mg-hornblende and unequilibrated zoned plagioclase (Ab core with inclusions and Olg rim free of inclusions), titanite, ilmenite and calcite.

The zonal composition of epidote IIA, marked by Fe-decrease toward the rims, may be explained by a concurrent reaction (Strens, 1965) that might also produce magnetite which is scattered (Fig. 6A):

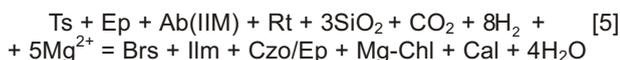


During metamorphism, Fe^{2+} ions could be easily oxidized to Fe^{3+} .

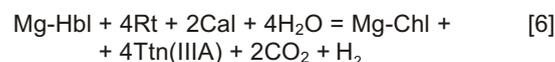
In layer 3, tschermakite may have been produced at the expense of actinolite, rutile and low-iron epidote (Fig. 6A) due to the reaction (Bucher and Grapes, 2011):



The barrosite rims on the tschermakitic cores in the bluish amphibole porphyroblasts might develop by the reaction of tschermakite and epidote (Ps_{35-40}) in the presence of albite (IIM) as a source of Na (Ernst, 1979; Poli, 1991) derived from the immediate marble surroundings. Among further products would be low-iron epidote/clinozoisite, Mg-chlorite, ilmenite and calcite (Fig. 6A–C):



Furthermore, chlorite and titanite of assemblage IIIA may have been produced due to retrogression in the presence of H_2O while Mg-hornblende rims IIA were altered to Mg-chlorite IIIA (Hunt and Kerrick, 1977; Ahn and Cho, 1998; Fig. 3D–F). Relict rutile IA preserved in the titanite of assemblage IIIA suggests that the former was used to generate the latter (Fig. 3C):



In the marble lenses, calcite appears amongst products of the reactions that formed assemblages IM and IA. However, most calcite in the dominant assemblage IIM was likely derived from the directly recrystallized original calcite.

THERMOBAROMETRIC CALCULATIONS

Isopleth thermobarometry. Having identified mineral assemblages in a rock of known bulk chemical composition and determined realistic metamorphic reactions that produced them, one can draw specific isopleths for minerals in a given thermodynamic system. By means of pseudosections, possible P-T paths can be reconstructed (Fig. 9).

Our analysis shows (Table 1) that the massive amphibolite represents system NCFMASHTO ($\text{Na}_2\text{O-CaO-FeO-MgO-Al}_2\text{O}_3\text{-SiO}_2\text{-TiO}_2\text{-H}_2\text{O-CO}_2$). For the rock analysed the whole-rock data were calculated to mole fraction and finally mole numbers of a particular oxide. Water content was set to excess. K_2O and MnO were omitted due to their low contents. The following isopleths were used to perform calculations: Al_{tot} (pfu), Al^{M1} (pfu) and Si^{T5} (pfu) for amphiboles; An% (mol) for plagioclases; vol.% of solids for chlorite. Omphacite (Cpx) has been also calculated by the software but such prediction is likely due to Ca and Mg overabundance in the bulk-rock composition (Fig. 9). *Theriak Domino* software interprets elevated Ca and Mg contents and SiO_2 , released during reaction [4] as omphacite.

In the massive amphibolites, the earliest protolith transformation occurred probably under very low-grade conditions by reaction [1], the product of which has not been preserved in the rocks studied. The oldest recognizable mineral assemblage IA had to be generated already under greenschist facies conditions. An intersection of Si^{T5} , Al_{tot} of amphiboles, chlorite vol% of solids and An% (mol) in plagioclase, in the presence of clinozoisite but absence of laumontite, would indicate $310 \pm 30^\circ\text{C}$ and $3\text{--}3.5 \pm 0.5$ kbar (Fig. 9). The textural evidence coupled with EMPA data show that Mg-Hbl is in contact with both Ab and Olg. Isopleths of $\text{Si}^{\text{T5}} = 6.24\text{--}6.35$ and $\text{Al}^{\text{M1}} = 0.45$ for the amphibole intersect with isopleths for albite (An% (mol) < 0.06) at 460°C and 9.5 kbar. For the Mg-Hbl-Olg pair an intersection of the due isopleths falls at $T = 530^\circ\text{C}$ and $P = 6$ kbar. The lack of garnet excludes temperatures $> 520^\circ\text{C}$ for the pressure range mentioned, thus the P-T peak for layer 3 can be set at maximum 520°C and $9\text{--}10.5 \pm 0.5$ kbar. Following reaction [2], Mg-Hbl+Olg pairs of assemblage IIA may have developed in the presence of titanite and ilmenite, which concurs with the coexistence of the two over the tight miscibility field (Fig. 9). This confirms the estimations of P-T conditions at $T = 500\text{--}550^\circ\text{C}$ and $P = 5\text{--}6.5$ kbar for these rocks.

The amphibolite layers 2 and 3 in the transitional zone were too thin to be analysed chemically. They had most likely to follow the same P-T path as the surrounding massive amphibolite. In these layers, plagioclase also remained unstable and the textural relationships indicate the coexistence of albite and

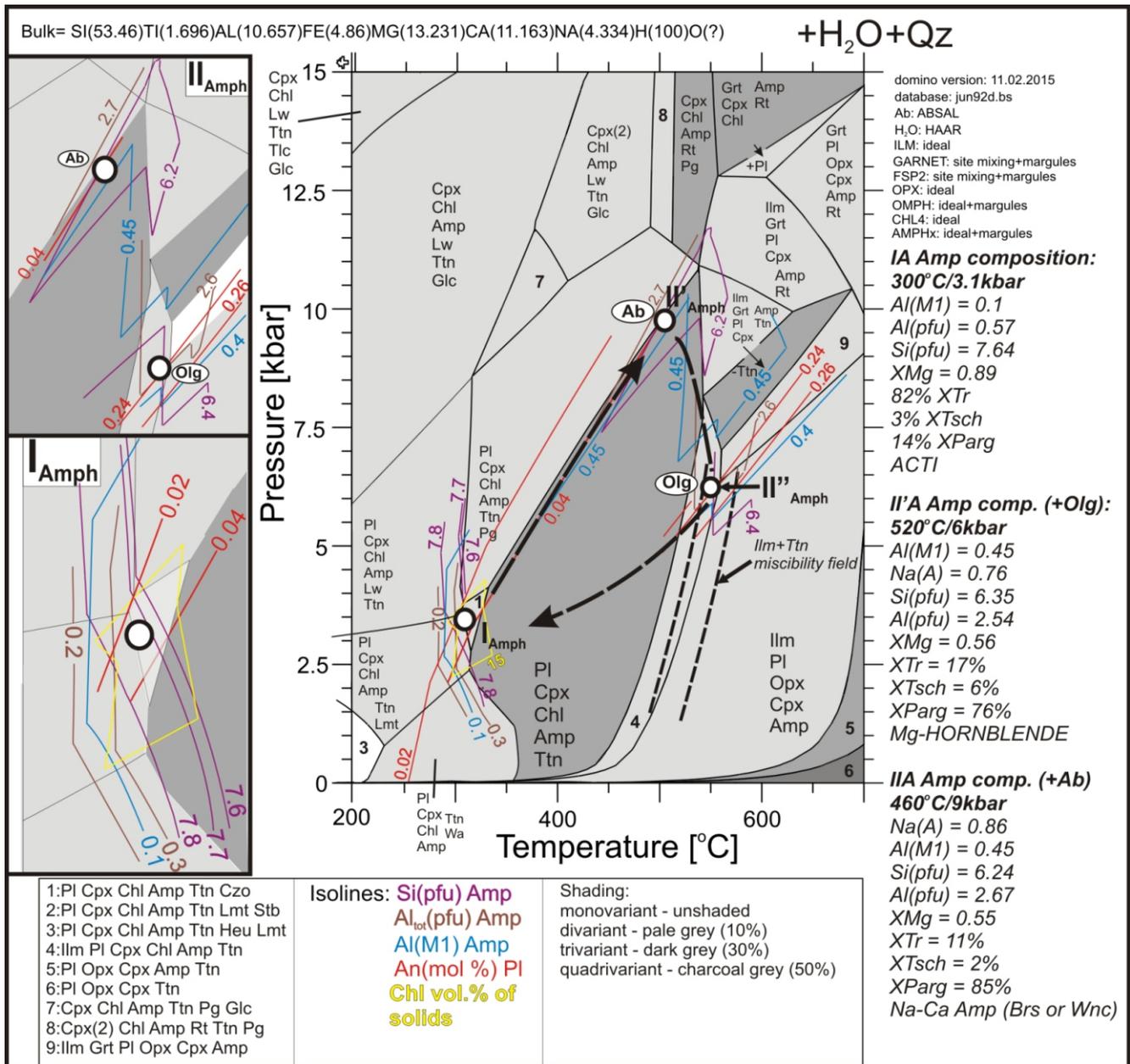


Fig. 9. P-T pseudosection calculated for the bulk-rock composition of the massive amphibolites studied in the NCFMASHTO system (with normalized mole proportions and excess of water)

The arrows in bold show the P-T path for amphibolites, whereas points II'A (Ab) and II'A (Olg) denote changing P-T conditions during ongoing reactions [2] and [3]; all points are determined based on the amphibole and plagioclase isopleths and additionally chlorite (% volume of solids) for the first stage IA; abbr. after Whitney and Evans (2010)

hornblende. Being aware of all limitations, we tentatively place the mineral assemblages of layers 2 and 3 against the pseudosection grid calculated for the massive amphibolite. The albite isopleths (An% (mol) <0.06) that intersect the Si(pfu) and Al(pfu) isopleths of the amphiboles indicate markedly higher pressure than for the Hbl+Olg pairs at a slightly lower temperature range of 480–530°C. Utilizing these isopleth readings, the peak pressure experienced by layers 2 and 3 may be estimated at 10.5 ± 0.5 kbar, which is identical with values obtained for the massive amphibolites. This shows that the data for the mafic layers in marbles and the amphibolites are compatible and the rocks underwent the same P-T path. Therefore the mineral assemblage of layer 3 which includes Mg-chlorite texturally equi-

brated with Si-depleted, Na+Al-enriched barroisite rims on the tschermakite cores can also be used, and then it indicates a pressure of 9–10.5 ± 0.5 kbar too.

Conventional thermobarometry. The above estimations were checked against those obtained utilizing conventional thermobarometry for most phases identified in the rocks studied. In mafic rocks, the pair Ep(PS₃₀₋₄₀)+Act is generally stable in the range of 300–500°C and 3–5 kbar (Ernst and Liu, 1998; Perraki et al., 2002; Bucher and Grapes, 2011; Baziotis et al., 2014). Within such a P-T range, the Al^{IV} increase in actinolite indicates temperature growth, whereas increasing Al^{VI} and Na^(A+B) suggest pressure rise still in the greenschist facies (e.g., Laird and Albee, 1981; Triboulet, 1992; Baziotis et al., 2014).

Assemblage Mg-Hbl+Olg+Ep_{10–20}+Cal is stable under amphibolite facies conditions (5–8 kbar and 500–700°C). In this assemblage, the increasing metamorphic grade is matched by rising XMg in amphiboles but decreasing pargasite component (Fe³⁺>Al^{VI}) and Si⁴⁺ decrease (Fig. 8A). Barroisitic, winchitic and tschermakitic amphiboles are more compatible with elevated pressures.

In the massive amphibolites, using the geobarometer of Bhadra and Bhattacharya (2007), the pair Mg-Hbl and Ab (X_{Ab} = 0.95) yielded a pressure of 10.6–10.7 kbar at 480°C whereas for the pair Mg-Hbl and oligoclase (X_{Ab} = 0.73) P-T conditions were determined at 6.6–7.4 kbar at a temperature of 520–540°C (Table 2).

Assemblage IA. For assemblage IA, Act+Ep paired with albite, the amphibole-epidote-chlorite geothermometer (Triboulet, 1992) was used because it relies on the specific calculation approach that requires <10 mol% of An in plagioclase and is valid for low-Ti (Ti < 0.1 pfu) amphiboles (Appendix 1), which fits the case studied. Knowing the Si^{T1}, Al^{IV}, Al^{VI}, Fe³⁺, Fe²⁺, Mg^{A,B}, Ca^B, Na^{M4}, Na^A values, the equilibrium constant [ln(K_d)] of Act and Mg-Hbl was calculated at 345 ± 5°C and 3.35 ± 0.5 kbar. Such a result agrees perfectly with that suggested by the isopleth method for pressure but yields slightly higher values for temperature, yet within the error limits.

Requirements set by other geothermobarometers make them inapplicable for assemblage IA. For instance the method proposed by Plyusnina (1982) requires the presence of hornblende not actinolite. Furthermore, actinolite is considered an intermediate member of the solid solution of Mg-rich tremolite and Fe-rich ferroactinolite (Leake et al., 2004; Schumacher, 2007). In Ca-rich rocks, actinolite may contain increased Mg-rich tremolite and decreased Fe-actinolite end-members as the result of free FeMg₁ substitution on the C site (e.g., Najorka and Gottshalk, 2003; Schumacher, 2007). In the case of the thermobarometric calculations, increased tremolite end-member in actinolite results in obtaining an overestimated temperature at the estimated corresponding pressure for the known bulk rock (Ilnicki, 2013). EPMA analyses revealed that actinolite inclusions in albite blasts both in marbles and amphibolites have elevated concentrations of the tremolite end-member (X_{Tr} > 0.8).

Assemblage IIA. In the case of the unequilibrated plagioclase of assemblage IIA in garnet-free amphibolites, metamorphic conditions were determined for the Mg-Hbl+Pl pairs (Table 2). The activity coefficients were calculated in proportion for each end-member in Mg-hornblende. For the esti-

mated minimum and maximum temperatures of 450°C and 550°C, relative pressures of Mg-Hbl+Olg pairs were calculated (assuming a constant volume of rock, V = 0) at 4.24 kbar and 6.59 ± 0.1 kbar, respectively. Pressure estimations follow the empirical reactions given by Bhadra and Bhattacharya (2007), with W_{C1} set typically at ~1.0 kJ after Holland and Powell (1992) (see Appendix 2). The obtained P-T range of 450°C/4.24 kbar and 550°C/6.59 kbar is wide, therefore the geothermobarometer devised by Gerya et al. (1997) was used to get better constraints. This is based on the estimations used by Zenk and Schulz (2004), namely An% (mol) > 10 and Amp = Hbl, which fits assemblage IIA (see Appendix 3). This method does not depend on plagioclase composition. Using the same EPMA analyses, the results obtained are 549–589°C and 6.2–7.5 kbar, persistently giving higher temperatures but significantly lower pressure values than the isopleth method. However the values correspond well with those obtained by Szczepański and Ilnicki (2014) for the P-T peak conditions. On the other hand, the geothermobarometer after Plyusnina (1982), which is based solely on Al_{tot} in hornblende and Ca (mol%) in plagioclase, yielded temperatures of 520–540°C, hence similar to those inferred from the isopleths, and somewhat higher pressures between 6 kbar and 8 kbar, yet still 2–4 kbar lower than the isopleth data.

INTERPRETATION AND DISCUSSION

THERMOBAROMETRY: P-T PATH

In the mafic rocks studied, an intersection of the amphibole and plagioclase isopleths (Fig. 9) indicates that assemblage IA was formed at 310 ± 30°C and 3.5 ± 0.5 kbar. In assemblage IIA, significantly variable composition of the unequilibrated plagioclase, expressed by highly concentrated isopleths, when coupled with the modestly variable Al and Si content in amphiboles, allows us to estimate a temperature of ~500–550°C and pressure in the range of 5–10 kbar for Amp-Ab pairs and 5–7 kbar for Amp-Olg pairs. However, the intersection of the isopleths for coexisting albite and Mg-amphibole (Fig. 9) yields pressures of 9–10.5 kbar and temperatures of 480–530°C. Such higher pressure conditions must have been attained while the rocks were undergoing transformation by reaction [2] which produced Mg-Hbl IIA that evidently coexisted with Ab for an unknown duration. The reaction continued to operate in mafic

Table 2

Results of conventional geothermobarometric determinations

M1	Analysis	Position	As.	T [°C] Triboulet	P [kbar] Triboulet	
	13.2.4	inclusion in Ab – core	IA	340	3.2	
13.2.5	inclusion in Ab – rim	IA	350	3.5		
13.4.1	inclusion in Ab – core	IM	366	3.9		
13.4.5	inclusion in Ab – rim	IM	357	4.0		
13.4.11	inclusion in Ab – core	IM	321	2.8		
13.4.12	inclusion in Ab – rim	IM	309	2.7		
M2	Analysis	Position	As.	T/P (Bhadra and Bhattacharya)	T/P (Zenk & Schulz)	T/P (Plyusnina)
	13.7.7	Amp rim, pair with Ab (X _{Ab} = 0.95)	IIA	480/10.6–10.7	480/10.24	540/8
	13.7.8	Amp rim, pair with Olg (X _{Ab} = 0.73)	IIA	520–540/6.6–7.4	495/5.7	520/6
	13.3.3	Amp rim, pair with Olg	IIA	450–550/4.24–6.4	589/7.5	540/6.5
	13.3.5	Amp rim, pair with Olg	IIA	450–550/4.15–6.59	561/7.1	530/6
	13.7.9	Amp rim, pair with Olg	IIA	450–550/4.14–6.48	549/6.2	520/6
	13.7.1	Amp rim, pair with Olg	IIA	450–550/4.24–6.59	584/7.4	520/6

rocks that were being exhumed to shallower depths of ~6 kbar but failed to equilibrate their minerals. A similar P-T path is demonstrated by the Na-amphibole rims on the tschermakite cores in layer 3 as well as the rimward increase of Al and decrease of Si in amphiboles occurring both in amphibolites and mafic layers at their transition to marbles. The increasing metamorphic grade is also corroborated by epidote with a discrete rimward decrease of the pistacite molecule content, yet growing oxygen fugacity cannot be excluded (Rötzler et al., 1999). Similar compositions of Act and Ep/Czo inclusions hosted by the albite blasts in marbles and in amphibolites further suggest that these phases were formed under similar conditions.

In layer 3, in the Brs+Wnc outgrowths on the tschermakitic cores, an increase of Al and decrease of Si toward the rims can be observed. Such outgrowths may have developed during a nearly isothermal episode of transient maximum burial to depths equivalent to a pressure of 9–10.5 kbar. The gradual transition T_s Wnc Brs may have been accomplished by the release of the cation Al^{3+} from tschermakite ($Al_{tot} = 4$ pfu), which replaced the Mg^{2+} cation due to ion substitution. This gave rise to the formation of transitional winchite ($Al_{tot} = 1$ pfu) and final barrosite ($Al_{tot} = 3$ pfu) with no assistance of other phases and eventually resulted in compositional zonality of the Na-amphibole blasts and their coexistence with magnetite (Fig. 6A). Considering the undersaturation of iron in the bulk rock composition, calculated P-T conditions and high X_{Mg} values in the amphiboles and chlorite, we assume that in layer 3, the formation of lawsonite or glaucophane and Fe-amphiboles such as crossite or riebeckite was prevented. Therefore, it is assumed that the phases observed in layer 3 actually developed at a pressure range of 9–11 kbar.

The pressure and temperature values obtained by means of conventional thermobarometry are comparable with the estimates obtained by means of the isopleth method. Some differences are possibly caused by a combination of microprobe inaccuracy and approximation errors.

In general, the reproducibility of P-T values justifies the results obtained. The determinations for the Mg-Hbl-Olg pairs in the rocks studied are practically identical with the estimations of the peak conditions of 560–570°C and 6.4 kbar reported by Szczepański and Ilnicki (2014). These authors also noted early metamorphism of pillow metabasites at 450–500°C and low pressures of 1.8–4.2 kbar. Our estimations for assemblage IA suggest $310 \pm 30^\circ\text{C}$ and 3.5 ± 0.5 kbar. The numbers are somewhat different, yet the data do indicate the lower section of the proposed P-T path (Fig. 9).

Our results determine the apex of the P-T path at 10.5 kbar (II'A) attained while the mafic rocks were being transformed by the ongoing reactions [2] and [3]. Reaction [2] was impeded by a transient pressure increase that allowed a growth of Mg-Hbl in the presence of albite while still preserving inclusions of assemblage 1A. Metamorphism went on as pressure was dropping at an almost steady temperature and Mg-Hbl became accompanied by blastesis of Olg and low-Fe epidote (II'A) during the main metamorphic episode climaxing at $540 \pm 20^\circ\text{C}$ and 6 ± 0.5 kbar (Fig. 9). However, reactions [2] and [3] failed to fully equilibrate the minerals in these rocks, which is shown by the incomplete transformation of Ab to Olg. Such a situation may have possibly occurred because of the short duration of this episode or because of a shortage of metamorphic fluids to assist the reactions.

Utilizing the data for the Mg-Hbl-albite pair a relevant geotherm is set to $\sim 17^\circ\text{C}/\text{km}$, which is typical of subduction. This controlled the metamorphism of the supracrustal rocks of the

Stronie Formation in the present-day western limb of the Orlica–Śnieżnik Dome. These were pulled down to depths corresponding to the lower crust for a time span of unknown duration and then almost isothermally exhumed to shallower depths where they remained moderately heated in the course of a regional metamorphic episode at the peak temperature conditions.

LOCAL AND REGIONAL CONTEXT

In rocks of the Stronie Formation (SF) throughout the Orlica–Śnieżnik Dome, zoned garnet in mica schists reveals a temperature range of 485°C–625°C and pressures of 3–7.5 kbar (Murtezi, 2006; Jastrzębski, 2009; Szczepański, 2010; Skrzypek et al., 2011a, b; 2014; Štípská et al., 2012). Less scattered values at the upper limit of this range have been obtained for both mafic and felsic metavolcanic rocks (Nowak and Żelaźniewicz, 2006; Murtezi, 2006) and for marbles (Jastrzębski, 2009). Such data would suggest that different parts of the SF were metamorphosed at different crustal depths. Nevertheless, the overall metamorphic grade in rocks of the SF increases easterly towards the Variscan collisional suture, where the SF rocks were brought into contact with the Brunovistulian terrane and heated up to 680°C at a pressure of 9–10 kbar (Schulmann and Gayer, 2000; Murtezi, 2006; Jastrzębski, 2009, 2012).

However, in the western part of the OSD, quite a distance from the suture, rocks of the SF were also metamorphosed at a pressure of ~10 kbar yet at a temperature lower by 100–120°C. Szczepański (2010), based on the crossing garnet isopleths and subdivision of garnet porphyroblasts into three concentric zones (inner cores, outer cores and rims), found that mica schists in the Góry Bystrzyckie (Fig. 1) underwent the following peak pressure conditions: (1) 580°C/10.2 kbar for the garnet rims in the northern part of the unit, (2) 560°C/8 kbar for the outer garnet cores in the central part of the unit and (3) 552°C/11.3 kbar for the outer garnet cores in the southern section of the unit. This author also noticed that classical thermobarometry [garnet-plagioclase-muscovite-biotite geobarometer after Hoisch (1990) modified by Wu and Zhao (2006)] yielded slightly higher pressure values for all samples studied by him (up to even 10.8 kbar on average at 550°C in the Gniewosów area). However, in mica schists further west in the OSD, in another local tectonic unit, the Zakletý Ridge (Fig. 1), Faryad and Kachlík (2013) were able to identify garnets for which the isopleth intersections indicated P/T conditions of 550°C/20–21 kbar, conditions that must have occurred prior to a subsequent medium pressure overprint. Moreover, in metatuffite hosted by phyllites of the Nové Město Fold Belt west of the OSD (Fig. 1), they found zoned garnets the rims of which are characterized by isopleths that also allow one to infer HP/LT metamorphism in a range of 11–15 kbar/350–450°C. Such results obtained from the two adjacent tectonic units would require that the phyllites were likely subducted to depths of 35–50 km and the mica schists as deep as ~65 km, all being metamorphosed along a steep geothermal gradient of $\sim 10^\circ\text{C}/\text{km}$. Although different by $\sim 7^\circ\text{C}/\text{km}$ from our estimate, both data sets testify to subduction of sedimentary-volcanogenic protoliths of the entire MSG, in particular the Stronie Formation, in the western part of the OSD.

Surprisingly, no evidence of HP metamorphism was reported from rocks of the MSG before Faryad's and Kachlík's (2013) findings. Winchite and barrosite identified by us in layer 3 are the first Na-amphiboles found in rocks of the western OSD. They are not original HP minerals, yet there is no evi-

dence that they may have replaced earlier glaucophane that overgrew a tschermakite inner core. Texturally winchite and barroisite are the youngest outer part of amphibole blasts that have older tschermakitic cores. Such zonation does not match those observed in minerals of the other mafic rocks studied. Perhaps this can be explained by the peculiar location of layer 3 which has been sheltered within marbles. In an early metamorphic episode, the carbonate host may have promoted a growth of Ca-amphibole instead of actinolite. Unless one assumes that, in the marbles and amphibolites, albite grains with actinolite+epidote±chlorite inclusions were derived from transformed glaucophane, any direct evidence of an early (U)HP episode in the evolution of these rocks is lacking. The textural relationships do not seem to support such transformation and the presence of glaucophane in mafic rocks. The observed mineral assemblages IIA, IIM and that of layer 3 in the transitional zone from marble to massive amphibolite all consistently point to a pressure of 9–10.5 ± 0.5 kbar at a temperature of 480–520°C, the conditions attained prior to the metamorphic peak at 520–530°C/6–6.5 kbar.

Mineral phases observed in metabasites throughout the region allow one to decipher only the latter peak conditions whereas potential relicts of any earlier event are lacking. As mentioned above, maximum P-T conditions estimated for metasedimentary and metavolcanogenic rocks of the MSG in the western OSD vary between 500–630°C and 5–11.3 kbar (Murtezi, 2006; Szczepański, 2010; Szczepański and Ilnicki, 2014). An estimate of 552°C/11.3 kbar comes from the outer cores in garnet grains (Szczepański, 2010) and is broadly similar to our data (10.2 kbar at 580°C). A considerable scatter of P-T values was explained by Szczepański (2010) by invoking three different metamorphic zones in the region. However, our data come from one exposure and cannot be explained by any metamorphic zonation. Moreover, such zonation would require systematic metamorphic mapping which is still to be done in the region. Nevertheless, Szczepański's (2010) observation in metapelites is compatible with our record for the marble-metabasite association and is in line with our estimates of higher pressure up to ~11 kbar attained prior to the temperature peak. Faryad's and Kachlik's (2013) results also corroborate an early HP event experienced by the MSG. However, Szczepański and Ilnicki (2014) claimed early metamorphism of the pillowed metabasalts at 450–500°C and 1.8–4.2 kbar, which is in striking contrast to the above-mentioned findings but may be easily explained by shuffling of tectonic slices exhumed from various structural levels, which is an option advocated by us. Therefore, we consider all of the reported data to be valid and scattered P-T estimates in a range of 300–630°C at 1.8–11.3 kbar do not negate them. This is in line with our observations of intense tectonic deformation of the Stronie Fm. (SF) supracrustal rocks that underwent multiple folding and shearing which eventually brought into contact rock units/tectonic slices that were deformed and metamorphosed at different depths within a subducted complex. Therefore, no uniform P-T data can be determined for the MSG.

The P-T path constructed for the mineral assemblages identified in the marble-amphibolite association (Fig. 9) predicts (1) a low-grade event at ~310°C /3.5 kbar (IA), (2) subsequent subduction down to ~500°C /10.5 kbar (II'A) followed by (3) nearly isothermal decompression (II''A), and then (4) retrogression back to greenschist facies. It may be assumed that Szczepański and Ilnicki (2014) were able to spot our events 1, 3

and 4, whereas Szczepański (2010) and Faryad and Kachlik (2013) managed to record events 2 and 3, still on a similar P-T path followed by the Stronie Fm. rocks in the western OSD. Such a steep path (Fig. 9) clearly indicates that the MSG rocks on the western limb of the OSD were subducted to a depth of 30–35 km, some of them possibly even more deeply but for a relatively short time span, and then became nearly isothermally exhumed to middle crustal depths, where they were deformed/sheared and heated long enough to obliterate almost all earlier mineral assemblages, and the obliteration continued during a subsequent greenschist overprint. Our results together with Szczepański's (2010) data seem to suggest that the subduction in the Góry Bystrzyckie rocks was possibly neither as cold nor as deep as revealed by Faryad's and Kachlik's (2013) data from the Zakletý Ridge further west in the OSD.

However, relicts of early events, if very scarce in the rocks, may have been overlooked in the course of microscopic examination or else eliminated by extensive recrystallisation and equilibration during the peak metamorphism. The latter possibility seems to be prompted by our study. In layers 2 and 3 enclosed in the marble lenses and in the adjacent massive amphibolite, such recrystallisation was evidently less effective, which left the rocks unequilibrated with legible records of earlier (pre-peak) mineralogy. This may have occurred because the marbles acted as a shelter that prevented mafic material enclosed in them from being exposed to the activity of metamorphic fluids which apparently assisted the peak event and helped to complete mineral transformation elsewhere. Furthermore, as the marbles are surrounded by amphibolites which also possess unequilibrated mineral assemblages, it follows that metamorphic transformations during the peak event were not uniform throughout the region and left some rock domains less thoroughly recrystallised. In the area studied, the metabasites are in contact with schistose metarhyolites that only carry a record of the peak conditions at 500–550°C/6–7 kbar (Murtezi, 2006). The strong foliation in the latter certainly promoted the activity of penetrating metamorphic fluids, the circulation of which was, however, markedly impeded in the massive mafic domain that enclosed the marbles. Such domains survived as less deformed and less metamorphosed pods within intensely schistose and sheared bulk surroundings, which agrees with the earlier-mentioned observations of the structural relationships in the area. Microstructures in the rocks studied show that the low deformation of the rocks was accomplished in a constrictional regime (L>S tectonite) at the time of albite blastesis (IA, IIA), which was followed by slightly stronger S>L type deformation when oligoclase and Mg-hornblende (II''A) recrystallised. These observations suggest that the HP event was likely associated with constrictional strain which may have been related to subduction channel flow. Records of such an event may have been also preserved in rocks at the hinge zones of larger-scale folds where the constrictional strain occurred (Żelaźniewicz et al., 2013).

The unique mineral composition of layer 3 is thought to reflect a low activity of metamorphic fluids in the carbonates, which were protected by the massive metabasite. In the Ca-rich domain, Ca-amphibole-like tschermakite grew in an early metamorphic phase and then was replaced by Na-bearing amphiboles when pressure increased. Mg-chlorite in layer 3 was not a retrograde phase but developed by reaction [6] and thus accompanied the Na-amphiboles.

In the region, the greenschist overprint is ubiquitous, yet in the rocks studied it is insignificant, being marked only by single flakes of chlorite of assemblage IIIA. Such weakly expressed retrogression concurs well with the limited influence of metamorphic fluids on the rocks studied which formed a less deformed and metamorphosed pod, relatively tightly isolated from the schistose surroundings. Beyond the studied pod, in most rocks of the region, recrystallization during the peak metamorphism was likely efficient enough to erase records of earlier metamorphic events.

In the mafic layers studied in marbles as well as in the host marbles and adjacent amphibolites, similar mineral assemblages developed under the early greenschist facies conditions. Such similarity suggests that the carbonate-basic protoliths of these rocks were originally intermingled and had a common metamorphic history. This is in line with a reef-related origin of marbles in the OSD proposed by Koszela (1997) whose lithostratigraphic reconstructions allowed an inference that the reefs developed on slopes of repeatedly active volcanoes. At Gniewoszczów, the marble lenses likely represent reef debris embedded in a volcanic edifice.

The carbonate rocks might have been originally incorporated into basalts in few ways: (1) on extruding basaltic lava entrapped carbonate xenoliths from wall rocks, (2) outpouring lava engulfed reef fragments detached from underwater biostromes, or (3) carbonate precipitating from seawater were being deposited simultaneously with extrusions of basaltic lava. In the case of mechanisms (1) or (2), which we prefer, the marble-amphibolite border would be shaped by thermal contact/metasomatic processes and such transformations happened directly after the incorporation of carbonate lumps into basaltic lava whereas subsequent transformations of the two contrasting rock types took place during regional metamorphism, including the inferred higher pressure event. We are in favour of option (2) because of the presence of scattered quartz grains which suggests some admixture of clastic matter between lava and carbonate lumps whereas actinolite and epidote grains scattered in the outer portions of the marble bodies as well as tiny isolated Act-Ep-Cal pods (Fig. 5) suggest some input of mafic volcanogenic matter collected by porous reef carbonate. Incorporation of such matter was most likely further assisted by metasomatic exchange of elements at the border zone between the carbonate reef lumps and hot mafic lava in an aqueous submarine environment. It is even possible to speculate that sodium in the marginal unique layer 3 came from the seawater yet details of the process are beyond the scope of this paper. Nevertheless, rocks of the marble/amphibolite interface described then underwent the regional metamorphism and were prone to record and preserve the mineral assemblages that developed at higher pressure conditions and survived later overprints.

CONCLUSIONS

The marble-amphibolite interface studied recorded contact/metasomatic and regional metamorphic processes that occurred at the boundary between carbonate reef debris and basaltic pillow lava/pyroclastic flow rocks. The interface embraces up to three mafic layers at the margin of the marble lens and the Mg-Hbl massive amphibolite which is transitional to the country pillowed Ca-amphibole metabasites. In the interface rocks, zoned plagioclase and amphibole occur with inclusions, which allowed identification of three consecutive mineral assemblages. These point to three P-T stages on the progressive metamorphic path at: (1) 310°C/3–4 kbar, (2) 480–500°C/10.5 kbar, (3) 500–530°C/6–6.5 kbar, based on isopleth intersections and checked against conventional thermobarometry. The data obtained define a steep clockwise P-T path and geothermal gradient of ~17°C/km, which suggest the subduction of metavolcanosedimentary rocks (Stronie Formation of the MSG) of the western limb of the Orlica–Śnieżnik Dome, with a transient yet discrete higher pressure episode. Mineral evidence for such an episode was also reported from mica schist in part of the dome and from metabasite of the adjacent Nové Město Belt (Faryad and Kachlík, 2013). The scarcity of mineral relicts capable of showing a HP event in the supracrustal rocks of the dome is presumably caused by more thorough equilibration and obliteration during the temperature peak at mid-amphibolite facies conditions and a subsequent ubiquitous greenschist facies overprint. Less equilibrated relicts can be expected in less strained pods, the rocks of which were thus prevented from more intense deformation that is evident in the zonally sheared surroundings. The rocks studied with the marble-amphibolite interface at Gniewoszczów belong to one of such pods and these are identifiable elsewhere in the dome. Searching for (U)HP mineral relicts in supracrustal rocks of the Młynowiec–Stronie Group should undoubtedly be continued for better understanding of the evolution of the Orlica–Śnieżnik Dome and adjacent units in the Sudetes.

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