Fluid influence on mineral reactions in ultrahigh-pressure granulites: a case study in the Śnieżnik Mts. (West Sudetes, Poland)

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Abstract Small tectonic slices of undeformed eclogites and ultrahigh-pressure granulites occur in three tectonic units of the Śnieżnik Mts. (SW Poland). Ultrahigh-pressure granulite/eclogite transitions with peak metamorphic conditions between 21 and 28 kbar at 800 to 1000 °C occur only in the Złote unit. Conventional U-Pb multigrain analyses of zircons from a mafic granulite provided 207Pb/206Pb ages between 360 to 369 Ma which are interpreted to approximate timing of original crystallisation from a melt. Diffusion kinetics and the restricted availability of a fluid phase mainly controlled the conversion from granulite to eclogite, although some bulk-chemical differences were also recognised. The ultrahigh-pressure granulites from the Złote unit exclusively contain H2O-rich inclusions with variable salinities which distinguishes them from high-temperature (HT)-granulites world-wide. This is also in contrast to the fluid regime (H2O-N2-CO2) recognised in the lower-temperature eclogites (600–800 °C) from the closely associated Międzygórze and Śnieżnik units. The variation in fluid composition between the lower-temperature eclogites and ultrahigh-pressure granulites on the one hand and ultrahigh-pressure granulites and HT-granulites on the other hand probably indicates contrasting P-T-t paths as a result of different tectonic environments.

Introduction

Metamorphic fluids play an important role in high-grade metamorphic processes, since reactions involving a free fluid phase cannot be used to constrain metamorphic P-T conditions without knowing the fluid composition. Fluid inclusions in metamorphic minerals may preserve the composition and/or physical properties of such fluids (e.g. Crawford and Hollister 1986; Touret 1981, 1992; Andersen et al. 1993). In combination with mineralogical constraints and textural data, fluid inclusions are an important tool that gives insight into high-grade fluid-rock interaction. Many studies have been undertaken on high-grade metamorphic rocks, in granulites as well as on eclogites. In general, HT-granulites are dominated by CO2-rich fluids (e.g. Touret 1981, 1992; Coolen 1982; Lamb et al. 1987; van Reenen and Hollister 1988; Ellevold and Andersen 1993; Herms and Schenk 1998). Some of them clearly originated during prograde or peak metamorphism (e.g. Vry and Brown 1991; Srikantappa et al. 1992; Herms and Schenk 1998). However, the importance of high-salinity brines in HT-granulites has been outlined by several authors (e.g. Aranovich et al. 1987; Newton 1995; Aranovich and Newton 1995; Fonarev et al. 1998). On the other hand, fluid inclusion studies of eclogite-facies rocks indicate a different fluid regime consisting of aqueous fluids with variable salinities, and N2 ± CO2 (Andersen et al. 1989, 1993; Klemd 1989; Klemd et al. 1992, 1994; Selverstone et al. 1992; Giaramita and Sorensen 1994; Philippot et al. 1995; Perchuk 1995). This apparent contrast in fluid regimes of granulite- and eclogite-facies rocks is supported by recent studies in the Western Gneiss Region (WGR) of Norway, where Grenvillian granulites were subjected to eclogite-facies metamorphism during the Caledonian orogeny 600 Ma later (Austrheim 1990). Metastable HT-granulites still contain high-density CO2-rich inclusions without evidence of free H2O, while fluids in eclogitised rocks and high-pressure granulites are dominated by H2O and N2 ± CO2. Furthermore,
the difference in fluid regimes between HT-granulites and eclogite-facies rocks is probably related to different tectonic settings (Andersen et al. 1993).

In order further to explore the fluid regimes associated with very high pressure rocks, we have studied granulites and eclogites from the Śnieżnik Mts. (Fig. 2), where two types of ultrahigh-pressure metamorphic rocks occur: (1) eclogites associated with amphibolite-facies gneisses (Śnieżnik- and Międzygorze units); (2) granulites interbedded with eclogite layers and domains (Złote unit). In a previous study, the fluid regime in the Śnieżnik- and Międzygorze units was documented (Klemd et al. 1995). Our present contribution covers the intimate association of eclogites, mafic and felsic granulites of the Złote unit, which was first described in meticulous detail by Smulikowski (1967). We shall show that this rock association was subjected almost simultaneously to P-T conditions of at least 21 to 28 kbar at 800 to 1000 °C, in contrast to the eclogites and granulites of the Western Gneiss Region (WGR) in Norway. Only limited amounts of H2O, and no CO2 were present during high- to ultrahigh-pressure metamorphism of the studied ultrahigh-pressure granulites which is supported by the fact that several prograde “frozen in” net transfer reactions survived ultrahigh-pressure metamorphism. This contradicts fluid investigations in HT-granulite terrains world-wide, but is in accordance with eclogite-facies fluid regimes.

**Geological setting of the Śnieżnik Mts.**

The West Sudetes are located along the north-eastern margin of the Bohemian Massif (Fig. 1). This tectonically very complex region encompasses a number of structural-stratigraphic domains, one of them being the Orlica-Śnieżnik dome which hosts the Śnieżnik Mts. (Fig. 1). Ages and deformation histories of this complex are still under discussion (e.g. Zelaźniewicz et al. 1995), as is the relationship to the Saxothuringian and Moldanubian Zones (e.g. Aleksandrowski 1993; Zelaźniewicz and Franke 1993; Oliver et al. 1993).

Eclogites occur in at least three tectonically bounded subunits (Fig. 2): (1) the Śnieżnik unit; (2) the Międzygorze unit; (3) the Złote unit (e.g. Bakun-Czubarow 1992; Bröcker and Klemd 1996). Eclogites of the Śnieżnik and Międzygorze units occur as lenses and blocks within amphibolite-facies gneisses (Don et al. 1990). In the Złote unit, very small amounts of eclogites occur intimately associated with mafic and felsic granulites (Smulikowski 1967; Pouba et al. 1985; Bakun-Czubarow 1991a, b; Kryza et al. 1996).

Smulikowski (1967), Brueckner et al. (1991) and Bröcker and Klemd (1996) interpreted the eclogite-facies metamorphism as an in situ process. In contrast, Don et al. (1990) considered the eclogites and granulites to be exotic tectonic inclusions within gneisses as the biotite lineation (L3) is only present in the amphibolitic rims of eclogite lenses but not in the interior parts. Estimated peak pressures in the Śnieżnik and Międzygorze units lie in the coesite stability field with pressures in excess of 27 kbar at temperatures between 700 and 800 °C (Bröcker and Klemd 1996). The Złote unit contains thin (up to 20 cm) bands of eclogites which are interlayered with intermediate to felsic granulites (Smulikowski 1967). This interlayering was interpreted by Bakun-Czubarow (1991b) to be the result of bimodal volcanism with small contaminations of sedimentary material (Bakun-Czubarow 1991b; Brueckner et al. 1991). The peak metamorphic P-T conditions attained in the Złote unit were estimated at 800 to 900 °C and at 16 to 22 kbar by Bakun-Czubarow (1991b) and Steltenpohl et al. (1993), while Pouba et al.
(1985) reported temperatures of up to 1000 °C. Bakun-Czubarow (1991a) described the presence of former coesite and peak metamorphic temperatures between 860 and 960 °C, which suggest that peak-pressure conditions for the Snieżnik, Międzygórze and Złoty units are not significantly different (Bröcker and Klemd 1996).

Using the Sm-Nd method (grt-cpx-whole rock), three eclogites from the Snieżnik and Międzygórze units were dated at 341 ± 7 Ma, 337 ± 4 Ma, and 324 ± 6 Ma (Brueckner et al. 1991). For one eclogite from the Złoty unit this method yielded an age of 352 ± 4 Ma.

### Sample location and petrography

Due to the extremely poor level of exposure in the Złoty unit, suitable samples could only be collected at Stary Gierałtow (Fig. 2), where eclogites occurs as 1 to 20 cm thick intercalations and/or patchy domains within granulites. This outcrop (up to 3 m in length and 1 m in width) is described in detail by Smulikowski (1967), Smulikowski and Bakun-Czubarow (1973) and Bakun-Czubarow (1991b). They mainly distinguished three rock types: (1) felsic granulites; (2) mafic granulites with small domains of (3) eclogite. Individual granulite layers range in thickness from several decimetres to one centimetre or millimetre. Six samples from this succession were studied in detail (Table 1).

The mafic granulites exhibit the subassemblage garnet, omphacite, quartz, and plagioclase, with biotite, rare white mica as inclusions in omphacite and kyanite as additional phases in varying modal proportions (400 points/section; Table 1). Kyanite either occurs as inclusions in garnet or in the matrix, often showing a thin corona of garnet (Fig. 3). Typical accessories are apatite, rutile, zircon and Fe-oxides. Zoisite, present in most eclogites of the other tectonic units (Bröcker and Klemd 1996) is absent in all investigated samples. The presence of coesite pseudomorphs is inferred from radial fractures around polycrystalline quartz inclusions in garnet, which were recognised in samples 1107, 1106 and 1041.

#### Table 1 Modal mineralogy (vol.%) of the Snieżnik Mts. (± present); 1107 contains additional antiperthite. Abbreviations according to Kretz (1983)

<table>
<thead>
<tr>
<th>Sample</th>
<th>1041</th>
<th>1041X</th>
<th>1041Xe</th>
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<th>1106a</th>
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<td>15</td>
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<tr>
<td>Bt</td>
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<td>2</td>
<td>-1</td>
<td>-1</td>
<td>+</td>
</tr>
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<td>+</td>
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**Fig. 2** Simplified geological map of the Snieżnik Mts. in SW Poland after Don et al. (1990) and Smulikowski (1967). I Międzygórze unit, II Snieżnik unit, III Złoty unit, I Mesozoic sediments, 2 Snieżnik augen gneisses, 3 Gierałtow gneisses, 4 Stroń-Młynowiec group, 5 granulites, 6 Variscan granitoids. SB Stronie Słąskie-Bielice fault, SM Stare Mesto fault, SK Stare Kletno fault, RU Ramzovian/Moldan-ubian thrust. Filled dots eclogite localities.

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**Fig. 3** Kyanite displaying a slightly kinked cleavage and rutile inclusion (dark, high relief) is surrounded by a garnet corona (Grt) also with rutile inclusions. Omphacite relics (arrows) surround the garnet coronas. Sample 1106; scale = 0.07 mm; 1 Nicol.

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**Fig. 4** Inclusions of coarse grained polycrystalline quartz and relictic palisade quartz (arrow) after coesite in garnet. Sample 1041; scale = 0.1 mm; crossed Nicols.
Important microstructural and mineral compositional characteristics

Many details of the petrography and mineral chemistry were already described in previous studies of Smulikowski (1967), Smulikowski and Bakun-Czubarrow (1973), Bakun-Czubarrow (1991b) and Kryza et al. (1996) and are thus not repeated here. Major compositional characteristics of the investigated samples are summarised in Table 2.

For our interpretation, the following observations are important: The jadeite content of omphacite in mafic granulites and eclogite varies between 25 and 35 mol%, in some retrograde rims it decreases to 18 mol%. The subhedral omphacite in the granulites occurs either poikiloblastic with plagioclase (Pl 1) and biotite inclusions, or as spatial vermicular-like intergrowth always surrounding Pl 1 and quartz (Fig. 5c, d). The Pl 1 and quartz inclusions do not show a preferred orientation nor a common extinction thus negating an exsolution from the omphacite host. The Pl 1 displays an An-content between 10 and 14 mol%. The Pl 1 and quartz inclusions are often intimately intergrown with biotite and phengite, the latter being absent from the matrix. Since Pl 1 is formed at the expense of omphacite and quartz, a retrograde formation of Pl 1 and quartz inclusions is unlikely, because both occur in almost equal volume amounts. Consequently, these inclusions are interpreted to belong to an earlier lower-pressure stage. Almost all relict Pl 1 is consumed by the formation of inclusion-free, sub- to idioblastic omphacite with straight grain boundaries close to the eclogite domain. The replacement of Pl 1 by inward-directed growth of omphacite indicates that the central parts of the omphacite may postdate the rims. Further recrystallisation produced inclusion-free, subidio- to idioblastic omphacite in the eclogite domains (Fig. 5a, b). As already pointed out, in all mafic granulites some garnet texturally occurs as coronas around kyanite separating it from omphacite, indicating that this garnet crystallised at the expense of kyanite and omphacite (Fig. 3). However, it should be noted that the mineral inclusions in omphacite are quite different than those ones in garnet. Subidioblastic matrix Pl 2 occurs in almost all samples and shows a weak zonation with an An-content ranging from 12 to 15 mol%. In contact with garnet or garnet coronas the An-values of plagioclase (Pl 3) are somewhat higher, between 15 and 22 mol%, and the zoning is inverted showing the highest An-values towards the rim.

The observed textural features indicate that omphacite crystallised at the expense of pre-eclogite-facies Pl 1. The recrystallised subidioblastic omphacites are generally unzoned, while the xenoblastic inclusion-rich omphacites display unsystematic zoning patterns. On rare occasions, an uniform increase in Na at the expense of Ca towards the fringed Pl 1 domain can be observed. The homogenisation of the subidioblastic omphacites indicates that volume diffusion was enhanced during recrystallisation and grain boundary migration. Apart from the thin eclogite domains where mineral reactions were completed, we interpret the textural relations between omphacite and Pl 1 inclusions to represent an incomplete prograde omphacite crystallisation. Consequently, the main question arises why the eclogite-forming reactions were completed in the eclogite domains and not in the surrounding coronitic granulites. Bulk chemical differences allow for three compositional garnet groups to be distinguished: (a) almandine-rich garnets (felsic granulite) with the compositional range Alm$_{55-65}$Grs$_{23-33}$Prp$_{10-20}$; (b) garnets with intermediate almandine content (mafic granulite) in the range Alm$_{50-55}$Grs$_{23-33}$Prp$_{13-25}$; (c) almandine-poor garnets

Table 2 Condensed analyses of minerals used in the geothermobarometric calculation. Micas based on 22 oxygens, clinopyroxene on 6 oxygens. Abbreviations according to Kretz (1983)

<table>
<thead>
<tr>
<th></th>
<th>Matrix garnet</th>
<th>Matrix biotite</th>
<th>Inclusions in garnet</th>
<th>Pl 2</th>
<th>Feldspar</th>
<th>Omphacite</th>
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<td>$X_{\text{Mg}}$</td>
<td>$X_{\text{Na}}$</td>
<td>$X_{\text{Mn}}$</td>
<td>$X_{\text{Fe}^{2+}}$</td>
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<td>0.20</td>
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$^a$ Inclusion in garnet
with almost equal proportions of grossular and pyrope (mafic granulate and eclogite) which display the compositional range Alm_{43-50}Grs_{22-32}Prp_{22-30}. The garnets from the eclogite domains have often a lower pyrope content than those from the mafic granulites (Table 2). This overall pyrope decrease is a result of the increased modal amounts of garnet and omphacite in the eclogite domains. The garnet in sample 1107 displays an extensive prograde zoning pattern on a millimetre scale, which is revealed by a smooth decrease in the pyrope/almandine ratios and an increase of the grossular and spessartine content from rim to core domain (Fig. 6a). Similar zoning patterns for the mafic granulates were observed by Bakun-Czubarow (1991b) and Kryza et al. (1996). However, the garnet coronas around kyanite in the eclogitised samples (1041, 1106) usually display no or just a weak zonation as is shown by the decreasing grossular content and an increasing almandine and pyrope content towards the rims adjacent to Pl 3. The Fe/(Fe + Mg) ratio remains almost constant or slightly decreases towards the rim (Fig. 6b). This implies only a small temperature change during corona growth of the garnet. Some inclusion-free garnets of the eclogites and mafic granulates display a similar zonation towards the outer rim, while the core domain is largely homogeneous (Fig. 6c). In some instances the outermost rims of garnet in contact with biotite show weakly developed reversed zoning profiles with decreasing pyrope/almandine ratios and increasing grossular contents (Fig. 6d). This pattern is considered to evidence limited re-equilibration as a result of Mg-Fe exchange between garnet and biotite during exhumation. No zonation pattern was recognised regarding the spessartine component.

The preservation of some growth zoning in the garnets of the felsic granulate is very important for the understanding of the reaction kinetics during the \( P-T \) evolution of these rocks, since intergranular diffusion (causes the homogenisation of the garnet composition) usually starts at temperatures of around 750 °C (Spear 1988). It is furthermore unusual for the growth zonation in the garnets in the ultrahigh-pressure granulites to be preserved better than in the garnets of eclogites from the

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**Fig. 5**

a) Eclogite domain consisting exclusively of inclusion-free garnet (grey) and omphacite (dark) with rare kelyphitic rims (black). Only locally omphacite contains vermicular quartz and plagioclase inclusions (arrows). Sample 1041 Xe; scale = 1.4 mm; 1 Nicol.

b) Eclogite-textured mafic granulate with Pl 2 and quartz (light, low relief) and biotite (arrows), which occurs in vein-like structures between omphacite (dark) and garnet (grey). Sample 1041; scale = 1.4 mm; 1 Nicol.

c) Subidioblastic omphacite (dark) with thin kelyphitic rims (black), and few vermicular Pl 1 and quartz inclusions (arrows). Omphacite and garnet (grey) occur intimately intergrown with Pl 2 and quartz (light). Sample 1041; scale = 0.4 mm; 1 Nicol.

d) Poikiloblastic, anhedral omphacite (dark) with vermicular Pl 1 and quartz inclusions (light) as well as anhedral, round inclusion-free garnet (grey) are surrounded by subidioblastic matrix quartz and Pl 2 (light). Sample 1106; scale = 1.4 mm; 1 Nicol.
other tectonic units, although these rocks display much lower temperatures when compared to the granulites (see below).

Antiperthitic *feldspars* mostly occur as small inclusions in the rim zone of garnet and only rarely within the matrix of the felsic granulite. Potassium-feldspar on the other hand occurs mostly intergrown with Pl 2. The antiperthite comprises regular rod- to spindle-like K-feldspar exsolution lamellae (<20 μm wide). To determine the bulk composition of the antiperthite, grains with homogeneous exsolution lamellae were analysed with a broad electron beam. The resulting alkali feldspar composition was Ab$_{65.4}$Or$_{26.2}$An$_{8.4}$. Only homogeneous albite (Ab$_{96-99}$Or$_{1-2}$An$_{10-12}$) next to K-feldspar (Ab$_{12-15}$Or$_{85-88}$An$_{0.4-0.6}$) occurred as inclusions in singular garnets. Obviously albite and K-feldspar inclusions are breakdown products during cooling of the initial ternary feldspar. Their composition is similar to those of the matrix K-feldspar (Ab$_{9-12}$Or$_{88-91}$An$_{0.1-0.5}$) and Pl 2 (Ab$_{94-88}$Or$_{2-3}$An$_{10-14}$) suggesting that these feldspars were also formed during the ternary feldspar breakdown.

The composition of the ternary feldspar is in excellent agreement with the composition of an original ternary feldspar (Ab$_{67}$Or$_{27}$An$_{6}$) in similar felsic granulites from the Złote unit as deduced by integrated line scan analysis from an anthiperthitic plagioclase (Kryza et al. 1996). Therefore, the ternary feldspar is interpreted to be a part of the peak metamorphic mineral assemblage.

**P-T conditions during ultrahigh-pressure metamorphism**

The textural-equilibrium mineral assemblage of the eclogite includes garnet-omphacite-rutile-quartz. The mafic granulites display the peak metamorphic mineral

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**Fig. 6a–f** Compositional zoning patterns in garnets of the mafic granulites of the Złote unit (for details see text)
assemblage garnet-kyanite-omphacite-quartz-Pl 2, where-as garnet(rim)-Pl 3-omphacite rim (when zoned) belong to a lower-pressure stage. Retrograde mineral phases are biotite and adjacent garnet (outermost rims). The felsic granulite displays the textural-equilibrium assemblage garnet(rim)-kyanite-ternary alkali feldspar-quartz. Retrograde phases are Pl 2 and K-feldspar as well as minor biotite.

Temperature conditions

The onset of prograde omphacite crystallisation probably proceeded somewhere between 500 and 700 °C, as is indicated by the presence of pre-eclogitic biotite, phengite (Si up to 3.33 pfu) and Pl 1 inclusions in omphacite in mafic granulites (see Goldsmith 1982; Spear 1989). As is outlined below, garnet nucleated at higher P-T conditions, which is probably due to the fact that the activation energy for garnet nucleation is much higher than that for clinopyroxene (Rubie 1990). This explains the different mineral inclusions in garnet and omphacite. To avoid the large uncertainty when considering appropriate activity models for omphacite and garnet by using thermodynamic datasets, we obtained absolute temperature estimates for the eclogite and mafic granulites by Mg-Fe exchange equilibrium, using the garnet-clinopyroxene geothermometers (Ellis and Green 1979; Krogh 1988; Fig. 7). All temperatures were calculated using Fe after correction for Fe3+, which was based on charge-balance criteria. Calculated temperatures show a large range due to the presence of variable amounts of Fe3+ and/or some re-equilibration during retrogression. The temperature estimates for the eclogite and mafic granulites were almost identical. Temperatures based on the Krogh calibration are 5 to 60 °C lower than the Ellis and Green calibration. Using rim-rim compositions of homogeneous garnets and unzoned omphacites, the Ellis and Green calibration indicates peak temperatures that range between 790-980 °C and 870–1040 °C for assumed pressures of 20 and 30 kbar (18 mineral pairs) respectively. However, the Fe/(Fe+Mg) ratio in garnets almost stays constant, indicating only a small temperature change during growth of the core and rim of the garnet. The absence of a prograde growth zonation in most of the garnets either implies that volume diffusion obliterated all zoning features, or that these garnets grew entirely during the peak of metamorphism at temperatures >800 °C.

The two-feldspar geothermometer of Fuhrman and Lindesly (1988) was applied to the felsic granulite 1107. The high-temperature ternary feldspar should lie on the same solvus as a hypothetical coexisting feldspar with equal An-, Ab- and Or-activities of all three components at a given temperature and pressure (Kryza et al. 1996; O’Brien et al. 1997). The composition of the hypothetical feldspar on the opposite limb of the solvus is composition Ab56Or38An6 yielding a minimum temperature of 990 °C at 27 kbar. Recrystallised matrix K-feldspar and plagioclase which show triple junctions with adjacent quartz display much lower temperatures between 540 and 660 °C at 10 to 15 kbar.

Pressure conditions

The occurrence of possible coesite pseudomorphs as inclusions in garnet and signs for the former presence of non-stoichiometric clinopyroxenes (Bakun-Czubarow 1991a, 1992) even indicate ultrahigh-pressure conditions in the Śnieżnik Mts. However, it is uncertain at what specific equilibrium temperature coesite was formed. The coesite pseudomorphs always occur as inclusions in garnet and, thus, may indicate passage through the coesite stability before peak-temperature conditions were reached (also see Bakun-Czubarow 1991a).

During prograde metamorphic conditions at least six plagioclase breakdown reactions can be proposed for eclogites and mafic granulites (Bucher and Frey 1994; abbreviations according to Kretz, 1983):

1. Ab = Jd + Qtz;
2. An = CaTs + Qtz;
3. 3 An = Grs + 2 Ky + Qtz;
4. 3 An + 4 Tr = 3 Prp + 11 Di + 7 Qtz + H2O;
5. 3 An + 3 Di = 2 Grs + Prp + 3 Qtz; or
6. 4 En + An = Prp + Di + Qtz.

A pressure estimate independent of the H2O-activity can be deduced from reaction 1 (Holland 1980, 1983). The plagioclase-breakdown reaction produces a Ca-tschermak component which is accommodated by the omphacite (reaction 2). As is shown above, this plagioclase-breakdown reaction is clearly an early prograde reaction which only proceeded to completion in the eclogite domains (plagioclase absence). Therefore in the

![Fig. 7 P-T estimates for mafic granulites and eclogites. Sample numbers in parentheses. See text for detailed description of the different equilibria and references. (HP high-pressure stage, Si = 3.33 phengite barometry)](image-url)
absence of plagioclase, the omphacite_{4035}-quartz assemblage indicates only minimum pressures of about 20 kbar at 900 °C for the eclogite. Due to the lack of other pressure sensitive mineral equilibria no further pressure estimates are possible for the eclogite. In the mafic granulites omphacite with a jadeite content of 35 mol% immediately adjacent to relictic Pl 1 (An_{14}) inclusions indicates equilibrium pressures of 10.5–14.5 kbar (Fig. 7), at temperatures between 500–700 °C. The Si-content of the phengite inclusions in omphacite indicates minimum pressures between 7 and 11 kbar (Massonne and Schreyer 1987). A pressure of 21.5 kbar at 1000 °C is given by the maximum jadeite content (35 mol%) of omphacite and Pl 2 (An_{15}) for samples 1041 and 1106. Further, the plagioclase-garnet-kyanite-quartz (GASP) barometer based on reaction 3 or the plagioclase-clinopyroxene-garnet-quartz (GADS) barometer based on reaction 5 can be used to deduce pressures. As discussed in the previous section, homogeneous garnet cores and homogeneous garnets seem to have entirely grown during the peak of metamorphism and are interpreted to have been equilibrated with adjacent omphacite, kyanite and Pl 2. The internally consistent data set of Berman (1988) and mixing models outlined below were used for the calculations. The activity model of Berman (1990) was applied for garnet, while the mixing model of Fuhrman and Lindsley (1988) was used for plagioclase. The mixing model of Holland (1983) was used for omphacite. Quartz and kyanite activity are 1. Pressures calculated by the GASP barometer for samples 1106 and 1041 are 26 and 28 kbar at 1000 °C, while pressures derived by the GADS barometer are significantly lower, ranging between 21 and 23 kbar at the same temperature. Pressures are almost identical when using the formulation of Newton and Haselton (1981) and Newton and Perkins (1982). The GASP barometer for the felsic granulite (sample 1107) which contains high-T ternary feldspar but no omphacite, reveals a pressure of 27 kbar at 1000 °C using garnet rim composition (highest Mg/Fe) and the anorthite component of the ternary feldspar. The corresponding pressure estimates derived by the GASP barometer for the different samples strongly advocate equilibrium conditions for the chosen mineral assemblages (Fig. 7).

In summary, geothermobarometric evaluation for matrix assemblages of the eclogites and granulites reveals peak metamorphic conditions between c. 21 to 28 kbar at 800 to 1000 °C.

**Post eclogite-facies conditions**

The investigated granulites and associated eclogite domains from the Złote unit have undergone remarkably little retrograde re-equilibration during post eclogite-facies conditions. Almost no retrograde overprint is displayed by the breakdown of primary silicates in all samples. In some mafic granulites matrix biotite appears to be in textural equilibrium with garnet and omphacite, however, it frequently occurs as vein-like intergrowths along the grain boundaries of these minerals (Fig. 5b). This suggests that this biotite formed during retrograde conditions through the consumption of garnet. Although no ternary feldspar and/or K-feldspar were found in this rock type, the biotite may be a breakdown product of garnet and former potassium-rich feldspar.

In rare cases, omphacite displays a thin retrograde rim. The jadeite content decreases to 18 mol% indicating pressures of 8–12 kbar at temperatures between 500 and 740 °C. Steltenpohl et al. (1993) calculated temperatures of 630 ± 50 °C at pressures of 11 ± 1.5 kbar for secondary garnet-biotite and garnet + hornblende + plagioclase + quartz assemblages in the granulites. Application of the garnet-biotite geothermometer (Perchuk and Lavrent’eva 1983) for garnet rims in contact with biotite in samples 1041Xe and 1041 revealed somewhat higher temperatures of 710 ± 45 at 10 kbar. The biotite composition was not corrected for Fe^{3+}.

Bröcker and Klemd (1996) calculated various P-T points reflecting the different stages of recrystallisation during the exhumation process of eclogites from the Śnieżnik and Międzygórze units, which range between 4 and 11 kbar at 600 and 650 °C.

**Timing of metamorphism**

For geochronology (U-Pb, Sm-Nd) we have selected two mafic granulites (samples 1106 and 1041). Isotope analyses were carried out at the “Zentrallaboratorium für Geochronologie” in Münster. Laboratory methods and details of data evaluation are described in Bröcker et al. (1998) and Witt-Eickschen and Kramm (1998). Analytical results are listed in Tables 3 and 4.

Previous work by Brueckner et al. (1991) yielded a Sm-Nd (Grt-Cpx-WR) age of 352 ± 4 Ma for an eclogite from Stary Gieraltów. For mafic granulites, we obtained broadly similar Sm-Nd garnet – whole rock ages (341 ± 10 Ma and 343 ± 11 Ma), with high errors due to the low spread in ^{147}Sm/^{144}Nd ratios (Table 3). The closure temperature for the Sm-Nd system in garnet is estimated at c. 600 °C (Mezger et al. 1992). Thus, the Sm-Nd garnet ages most likely correspond to a post-peak metamorphic stage of the cooling path, but still at eclogite-facies conditions. By use of the U-Pb zircon method, we have attempted to constrain timing of peak metamorphic conditions, because the high isotopic closure temperature of zircon (>900 °C, e.g. Mezger and Krogstad 1997) closely corresponds to the estimated maximum metamorphic temperatures (800–1000 °C). Conventional U-Pb multigrain dating of zircon was applied to six unabraded size fractions of sample 1106. The studied zircon population consists mostly of prismatic grains with rounded terminations and very small amounts of rounded zircons (Fig. 8a–d). The morphological characteristics most likely are related to various
Table 3 Sm-Nd analytical results for whole rock samples and garnet from Stary Gieraltów

<table>
<thead>
<tr>
<th>Sample</th>
<th>Type</th>
<th>Sm (ppm)</th>
<th>Nd (ppm)</th>
<th>(^{147}\text{Sm}/^{144}\text{Nd})</th>
<th>(^{143}\text{Nd}/^{144}\text{Nd})</th>
<th>Age (Ma, 2(\sigma))</th>
</tr>
</thead>
<tbody>
<tr>
<td>1106</td>
<td>Whole rock</td>
<td>6.003</td>
<td>24.22</td>
<td>0.1498</td>
<td>0.512322 (12)</td>
<td>341 ± 10</td>
</tr>
<tr>
<td>1041</td>
<td>Whole rockd</td>
<td>9.134</td>
<td>10.32</td>
<td>0.5457</td>
<td>0.513210 (19)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Garnet</td>
<td>7.243</td>
<td>25.61</td>
<td>0.1710</td>
<td>0.512491 (12)</td>
<td>343 ± 11</td>
</tr>
<tr>
<td></td>
<td>Whole rock</td>
<td>7.258</td>
<td>25.71</td>
<td>0.1707</td>
<td>0.512479 (13)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Garnet</td>
<td>10.22</td>
<td>11.13</td>
<td>0.5551</td>
<td>0.513347 (19)</td>
<td></td>
</tr>
</tbody>
</table>

\(^{147}\text{Sm}/^{144}\text{Nd}\) ratios were assigned uncertainties of 0.3%.

Numbers in parentheses indicate uncertainties in ratios quoted at the 2\(\sigma_m\) level.

Ages were calculated using linear least square regression (York 1969).

Duplicates were prepared from a different aliquot of the same whole rock powder. Frequent measurement of the LaJolla standard yielded a \(^{143}\text{Nd}/^{144}\text{Nd}\) ratio of 0.511844 ± 30 (n = 24).

Table 4 U-Pb analytical results for zircon of the mafic granulite 1106 from Stary Gieraltów

<table>
<thead>
<tr>
<th>Size ((\mu)m)</th>
<th>Weight (mg)</th>
<th>U (ppm)</th>
<th>Pb(^{total}) (ppm)</th>
<th>(^{206}\text{Pb})/(^{238}\text{U})</th>
<th>(^{207}\text{Pb})/(^{235}\text{U})</th>
<th>(^{207}\text{Pb})/(^{206}\text{Pb})</th>
<th>Correlation coefficient</th>
<th>(^{206}\text{Pb})/(^{208}\text{Pb})</th>
<th>(^{207}\text{Pb})/(^{208}\text{Pb})</th>
</tr>
</thead>
<tbody>
<tr>
<td>250–200</td>
<td>2.154</td>
<td>191</td>
<td>10.6</td>
<td>9.3</td>
<td>0.08752</td>
<td>15880</td>
<td>0.05663 (17) 0.4200 (13)</td>
<td>0.05379 (3)</td>
<td>0.987</td>
</tr>
<tr>
<td>200–160</td>
<td>1.434</td>
<td>183</td>
<td>10.2</td>
<td>8.9</td>
<td>0.09316</td>
<td>11741</td>
<td>0.05675 (17) 0.4204 (17)</td>
<td>0.05373 (15)</td>
<td>0.744</td>
</tr>
<tr>
<td>160–125</td>
<td>1.500</td>
<td>197</td>
<td>9.6</td>
<td>9.6</td>
<td>0.08782</td>
<td>14057</td>
<td>0.05694 (17) 0.4228 (14)</td>
<td>0.05385 (7)</td>
<td>0.928</td>
</tr>
<tr>
<td>125–100</td>
<td>1.109</td>
<td>209</td>
<td>11.5</td>
<td>10.1</td>
<td>0.08722</td>
<td>10568</td>
<td>0.05627 (17) 0.4180 (13)</td>
<td>0.05388 (4)</td>
<td>0.975</td>
</tr>
<tr>
<td>80–62</td>
<td>0.921</td>
<td>305</td>
<td>16.7</td>
<td>14.7</td>
<td>0.08461</td>
<td>11311</td>
<td>0.05607 (17) 0.4157 (13)</td>
<td>0.05376 (4)</td>
<td>0.974</td>
</tr>
<tr>
<td>&lt;40</td>
<td>1.448</td>
<td>370</td>
<td>20.4</td>
<td>18.0</td>
<td>0.08083</td>
<td>23977</td>
<td>0.05641 (17) 0.4196 (13)</td>
<td>0.05395 (2)</td>
<td>0.974</td>
</tr>
</tbody>
</table>

a Radiogenic lead only
b Measured ratios
c Corrected for fractionation, spike, blank and initial common lead. Numbers in parentheses indicate uncertainties in ratios quoted at the 2\(\sigma_m\) level.
d Ages in Ma (2\(\sigma\))

degrees of zircon dissolution during high-grade metamorphism (e.g. Kröner et al. 1998). All grains were clear, without visible inclusions. Cathodoluminescence (CL) studies provided no indication for magmatic zonation or the presence of inherited cores; however, the grains studied are not homogeneous but exhibit irregular shaped patches with light and dark grey CL colours.

The uranium concentrations of the four grain size fractions > 100 \(\mu\)m are little different and range between 183–209 ppm; the smaller size fractions (80–62 \(\mu\)m and < 40 \(\mu\)m) have higher uranium contents of 305 ppm and 370 ppm, respectively (Table 4). All fractions are discordant and plot very close to the concordia curve near to a poorly defined upper intercept around 370 Ma (Fig. 9). The small spread in \(^{206}\text{Pb}/^{238}\text{U}\) and \(^{207}\text{Pb}/^{235}\text{U}\) does not allow to calculate a geological meaningful discordia chord. As a result, an upper concordia intercept cannot clearly be established, but a date of 370 to 375 Ma is broadly constrained. The \(^{207}\text{Pb}/^{206}\text{Pb}\) ages range between 360 to 369 Ma. Since internal structures and morphology are typical for zircons modified during high-grade metamorphism, we interpret the zircon dates as best approximation for primary crystallisation from a melt, which took place either before or during early stages of high-grade metamorphism. Due to the fact that unambiguous indications for newly grown metamorphic zircons were not found, timing of peak metamorphic conditions could not be constrained. In order to find the answer to this question, a more detailed U-Pb zircon study, using both conventional dissolution and the vapour-digestion technique, is currently in progress.

Fluid inclusions

Fluid inclusions were examined using the U.S. Geological Survey gas flow heating/freezing stage (for description see Roedder 1984) at the Universität Bremen. A detailed description of the microthermometry and calibration procedure is given in Klemd (1989) and Klemd et al. (1992). Densities and isochors for the H\(_2\)O-salt inclusions were calculated using the equation of state for H\(_2\)O-NaCl (Zhang and Frantz 1987). Salinities were determined by ice-melting depressions using the experimental data of Potter et al. (1978). Raman microspectroscopy on representative fluid inclusions as performed at the University of Göttingen. A detailed description of the method used is given in Klemd et al. (1992).

Only 120 fluid inclusions were found in 15 thick sections of the samples 1041, 1041X, 1041Xe, 1107 and 1106 from Stary Gierałtów, although more than 40 thick sections were investigated. They exclusively occur in matrix quartz. Despite an intensive search no inclusions were found in omphacite and garnet. The inclusions are usually ≤12 \(\mu\)m in diameter, four have a size of up to 24 \(\mu\)m. The shape of all inclusions ranges from irregular to spherical, and negative crystal shapes also occur. All inclusions are aqueous, simple two-phase inclusions consisting of an aqueous liquid and a vapour bubble. They mostly occur along healed fractures which may or may not crosscut grain boundaries irrespective of the fluid composition. The fluid inclusions are extremely rare and no fracture intersections are present. It was therefore
impossible to establish a relative age relationship using petrographic observations. All of these inclusions are consequently regarded as secondary or pseudosecondary. Carbondioxide-bearing inclusions were not observed. Apart from sample 1107 which exclusively contains high-salinity inclusions, all other samples contain both low- as well as high-salinity fluid inclusions.

The aqueous inclusions constitute two compositional groups. The first group (low-salinity inclusions) has final melting temperatures (ice melting) between

Fig. 8a–d Scanning electron photomicrographs showing zircon morphologies in sample 1106: a, b Short-prismatic grains with rounded, recrystallised terminations; c, d irregular rounded grains related to metamorphic corrosions, surface pitting and recrystallisation. In all cases, crystal faces are not (a, c, d) or only poorly preserved (b).

Fig. 9 Concordia diagram for zircons of sample 1106 from Stary Gieraltów. Error ellipse of grain-size fraction A is hidden behind B and not labelled.

Fig. 10 Relationship between homogenization temperatures (T_h) and final melting temperatures (T_{mf}) of fluid inclusions.
that, apart from Na+ and K+, divalent ions such as Ca2+ and Mg2+ are also present. Cotectic and clathrate melting were not observed. Homogenisation temperatures (Th) between 100 and 290 °C are essentially the same as for the first group (Fig. 10). Densities range between 0.92 and 1.08 g/cm³. For both groups of inclusions homogenisation is always into the liquid phase. Raman spectroscopy on representative inclusions did not reveal the presence of minor gases such as CO2 and/or N2. The variation of homogenisation and melting temperatures in the low-salinity and high-salinity inclusion group could be due to autodecuppitation during exhumation conditions of at least two generations of fluid inclusions, namely low- and high-salinity inclusions. Such a process would explain the variation of melting and homogenisation temperatures and, hence, the resulting density range. This is supported by the fact that fluid inclusions along healed fractures and in clusters do not show compositional differences, as would have been expected (e.g. Touret 1981). However, modified fluid inclusions of the same shape should reveal a correlation between Th and inclusion size, since larger inclusions, as a result of pressure differences between internal pressure (within the inclusion) and confining pressure, re-equilibrate before smaller ones during exhumation (e.g. Bodnar et al. 1989; Hall et al. 1991; Klemd et al. 1995). Such a correlation was not observed in the here investigated samples. However, it should be noted that recent theoretical findings of Küster and Stöckhert (1997) seriously challenge this interpretation. They consider decrepitation (failure by fracturing) of fluid inclusions in natural rocks not to be feasible above 300–350 °C and thereby question all former experimental results. Also, a preferential loss of H2O during re-equilibrium of the smaller inclusions as described by Hall and Sterner (1993) cannot be proved, because there is no correlation between inclusion size and salinity. Another possible explanation for the salinity range is the entrapment under different P-T conditions of more than two generations of fluids. The fact that some inclusions seem to have decrepitated or stretched during the post-amphibolite-facies conditions and others not, is probably a result of the individual fluid density, geometry as well as size of inclusions.

Discussion and conclusions

Metamorphic evolution

The eclogite mineral assemblage omphacite-garnet-quartz-rutile indicates temperatures of 800 to 1000 °C at minimum pressures of about 20 kbar. Mafic granulites contain the mineral parageneses omphacite-garnet (cores)-kyanite-quartz-rutile-Pl 2 which constrain equilibrium pressures of 20–28 kbar for this temperature range. Pressures of at least 27 kbar at 990 °C were estimated for the felsic granulite with the peak metamorphic mineral assemblage garnet(rim)-kyanite-antiperthite-quartz. Pseudomorphs after coesite indicate that even higher pressures were reached at some stage of the metamorphic evolution. The absence of garnet-breakdown reactions to orthopyroxene and sillimanite-bearing assemblages indicate an overstepping of such reactions due to unfavourable reaction kinetics during rapid isothermal decompression. These features and the lack of melting processes are remarkable when considering temperatures of up to 1000 °C during peak metamorphism.

The Pl 1, phengite, quartz, and biotite inclusions in omphacite are interpreted as relic phases related to an early stage of the prograde metamorphic evolution. As is indicated by the petrographical and mineral chemical evidence, the garnet coronas around kyanite in the mafic granulite are the result of the prograde reaction: 3 Di + 2 Ky = Grt + 2 Qtz close to peak metamorphic conditions (Fig. 11). In accordance with the prograde garnet zoning in the felsic granulite, this implies a clockwise P-T path (Fig. 12), which is characterised by the presence of kyanite and the absence of sillimanite.

![Fig. 11 Schreinemakers grid depicting a CMAS invariant point and related univariant reaction curves relevant to the reaction textures preserved in the mafic granulites. The generalised prograde P-T path is shown by the large arrow. The diagram was constructed using the thermodynamic dataset of Berman (1988) and calculated mineral phase activities (sample 1041)](image-url)
domains. Granulites during ultrahigh-to bulk chemical differences compared to the eclogite (Zuborow 1991b). Therefore, we assume that the persistence of matrix Pl 2 and ternary feldspar in the granulites, while the composition of the felsic granulites is as well as higher SiO2-content than the associated eclogites, which occurred during amphibolite-facies conditions at temperatures as low as 550 to 660 °C. Densities of fluid inclusions which were trapped or retrapped during recrystallisation of the quartz grains caused by infiltration of a CO2 ‹ N2-rich fluid phase (e.g. Johnson and Hollister 1995). Fluid inclusions observed in the lower-temperature (600–800 °C) eclogites in the Śnieżnik and Międzygorze unit (Klemd et al. 1995), and with the fact that all observed fluid inclusions in the granulites and eclogites of the Złote unit are aqueous. Furthermore, some H2O-rich fluid must have been present locally in the recrystallised eclogite domains where volume diffusion and thus grain boundary diffusion were more efficient. It is however highly unlikely that those fluid inclusions, if formed at ultrahigh-pressure conditions, would have survived the isothermal decompression during exhumation, and the subsequent static recrystallisation of the quartz grains which occurred during amphibolite-facies conditions at temperatures as low as 550 to 660 °C. Densities of fluid inclusions which were trapped or retrapped during recrystallisation will probably reflect those P-T conditions (e.g. Johnson and Hollister 1995).

Considering the anhydrous mineral assemblage of the investigated rocks, a fluid phase causing the stabilisation

**Fig. 12 P-T path for eclogites and mafic granulites from the Śnieżnik Mts.** Generalized exhumation path of the ultrahigh-pressure rocks from the Złote unit after Steltenpohl et al. (1993). H2O-isochores for minimum and maximum density (g/cm3) of high-salinity (continuous line) and low-salinity (interrupted line) inclusion fluids. HP = high-pressure stage, as taken from Fig. 9 (Jd18 + Qz = Ab10, 2 amphibolite-facies conditions after Steltenpohl et al. (1993), and Borkowska (1996)]. For further references and details see text

This is typical for a continent-continent collision setting, not for subduction or collision-magmatism tectonics (Bucher and Frey 1994).

A considerable problem concerns the postulated stability of plagioclase and antiperthite at pressures between 21 and 28 kbar in the mafic and felsic granulites. A major constraint concerning the plagioclase stability is the bulk rock chemistry of the igneous precursor rock (Green and Ringwood 1967). Plagioclase (Ab-rich) may be stable up to 29 kbar at 1000 °C in eclogite-facies rocks with low Ca/Na ratios, high SiO2- and Al2O3-contents, and high Na2O-contents (Green and Ringwood 1967; Ringwood 1975). The mafic granulites of the Złote unit have a lower Ca/Na ratio and Na2O-content as well as higher SiO2-content than the associated eclogites, while the composition of the felsic granulites is similar to rhyodacite (Smulikowski 1967; Bakun-Czubarow 1991b). Therefore, we assume that the persistence of matrix Pl 2 and ternary feldspar in the granulites during ultrahigh-P-T conditions is mainly due to bulk chemical differences compared to the eclogite domains.

Fluid inclusions

Maximum and minimum isochores were constructed for the highest- and lowest-density aqueous inclusions (Fig. 12). The isochores of the highest densities of both groups pass through the P-T field of the estimated amphibolite-facies conditions and through or just below the peak P-T estimate. There are at least three possible explanations: (1) the high-density inclusions are unmodified relics of fluids that were present during or close to the peak of metamorphism; (2) the high-density inclusions were trapped or retrapped at amphibolite-facies conditions; (3) most high-density inclusions were trapped after amphibolite-facies conditions. We regard the second option as the most probable one, because fluids at peak temperatures (800–1000 °C) (option 1) should reveal a low H2O-activity. Otherwise, the granulites and eclogite domains should show some evidence for partial melting. Further, a low H2O-activity explains the presence of the almost completely anhydrous ultrahigh-pressure mineral assemblage of the Złote unit. Another cause of a low H2O-activity would be the dilution of H2O by a free non-aqueous fluid phase such as CO2 and/or N2. Many HT-granulites world-wide contain high-density CO2 ‹ N2-fluid inclusions thus implying the development of anhydrous mineral assemblages by means of dehydration caused by infiltration of a CO2 ‹ N2-rich fluid phase (e.g. for extensive discussion and reference list see Touret 1992, 1993 and “Introduction” chapter). However, almost all of those granulites show peak-metamorphic conditions between 750–900 °C at 4–12 kbar and many furthermore show a P-T-t path suggesting magmatic accretion. The very high pressure granulites of the Złote unit on the other hand were subjected to much higher eclogite-facies conditions in the range of 800 to 1000 °C at 21 to 28 kbar and display a clockwise P-T-t path (see above). Eclogite-facies metamorphism is typically accompanied by a fluid regime which is dominated by H2O and in places by N2 ‹ CO2 (e.g. Touret 1992; Andersen et al. 1993). This is in accordance with H2O-N2-CO2-rich fluid inclusions observed in the lower-temperature (600–800 °C) eclogites in the Śnieżnik and Międzygorze unit (Klemd et al. 1995), and with the fact that all observed fluid inclusions in the granulites and eclogites of the Złote unit are aqueous. Furthermore, some H2O-rich fluid must have been present locally in the recrystallised eclogite domains where volume diffusion and thus grain boundary diffusion were more efficient. It is however highly unlikely that those fluid inclusions, if formed at ultrahigh-pressure conditions, would have survived the isothermal decompression during exhumation, and the subsequent static recrystallisation of the quartz grains which occurred during amphibolite-facies conditions at temperatures as low as 550 to 660 °C. Densities of fluid inclusions which were trapped or retrapped during recrystallisation will probably reflect those P-T conditions (e.g. Johnson and Hollister 1995).
of retrograde biotite must have had an external source of derivation. However, fluid inclusions in microfractures are very rare, which indicates that only a few microcracks acted as conduits for fluid transport. This is distinctively different to the behaviour of the country-rock gneisses which responded to fluid infiltration by partial melting (Fischer 1935; Smulikowski 1967). Although it is highly unlikely that a considerable amount of fluid was able to infiltrate the ultrahigh-pressure rocks at post-amphibolite-facies conditions (option 3), there is no unambiguous evidence that all infiltration occurred during amphibolite-facies conditions. However, if some of the fluid inclusions were trapped or retrapped at amphibolite-facies conditions ($P_{o_b} = 9.5–12.5$ kbar, $T = 550–750$ °C) their preservation would require an exhumation of the host rocks almost parallel to the highest density inclusions (Fig. 12), otherwise those inclusions would stretch or re-equilibrate as a result of internal over- or underpressure (e.g. Bodnar et al. 1989; Touret 1992; Giaranamita and Sorensen 1994).

In summary, the presence of a discrete and free fluid phase during eclogite-facies metamorphism cannot unequivocally be substantiated by our fluid inclusion study. The absence of any CO$_2$ and/or N$_2$, which would lower the H$_2$O-activity, suggests that only limited amounts of H$_2$O were present during high- to ultrahigh-pressure metamorphism. We assume that the preservation of local pre-eclogitic reaction textures in the mafic granulites as well as prograde growth zonations of garnets in the felsic granulites is largely controlled by limited volume diffusion. The preservation of pre-eclogitic reaction textures and prograde-zonation patterns in garnet (sample 1107), instead of a homogenisation by intragranular diffusion, indicates kinetic barriers to polymorphic inversion in response to pressure variations (Rubie and Thompson 1984). The anhydrous ultrahigh-pressure rocks escaped deformation and subsequent pervasive fluid infiltration. Interface kinetics and diffusion were too sluggish for net-transfer reactions to proceed, thereby promoting the preservation of the pre-eclogite and eclogite-facies mineral assemblages. However, some aqueous fluid must have been present in the more recrystallised domains, since volume diffusion was much more efficient here than in the associated domains of disequilibrium. This is also supported by the presence of exclusively H$_2$O rich inclusions in the ultrahigh-pressure rocks of the Złote unit. Textural observations indicate that large fluid activity gradients persisted during high-pressure metamorphism which suggests a local fluid buffering as well as short-distance transport during eclogite-facies metamorphism. Furthermore, our results suggest limited fluid infiltration during amphibolite-facies conditions.

Geodynamic consequences

The ultrahigh-pressure rocks from the Złote unit apparently have experienced a different metamorphic history than the eclogites from the Międzygorze and Śnieżnik units (Bröcker and Klemd 1996). The lower peak metamorphic temperatures ($< 800$ °C) of the latter indicate that these eclogites are the product of subduction, since the oceanic and the continental thermal gradient away from subduction zones imply temperatures between 1000 and 1120 °C at about 100 km depth (Hacker and Peacock 1995). The U-Pb zircon dating of a mafic granulite indicates primary crystallisation around 365 to 375 Ma, shortly before or during early stages of metamorphism. The Sm-Nd mineral-whole rock data for an eclogite and granulites from Stary Gieraltów yielded dates between 340 to 350 Ma (Brueckner et al. 1991; this study) which are interpreted as a lower time limit for eclogite-facies metamorphism.

The exhumation of the ultrahigh-pressure rocks from the Złote unit (Fig. 12) which involved simultaneous cooling and decompression, may have occurred after a continent-continent collision i.e. after the Carboniferous collision of Gondwana and Baltica. The almost iso-thermal exhumation of the eclogites in the Śnieżnik and Międzygorze units must have occurred after exhumation of the ultrahigh-pressure rocks from the Złote unit, since the eclogites were subducted before collision. This conclusion is in accord with the assumption that the range of Sm-Nd ages from the Śnieżnik and Międzygorze units (341–329 Ma, Brueckner et al. 1991) is related to different episodes of exhumation of high-pressure tectonic slabs within a relatively short time span. This model of sequential uplift, as first suggested by Brueckner et al. (1991), best explains the differences in the $P$-$T$ paths (see Bröcker and Klemd 1996). We regard the Śnieżnik Mts. as a collage of several ultrahigh-pressure units which were tectonically exhumed and juxtaposed after collisional tectonics. The $P$-$T$ differences between eclogite-facies rocks from different locations of the study area indicate amalgamation of a composite framework of tectonic slabs, derived from various depths.

The importance of contrasting eclogite-facies fluid regimes

One of the problems when dealing with $P$ and $T$ estimates of metamorphic rocks is the presence or absence of a free fluid phase, i.e. the influence of the fluid on intergranular diffusion and reaction kinetics (see discussion and references in Rubie 1990; Rubie and Thompson 1984). The results of this study are in accord with findings of Selverstone et al. (1990), Philippot and Selverstone (1991) and Klemd et al. (1992) who presented evidence from fluid inclusion studies indicating fluid segregation and limited fluid transport rather than pervasive fluid flow during eclogite-facies metamorphism. The HT-granulites (750–900 °C, 4–12 kbar), which are dominated by CO$_2$-rich fluids and in places by high-salinity brines (e.g. Aranovich et al. 1987; Touret 1992; Andersen et al. 1993; Newton 1995; Fonarev et al. 1998), have a different fluid
regulated prior to eclogite-facies metamorphism. The rocks formed by subduction tectonics and eclogite-facies magmatic accretion. However this difference in fluid regimes in different ultrahigh-dehydration, while rocks from continent-continent collision tectonics is suggested to be a consequence of this concept (Andersen et al. 1993). The lower-temperature eclogite terrains of the Alps and the Caucasus (Philippot and Selverstone 1991; Philippot et al. 1995; Perchuk 1995). An important factor controlling the fluid composition is possibly the general tectonic setting. For example, many HT-granulites are the result of magmatic accretion (supplying a CO₂-rich fluid phase) beneath already existing crust and thus may display counterclockwise P-T-t paths (see Bohlen 1987; Touret 1993; cf. Santosh and Yoshida 1992). However, the geometry of prograde P-T-t paths of HP-granulites are commonly clockwise, as is shown in the present case, thus favouring different tectonic mechanisms instead of magmatic accretion (e.g. Harley 1989). This consequently explains the lack of CO₂-rich inclusions and associated side effects such as incipient charnockitisation and K-feldspar metasomatism (see Newton 1995, for an extensive discussion and references).

The difference in fluid regimes between eclogite-facies rocks formed by subduction tectonics and eclogite-facies rocks formed by continent-continent collision tectonics is suggested to be a consequence of this concept (Andersen et al. 1993). In subduction zone environments, for example, besides hydrated oceanic crust large volumes of clastic sediments are subducted, thus, representing a major source of nitrogen and water during dehydration, while rocks from continent-continent collision environments may have been thoroughly dehydrated prior to eclogite-facies metamorphism. The differences in the fluid regimes in different ultrahigh-pressure rocks of the Śnieżnik Mts. (Śnieżnik and Międzygórze units versus Złote unit) support the model of Andersen et al. (1993). The lower-temperature eclogites from the Międzygórze and Śnieżnik units, which formed as a result of subduction are dominated by a H₂O-salt-N₂- CO₂ fluid regime. In contrast, the ultrahigh-pressure rocks from the Złote unit, which are related to continent-continent collision tectonics, are dominated by a H₂O-salt fluid system. This clearly distinguishes high-pressure (HP)-granulites from HT-granulites which are believed to have been generated by magmatic accretion. However this difference in fluid regimes between HP- and HT-granulites should be verified by further fluid inclusion studies on other ultrahigh-pressure and HP-granulites.

The results of our studies in the Śnieżnik Mts. (Klemd et al. 1995; this study) demonstrate the general petrogenetic potential of fluid inclusion studies in metamorphic rocks. Besides constraints for fluid activities and P-T conditions, an increasing body of evidence suggests striking differences in fluid compositions related to different tectonic settings and resulting P-T-t paths. As a consequence, fluid inclusion studies in high-pressure rocks may assist in identification of geodynamic processes and thus help to unravel the metamorphic evolution in complex tectonometamorphic settings.

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