Sm–Nd and U–Pb dating of high-pressure granulites from the Złote and Rychleby Mts (Bohemian Massif, Poland and Czech Republic)

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ABSTRACT In the Orlica-Śnieżnik complex at the NE margin of the Bohemian Massif, high-pressure granulites occur as isolated lenses within partially migmatized orthogneisses. Sm–Nd (different grain-size fractions of garnet, clinopyroxene and/or whole rock) and U–Pb [isotope dilution-thermal ionization mass spectrometry (ID-TIMS) single grain and sensitive high-resolution ion microprobe (SHRIMP)] ages for granulites, collected in the surroundings of Červený Důl (Czech Republic) and at Stary Gierałtów (Poland), constrain the temporal evolution of these rocks during the Variscan orogeny. Most of the new ages cluster at *c.* 350–340 Ma and are consistent with results previously reported for similar occurrences throughout the Bohemian Massif. This interval is generally interpreted to constrain the time of high-pressure metamorphism. A more complex evolution is recorded for a mafic granulite from Stary Gierałtów and concerns the unknown duration of metamorphism (single, short-lived metamorphic cycle or different episodes that are significantly separated in time?). The central grain parts of zircon from this sample yielded a large spread in apparent $^{206}Pb/^{238}U$ SHRIMP ages (*c.* 462–322 Ma) with a distinct cluster at *c.* 365 Ma. This spread is interpreted to be indicative for variable Pb-loss that affected magmatic protolith zircon during high-grade metamorphism. The initiating mechanism and the time of Pb-loss has yet to be resolved. A connection to high-pressure metamorphism at *c.* 350–340 Ma is a reasonable explanation, but this relationship is far from straightforward. An alternative interpretation suggests that resetting is related to a high-temperature event (not necessarily in the granulite facies and/or at high pressures) around 370–360 Ma, that has previously gone unnoticed. This study indicates that caution is warranted in interpreting U–Pb zircon data of HT rocks, because isotopic rejuvenation may lead to erroneous conclusions.

Key words: granulite; Orlica-Śnieżnik complex; Sm–Nd dating; Sudetes; U–Pb dating.

INTRODUCTION

A characteristic feature of many collisional orogens are high-pressure (HP) granulites that formed at the base of overthickened crust, or as a result of subduction to mantle depths (e.g. O'Brien & Rötzler, 2003). The *P–T* field of HP granulites overlaps with parts of the eclogite facies and the kyanite stability field of the granulite facies (e.g. O'Brien & Rötzler, 2003). Mineral assemblages of such rocks are orthopyroxene-free and comprise garnet–clinopyroxene–plagioclase–quartz in metabasic rocks and are kyanite–K-feldspar-bearing in metapelitic and felsic bulk rock compositions (e.g. O'Brien & Rötzler, 2003; Pattison, 2003). HP granulites are widespread in the Variscan orogenic belt (e.g. O'Brien & Rötzler, 2003, and references therein) and a notable example occurs in the Orlica-Śnieżnik complex (OSC) in the West Sudetes (Fig. 1).

The OSC represents a crustal segment with a complex geological history within the Bohemian Massif and structurally occupies a position in the hangingwall of the Moldanubian Thrust Zone (Żelaźniewicz *et al.*, 2002). The OSC was strongly influenced by Cambro-Ordovician granitic magmatism and medium-pressure/high-temperature metamorphism related to the Variscan orogeny (e.g. Don *et al.*, 1990; Żelaźniewicz *et al.*, 2002). The significance of pre-Variscan metamorphism is actively debated (e.g. Kröner *et al.*, 2001; Żelaźniewicz *et al.*, 2002; Lange, 2004). A prominent feature in the OSC are remnants of eclogites and HP granulites, preserved as isolated occurrences (several metres- to tens of metres in size) within orthogneisses (e.g. Smulikowski, 1967; Bakun-Czubarow, 1991a,b, 1992; Bröcker & Klemd, 1996; Kryza *et al.*, 1996). Some authors interpreted these rocks as exotic tectonic inclusions (e.g. Don *et al.*, 1990), whereas others argued for an *in-situ* origin (Brueckner *et al.*, 1991; Bröcker & Klemd, 1996).

This study focuses on the geochronology of the granulites, which only are known from the Złote and Rychleby Mts in the eastern part of the OSC (Smulikowski, 1967; Pouba *et al.*, 1985; Smulikowski & Smulikowski, 1985; Bakun-Czubarow, 1991a,b, 1992; Bröcker & Klemd, 1996; Kryza *et al.*, 1996; Klemd &

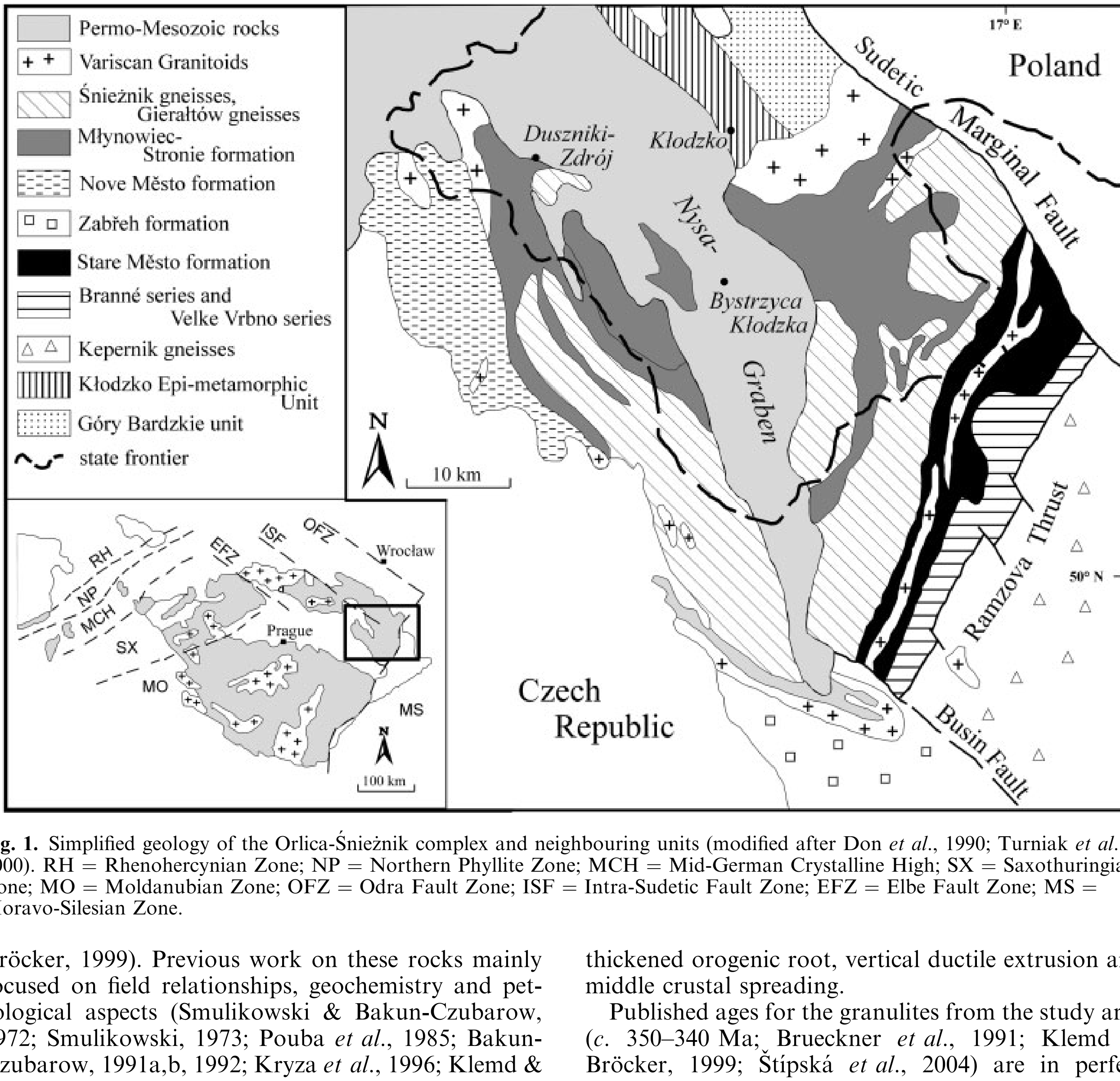

Fig. 1. Simplified geology of the Orlica-Śnieżnik complex and neighbouring units (modified after Don *et al.*, 1990; Turniak *et al.*, 2000). RH = Rhenohercynian Zone; NP = Northern Phyllite Zone; MCH = Mid-German Crystalline High; SX = Saxothuringian Zone; MO = Moldanubian Zone; OFZ = Odra Fault Zone; ISF = Intra-Sudetic Fault Zone; EFZ = Elbe Fault Zone; MS = Moravo-Silesian Zone.

Bröcker, 1999). Previous work on these rocks mainly focused on field relationships, geochemistry and petrological aspects (Smulikowski & Bakun-Czubarow, 1972; Smulikowski, 1973; Pouba *et al.*, 1985; Bakun-Czubarow, 1991a,b, 1992; Kryza *et al.*, 1996; Klemd & Bröcker, 1999) and established a general geological framework. The granulites are considered to be derived from bimodal volcanic rocks of unknown protolith age, which were slightly contaminated by sedimentary material (Pouba *et al.*, 1985; Bakun-Czubarow, 1991b). The peak metamorphic *P–T* conditions were estimated at 800–900 °C and at 16–22 kbar by Bakun-Czubarow (1991b), while Pouba *et al.* (1985) reported temperatures of up to 1000 °C. Bakun-Czubarow (1991a) described the presence of quartz pseudomorphs possibly after coesite and peak metamorphic temperatures between 860 and 960 °C. Other studies suggested pressures of at least 21–28 kbar (up to 35 kbar) at temperatures between 800 and 1000 °C (Kryza *et al.*, 1996; Klemd & Bröcker, 1999). The exhumation of these rocks was placed by Štípská *et al.* (2004) in the context of syn-convergent processes, which include initial buckling of crustal layers in a thickened orogenic root, vertical ductile extrusion and middle crustal spreading.

Published ages for the granulites from the study area (*c.* 350–340 Ma; Brueckner *et al.*, 1991; Klemd & Bröcker, 1999; Štípská *et al.*, 2004) are in perfect agreement with the ages reported for similar occurrences throughout the Bohemian Massif (e.g. O'Brien & Rötzler, 2003, and references therein). However, a problematic issue still is the difficulty to date precisely distinct stages of the *P–T–t*-deformation path. Unresolved issues concern the duration of high-pressure metamorphism (single, short-lived metamorphic cycle or different episodes that are significantly separated in time?) and the temporal and structural relationships between orthogneisses and granulites.

A previous study of the Złote granulites has indicated that geochronological information related to earlier events is not completely erased by the last thermal overprint (Klemd & Bröcker, 1999). U–Pb zircon dating of multigrain separates yielded $^{207}Pb/^{206}Pb$ ages between 369 and 360 Ma, which were considered to approximate the time of crystallization from a melt, either before or during early stages of high-temperature metamorphism (Klemd & Bröcker, 1999). Interpretation of these zircon ages is hampered by general limitations of multigrain dating, which may obscure correct identification of complex age patterns. Despite this restriction, the available data set has indicated the interesting prospect that the geochronological record stored in the granulites may document a more detailed picture of the magmatic and metamorphic history. If this hypotheses can be substantiated, the geochronology of the Złote and Rychleby granulites may contribute significantly to a better understanding of the geodynamic evolution of the OSC and similar occurrences of HP granulites elsewhere in the Variscides. In order to shed more light on this issue, we have studied felsic and mafic granulites from the OSC by means of Sm–Nd (garnet, clinopyroxene, whole rock) and U–Pb zircon geochronology (ID-TIMS single grain and SHRIMP dating).

GEOLOGICAL SETTING AND PREVIOUS GEOCHRONOLOGY

The OSC in the West Sudetes (Figs 1 & 2) is bordered by two NW-trending fault zones, the Sudetes Marginal Fault in the north-east and the Busin Fault in the south-west. Both fault zones extend parallel to the Odra and Elbe Fracture Zones, the major lineaments dividing the West Sudetes from other crustal segments within the Variscides (Fig. 1). The OSC can be described as a dome structure with a meta-granitic to migmatitic core, including tectonic bodies of mafic and acidic HP to UHP rocks, mantled by metasediments and metabasic rocks metamorphosed under middle to lower amphibolite facies *P–T* conditions, and an outer envelope of similar lithologies with mineral assemblages indicating greenschist facies metamorphism (e.g. Don *et al.*, 1990; Żelaźniewicz *et al.*, 2002).

Orthogneisses represent the dominant rock type of the study area and traditionally are subdivided into two groups: the variably deformed, porphyritic Śnieżnik gneisses and the migmatitic Gierałtów gneisses (e.g. Don *et al.*, 1990; Żelaźniewicz *et al.*, 2002). Geochemical studies (including REE and Sr–Nd isotope characteristics) indicated no systematic differences between both rock types, which lead to the conclusion that Śnieżnik and Gierałtów gneisses represent different textural variants of the same protolith, variably modified by deformation and migmatization (Turniak *et al.*, 2000; Lange *et al.*, 2002; Lange, 2004).

Additional major constituents of the OSC are metasediments and metavolcanic rocks (Młynowiec–Stronie group; Fig. 1), which comprise plagioclase gneisses and mica schists (Młynowiec formation, >2000 m thick), and a variegated succession of biotite- to staurolite-grade mica schists, paragneisses, quartzites, marbles and amphibolites (Stronie formation, *c*. 3000 m thick; Don *et al.*, 1990; Żelaźniewicz *et al.*, 2002).

Variable degrees of migmatization in the orthogneisses and staurolite-bearing gneisses indicate at least upper amphibolite facies *P–T* conditions (e.g. Lange *et al.*, 2002; Żelaźniewicz *et al.*, 2002). The peak-pressure conditions for the orthogneisses may have been within the eclogite or HP granulite facies, as the HP metamorphism recorded in eclogite and granulite lenses is interpreted by some authors as an *in situ* process (e.g. Brueckner *et al.*, 1991; Bröcker & Klemd, 1996; for a contrasting view see Don *et al.*, 1990). This would imply that the currently preserved mineral assemblages in the orthogneisses are due to a later overprint, that did not completely re-equilibrate the eclogite and granulite parageneses.

The *P–T* conditions are only well constrained for eclogites and granulites, which occur as isolated blocks and lenses within Gierałtów gneisses. For these rocks, estimated peak-pressures lie in the coesite stability field with pressures in excess of 27 kbar at temperatures between 700–800 and 800–1000 °C, respectively (Bakun-Czubarow, 1991a,b, 1992; Bröcker & Klemd, 1996; Kryza *et al.*, 1996; Klemd & Bröcker, 1999).

The deformation history of the OSC is complex and comprises at least three stages. Some workers interpret the main deformation (D2 & D3) as a Late Caledonian process, whereas others emphasize the importance of Early to Late Carboniferous events (for overviews see Don *et al.*, 1990 and Żelaźniewicz *et al.*, 2002).

There seems to be a general consensus that most orthogneisses have Cambrian–Ordovician protolith ages (*c*. 520–490 Ma; e.g. van Breemen *et al.*, 1982; Oliver *et al.*, 1993; Kröner *et al.*, 1997, 2001; Turniak *et al.*, 2000; Štípská *et al.*, 2004) and that all rock types underwent at least one episode of high-temperature (HT) metamorphism during Variscan times at *c*. 340 Ma (e.g. Borkowska *et al.*, 1990; Turniak *et al.*, 2000; Lange *et al.*, 2002). Younger apparent protolith ages indicated by Rb–Sr whole-rock dating (464 $\pm$ 18 and 395 $\pm$ 35 Ma, Borkowska *et al.*, 1990) are most likely due to disturbance of the Rb–Sr system during metamorphic overprinting at *c*. 340 Ma (Lange, 2004). Rb–Sr and ^{40}Ar–^{39}Ar dating of phengite, biotite and hornblende mostly provided Carboniferous cooling ages for orthogneisses (*c*. 340–330 Ma; Borkowska *et al.*, 1990; Steltenpohl *et al.*, 1993; Lange *et al.*, 2002; Marheine *et al.*, 2002; Lange, 2004), but also indicated younger age groups at *c*. 320 Ma (Marheine *et al.*, 2002) and *c*. 294 Ma (Lange *et al.*, 2002). A SHRIMP U–Pb age of 342 $\pm$ 6 Ma for zircon overgrowths was interpreted to closely approximate the time of HT metamorphism (Turniak *et al.*, 2000).

The protolith ages of eclogites and HP granulites are not known. Štípská *et al.* (2004) reported a ^{206}Pb/^{238}U SHRIMP Zircon age of 473 $\pm$ 8 Ma for a felsic granulite from the Rychleby Mts. This age was only recognized in the core of one zircon grain and is interpreted to date an event in the history of the protolith. These authors provided no further explanation for their conclusion (xenocryst? inheritance?

protolith age?) and at present the database is too small for a reliable interpretation. Note also that a SHRIMP U–Pb zircon age of *c.* 525 Ma for an eclogite (D. Gebauer, unpublished data) was repeatedly quoted in the regional literature. Metamorphic ages of the HP rocks are well-constrained. Eclogites within orthogneisses yielded Sm–Nd ages (garnet, omphacite, whole rock) of 341 ± 7, 337 ± 4 and 329 ± 6 Ma (Brueckner *et al.*, 1991). For an eclogite intercalated with HP granulite in the Złote Mts, Brueckner *et al.* (1991) reported an age of 352 ± 4 Ma and mafic granulites from the same occurrence yielded Sm–Nd ages (garnet, whole rock) of 341 ± 10 and 343 ± 11 Ma (Klemd & Bröcker, 1999). Štípská *et al.* (2004) reported a SHRIMP $^{206}Pb/^{238}U$ age of 341.6 ± 4.7 Ma and a $^{207}Pb/^{206}Pb$ evaporation age of 341.4 ± 0.7 Ma for multifaceted, near-spherical zircon from two felsic granulites collected in the Rychleby Mts.

It is important to bear in mind that the studied granulites are eclogite facies rocks and that currently available geochronological data do not distinguish between HP and HT metamorphism recorded in the various rock types, as all ages cluster around 340 Ma (e.g. Brueckner *et al.*, 1991; Steltenpohl *et al.*, 1993; Klemd & Bröcker, 1999; Turniak *et al.*, 2000; Lange *et al.*, 2002; Lange, 2004; Štípská *et al.*, 2004).

FIELD RELATIONS AND SAMPLE DESCRIPTION

The granulites of the Złote and Rychleby Mts occur in a NE–SW trending zone (*c.* 0.5–2 km wide and *c.* 12 km long) within orthogneisses of the Gierałtów type in the border region between Poland and the Czech Republic (Fig. 2; Pouba *et al.*, 1985; Bakun-Czubarow, 1991b). Near Stary Gierałtów, the granulite belt is bounded to the SW by a cross-cutting fault. Along-strike, the granulites show continuous transitions towards the adjacent orthogneisses (Pouba *et al.*, 1985), with gradual contacts most likely resulting from severe overprinting by migmatization and amphibolite facies metamorphism. Due to bad outcrop conditions, field relationships are difficult to interpret. Some authors suggested that the granulites occur in the core of an antiform (e.g. Kryza *et al.*, 1996; Don, 2001). Other workers described the granulites as relics of a syn-form (e.g. Pouba *et al.*, 1985; Dumicz, 1993; Szczepanski & Anczkiewicz, 2000). On the Polish side of the border, only one very small outcrop of well-preserved granulites is known. This occurrence at Stary Gierałtów (*c.* 3–4 m long and *c.* 1 m high) has attracted much attention, due to findings of presumed pseudomorphs after coesite, suggesting ultrahigh-pressure metamorphism (Bakun-Czubarow, 1991a,b, 1992; Kryza *et al.*, 1996; Klemd & Bröcker, 1999). This outcrop is located close to the Biała Ladecka river and was damaged in the last years, during flooding periods. The intimate interlayering of various granulites and eclogites described by Smulikowski (1967), Smulikowski & Bakun-Czubarow (1972) and Bakun-Czubarow (1991b) can no longer be recognized in detail, but with the exception of eclogites, good samples of the main rock types (felsic and mafic granulites) still can be found. On the Czech side of the border (Rychleby Mts), small occurrences and isolated blocks of granulites are more common, but also here outcrops

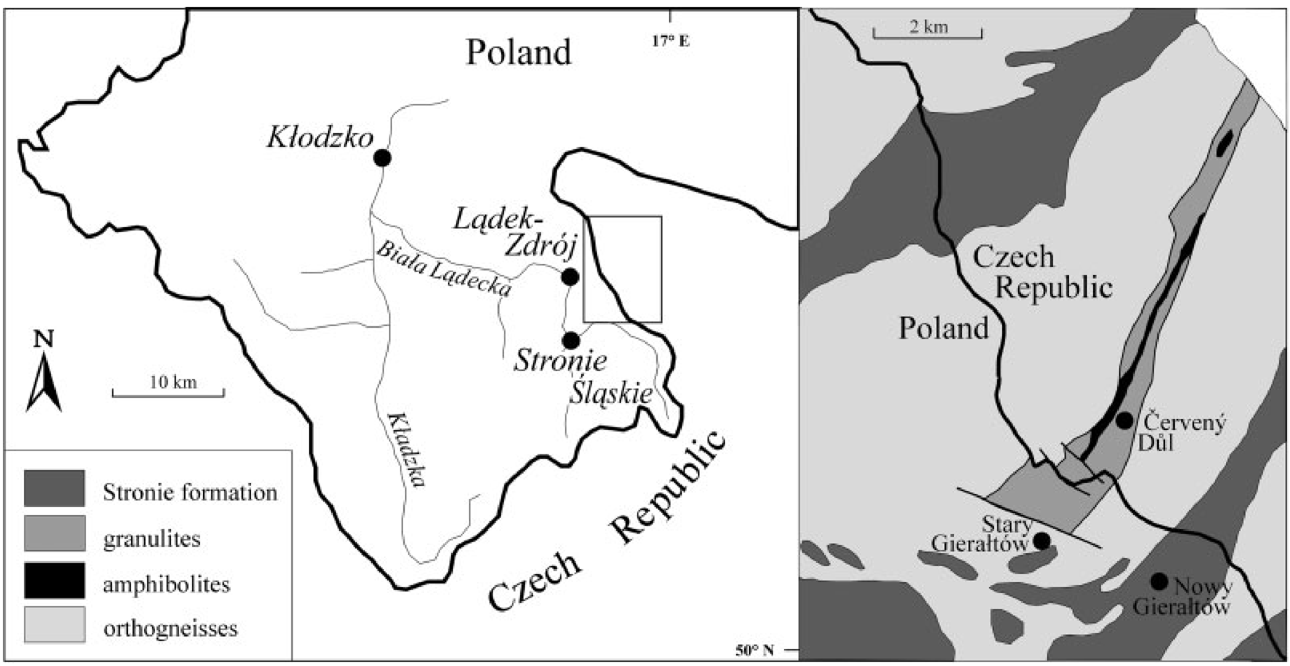

Fig. 2. Location of the study area and simplified sketch map of the granulite belt (modified after Bakun-Czubarow, 1991a) with sample occurrences discussed in the text.

mostly expose the country rock orthogneisses (Pouba *et al.*, 1985). For details of the petrography and mineral chemistry of the OSC granulites see previous work by Smulikowski (1967), Smulikowski & Bakun-Czubarow (1972), Bakun-Czubarow (1991a,b, 1992), Kryza *et al.* (1996) and Klemd & Bröcker (1999).

Sample 1106 is a mafic granulite from Stary Gierałtów (Fig. 2) with the mineral assemblage garnet–omphacite–quartz–plagioclase–biotite–kyanite. Rare white mica occurs as inclusions in omphacite and kyanite. Typical accessory minerals are apatite, rutile, zircon and Fe-oxides. The presence of pseudomorphs after coesite is inferred from radial fractures around polycrystalline quartz inclusions in garnet (see fig. 4, p. 360 of Klemd & Bröcker, 1999). Zircon multigrain separates of this sample were previously dated with the ID-TIMS U–Pb method (Klemd & Bröcker, 1999). Six unabraded size fractions yielded discordant results and plot very close to the Concordia near to a poorly defined upper intercept around 370 Ma (see fig. 9, p. 367 of Klemd & Bröcker, 1999). The small spread in $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ did not allow the calculation of a geological meaningful discordia line, but the $^{207}Pb/^{206}Pb$ ratios range from 369 to 360 Ma. Zircon from this sample was studied in greater detail using the U–Pb SHRIMP method.

From the Czech side of the border, three samples were collected NE of the location Červený Důl (Fig. 2; for a more detailed geological map see Pouba *et al.*, 1985). Two samples (204, 209) represent mafic granulites with a mineral assemblage similar to sample 1106. Sample 205 is an omphacite-free felsic granulite mainly consisting of the assemblage K-feldspar–plagioclase–quartz–garnet–biotite. Minor constituents comprise titanite, zircon, apatite and rutile.

RESULTS

Sm–Nd dating

Samples 204, 205 and 209 from the surroundings of Červený Důl were selected for Sm–Nd dating. Analytical results are reported in Table 1. Age calculations are based on different grain-size fractions of garnet, omphacite and/or whole rock. Isochron diagrams are displayed in Fig. 3. Garnet from the mafic granulites has lower Sm (5.1–7.3 ppm) and Nd (6.4–9.9 ppm) concentrations than garnet from the felsic sample (Sm: *c*. 32 ppm; Nd: 28.9–30.3 ppm). Garnet of both rock types also differ in their $^{147}Sm/^{144}Nd$ ratios (mafic samples: 0.44 and 0.49; felsic sample: 0.64–0.67). $^{147}Sm/^{144}Nd$ ratios of whole rocks and omphacite are between 0.14 and 0.15. Samples 204 and 209 (mafic granulites) yielded Sm–Nd ages of 357 ± 10 and 351 ± 10 Ma, respectively (Fig. 3a,b). Sample 205 (felsic granulite) gave an Sm–Nd age of 337 ± 4 Ma (Fig. 3c).

U–Pb dating (ID-TIMS)

Samples 204 and 205 were selected for ID-TIMS U–Pb dating of single zircon grains. Zircon populations of these samples consist of grains with different morphologies, showing a complete range from prismatic to spherical grains. Prismatic grains generally have rounded terminations, suggesting zircon dissolution during high-grade metamorphism. Variable degrees of rounding have produced spherical grains, which occur in addition to multifaceted zircons, commonly interpreted to have crystallized during granulite facies metamorphism.

Cathodoluminescence (CL) studies indicated three types of internal patterns. Type 1 was recognized in strongly rounded to short prismatic and in multifaceted grains and is characterized by the absence of magmatic zonation or inherited cores; however, the grains are not homogeneous but exhibit irregular patches with different CL intensity. Type 2 (prismatic zircon) is characterized by variable degrees of oscillatory zoning. Type 3 (only recognized in the felsic sample 205) consists of inherited cores surrounded by homogenous to oscillatory zoned zircon. Small and discontinuous overgrowths are common for all types.

It is important to note that multifaceted zircon and spherical grains resulting from partial dissolution during metamorphism could not be distinguished

Table 1. Sm–Nd isotopic results of mafic and felsic granulites from Červený Důl.

Sample no.	Rock type	Sample	Grain size (μm)	Sm (ppm)	Nd (ppm)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	Age in Ma ± 2σ
204	Granulite (mafic)	Whole rock[a]		18.85	4.794	0.1538	0.512506 (16)	351 ± 10
		Garnet[b]	355–250	7.020	9.879	0.4409	0.513176 (15)	356 ± 11
		Garnet[b]	250–180	7.331	9.644	0.4596	0.513201 (13)	347 ± 10
205	Granulite (felsic)	Whole rock[a]		18.33	4.242	0.1399	0.512224 (10)	337 ± 4
		Garnet[b]	355–250	32.23	30.30	0.6432	0.513335 (17)	337 ± 6
		Garnet[b]	250–180	32.07	28.91	0.6708	0.513397 (14)	338 ± 5
209	Granulite (mafic)	Whole rock[c]		13.63	3.378	0.1498	0.512461 (12)	357 ± 10
		Garnet[b]	355–250	5.077	6.442	0.4765	0.513222 (9)	356 ± 7
		Garnet[b]	250–180	5.278	6.529	0.4887	0.513273 (17)	366 ± 9
		Clinopyroxene[d]	355–250	3.632	14.71	0.1492	0.512462 (16)	355 ± 8

Age calculations based on: [a]whole rock–garnet–garnet, [b]whole rock–garnet, [c]whole rock–garnet–garnet–clinopyroxene, [d]whole rock–clinopyroxene.
See Appendix for analytical details.

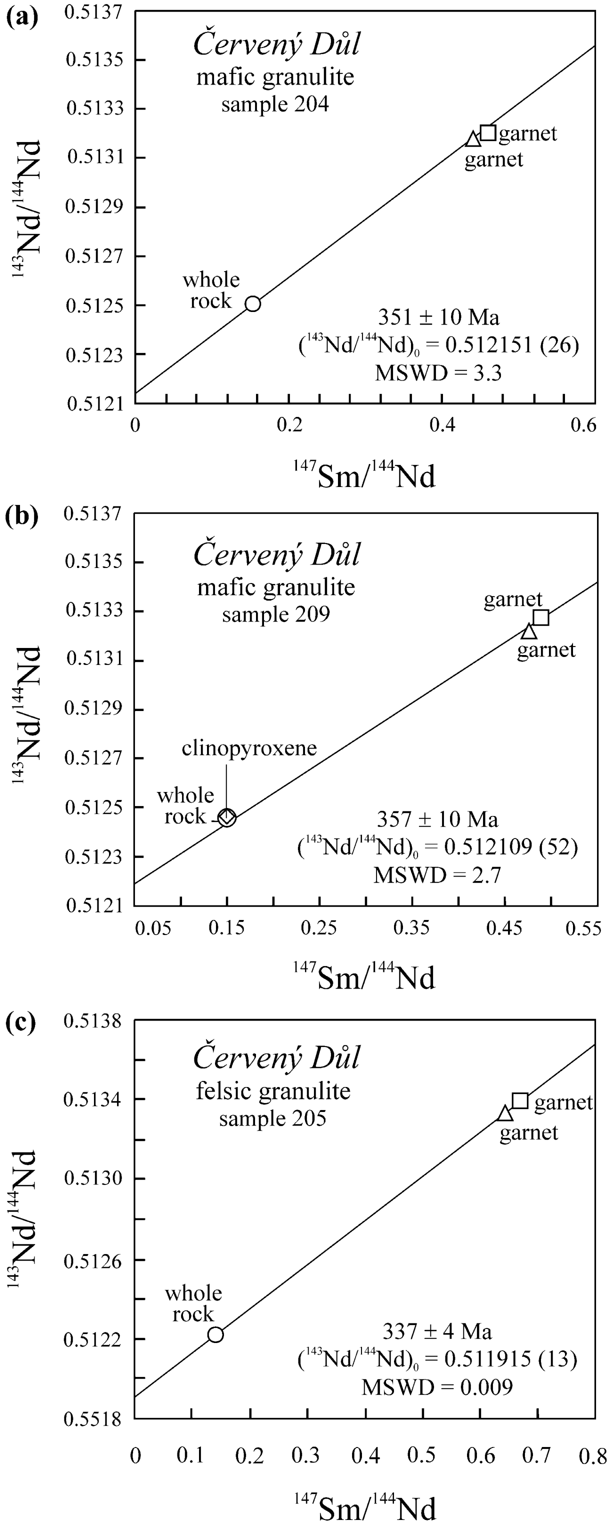

Fig. 3. Sm–Nd isotope diagrams for granulites from Červený Důl: (a) sample 204, (b) sample 209, (c) sample 205.

under the stereomicroscope. The same holds true for prismatic zircon with different internal structures. Zircon selected for U–Pb analyses was clear and colourless, without visible inclusions or microfractures. All grains were air-abraded before dissolution in order to remove the outermost overgrowths. Analytical results are reported in Table 2. Concordia diagrams are displayed in Fig. 4.

Zircon of the mafic granulite provided three groups of concordant ages (Fig. 4a): (a) Two prismatic zircons yielded $^{206}Pb/^{238}U$ ages of 411.4 ± 1.5 and 413.5 ± 1.6 Ma. (b) Four spherical grains gave a weighted mean average of 359.3 ± 0.8 Ma. (c) Two grains (spherical and strongly rounded short-prismatic) yielded $^{206}Pb/^{238}U$ ages of 340.2 ± 1.2 and 341.1 ± 1.3 Ma. Zircon of the felsic granulite provided $^{206}Pb/^{238}U$ ages between *c.* 393 and 323 Ma (Fig. 4b). Prismatic grains yielded 393.3 ± 2.8 and 347.2 ± 1.4 Ma. One spherical grain has a $^{206}Pb/^{238}U$ age of 338.1 ± 1.3 Ma; three other spherical grains provided a pooled $^{206}Pb/^{238}U$ age of 326.1 ± 1.5 Ma.

U–Pb SHRIMP dating

Two size fractions (200–160 & 160–125 μm) of zircon from sample 1106 were analysed. Almost all zircon is composed of a clear central area, but occasionally fractured and surrounded by a thin rim. CL imaging shows the central areas to be composed of rather homogeneous zircon with faint remnants of sector zoning. The zircon is anhedral with variable shapes from rounded types with broad sector zoning, to more elongate grains with only very faint oscillatory or complex zoning preserved. The rims show up as dark (in CL) overgrowths. Embayments and internal patches of this dark CL zircon are also observed. No irrefutable cores were identified, although a few possible inherited cores, as indicated by different zoning patterns, were analysed. Figure 5 shows CL images of a selection of zircon from this sample.

Thirty-three spots were analysed on 28 different grains. The data are reported in Table 3 and shown on a conventional Wetherill Concordia diagram (Fig. 6) as well as on combined cumulative probability and histogram plots (Fig. 7a,b). The data spread along concordia, with the majority tending to cluster at the younger end of the spectrum. A number of relatively dark CL 'cores' with pronounced zoning were also analysed, but did not yield older ages than observed for the sector-zoned parts. Analyses of the dark CL rims plot as a group, albeit with some scatter, with $^{206}Pb/^{238}U$ ages of *c.* 350–361 Ma. Four of six spots define a cluster at *c.* 350 Ma.

There is no obvious correlation between U concentration and $^{206}Pb/^{238}U$ age (Table 3). For example, the analysed spots that yielded the highest $^{206}Pb/^{238}U$ ages (462–458 Ma) are characterized by low U concentrations (33–61 ppm). Low U concentrations (46–80 ppm) were also determined for spots with $^{206}Pb/^{238}U$ ages between 364 and 428 Ma. However, there is also a group of similar ages (358–440 Ma) that

Table 2. U–Pb isotopic data of mafic and felsic granulites.

Sample	Grain	Isotopic ratios				Age in Ma ± 2σ		
		$^{206}Pb/^{204}Pb^{a}$	$^{207}Pb/^{206}Pb^{b}$	$^{207}Pb/^{235}U^{b}$	$^{206}Pb/^{238}U^{b}$	$^{206}Pb/^{238}U^{b}$	$^{207}Pb/^{235}U^{b}$	$^{207}Pb/^{206}Pb^{b}$
204	45-0	182.9	0.05381 (58)	0.4234 (53)	0.05707 (25)	357.8 ± 1.6	358.5 ± 4.5	363.1 ± 23.9
	45-1	207.9	0.05383 (43)	0.4266 (40)	0.05748 (22)	360.3 ± 1.4	360.8 ± 3.4	363.8 ± 17.8
	45-3	183.9	0.05381 (70)	0.4252 (62)	0.05731 (24)	359.2 ± 1.5	359.7 ± 5.2	363.2 ± 29.2
	45-4	161.4	0.05374 (67)	0.4249 (61)	0.05735 (26)	359.5 ± 1.6	359.6 ± 5.2	360.0 ± 28.3
	K17-2	1439	0.05513 (22)	0.5036 (28)	0.06625 (25)	413.5 ± 1.6	414.1 ± 2.3	417.3 ± 9.1
	K17-3	1096	0.05513 (16)	0.5010 (24)	0.06591 (23)	411.4 ± 1.5	412.4 ± 2.0	417.6 ± 6.6
	K17-4	1941	0.05335 (17)	0.3986 (20)	0.05419 (20)	340.2 ± 1.2	340.7 ± 1.7	344.1 ± 7.0
	K17-5	2409	0.05333 (15)	0.3995 (18)	0.05433 (20)	341.1 ± 1.3	341.3 ± 1.6	342.9 ± 6.2
205	B-0	712.0	0.05294 (39)	0.3795 (42)	0.05198 (41)	326.7 ± 2.6	326.6 ± 3.6	326.3 ± 16.1
	B-1	793.0	0.05305 (29)	0.3827 (41)	0.05232 (47)	328.7 ± 3.0	329.0 ± 3.5	330.9 ± 11.9
	B-2	897.5	0.05347 (20)	0.4080 (23)	0.05533 (22)	347.2 ± 1.4	347.4 ± 1.9	349.0 ± 8.3
	B-3	2099	0.05458 (17)	0.4734 (36)	0.06291 (44)	393.3 ± 2.8	393.6 ± 3.0	395.1 ± 7.0
	A21-0	336.5	0.05295 (55)	0.3754 (51)	0.05142 (41)	323.2 ± 2.6	323.6 ± 4.4	326.6 ± 22.7
	A21-1	556.6	0.05326 (36)	0.3955 (32)	0.05386 (21)	338.1 ± 1.3	338.4 ± 2.8	339.7 ± 14.9

[a]Measured value corrected for mass fractionation.
[b]Ratios and ages corrected for mass fractionation, spike, Pb blank and initial common Pb (at 360 Ma from Stacey & Kramers, 1975).
See Appendix for analytical details.

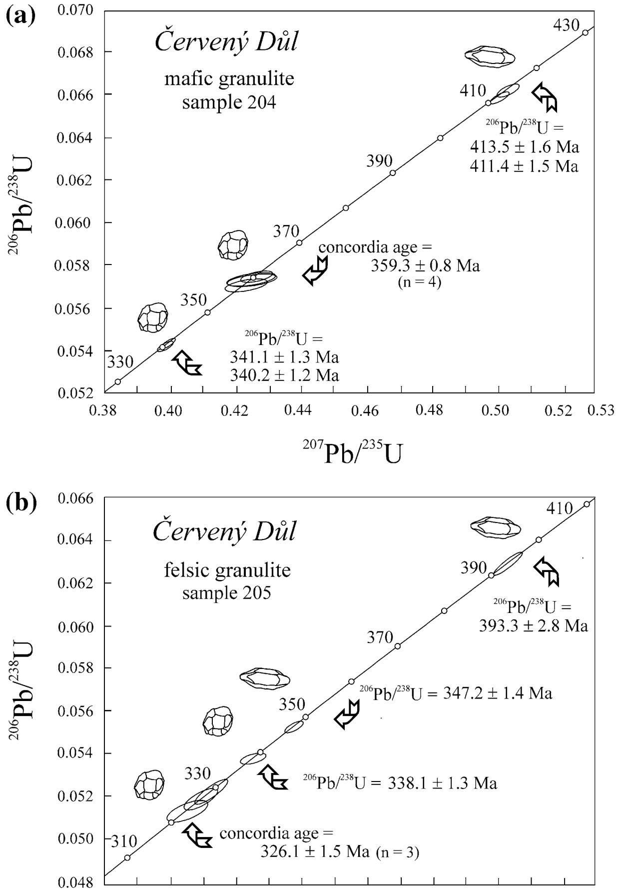

Fig. 4. U–Pb Concordia diagrams showing results of ID-TIMS dating of single zircon grains: (a) sample 204, (b) sample 205.

shows high U concentrations (340–776 ppm). All zircon rims (*c.* 350 Ma) show high U concentrations (440–830 ppm).

DISCUSSION

UHP to HP metamorphism at 350–340 Ma and subsequent processes

Until recently, age information related to the HP granulites from the OSC was based on only a few Sm–Nd results from a single outcrop at Stary Gierałtów (Brueckner *et al.*, 1991; Klemd & Bröcker, 1999), indicating a metamorphic age of *c.* 350–340 Ma. The geological significance of this age group is further substantiated by U–Pb zircon and Sm–Nd geochronology of samples from the Rychleby Mts (Štípská *et al.*, 2004; this study). Published and new Sm–Nd ages from the Złote and Rychleby Mts overlap within error, but the ages obtained for mafic granulites from Stary Gierałtów (341 ± 10, 343 ± 11 Ma; Klemd & Bröcker, 1999) and Červený Důl (357 ± 10, 351 ± 10 Ma; this study) have relatively high uncertainties, due to a small spread in $^{147}Sm/^{144}Nd$ ratios. The age of 352 ± 4 Ma (garnet, omphacite, whole rock) for an eclogite within granulites from Stary Gierałtów (Brueckner *et al.*, 1991) and the age of 337 ± 4 Ma (two-size fractions of garnet, whole rock) for a felsic granulite from Červený Důl (Fig. 3c) are considered here to approximate the upper and lower time limit of UHP to HP metamorphism in the study area. This time bracket is in perfect agreement with results reported for other granulite occurrences in the Bohemian Massif (for a compilation of the relevant references see Kröner *et al.*, 2000a and O'Brien & Rötzler, 2003). However, although the Sm–Nd ages are corroborated by U–Pb zircon dating, it is currently not completely clear whether or not the *c.* 340 Ma ages record the time of the metamorphic climax or a later HP stage.

ID-TIMS dating yielded a pooled $^{206}Pb/^{238}U$ age of 326 ± 2 Ma for three spherical zircon from the felsic granulite 205. This age might indicate anatectic processes post-dating the pressure peak, but no field or textural observations support this hypothesis. More

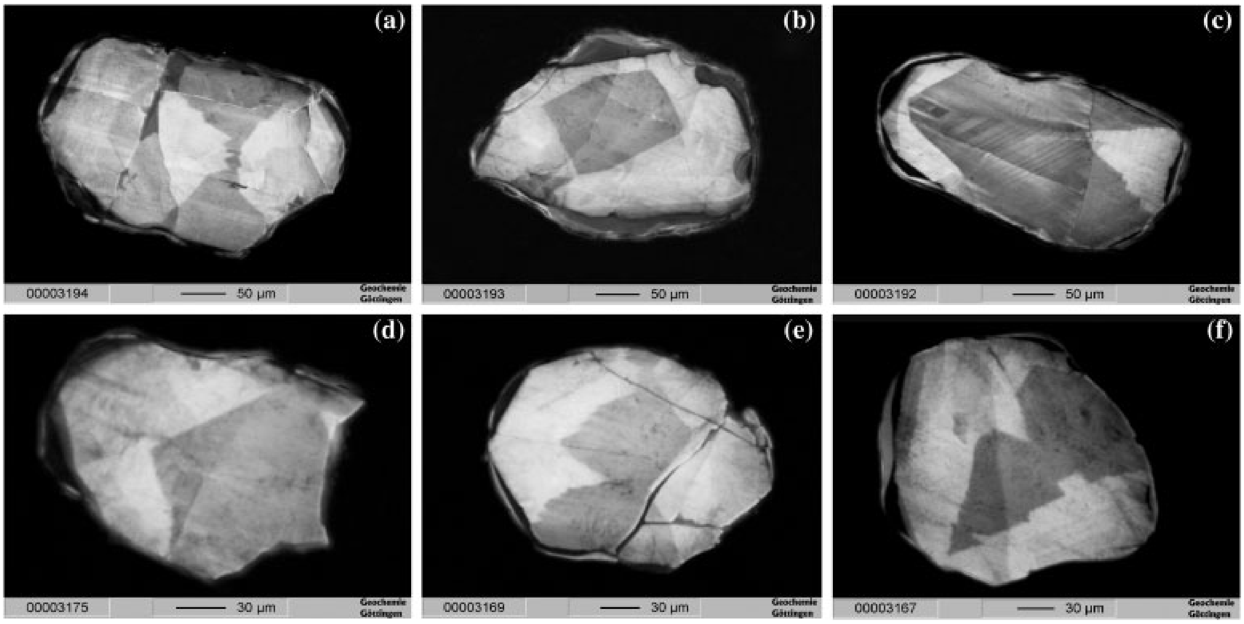

Fig. 5. CL images of representative zircon from sample 1106, Stary Gierałtów.

Table 3. Summary of SHRIMP U–Pb zircon data for sample 1106.

Grain Spot*	U (ppm)	Th (ppm)	^{232}Th/^{238}U	^{206}Pb$_c$ (%)	^{206}Pb* (ppm)	^{207}Pb*/^{206}Pb* ± %	^{207}Pb*/^{235}U ± %	^{206}Pb*/^{238}U ± %	^{206}Pb/^{238}U Age (Ma)
1.1a	86	42	0.51	2.23	4.33	0.0575 ± 10	0.455 ± 10	0.0574 ± 2.5	360 ± 9
2.1a	55	18	0.33	2.36	3.00	0.0558 ± 17	0.475 ± 17	0.0617 ± 2.7	386 ± 10
3.1a	56	21	0.39	2.52	2.91	0.0516 ± 17	0.419 ± 17	0.0588 ± 3.4	368 ± 12
4.1a	612	100	0.17	0.28	29.4	0.0537 ± 2.4	0.413 ± 2.6	0.05578 ± 0.97	350 ± 3
4.2a	544	88	0.17	0.18	26.2	0.0529 ± 1.8	0.408 ± 2.1	0.05591 ± 0.96	351 ± 3
4.3a	394	157	0.41	0.21	23.9	0.0573 ± 1.2	0.558 ± 1.5	0.07056 ± 1.0	440 ± 4
4.4a	33	10	0.33	2.17	2.12	0.0554 ± 23	0.562 ± 23	0.0736 ± 2.6	458 ± 11
5.1a	74	37	0.52	1.70	4.42	0.0560 ± 9.5	0.530 ± 9.6	0.0687 ± 1.6	428 ± 7
6.1a	125	59	0.49	0.97	6.18	0.0542 ± 6.7	0.425 ± 6.8	0.05688 ± 1.4	357 ± 5
7.1a	58	24	0.42	2.38	3.24	0.0566 ± 14	0.494 ± 14	0.0633 ± 2.0	396 ± 8
8.1m	349	62	0.18	0.26	19.2	0.0568 ± 1.3	0.503 ± 1.6	0.06419 ± 1.0	401 ± 4
9.1a	78	35	0.47	1.48	4.14	0.0538 ± 14	0.425 ± 14	0.0610 ± 1.7	382 ± 7
10.1b	830	49	0.06	0.08	39.9	0.0542 ± 1.6	0.418 ± 1.8	0.05595 ± 0.92	351 ± 3
11.1b	439	77	0.18	0.15	21.0	0.0547 ± 1.2	0.420 ± 1.6	0.05573 ± 0.99	350 ± 3
12.1m	776	55	0.07	0.12	39.9	0.0550 ± 1.0	0.454 ± 1.4	0.05981 ± 0.93	375 ± 3
13.1a	111	45	0.42	0.95	5.48	0.0540 ± 9.7	0.423 ± 9.8	0.0569 ± 1.5	357 ± 5
14.1a	79	27	0.36	1.68	4.16	0.0557 ± 16	0.466 ± 16	0.0607 ± 1.8	380 ± 7
15.1a	61	29	0.50	1.66	3.96	0.0568 ± 14	0.582 ± 14	0.0743 ± 1.8	462 ± 8
16.1a	69	29	0.44	1.43	3.93	0.0578 ± 10	0.522 ± 10	0.0655 ± 1.7	409 ± 7
17.1r	468	96	0.21	0.14	20.6	0.0532 ± 1.4	0.375 ± 1.7	0.05123 ± 0.92	322 ± 3
17.2	664	132	0.21	0.14	33.3	0.0536 ± 1.7	0.430 ± 1.9	0.05821 ± 0.93	365 ± 3
18.1a	73	26	0.37	0.74	3.80	0.0555 ± 3.7	0.461 ± 4.0	0.0602 ± 1.5	377 ± 5
19.1b	509	61	0.12	0.08	25.2	0.0536 ± 2.0	0.425 ± 2.2	0.05757 ± 1.1	361 ± 4
20.1a	80	33	0.43	0.75	4.01	0.0553 ± 5.1	0.443 ± 5.2	0.05805 ± 1.3	364 ± 5
21.1m	340	60	0.18	–	16.7	0.0524 ± 1.7	0.412 ± 2.0	0.05696 ± 1.1	357 ± 4
21.2a	46	20	0.44	0.62	2.41	0.0535 ± 5.7	0.449 ± 6.0	0.0608 ± 1.7	381 ± 6
22.1c	449	285	0.66	0.17	26.6	0.0568 ± 2.3	0.541 ± 2.7	0.06906 ± 1.3	431 ± 6
23.1a	63	29	0.48	0.99	3.26	0.0562 ± 4.5	0.460 ± 4.7	0.05941 ± 1.3	372 ± 5
24.1b	428	52	0.13	0.05	21.0	0.0541 ± 1.1	0.426 ± 1.4	0.05711 ± 0.91	358 ± 3
25.1a	120	68	0.59	0.60	6.07	0.0541 ± 5.1	0.438 ± 5.2	0.05863 ± 1.3	367 ± 5
26.1a	71	26	0.39	0.42	3.56	0.0536 ± 3.6	0.431 ± 3.8	0.05832 ± 1.3	365 ± 5
27.1a	53	24	0.47	0.71	3.43	0.0526 ± 12	0.538 ± 12	0.0742 ± 2.7	461 ± 12
28.1a	106	53	0.52	0.53	5.22	0.0542 ± 3.4	0.427 ± 3.6	0.05707 ± 1.2	358 ± 4

Errors are 1-sigma; Pb$_c$ and Pb* indicate the common and radiogenic portions, respectively. Isotopic ratios are common Pb corrected using measured ^{204}Pb.
*Zircon characterization: a = Type 1, sector-zoned interiors; b = metamorphic rims; c = zoned cores; m = possible mixed analyses; r = other rims.
Uncertainties are at the 1σ level. See Appendix for analytical details.

likely it seems to be a relationship to post-HT processes during exhumation, because the zircon age overlaps within error with ^{40}Ar/^{39}Ar muscovite and biotite ages (328.8 ± 1.7 and 328 ± 2 Ma, Steltenpohl *et al.*, 1993) as well as some Rb–Sr biotite-whole rock ages (*c.* 325–332 Ma, Lange *et al.*, 2002)

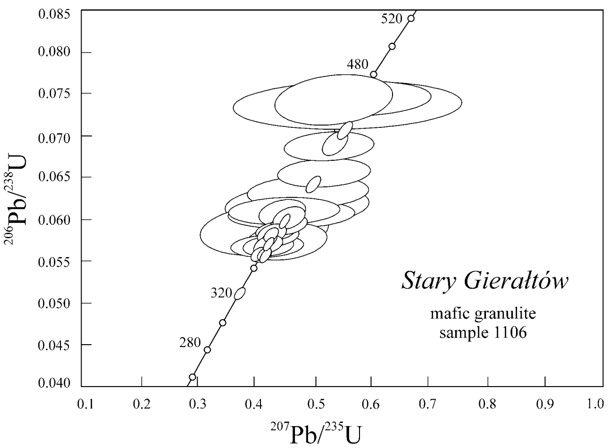

Fig. 6. U–Pb Concordia diagram of all SHRIMP data for zircon from sample 1106, Stary Gierałtów.

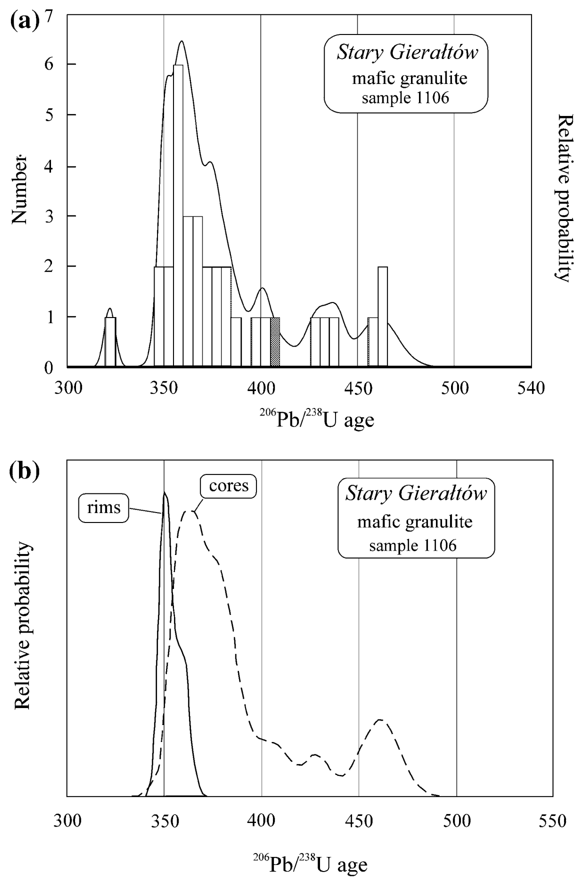

Fig. 7. (a) Combined cumulative probability plot and frequency histogram of $^{206}Pb/^{238}U$ ages for all SHRIMP data from sample 1106; (b) combined cumulative probability plots of $^{206}Pb/^{238}U$ ages for cores and rims of sample 1106.

reported for orthogneisses from the OSC. This suggests that this zircon dates a late stage on the cooling path (amphibolite to greenschist facies).

Possible geological significance of the *c.* 370–360 Ma ages

The main goal of the SHRIMP study was to unravel the geological significance of previously reported zircon multigrain dates of *c.* 370–360 Ma for granulites (Klemd & Bröcker, 1999). Although SHRIMP dating confirmed that these ages are not an artefact of multigrain dating, interpretation remains to be difficult, due to a large spread in $^{206}Pb/^{238}U$ ages (*c.* 462–322 Ma; Fig. 6; Table 3).

What is immediately obvious from the new data is that no single age can be calculated for the sector-zoned central areas (Table 3). The $^{206}Pb/^{238}U$ dates for these parts of the grains show a large variability. A cumulative probability plot of $^{206}Pb/^{238}U$ ages indicates a distinct maximum around 365 Ma, however, a spread towards higher ages is well-developed (Fig. 7b). Although these data could indicate a heterogeneous age population (i.e. the U/Pb ages are real), this is not convincing, because the studied domains appear to be all of the same type and generation. This suggests that the spread of apparent U/Pb ages must be a consequence of Pb-loss, although there is no correspondence between degree of resetting and U concentration (Table 3). Clearly the faint and 'washed-out' appearance of the zoning indicates some significant post-crystallization modification of these grains, which is also suggested by morphological criteria (prismatic grains with rounded terminations to strongly rounded grains) and internal features (broad homogenous domains, sector-zoning, relict oscillatory zoning), as well as the spread in $^{206}Pb/^{238}U$ dates. In contrast to the central areas, which show a large spread in $^{206}Pb/^{238}U$ ages, the zircon overgrowths yielded a narrow range between *c.* 350 and 361, but mostly cluster at *c.* 350 Ma (Table 3).

The closure temperature for diffusion of Pb in zircon is considered to be >900 °C (Lee *et al.*, 1997; Cherniak & Watson, 2001) and thus non-metamict zircon is considered to retain its original geochronological information under most crustal metamorphic conditions (e.g. Mezger & Krogstad, 1997). Ashwal *et al.* (1999) suggested that the U–Pb system in non-metamict zircon can experience resetting, due to volume and/or fracture-assisted diffusion, if the blocking temperature for zircon is exceeded for a long time span, or if granulite to amphibolite facies metamorphism repeatedly affected the zircon. Such a scenario is currently not documented for the granulites and their country rocks, but cannot easily be dismissed. Several studies have postulated a complex polymetamorphic evolution for the Sudetes, including both pre-Variscan and Variscan HT events (e.g. Don, 1990; Oliver *et al.*, 1993; Johnston *et al.*, 1994; Přikryl *et al.*, 1996; Cymerman *et al.*, 1997; Aleksandrowski *et al.*, 2000; Franke & Żelaźniewicz, 2000; Kröner *et al.*, 2000b, 2001; Żelaźniewicz *et al.*, 2002). Furthermore, it is well known that the metamorphic evolution of the European Variscides comprises at least two HP stages: an early pre-Late

Devonian episode (> 380 Ma) and a Carboniferous event (*c*. 340 Ma; e.g. O'Brien, 2000, and references therein). Unambiguous geochronological evidence for protracted or multiple high-grade metamorphism has yet not been found in the study area (Lange, 2004), but the record of pre-Variscan or pre-340 Ma HT metamorphism may have completely been erased during subsequent overprinting.

One interpretation is that zircon is affected by variable Pb-loss at the time the rims formed. The exact mechanism for this partial resetting – solid-state recrystallization due to radiation damage or expulsion of large-radius trace elements (e.g. Köppel & Sommerauer, 1974; Pidgeon, 1992; Hoskin & Black, 2000) or volume and/or fracture assisted diffusion (Ashwal *et al*., 1999) – remains to be determined. Important here is only the fact that pre-463 Ma igneous zircon was reset to variable degrees during metamorphism. It cannot be ruled out that this process is linked to prolonged and/or multiple heating events, however, at this stage solid-state recrystallization is considered as a more plausible interpretation.

At what stage of the metamorphic history did this occur? What is the significance of the age cluster at *c*. 365 Ma indicated by SHRIMP, ID-TIMS single and multigrain U–Pb methods? It is possible that there is a cause-and-effect link between eclogite to HP granulite facies metamorphism at *c*. 350–340 Ma and resetting of the U/Pb system in zircon. In this case, the 370–360 Ma dates would represent mixed ages without any geological relevance. Alternatively, it is possible that the 370–360 Ma dates mark the time of a distinct stage in the *P–T* evolution. In this context, it is interesting to note that a $^{207}Pb/^{206}Pb$ zircon evaporation date of 366.1 ± 1.1 Ma was also reported for an unfoliated granite of the study area, collected from a large melt patch within anatectic orthogneisses, adjacent to the granulite belt (Štípská *et al*., 2004). The similar ages obtained for zircon can be interpreted as supporting evidence that a distinct geological process affected the OSC at that time. However, at least three explanations must be considered: (a) Due to methodological limitations of the evaporation method (lack of evidence for concordance), the $^{207}Pb/^{206}Pb$ date strictly can only be interpreted as a minimum age. (b) This date provides valid geochronometric information and indicates zircon crystallization from a melt before or during early stages of high-temperature metamorphism (Klemd & Bröcker, 1999; Lange *et al*., 2003; Štípská *et al*., 2004). (c) The studied zircon represent xenocrysts derived from granulitic source rocks that were affected by partial Pb-loss within their host rocks. Subsequently, this zircon was entrained in anatectic melts that formed during metamorphism around 340 Ma.

Based only on available information it is difficult to ascertain which of these interpretations is most likely. SHRIMP dating and trace element studies of zircon from the melt patch and potential source rocks (orthogneisses and granulites) may contribute valuable data to solve this question. With the available data it cannot completely be ruled out that Pb-loss and incomplete resetting of zircon in the HP granulites is related to a HT event affecting the OSC around 370–360 Ma. This carries the implication that HP granulite to eclogite facies *P–T* conditions may have been attained over a much longer time span than generally accepted, however, it could also be argued that the 370–360 Ma ages constrain a distinct HT stage in the prograde evolution under medium pressures. Due to the absence of any supporting petrological observations or data, a relationship to early stages of HP–HT metamorphism can currently not be documented. Geochronological data for garnet–omphacite assemblages and multifaceted zircon suggest rather tight time brackets for the HP event (*c*. 350–340 Ma, Brueckner *et al*., 1991; Klemd & Bröcker, 1999; Štípská *et al*., 2004; this study). Models suggesting a distinctly longer time span for HP metamorphism can only be verified, if dating of unequivocal high-pressure assemblages, e.g. by means of Sm–Nd and Lu–Hf geochronology, confirms the 370–360 Ma ages.

CONCLUSIONS

This study complements the geochronological data set for the OSC and further substantiates the significance of UHP to HP metamorphism at *c*. 350–340 Ma. An important finding is the observation that in the granulites magmatic protolith zircon has experienced variable degrees of resetting during high-grade metamorphism. The initiating mechanism and the time of Pb-loss has yet to be resolved. A connection to eclogite to HP granulite facies metamorphism at *c*. 350–340 Ma is a realistic option, but this relationship is far from straightforward. An alternative interpretation suggests that resetting is related to a HT event around 370–360 Ma – not necessarily at high pressures – that has previously gone unnoticed. The results of this study suggest that caution is warranted in interpreting U–Pb data of zircon from HT rocks, because isotopic rejuvenation may lead to incorrect decoding of the geochronological record.

A key aspect for a better understanding of the OSC history and a challenge for future geochronological studies is the necessity to look through the metamorphic overprints at *c*. 350–340 Ma. Geochronology will play an important role in attempts to refine geodynamic interpretations, if distinct tectonometamorphic stages in the pre-350 Ma evolution can unambiguously be pin-pointed by precise ages.

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APPENDIX

Analytical methods

Sm–Nd and U–Pb ID-TIMS analyses were carried out at the *Zentrallabor für Geochronologie*, *Institut für Mineralogie*, *Universität Münster*, *Germany*. Sample digestion for Sm–Nd studies was carried out in Teflon® bombs within screw-top steel containers, similar to the method developed by Krogh (1973) for zircon. In a first step, whole-rock powders and mineral separates (garnet: *c.* 150–225 mg; omphacite: *c.* 220 mg) were mixed with a $^{149}Sm/^{150}Nd$ spike in Savillex® screw-top vials and dissolved in an $HF–HNO_3$ (5:1) mixture on a hot plate overnight. The solution was then reduced to a small volume by evaporation on a hot-plate. After adding fresh $HF–HNO_3$ (5:1), the sample containers were transferred into a Teflon® bomb within a Parr® steel autoclave and kept at 180 °C for a few days. After removal from the autoclave, a few drops of $HClO_4$ were added to break down fluorides during drying on a hot-plate. After evaporation, 6 N HCl was added to the residue to remove remaining fluorides and excess $HClO_4$ in a second evaporation step. Bulk REE were separated by standard ion-exchange procedures (AG 50W-X8 resin) on quartz glass columns using 2.5 and 6 N HCl as eluents. Sm and Nd were separated from the other REE fraction using Teflon® powder coated with 2-ethyl-hexyl phosphoric acid and 0.2 and 0.4 N HCl as eluting agent. For mass-spectrometric analysis, Sm and Nd were loaded with HCl on Re filaments using a triple filament configuration and measured on a VG Sector 54 multicollector mass spectrometer in dynamic mode. Correction for mass fractionation is based on $^{146}Nd/^{144}Nd = 0.7219$. The $^{147}Sm/^{144}Nd$ ratios were assigned uncertainties of 0.3% (2σ). Uncertainties of the $^{143}Nd/^{144}Nd$ ratios are reported on the $2\sigma_m$ level. Repeated runs of the LaJolla standard gave an average $^{143}Nd/^{144}Nd$ ratio of 0.5118602 ± 16 (2σ, $n = 11$).

Whole rock powders and mineral separates were prepared from fresh rock material. Samples were crushed in a jaw-crusher and an aliquot was ground in a tungsten carbide mill. For mineral separation, the remaining material was further reduced in size by grinding in a disc mill. After sieving, fines were removed by washing and minerals were enriched with a Frantz isodynamic magnetic separator. After additional hand-picking, mineral concentrates (optically pure > 99%) were washed in ethanol (p.a.) and ultrapure H_2O in an ultrasonic bath.

Morphological characteristics were documented by scanning electron microscopy (SEM) at the *Institut für Mineralogie*, *Universität Münster*, using a JEOL JSM 840A instrument. Catholuminescence imaging (CL) was carried out at the *Geochemisches Institut der Universität Göttingen* with a JEOL JXA 8900 RL electron microprobe.

For ID-TIMS U–Pb analyses, hand-picked zircon was washed in 7 N HNO_3 at 80 °C for 25 min. Subsequently, individual zircon grains were placed in multi-sample Teflon® microcapsules (Parrish, 1987) and dissolved for 4 days in HF-HNO_3 (4:1) at 180 °C. Dissolved zircon was spiked with a mixed ^{233}U–^{205}Pb tracer solution, dried at 80 °C, redissolved in 6 N HCl and equilibrated at 180 °C for one day. After drying at 80 °C, the samples were loaded on a single Re filament using a mixture of silica gel and 6 N HCl–0.25 N H_3PO_4. Isotope ratios of Pb and U were measured with a Daly type detector in ion-counting mode on a VG Sector 54 thermal ionization mass spectrometer (TIMS). Pb and U (as UO_2^+) were run sequentially on the same filament at temperatures of 1250–1350 and 1300–1450 °C, respectively. Average measured precision (2σ error) of Pb isotope ratios was 0.8% (^{206}Pb/^{204}Pb), 0.18% (^{207}Pb/^{206}Pb), and 0.06% (^{206}Pb/^{205}Pb). Ratios were corrected individually for mass fractionation by 1.4 ± 0.8‰/amu (^{207}Pb/^{206}Pb, ^{206}Pb/^{205}Pb), and 1.5 ± 0.8‰/amu (^{206}Pb/^{204}Pb), based on multiple analyses of NBS 982 standard. ^{233}U/^{238}U ratio was measured better than ±0.13% (2σ) and mass fractionation was corrected by 2.0 ± 0.8‰/amu based on analyses of NBS U-500 standard. For each charge of samples, the maximum Pb blank was assumed to be equivalent to the amount of the common Pb in the most radiogenic sample. The small sample size in combination with moderate U concentrations required the analysis of radiogenic Pb amounts as low as 4.8 pg. Consequently, analytical blanks had to be controlled at low level. In the course of this study, the Pb blank decreased from 3 to 1 pg due to improved reagents and sample handling. For initial lead correction, isotopic compositions were calculated according to the model of Stacey & Kramers (1975), assuming zircon crystallization at 360 Ma. In all samples $^{206}Pb_{rad}/^{206}Pb_{tot}$ is sufficiently high to obtain precise ^{206}Pb/^{238}U ages. The U blank was too small to be measured and was thus assumed to amount 20% of the individual Pb blank, based on experience with the analysis of mg-sized samples. Uncertainties in ^{206}Pb/^{238}U, ^{207}Pb/^{235}U and ^{207}Pb/^{206}Pb ages were calculated using the algorithm of Ludwig (1980) and errors are quoted at the 2σ level. Software Isoplot/Ex, rev. 2.49 (Ludwig, 2001) was used for the concordia plot.

For SHRIMP analyses, clean zircon from sample 1106 was mounted in epoxy at the Research School of Earth Sciences (RSES), together with the RSES reference zircons FC1 and SL13. Photomicrographs in transmitted and reflected light were taken, together with SEM CL images, in order to decipher the internal structures of the sectioned grains and to target specific areas within the zircon for analysis (spot size: *c.* 20 μm), e.g. metamorphic rims or inherited cores. U–Pb analyses were done in a number of sessions using SHRIMP 1I at the RSES. The data have been reduced in a manner similar to that described by Williams (1998, and references therein), using the SQUID Excel Macro of Ludwig (2000). For the zircon calibration the Pb/U ratios have been normalized relative to a value of 0.1859 for the ^{206*}Pb/^{238}U ratio of FC1 reference zircon, equivalent to an age of 1099 Ma (Paces & Miller, 1989). U and Th concentrations were determined relative to the SL13 standard. Uncertainties given for individual analyses (ratios and ages) are at the 1σ level in Table 3. Concordia and relative probability plots were carried out using Isoplot/Ex (Ludwig, 1999).