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Refining the granite, gneiss and schist interrelationships within the Lusatian–Izera Massif, West Sudetes, using SHRIMP U-Pb zircon analyses and new geologic data

Andrzej Żelaźniewicz[1, 2], Christopher M. Fanning[3] & Stanisław Achramowicz[1]

[1] *Instytut Nauk Geologicznych PAN, Podwale 75, 50-449 Wrocław, Poland; pansudet@pwr.wroc.pl*
[2] *Instytut Geologii, Uniwersytet Adama Mickiewicza, ul. Maków Polnych 16, 61-616 Poznań*
[3] *RSES, the Australian National University, Canberra*

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Abstract The eastern part of the Lusatian–Izera Massif, West Sudetes, comprises different types of gneissose rocks, collectively known as the Izera gneisses, with a subordinate component of petrographically varied mica schists. Coarse-grained gneisses and their protoliths have been dated at 515–480 Ma, but the lack of age data for other rocks has impeded accounts of their mutual relationships and, thus, the region's geological evolution. This paper reports new sensitive high-mass resolution ion microprobe (SHRIMP) U-Pb zircon data, and some new field and petrographic observations, for three representative rock types: 1) the Złotniki schist (a fine-grained quartz–albite–chlorite–sericite–biotite schist); 2) a fine-grained gneiss that grades to 'porphyroblastic' granite and which occurs on the slopes of Mt. Stóg Izerski; 3) a leucogranite found just the south of the village of Kotlina. A volcanogenic intercalation in the Złotniki Lubańskie schists developed at 560 Ma and contained xenocrystic zircons that grew in the source at 620 Ma and 600–580 Ma. The schists are interpreted as the metamorphosed equivalent of the Lusatian greywackes, which were derived from a dissected arc and deposited in a convergent-margin basin along northern peri-Gondwana. The zircons from the fine-grained gneisses yielded four age groups: 515 ± 7 Ma, 500 ± 12 Ma, 487 ± 13 Ma and 471 ± 8 Ma. Similar age groups of zircons can also be found in the coarse-grained metagranites. Rifting of Gondwana during the mid-Cambrian–early Ordovician was a protracted thermal event lasting ~30–45 m.y., with episodic attenuation of the mainland crust every ~5–10 m.y. before continental fragments finally became separated. Each episode successively promoted an increased heat flux from the mantle that facilitated melting of the crust, causing metamorphism and fusion of the Precambrian Lusatian–Izera basement and a final phase of S-type felsic magmatism. The leucogranite sample yielded zircons in two age groups, 508 ± 5 Ma and 483.1 ± 3.6 Ma, with low Th/U ratios, which is interpreted as a product of an anatectic melting at deeper crustal levels. These leucogranites are in close spatial relation with belts of mica schist, which could mean that these granites used some rheologically weak zones that were introduced into the Izera pluton where large fragments of country rocks were trapped within the ~500 Ma granites.

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INTRODUCTION

The eastern part of the Lusatian–Izera Massif (*sensu* Żelaźniewicz & Aleksandrowski, 2008), West Sudetes, is predominantly made of gneissose rocks known collectively as the Izera gneisses, with a subordinate component of petrographically variable mica schists. The origin of the Izera gneiss unit and its relationship to the regional geology, as well as internal gneiss–schist relationships themselves, has long been debated (Oberc, 1958, 1972; Kozłowska-Koch, 1965; Szałamacha & Szałamacha, 1968; Żaba, 1982, 1984; Oberc-Dziedzic, 1988; Smulikowski, 1972; Czapliński, 1998; Żelaźniewicz et al. 2003), not least because age data has been scarce and there have been insufficiently detailed petrologic and tectonic studies. This paper contributes to these discussions by providing new isotopic ages for hitherto undated rocks and discussing the regional implications of these dates.

A plethora of petrographic variants of the Izera gneisses have been distinguished by virtue of the observed textural, structural and compositional differences, with divergent genetic connotations given. Besides a variety of metagranites, also paragneisses, homophanous gneisses, anatectic granites, metasomatic granites and products of feldspathic granitization of mica schists, mylonites and blastomylonites were described by various authors. Some of the proposed ideas have today rather historical values, in particular the concepts promoting granitization and

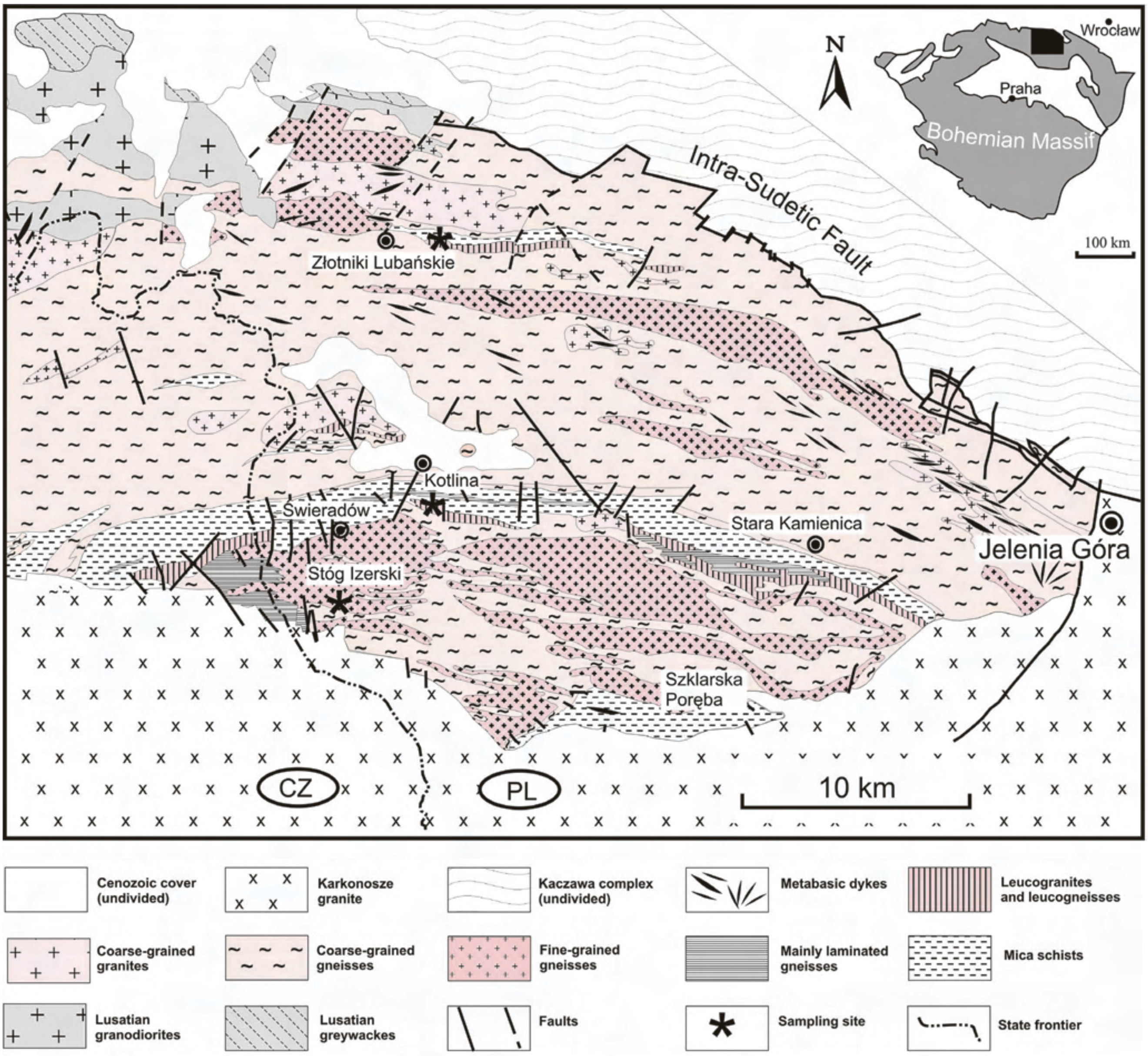

Fig. 1. Geological sketch map of the northeastern (Izera) part of the Lusatian–Izera Massif (modified after Milewicz *et al.* (1979) and Sawicki (1995). Inset shows map location (black box) within the Bohemian Massif.

metasomatic origin of the Izera gneisses (Kozłowska-Koch, 1965) or blastomylonitization (Szałamacha & Szałamacha, 1968).

Recent overview papers and small-scale maps on the Izera gneisses have tended to interpret the coarse-grained granites, augen gneisses, fine-grained gneisses and granites, and leucogranites as orthogneisses with minor cordierite, garnet and rare sillimanite (Oberc-Dziedzic, 2007). However, our literature and field studies show that part of the Izera gneisses are actually fine-grained paragneisses, which should be also taken into account.

The dominant orthogneisses range from almost undeformed granites via augen gneisses to thinly laminated mylonites and ultramylonites. To date, only the coarse-grained orthogneisses have been dated isotopically. A Rb-Sr mineral–whole rock isochron data (Borkowska *et al.*, 1980), U-Pb data on zircons (Korytowski *et al.* 1993; Oliver et al., 1993; Oberc-Dziedzic *et al.*, this volume) and Pb-Pb zircon evaporation method (Kröner *et al.*, 1994; Hegner & Kröner, 2000) yielded ages between 515 Ma and 480 Ma which have been interpreted as a time of intrusion of their granitic protolith. The fine-grained gneisses and leucogranites remain undated. Moreover, the ~500 Ma metagranites host a variety of gneissic enclaves, some of which contain relicts of mineral assemblages that are indicative of an earlier high pressure–high temperature episode (Achramowicz & Żelaźniewicz, 1998; Żelaźniewicz & Achramowicz, 1998).

The ~500 Ma metagranites enclose four narrow belts, along with smaller inliers, of mica schists (Fig. 1). These schists are generally interpreted to be remnants of older country rocks that have became reworked, alongside with their granitic hosts, by Variscan deformation and metamorphism (Oberc-Dziedzic, 2003; Żelaźniewicz *et al.* 2003). The schists differ significantly from each other in their mineral compositions and metamorphic grades (Smulikowski, 1972; Żelaźniewicz *et al.*, 2003). At least some of the rocks in the northernmost Złotniki Lubańskie schist belt, especially those rich in feldspar and chlorite, are very likely to have a volcanogenic origin, and this means that these could prove suitable in providing a maximum depositional age for the entire belt (Fig. 1).

In the present paper, we provide isotopic ages for the protolith of the Złotniki Lubańskie mica schists, leucogranite and fine-grained gneisses in order to shed more light on the currently disputed evolution of the Lusatian–Izera Massif and the relationships of its comprising lithologic and lithotectonic units. With this aim in mind, we analysed zircons from the Złotniki Lubańskie schsits, granites and gneisses, using the SHRIMP technique, and discuss the results within the framework of the available geological data.

GEOLOGICAL SETTING

In the eastern part of the Lusatian–Izera Massif, north of the Karkonosze pluton, the pre-500 Ma rocks are represented by the Lusatian granodiorite and the Lusatian greywacke, both of which occur within mica schists belts enclosed by the Izera gneisses (Fig. 1). There are at least four such E–W trending schist belts, each belt containing a variety of schistose rocks that display different mineral composition and metamorphic grades (Oberc, 1958, 1972; W. Smulikowski 1972; Kozłowski, 1974) but sharing a generally similar structural history (Żelaźniewicz *et al.*, 2003).

The northernmost belt is the Złotniki Lubańskie mica schist belt (Fig. 1) and is composed of a metasedimentary–volcanogenic series. These polygenic and multiply deformed rocks were metamorphosed under greenschist facies conditions.

In contrast, the Szklarska Poręba schist belt (Fig. 1) is composed of rocks that underwent metamorphism at significantly higher temperatures. Their metamorphism has been linked with the thermal influence of the adjacent Variscan-age Karkonosze granite (Borkowska, 1966; Smulikowski, 1972). However structural observations combined with petrologic data point to multiple tectono-thermal events and suggest that the Szklarska Poręba schists were affected at least twice by relatively high-temperature metamorphism, with an earlier episode coeval or preceeding the emplacement of the Izera granites at ~515–480 Ma (Achramowicz & Żelaźniewicz, 1998, unpublished data; Żelaźniewicz *et al.*, 2003). The earliest episode is marked by an older generation of cordierite that is structurally associated with early tight to isoclinal folds (F1); there is accompanying grain-scale evidence of melting along grain boundaries, including garnet partially molten in the presence of cordierite. High temperature conditions are also documented by the presence of a meionite–diopside–wollastonite–quartz assemblage in associated calc-silicate rocks. These textures, hitherto unreported, were overprinted by andalusite and a second generation of cordierite that developed syn- to post-tectonically with respect to local cascade-type folds (F3) that were superposed due to forcible intrusion of the Karkonosze granite.

The Stara Kamienica schist belt occurs between the Szklarska Poręba and Złotniki belts (Fig. 1) and is a polygenic unit that contains a variety of mica schists accompanied by quartzites, leptinites, granites and mylonitic augen gneisses, erlans and amphibolites. All these lithologies are spatially arranged into a few lithologically diverse assemblages that are separated by steeply dipping shear zones. These shear zones have brought into contact rocks that formed under variable P–T conditions, differing by ~200–300°C and 1–2 kbar (Żelaźniewicz *et al.*, 2003). The entire Stara Kamienica schist belt accommodated a major fault zone with significant normal component and helped to demarcate the Izera metagranite body into two domains (Fig. 1). In general, to the north of the belt, the coarse-grained orthogneisses dominate over the fine-grained gneisses, which occur as xenoliths and enclaves of various dimensions (Żelaźniewicz *et al.*, 2003). To the south of this belt, the fine-grained gneisses prevail and incorporate far smaller bodies of coarse-grained gneisses or granites. With respect to the inferred Stara Kamienica fault zone, the southern domain actually represents the footwall in which deeper levels of the Izera pluton have been exposed, possibly with more abundant fragments of metamorphic envelope.

In the southern domain, south of the Stara Kamienica schist belt, the metamorphic rocks are mainly fine-grained gneisses. New field evidence shows that the foliation planes of the gneisses are locally discordantly cut by the coarse-grained granite (orthogneisses): this helps determine the relative age relationships between the two major rock units. However, the fine-grained gneisses in this part of the massif are even more complex than has been previously reported. Some are represented by two-mica and two-feldspar gneisses, likely once metagreywackes; some are represented by chiefly quartzofeldspathic gneisses. The terminology by which previous authors discriminated between these mineralogically poor gneisses was reviewed by Żaba (1982); this emphasises the remarkable textural variations, despite the minor mineralogical differences. Żaba (1984) drew attention to the fact that fine-grained gneisses pass into granites with an almost ghostly preserved fabric, while these latter granites, which contain bluish quartz and feldspar megacrysts, have a so-called 'porphyraceous' texture that grades into porphyritic granites. And these porphyritic granites form veins with diffused boundaries within the coarser-grained gneisses (Żaba, 1984). Historically, similar quartz and feldspar megacrysts have been termed 'porphyroblasts' if observed in a rock with a gneissic fabric (M. Szałamacha, 1964; J.

Fig. 2. Quartz–albite–chlorite–sericite–biotite schist (metatuffite) from Złoty Potok, within the Złotniki Lubańskie mica schist belt.

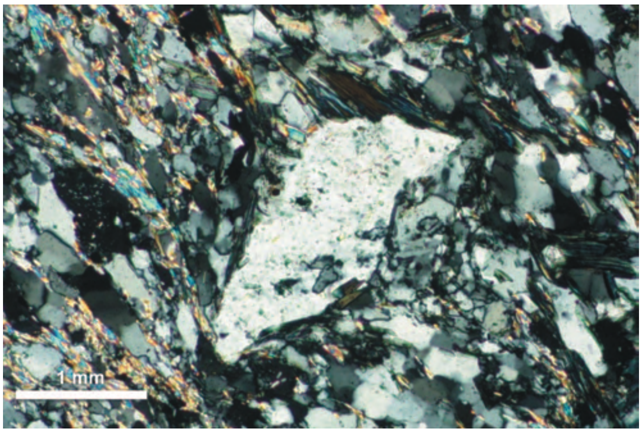

Fig. 3. Photomicrograph of the metatuffite shown in Fig. 2. Relict phenocryst is now a porphyroclast due to superposed shearing. Crossed polars.

Szałamacha, 1966) or 'porphyrocrysts' if present in granitic rocks (Oberc-Dziedzic, 2007). W. Smulikowski (1972) noted that euhedral zoned plagioclase phenocrysts and the presence of pinite pseudomorphs after cordierite and garnet signified crystallization from a melt. Indeed, the evidence for this process, as well as grain-scale melting, was written in the microstructure of both the fine-grained gneisses and granites. Our new textural observations include microcline phenocrysts with zonally arranged inclusions; rational faces of plagioclase against quartz, or cordierite against groundmass; euhedral grains of quartz and feldspars; and interstial plagioclase and the common breakdown of muscovite, both of which are features usually assigned to rocks undergoing melting or crystallizing from a melt (Vernon, 1999; Sawyer, 1999, 2008; Vernon & Paterson, 2008). Yet despite such small-scale evidence for melting, regular migmatites with leucosomes are absent in the region.

Some gneisses within the mica schist belts (Fig. 1) have been proposed to be evidence of anatectic granites (Żaba, 1984). We consider that the so-called intra-schist gneisses, with genuine augen fabric, can hardly be explained without invoking Izera granite veins and apophyses being injected into the schists and then being subjected to shear deformation. This interpretation is consistent with the overall geochemical similarities between the Izera gneisses as described by Oberc-Dziedzic *et al.* (2010), although detailed geochemical data from these gneisses is still lacking. The same 'intra-schist' position is locally also occupied by volumetrically less significant leucogranites, an observation that strengthens the hypothesis that there was an originally close spatial relationship between the Izera granites and the metasedimentary rocks at the time these granites were intruded.

The leucogranites, or leucogneisses whenever the former have been deformed, also occur close to or along the contacts with mica schist belts (Fig. 1). These leucogranites/leucogneisses have been interpreted in several different ways: (1) as a marginal facies of the Izera pluton (Berg, 1923); (2) as a product of alkali metasomatism bound to the fault zones (K. Smulikowski, 1958; K. Kozłowski, 1974); or (3), as the effects of alteration caused by hydrothermal influence of the Karkonosze granite (A. Kozłowski, 1978). The leucogranite at the southern border of the Stara Kamienica schists actually occurs in the gneissic footwall of a fault zone which coincides with the schist belt. Such a position has been used to support the idea of tectonic control on the origin and spatial distribution of the leucogranite/leucogneiss bodies (K. Smulikowski, 1958; M. Szałamacha, 1964; J. Szałamacha, 1966). Mineral and chemical compositions of the leucogranites differ slightly from other variants of the Izera gneisses (K. Kozłowski, 1974; Oberc-Dziedzic *et al.*, 2005). They locally contain tourmaline either as characteristic suns or as single grains, and the leucogranite tourmaline seems to resemble tourmaline aggregates that are occasionally seen in the fine-grained gneisses that grade to granites.

SAMPLED ROCKS

Zlpot Sample

The Złotniki schist belt was sampled East of Złotniki at Złoty Potok (N51°00.93', E15°22.28'), here abbreviated to the Zlpot Sample . The rock is a fine-grained quartz–albite–chlorite–sericite–biotite schist (Fig. 2) that had experienced three deformational episodes of which the two younger ones also affected the adjacent coarse-grained Izera gneisses. Syn- to intertectonic blastesis of rather randomly distributed biotite flakes overprints the earliest S1 axial planar foliation and locally follows the S0/S1 intersection lineation. This biotite grew due to a chlorite + white mica reaction that defines the peak conditions of metamorphism. The presence of still-recognizable quartz and albite phenocrysts, along with the mineral compo-

Fig. 4. Fine-grained gneiss with the character of a porphyrytic granite, from Mt. Stóg Izerski.

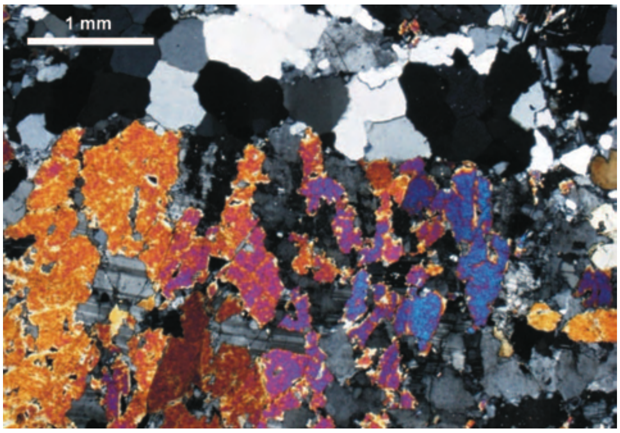

Fig. 5. Fine-grained gneiss displaying tourmaline intergrown with plagioclase; the fragments of the two minerals retain optical continuity: from Stóg Izerski. Crossed polars.

sition of the schists generally, point to a volcanogenic protolith. Deformational overprinting has converted these phenocrysts to porphyroclasts (Fig. 3).

The leucogranite veins that occur close to this schist follow the S1 planes. These veins are probably apophyses linked to the larger leucogranitic body that lies adjacent to the south (Fig. 1). Such an interrelationship suggests that the veins are younger than the metasedimentary–volcanogenic schists in the belt and, thus, they constrain the maximum age of these schists. Zircons were retrieved from the metavolcanogenic portion of the Zlpot sample for isotopic dating and to get more detailed constraints on a maximum depositional age of the protolith.

Istog Sample

The fine-grained gneisses that grade to 'porphyroblastic' granites (M. Szałamacha & J. Szałamacha, 1982) occur on the slopes of Mt. Stóg Izerski (Fig. 1), here shortened to 'Istog'. The sampled rock (N50°54.55', E15° 25.20'), the Istog sample, has the appearance of a porphyrytic granite with weak gneissic fabric (Fig. 4). It is composed mainly of two feldspars and quartz, with minor muscovite, cordierite and scarce biotite. Accessory tourmaline is locally abundant and tends to form nests in which large crystals are intergrown with plagioclase, fragments of the two minerals retaining optical continuity (Fig. 5). In this granite are apparently random megacrysts of euhedral to subhedral quartz (up to 1.5 cm) and K-feldspar (up to 5 cm), the latter containing zonally arranged quartz and plagioclase inclusions. In textural domains of no discernable fabric, rectangular plagioclase grains were able to develop their own faces against quartz or crystallize in the interstices between large grains of quartz and K-feldspar (Fig. 6). These are newly described textures. The Istog sample was collected to temporally constrain the formation of the protolith of these rocks and its further evolution.

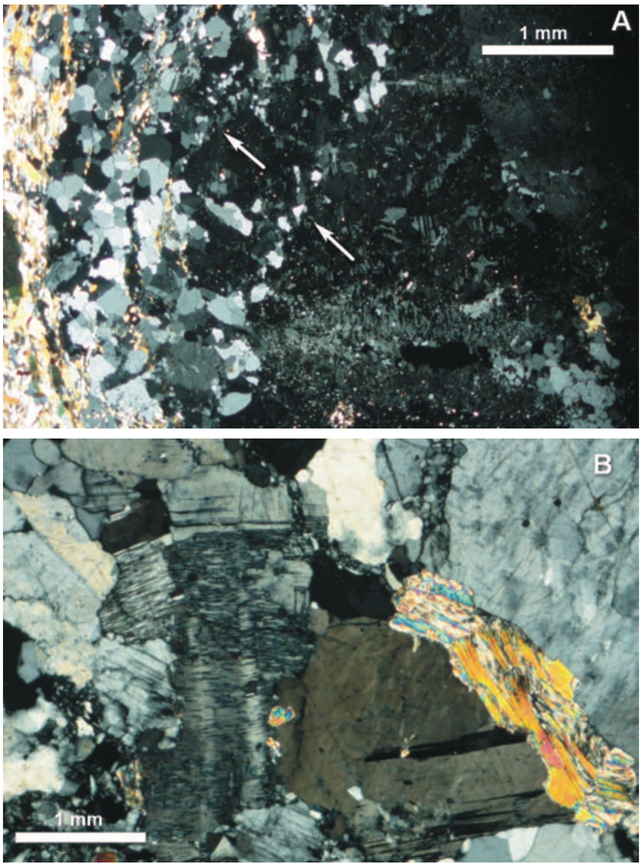

Fig. 6. Photomicrographs of the fine-grained gneiss shown in Fig. 4. (A) K-feldspar (dark) with zonally arranged quartz and plagioclase inclusions. (B) Subhedral feldspar grains in contact. Crossed polars.

Kotli Sample

Leucogranite was sampled in an abandoned quarry (N50°54.96', E15°23.09') to the south of the village of Kotlina (Fig. 1), the sample here being termed the Kotli sample. The rock is whitish and almost undeformed and contains quartz–microcline (perthite)–albite, locally with

Fig. 7. Leucogranite from Kotlina.

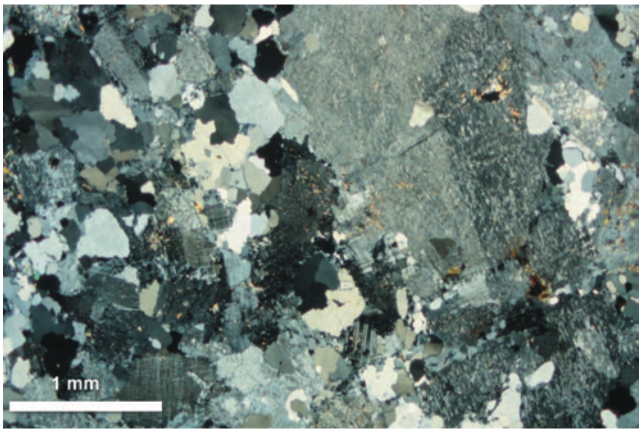

Fig. 8. Photomicrograph of the Kotlina leucogranite from Fig. 7. Crossed polars.

scarce single or somewhat aggregated biotite flakes (Figs. 7, 8). Microcline is partly replaced by chessboard albite or is, occasionally, overgrown by younger, grid-twinned microcline. Tourmaline is an accessory phase, which locally appears as randomly distributed nests or 'suns'. The tourmaline–feldspar nests present in the rock resemble analogous associations observed in the fine-grained gneisses, which may imply some ultimate genetic links between the two lithologic units.

SHRIMP ZIRCON STUDIES

Methodology

Zircon grains from the Zlpot, Istog, and Kotli samples were hand picked from heavy mineral concentrates and mounted in epoxy with chips of the FC1 (Duluth Gabbro) and SL13 (Sri Lankan gem) reference zircons. The grains were, approximately, sectioned in half and polished. Reflected and transmitted light photomicrographs and cathodoluminescence (CL) scanning electron microscope (SEM) images were taken for all zircons. The CL images were used to decipher the internal structures of the sectioned grains and to target specific areas within the zircons for isotopic analysis. The U-Pb analyses of the zircons were made using SHRIMP II, each analysis consisting of 6 scans through the mass range. The data were reduced in a manner similar to that described by Williams (1998, and references therein), using the SQUID Excel Macro of Ludwig (2001). The Pb/U ratios were normalised relative to a value of 0.1859 for the $^{206}Pb/^{238}U$ ratio of the FC1 reference zircon, equivalent to an age of 1,099 Ma (Paces & Miller, 1993). Uncertainties given for individual analyses (ratios and ages) are at the one σ level. The Tera and Wasserburg (1972) concordia plots and the relative probability plots were prepared using ISOPLOT/EX (Ludwig, 2005). Some examples of the analyzed zircons are shown on Fig. 9, 11 and 13, the analytical data are presented in Table 1 and concordia diagrams are given as Fig. 10, 12 and 14.

Results and interpretations

Zlpot Sample

Most zircons from the Zlpot sample are euhedral to subhedral, prismatic (200–250 mm long), usually with a low aspect ratio of 1:1.5 to 1:3, and show oscillatory zoning that is characteristic of igneous rocks (Fig. 9). Their internal structures vary. One group of grains shows simple and uniform zonation pattern (grains 4, 7, 6, 9, 11 and 12) so that they might have crystallized during one event only. The zircons of the second group are complex, with variably structured cores and rims (grains 1, 2, 3, 8 and 10), suggesting at least two growth episodes. A third, but minor, group is represented by zircons that are anhedral, often rounded, and may or may not be zoned (No. 13): these are possibly detrital.

Most analyses form two distinct groups (Fig. 10, Table 1). A younger group, (A), yields a mean age of 559.1 $\pm$ 6.4 Ma and is the age of either planar or oscillatory zoned euhedral/subhedral rims, which overgrew cores of variable structure. In one case (grain 10), the core is unzoned. The Th/U ratio in group A varies from 0.38 to 0.61. An older group, (B), yields a mean age of 620.6 $\pm$ 6 Ma and derives from both rims and cores of subhedral grains. Zircons of this groups show either simple oscillatory zoning or posses a more complex structure (Figs 9, 10). The scatter of the Th/U ratios is even greater than in group A and ranges from 0.05 to 1.02.

The remainder of zircons in the Zlpot sample represent Meso/Palaeoproterozoic and Archaean inheritance, mainly in the $\sim$1.9–2.1 Ga interval, as was recorded by rims (grain 3) or cores and rims in single grains (grains 6, 15). The latter are concordant, so have not suffered Pb-loss until recent times, and their zonal structure suggests a magmatic event around 2.0–2.1 Ga followed by a metamorphic one at $\sim$1.9 Ga. A slightly negatively discordant analysis (grain 2) yielded a $^{207}Pb/^{206}Pb$ age of 2.6 Ga,

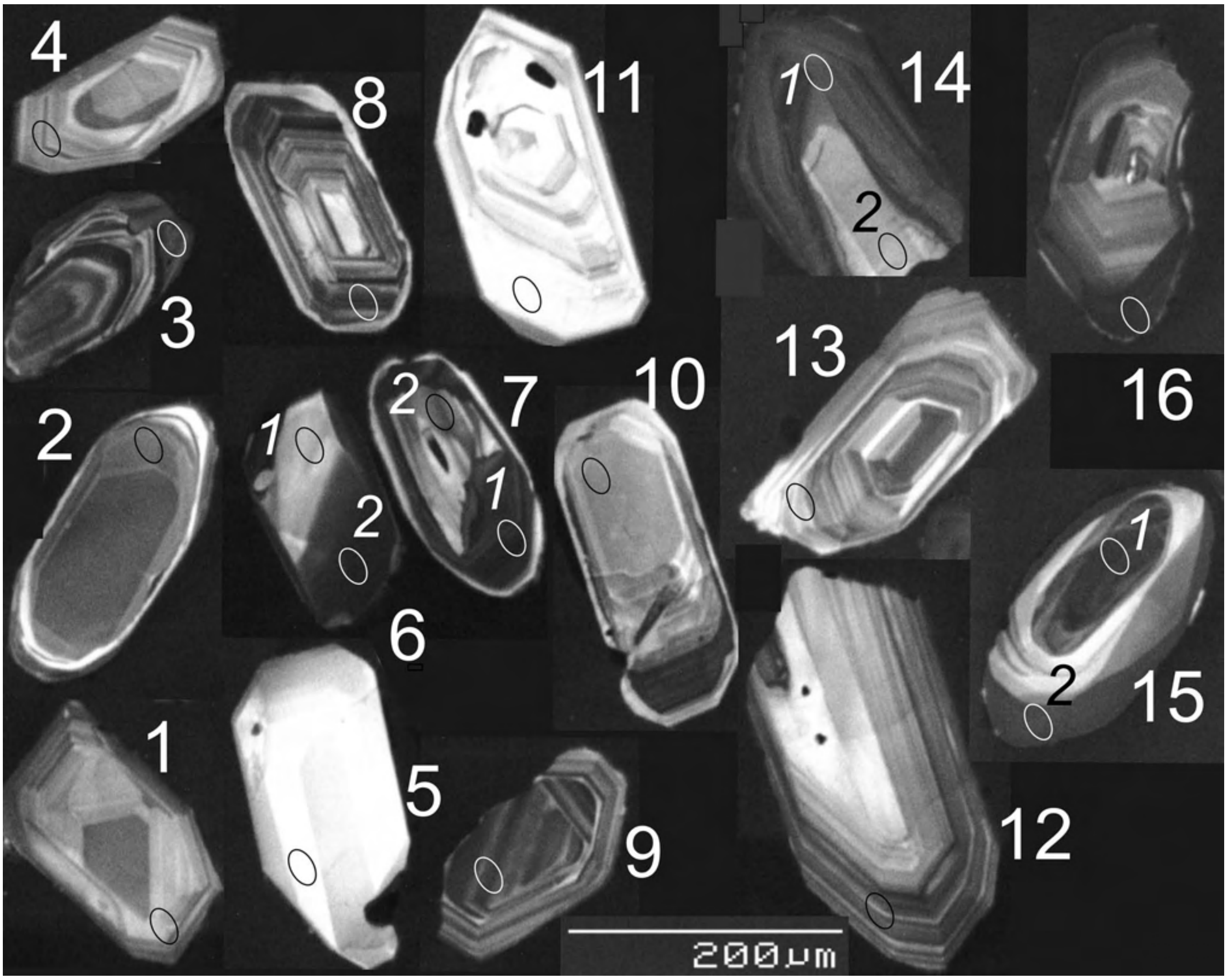

Fig. 9. Cathodoluminescence images of zircons from the Zlpot sample. Location of spot analyses are shown (data in Table 1).

revealing a still older zircon-growing event during the late Archaean. All inherited grains have more or less rounded shapes, which point to a detrital provenance.

The CL images of the Neoproterozoic zircons (groups A and B) show that some have oscillatory zoning and others display sector zoning. These observations suggest that the two groups were essentially grown during magmatic, probably anatectic, event(s). The euhedral morphologies of zircons from group A (grains 4,5 and 10–12) and their relatively high Th/U ratios (Table 1) are consistent with such interpretation. Because the Zlpot zircons come from quartz–albite–chlorite–biotite schists with still recognizable relics of former phenocrysts, the zircon data are in line with the conclusion about the volcanogenic (tuffogenic) origin of the protolith. Slightly rounded zircon morphologies (grains 7 and 8) of group B (~620 Ma) would imply that such grains passed through a hypergenic stage before they became finally incorporated into the ~560 Ma protolith. Taken together, the various lines of evidence support a volcanogenic origin for the schistose rocks in the Złotniki belt. The Variscan overprint occurred under greenschist facies conditions and caused strong fabric development in the Złotniki schists, but it was not capable of remobilizing Zr, therefore, no event younger than 560 Ma is likely to be recorded in the zircon population in these rocks.

In the western part of the Lusatian–Izera Massif, a widespread 570–540 Ma magmatic event has been repeatedly noted (Kröner *et al.*, 1994; Gehmlich *et al.*, 1997; Linnemann *et al.*, 2000; Linnemann & Romer, 2002; Gehmlich, 2003; Linnemann *et al.*, 2007). From this, the volcanic activity recorded in the Złotniki belt is most likely an event that can be interpreted as an igneous manifestation of the Cadomian orogeny (Linnemann *et al.*, 2007). Earlier papers did not report an older thermal event at ~620 Ma, as represented by group B in the Zlpot sample. Our new data suggest that the 560 Ma Cadomian event must have been preceded by an event ~60 Ma older. It may have been an early orogenic/magmatic phase of the Cadomian orogeny or else represent another, possibly independent, event associated with the break-up of the Rodinia supercontinent during the Neoproterozoic. Our data suggest that the ~620 Ma igneous rocks must have been exposed at the surface, eroded and clastic material delivered to thickening sedimentary–volcanogenic deposits at around 560 Ma. This is in line with observations in

Table 1

Summary of SHRIMP U-Pb zircon results for all analysed samples

							Total Ratios						Radiogenic Ratios						Age (Ma)				
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{238}U/^{206}Pb$	±	$^{207}Pb/^{206}Pb$	±	$^{206}Pb/^{238}U$	±	$^{207}Pb/^{235}U$	±	$^{207}Pb/^{206}Pb$	±	r	$^{206}Pb/^{238}U$	±	$^{207}Pb/^{206}Pb$	±	% Disc
Zlpot																						
1.1	181	68	0.38	14.3	0.000195	<0.01	10.833	0.137	0.0588	0.0008	0.0923	0.0012						569	7			
2.1	421	122	0.29	186.7	0.000014	0.02	1.936	0.020	0.1765	0.0004	0.5165	0.0055	12.560	0.137	0.1764	0.0004	0.973	2684	23	2619	4	-2
3.1	460	33	0.07	146.4	0.000037	0.05	2.699	0.029	0.1269	0.0004	0.3703	0.0040	6.454	0.073	0.1264	0.0004	0.950	2031	19	2048	6	1
4.1	263	76	0.29	21.2	0.000175	0.14	10.674	0.121	0.0604	0.0007	0.0936	0.0011						577	6			
5.1	83	41	0.49	6.5	0.000558	0.09	11.072	0.154	0.0595	0.0013	0.0902	0.0013						557	8			
6.1	123	82	0.67	35.8	0.000035	0.05	2.949	0.045	0.1157	0.0011	0.3389	0.0052	5.383	0.097	0.1152	0.0011	0.848	1881	25	1883	17	0
6.2	2140	1016	0.47	622.1	0.000020	0.03	2.955	0.030	0.1155	0.0002	0.3383	0.0034	5.374	0.056	0.1152	0.0002	0.984	1878	17	1883	3	0
7.1	880	112	0.13	76.8	0.000029	<0.01	9.852	0.103	0.0603	0.0004	0.1015	0.0011						623	6			
7.2	278	283	1.02	24.9	0.000106	0.10	9.596	0.108	0.0618	0.0007	0.1041	0.0012						638	7			
8.1	710	674	0.95	60.7	0.000050	<0.01	10.060	0.124	0.0593	0.0004	0.0995	0.0013						612	7			
9.1	581	459	0.79	50.4	0.000106	<0.01	9.903	0.106	0.0590	0.0005	0.1012	0.0011						621	7			
10.1	209	111	0.53	16.3	0.000135	0.11	11.029	0.243	0.0597	0.0009	0.0906	0.0020						559	12			
11.1	95	58	0.61	7.4	0.000537	0.02	11.034	0.150	0.0589	0.0012	0.0906	0.0013						559	7			
12.1	824	356	0.43	63.5	0.000006	0.09	11.142	0.117	0.0593	0.0004	0.0897	0.0010						554	6			
13.1	426	205	0.48	34.6	0.000025	<0.01	10.597	0.116	0.0591	0.0006	0.0944	0.0011						582	6			
14.1	796	42	0.05	69.7	-	<0.01	9.818	0.103	0.0603	0.0004	0.1019	0.0011						625	6			
14.2	217	227	1.05	18.7	-	<0.01	9.937	0.118	0.0600	0.0009	0.1007	0.0012						618	7			
15.1	1350	27	0.02	458.6	0.000001	<0.01	2.528	0.027	0.1414	0.0004	0.3956	0.0043	7.714	0.087	0.1414	0.0004	0.963	2149	20	2245	5	4
15.2	769	357	0.46	227.4	0.000019	0.03	2.906	0.033	0.1235	0.0004	0.3441	0.0039	5.849	0.068	0.123	0.000	0.966	1906	18	2004	5	5
16.1	1550	32	0.02	143.4	0.000010	<0.01	9.286	0.105	0.0603	0.0005	0.1079	0.0012						660	7			
Istog																						
1.1	560	40	0.07	39.7	0.000058	<0.01	12.125	0.150	0.0562	0.0005	0.0826	0.0010						511.7	6.2			
2.1	290	61	0.21	19.9	0.000122	0.11	12.551	0.142	0.0579	0.0008	0.0796	0.0009						493.7	5.5			
3.1	700	83	0.12	47.0	0.000069	0.03	12.785	0.136	0.0571	0.0005	0.0782	0.0008						485.3	5.0			
4.1	122	74	0.60	44.5	0.000021	0.03	2.362	0.029	0.1300	0.0009	0.4260	0.0052	7.911	0.115	0.1347	0.0011	0.834	2287	25	2160	14	-6
5.1	1229	61	0.05	85.6	0.000043	<0.01	12.328	0.138	0.0563	0.0004	0.0812	0.0009						503.4	5.5			
5.2	374	43	0.12	26.6	0.000003	0.04	12.086	0.134	0.0579	0.0007	0.0827	0.0009						512.2	5.6			
6.1	384	33	0.09	25.2	0.000111	0.09	13.086	0.146	0.0572	0.0007	0.0764	0.0009						474.3	5.2			
6.2	1241	347	0.28	102.0	0.000181	0.24	10.451	0.113	0.0615	0.0004	0.0955	0.0011						587.7	6.2			
7.1	761	69	0.09	50.9	0.001692	2.96	12.840	0.137	0.0804	0.0030	0.0756	0.0009						469.7	5.5			
8.1	212	58	0.27	17.2	0.000052	<0.01	10.586	0.147	0.0583	0.0008	0.0946	0.0013						582.7	7.9			
8.2	790	98	0.12	53.0	0.002948	4.86	12.802	0.149	0.0957	0.0087	0.0743	0.0013						462.1	7.9			
9.1	469	41	0.09	32.5	0.000168	<0.01	12.414	0.136	0.0568	0.0006	0.0806	0.0009						499.7	5.4			
10.1	701	53	0.08	50.7	0.000580	0.93	11.875	0.127	0.0653	0.0006	0.0834	0.0009						516.5	5.4			

10.1	872	111	0.13	63.4	0.000063	<0.01	11.813	0.135	0.0574	0.0005	0.0847	0.0010						524.1	5.9			
11.1	248	90	0.36	16.6	0.000128	<0.01	12.842	0.150	0.0564	0.0009	0.0779	0.0009						483.6	5.5			
11.2	716	305	0.43	60.0	0.000103	0.15	10.250	0.109	0.0611	0.0005	0.0974	0.0011						599.3	6.2			
12.1	849	55	0.06	58.2	0.000196	0.10	12.529	0.132	0.0579	0.0005	0.0797	0.0009						494.6	5.1			
13.1	754	53	0.07	52.8	0.000114	0.03	12.251	0.141	0.0576	0.0005	0.0816	0.0010						505.7	5.7			
14.1	464	39	0.09	34.7	0.000197	0.29	11.503	0.125	0.0605	0.0006	0.0867	0.0010						535.9	5.7			
15.1	220	52	0.23	14.7	0.000233	<0.01	12.877	0.175	0.0544	0.0009	0.0779	0.0011						483.5	6.4			
16.1	347	35	0.10	24.5	0.002515	3.86	12.172	0.155	0.0883	0.0076	0.0790	0.0013						490.0	8.0			
Kotli																						
1.1	370	95	0.26	24.9	0.000081	0.07	12.743	0.117	0.0575	0.0008	0.0784	0.0007						486.7	4.4			
2.1	455	210	0.46	35.6	0.000099	0.09	10.996	0.096	0.0596	0.0006	0.0909	0.0008						560.6	4.8			
3.1	835	40	0.05	58.4	0.000035	<0.01	12.277	0.103	0.0564	0.0004	0.0816	0.0007						505.4	4.2			
2.2	491	39	0.08	35.8	0.000000	0.05	11.763	0.104	0.0583	0.0008	0.0850	0.0008						525.8	4.6			
4.1	168	56	0.33	14.0	0.000155	<0.01	10.328	0.105	0.0587	0.0009	0.0970	0.0010						596.5	6.0			
4.2	1847	100	0.05	131.9	0.000119	0.08	12.029	0.095	0.0582	0.0003	0.0831	0.0007						514.4	4.0			
5.1	516	56	0.11	35.8	0.000078	<0.01	12.394	0.117	0.0557	0.0006	0.0808	0.0008						501.1	4.6			
6.1	232	214	0.92	20.8	-	0.21	9.565	0.136	0.0627	0.0010	0.1043	0.0015						639.7	8.9			
6.2	425	39	0.09	30.0	0.000057	<0.01	12.147	0.106	0.0568	0.0006	0.0824	0.0007						510.4	4.4			
7.1	470	17	0.04	33.3	0.000056	<0.01	12.119	0.106	0.0575	0.0006	0.0825	0.0007						511.1	4.4			
7.2	378	21	0.06	28.3	-	<0.01	11.480	0.122	0.0567	0.0007	0.0873	0.0009						539.4	5.6			
8.1	320	52	0.16	21.3	-	0.02	12.901	0.119	0.0569	0.0007	0.0775	0.0007						481.2	4.4			
9.1	217	46	0.21	14.4	0.000172	0.20	12.976	0.130	0.0583	0.0009	0.0769	0.0008						477.6	4.7			
9.2	209	75	0.36	59.8	0.000047	0.07	3.010	0.036	0.1149	0.0053	0.3320	0.0040	5.233	0.251	0.1143	0.0053	0.250	1848	19	1869	84	1
10.1	375	282	0.75	30.7	0.000071	<0.01	10.518	0.096	0.0591	0.0007	0.0951	0.0009						585.8	5.2			
10.2	334	19	0.06	22.5	-	0.01	12.735	0.117	0.0570	0.0007	0.0785	0.0007						487.3	4.4			
11.1	182	38	0.21	12.4	0.000101	<0.01	12.560	0.130	0.0549	0.0010	0.0798	0.0008						495.1	5.0			
12.1	623	204	0.33	41.8	0.000007	0.01	12.815	0.134	0.0569	0.0005	0.0780	0.0008						484.3	5.0			
13.1	482	52	0.11	32.2	0.000118	0.13	12.880	0.113	0.0578	0.0006	0.0775	0.0007						481.4	4.1			
14.1	1389	99	0.07	97.5	0.000022	0.03	12.247	0.110	0.0576	0.0003	0.0816	0.0007						505.8	4.4			

Uncertainties are given at the one σ level. The percentage of ^{206}Pb (common Pb) is given by f_{206} %. For % Discordance, 0% denotes a concordant analysis. For areas older than ~800 Ma, a correction for common Pb was made using the measured $^{204}Pb/^{206}Pb$ ratio. For areas younger than ~800 Ma, a correction for common Pb was made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ ratios following Tera and Wasserburg (1972) as outlined in Williams (1998). For the Zlpot and Istog samples, the error in the FC1 reference zircon calibration was 0.55% for the analytical sessions (not included). For the Kotli sample, the error in the FC1 reference zircon calibration was 0.34% for the analytical session (not included).

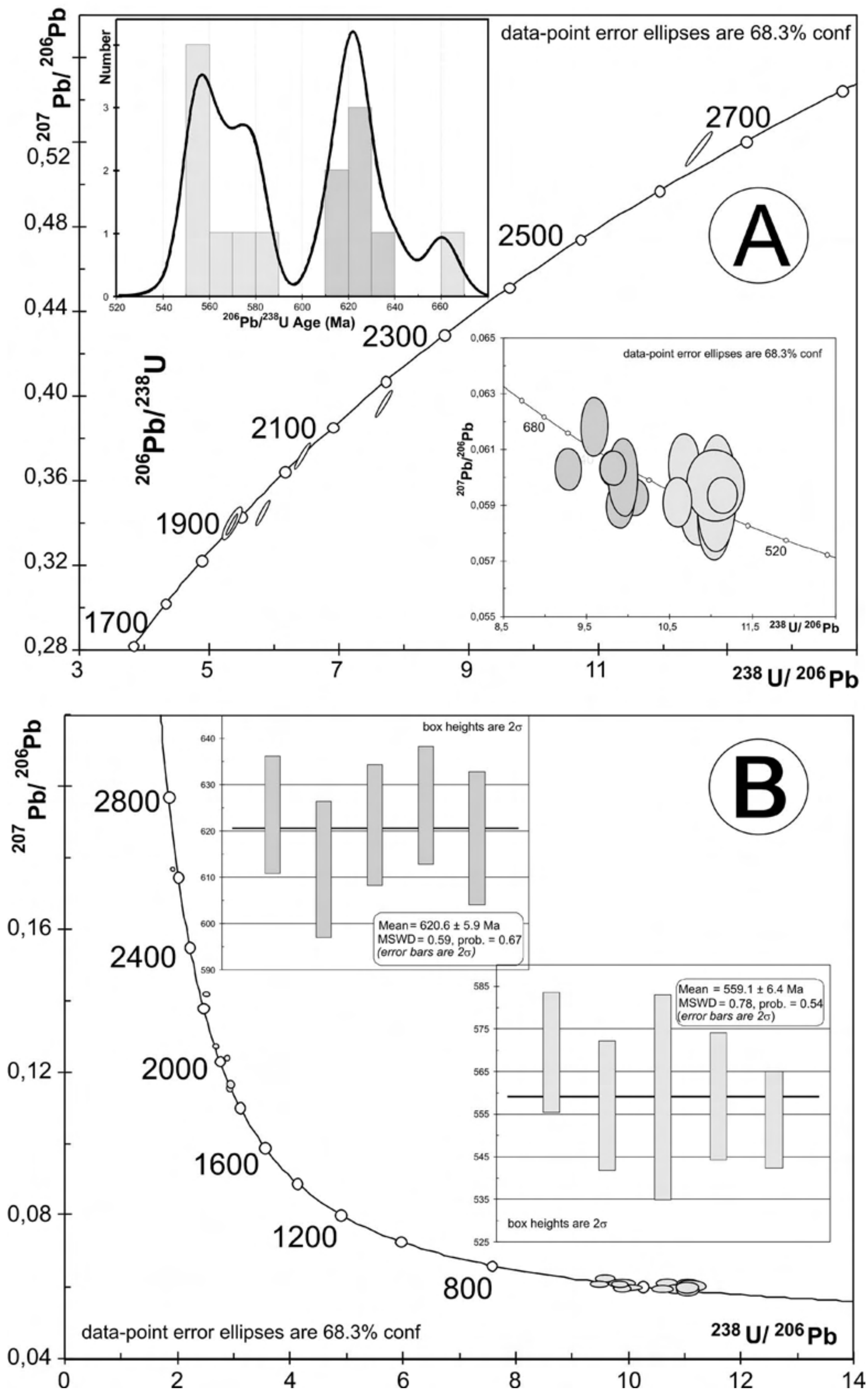

Fig. 10. Diagrams for SHRIMP U-Pb zircon analyses from the Zlpot sample. (A) Wetherill concordia plot for older zircons. Tera–Wasserburg concordia plot for younger zircons in lower right inset; relative probability plot with stacked histogram in upper left inset. Analyses are plotted as 1σ error ellipses. (B) Tera–Wasserburg concordia plot of all data; insets show weighted averages for the two age groups.

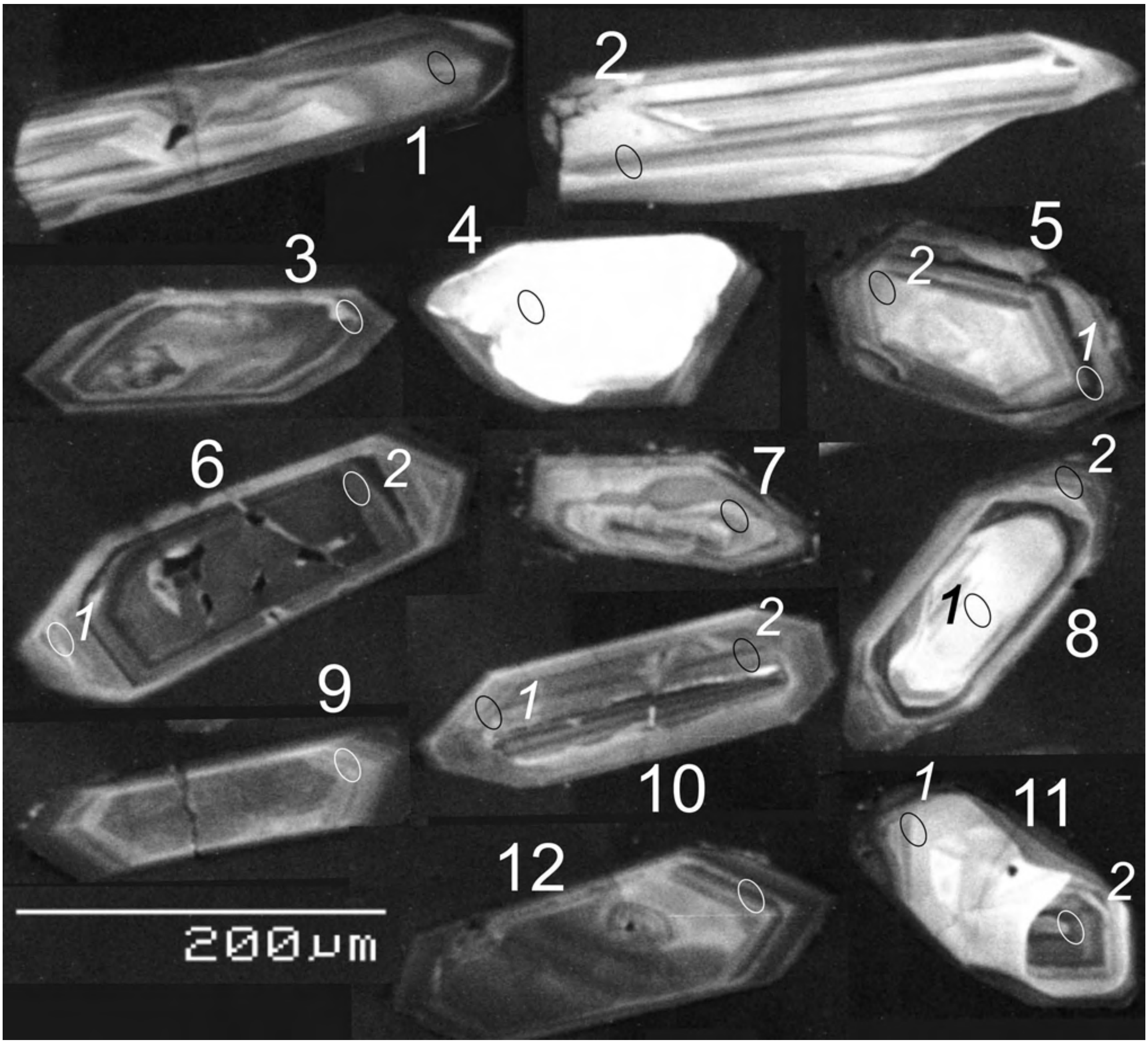

Fig. 11. Cathodoluminescence images of zircons from the Istog sample, with locations of analysed spots.

Lusatia, further west, where the Lusatian greywackes are intercalated with pyroclastic layers dated at 562 ± 4 Ma and conglomerates that contain pebbles of magmatic rocks dated at 577–573 Ma (Pb-Pb zircon, Gehmlich *et al.*, 1997). As the ~560 Ma volcanic part of the Złotniki Lubańskie schist belt also proves the existence of a sedimentary basin of that age, we conclude that rocks in this schist belt are equivalent to the Lusatian greywacke. A similar conclusion, based on geochemical similarities, is drawn by Oberc-Dziedzic *et al.* (this volume).

Kemnitz (2007) interpreted the Lusatian greywacke deposits as derived from a dissected magmatic arc and acccumulated in a convergent-margin basin along northern peri-Gondwana. We expect that at least the older part of that arc may have developed at ~620 Ma, although rocks of such age have yet to be found in Lusatia.

Istog Sample

Zircons from the Istog sample are also 200–250 mm long, with bimodal axial ratios of 1:4–5 and 1:2 (Fig. 11). When viewed under transmitted and reflected light, the grains are euhedral. In CL images, some Istog zircons show planar zoning, but most reveal a more or less complex core-and-rim structure. The cores may be remarkably uniform and unzoned or they may have planar to radial zoning, often with diffuse irregular boundaries. Grains with oscillatory zonation are relatively infrequent.

Sixteen analyses yielded a mean $^{206}Pb/^{238}U$ age of 497 ± 10 Ma, which coincides with an average age of ~500 Ma obtained for most U-Pb studies of the Izera gneisses. But there is a spread of analyses between 516 Ma and 462 Ma that reveals four age groups: 515 ± 7 Ma, 500 ± 12 Ma, 487 ± 13 Ma and 471 ± 8 Ma. This spread is typical in these

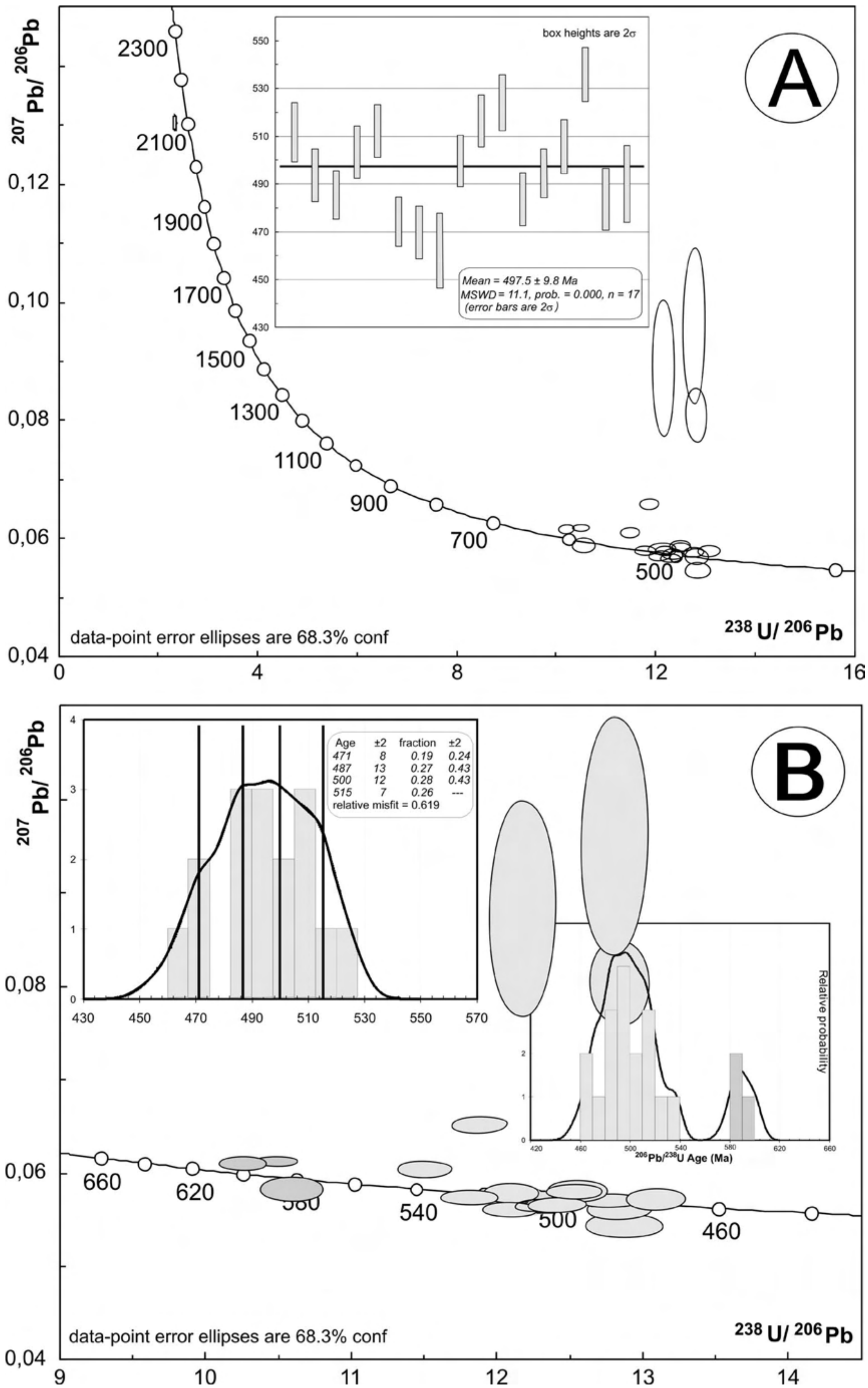

Fig. 12. Tera–Wasserburg diagrams for SHRIMP U-Pb zircon analyses from Istog sample, analyses plotted as 1σ error ellipses. (A) Plot of all data; inset shows weighted average. (B) Plot for younger zircons; inset shows relative probability plot for four age groups between 515 Ma and 470 Ma.

rocks and suggests a thermal event of ~35–30 m.y. duration (~515–480 Ma). There is also a group of ages spanning 600–580 Ma (Fig. 12, Table 1) only yielded by cores of euhedral crystals (grains 6, 8 and 11). The cores are almost unzoned, or have oscillatory zoning, and represent either euhedral outlines or fragments of resorbed grains. They became extensively overgrown by complex, metamorphic or anatexis-related rims that belong to an age group of 471 ± 8 Ma. Such zircon structures suggest strong remobilization of an unknown felsic protolith of Neoproterozoic age. Both magmatic and metamorphic processes led to a remarkable degree of dissolution of the older zircons and to re-growth as rims (grains 6,8 and 11) or as new grains (1–3,5,7 and 9–16) with euhedral shapes. Indistinct zoning, disturbed sector zoning and oscillatory zoning together suggest crystallization from a melt. One zircon of 2.2 Ga age points to early Palaeoproterozic crust, an inference similar to that drawn from the Zlpot sample.

The presence of the four age groups within an interval of ~515–470 Ma found in the Istog sample seems to indicate a tectonothermal event up to 45 m.y. long, though the details of this proposed event remain obscure. The sample itself is a felsic gneiss with a weakly preserved fabric, enclosing megacrystic quartz and feldspars, and with a microstructural record of partial melting. Its metamorphic history stopped just before it became a granite. The presence of many zircons with euhedral shapes and zonal structures suggest a melt-assisted crystallization. The inferred metamorphic conditions and anatectic nature of the melt are consistent with the very low Th/U ratios, between 0.06 and 0.12, that were measured: only a few were in the range of 0.27–0.36. Such somewhat higher Th/U ratios, which refer to younger zircons dated at 483 Ma, may possibly reflect the attainment of a higher volume of partial melt. The youngest ages of around 460 Ma, measured in outer parts of the grains, are apparently too low, probably be due to some Pb loss, and were discarded. The studied zircon population is small, which at this stage of the study impedes better resolution of the ~515–470 Ma thermal activity. The host rock may have undergone metamorphic transformation at ~515–500 Ma and some degree of melting at ~483 Ma. More extensive migmatiza- tion presumably occurred in deeper, largely unexposed, levels of the Lusatian–Izera basement as suggested by some field obervations.

Kotli Sample

In the Kotli sample, zircons are slightly smaller than in the other samples (120–200 mm long), with quite low aspect ratios of 1:1.5 to 1:2 (Fig. 13). Most grains have cores and rims, but are variably structured. Some cores are clearly xenocrystic; rims are of different breadth and can be either densely or broadly oscillatory zoned or show poor zoning. A few cores are oval and almost unzoned, others are irregularly and patchily zoned, while yet others display only weak, broad radial zonation. Grains are euhedral to subhedral, generally bright in CL.

In the Kotli leucogranite sample, most $^{206}Pb/^{238}U$ ages (13 out of 20) are spread between ~515 Ma and ~480 Ma, six ages bracket Neoproterozoic events (640–560 Ma) and 1 grain (core) proved to be ~1.85 Ga old. The dominant age cluster can be subdivided into two distinct groups: 508 ± 5 Ma (8 analyses), and 483.1 ± 3.6 Ma (six analyses; Fig. 14, Table 1). These age clusters are defined by zircons whose structures may indicate both magmatic and metamorphic or anatectic origins. An anatectic origin is consistent with low Th/U ratios of between 0.05 and 0.21, which is similar to values measured for the fine-grained gneiss of the Istog sample. Except for one whole acicular zircon crystal, the youngest measured ages were invariably from outgrowths that showed radial sector zoning or planar banded zoning that had developed on older grains with which some of the outgrowths may have been in crystallographic continuity. Therefore, a melt phase may have been in operation around 483 Ma. This, along with aggregated K-feldspar grains and the presence of tourmaline, resembles to some extent the fine-grained gneiss in the Stóg Izerski–Świeradów area and suggests that the two rock units are likely to have shared some genetic links. Such an impression is enhanced by the inherited zircons whose ages of 640–560 Ma are consistent with Neoproterozoic events described above. The inherited Mesoproterozoic component is compatible with that found in the Złotniki Lubański metatuffitic schists, and the spread of Neoproterozoic ages matches the zircons of groups A and B from these metatuffites. Thus, we suggest that the leucogranite protolith came from a source with a crustal history similar to that revealed by the schists.

DISCUSSION AND CONCLUSIONS

Previously, in the Lusatian–Izera Massif, two older magmatic events were known to have occurred before the Izera granite was emplaced in mid-Cambrian–early Ordovician times. One of them was represented by arc-related granitoids that developed at 587–560 Ma (Kröner *et al.*, 1994; Gehmlich, 2003). These were eroded and deposited in conglomerates intercalated with the Lusatian greywacke at 560 Ma (Gehmlich *et al.*, 1997). The other magmatic event was marked by post-tectonic granodiorites, dated at 540–530 Ma, which intruded the older granitoids and greywackes (Korytowski *et al.*, 1993; Białek, 2003; Żelaźniewicz *et al.*, 2004) and produced contact metamorphic effects in the greywackes (Kemnitz, 2007) at the end of the Cadomian orogeny (Linnemenn *et al.*, 2007).

New data reported in this paper suggest that at least one, but possibly two, additional thermal events contributed to the formation of the Precambrian Lusatian–Izera crust. Xenocrystic zircons from the Złotniki schists and from the fine-grained Izera gneisses clearly indicate that they grew at 620 Ma and 600–580 Ma, respectively. Although their host rocks are unknown, they must have been reworked by crustal processes that produced the

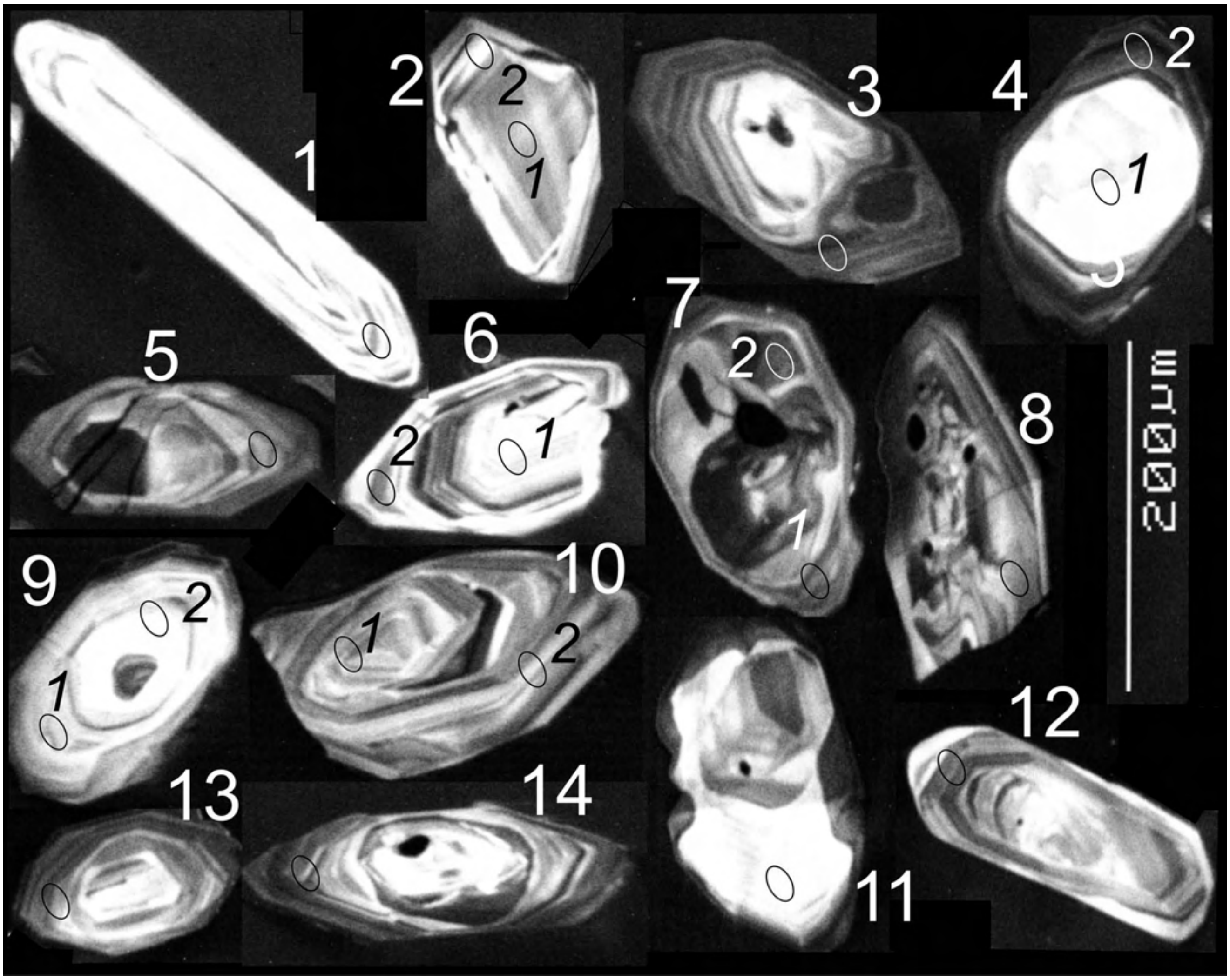

Fig. 13. Cathodoluminescence images of zircons from Kotli sample, with location of analysed spots.

~560 Ma volcanogenic rocks and 515–480 Ma S-type granites and leucogranites.

As far as can be judged from the xenocrystic zircons, a protolith of the fine-grained Izera gneiss may have originated around 600–580 Ma. Although older than the Złotniki volcanogenic rocks, the two units likely represent similar Cadomian crust. This crust was then involved in a protracted thermal episode at ~515–470 Ma. In the Variscan belt, an episode of this age is commonly connected with rifting of Gondwanan crust (e.g. Pin *et al.*, 2007) and the eventual formation of the Rheic Ocean (Linnemann *et al.*, 2007) and the associated seaways between Gondwana-derived terranes that ultimately accreted to Laurussia during the Late Palaeozoic, so forming Variscan Europe. Attenuation of the crust before continental fragments became separated may have occurred episodically, with significant increments at every ~5–10 m.y. Each episode successively promoted an increased heat flux from the mantle that facilitated melting of the crust, metamorphism and eventual fusion of the Precambrian Lusatian–Izera basement, which terminated with felsic magmatism in mid-Cambrian–early Ordovician times.

The same model can be applied to the formation of other ~515–470 Ma granitoid plutons elsewhere in the Sudetes, for instance in the Orlica–Śnieżnik Dome where migmatic gneisses and porphyrytic metagranite formed at a similar time between 515 Ma and 480 Ma (Franke & Żelaźniewicz, 2000). Intrusions of a porphyritic granite (Śnieżnik augen gneiss) during this period were accompanied by migmatization of Neoproterozoic rocks, which obliterated many earlier features and allowed a great number of zircons to dissolve and (re)crystallized from a melt. Plutonic processes were associated with effusive ones which gave rise to rhyolitic lava flows and associated volcanoclastic rocks alternating with metasedimentary rocks (Jastrzębski *et al.*, 2010).

A statistical subdivision of the age numbers yielded by such rocks into two, three or more groups depends on the sample itself and whether it is magmatic, migmatitic, or metamorphic. Whichever the case, Zr must have been available to ensure the growth of zircons over the entire period of ~35–40 m.y. duration. If the availability of Zr supposedly increased every 5–10 m.y., the statistically distinguishable age groups may have developed as it is actu-

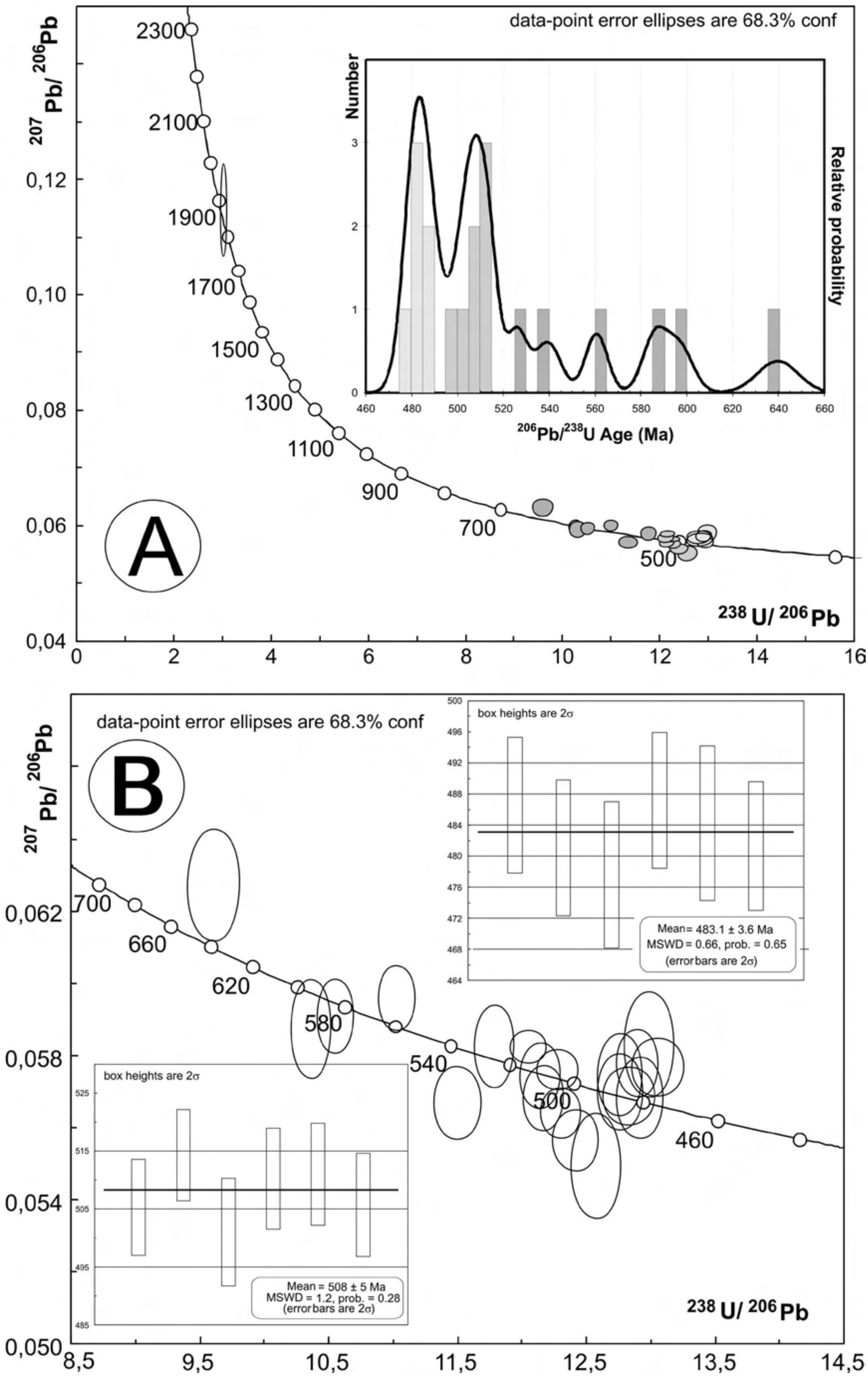

Fig. 14. Tera–Wasserburg diagrams for SHRIMP U–Pb zircon analyses in Istog sample; analyses plotted as 1σ error ellipses. (A) Plot of all data; inset shows relative probability plot. (B) Plot for younger zircons; inset shows weighted averages for the two age groups revealed.

ally observed in the Izera or Orlica–Śnieżnik rocks. Magmas derived from the deep crust during sequential episodes of increased heat flux from the mantle might have contributed to a magma chamber by forming a mix of successive batches of magma, each with its own zircon population. Trace element patterns and Hf-isotope compositions between populations and across single zircon grains from a wider selection of samples would be necessary to confirm this supposition.

The Istog sample, however, does represent a metamorphic rock which underwent thorough transformations toward a granite but not a granite that was derived from any magma chamber. According to the zircon data reported here, such strong reworking was accomplished contemporaneously with the intrusion of the Izera batholith. Based on the studied sample it is difficult to tell if and how separately injected magma batches into the magma chamber controlled the growth of zircons that were, probably, in adjacent metamorphic rocks.

The new data show that both the fine-grained gneiss and leucogranite have components inherited from ~640–580 Ma Neoproterozoic crust, similar to that outcropping in Lausitz. The leucogranite sample resulted from anatectic melting of such crust, melting that was also experienced by the fine-grained gneiss and which must have been widespread in deeper levels of the Izera basement. Taking into account some geochemical differences, it is possible that the leucogranites developed independently of the dominant coarse-grained Izera granites.

The close spatial association of leucogranites with the mica schist belts suggests that, during emplacement, these leucogranites utilized rheologically weak zones which were introduced into the Izera pluton in places where large fragments of the country rocks were entrapped by the ~500 Ma granites. Such zones of weaknesses may have channelized later fluids to initiate the more or less ubiquitous albitization and secondary microclinization observed in the leucogranites or leucogneisses. Our data generally confirm the old idea of K. Smulikowski (1958) who predicted the role of tectonics in the spatial distribution of leucogranite bodies in the Izera region.

Geochemistry and relicts of cordierite, garnet and sillimanite suggest an S-type granite as the protolith of the Izera gneisses (Oberc-Dziedzic *et al.*, 2005; Żelaźniewicz *et al.*, 2003). This type of granite is indicative of collisional to post-collisional geotectonic settings, as proposed by Barbarin (1996, 1999) and Sylvester (1998). However, the Izera orthogneisses might also be interpreted as having derived from A-type granites with little or none connection with a collisional event (Oberc-Dziedzic *et al.*, 2005; Pin *et al.*, 2007). The Neoproterozoic Lusatian greywackes, or similar rocks, have been considered the most likely source for the S-type Izera granites (Białek 2003; Oberc-Dziedzic *et al.*, this volume). But we suggest that other metasedimentary rocks as well as older felsic magmatic rocks should also be taken into account when considering the nature of the protolith for the Izera granites.

The new data indicates that caution is needed when discussing the protolith age of the mica schist belts in the eastern part of the Lusatian–Izera Massif. The extrapolation of age data from one belt to another may be invalid. Our data for the Złotniki schists proves that they are older than the ~500 Ma granite. Similarly, the Szklarska Poręba schist belt is older than the Izera granite because it was affected by the metamorphism of this granite's intrusion (Achramowicz & Żelaźniewicz, 1998), an opinion expressed previously by Oberc (1972). The Stara Kamienica schist belt is also apparently older than the Izera granite because one can find metagranite bodies inside the belt (Fig. 1), which have been dated at ~506 Ma (U-Pb zircons; Sylvie Philippe, pers. communication, 1999) and considered as granitic veins derived from the pluton and penetrating its envelope (Żelaźniewicz *et al.*, 2003). An alternative explanation may be that the augen gneisses were later tectonically intercalated into the mica schists (Oberc-Dziedzic *et al.*, 2010). This latter explanation seems less probable, however, because such an insertion would then also apply to the granites and leucogranites that occupy the 'intra-schist' position, too. Thus, tectonic insertion of the Izera granites and leucogranites into the schist belts seems, to us, doubtful.

In general, the age groups of zircons obtained in this study are quite similar to those obtained by Oberc-Dziedzic *et al.* (this volume) for the coarse-grained and fine-grained Izera metagranites, with the most common age clusters of 517–484 Ma and 509–468 Ma, respectively. Xenocrysts of 580–560 Ma were found in the coarse-grained variant and both variants contain zircons ~1.9 Ga old.

The new data, within limits, allows fresh speculation on the existence of ~2.1–1.9 Ga continental terrain which underwent significant accretion during a 2.1–1.9 Ga orogenic event. This event comprised both earlier igneous and late metamorphic transformations, as indicated by the analyzed zircon grains. The ~2.1–1.9 Ga continental fragment was eroded and provided clasts to a sedimentary sequence that underwent metamorphism and anatectic melting that terminated at ~620 Ma. Subsequent exhumation of these anatexites delivered detrital material to greywacke deposits of the Cadomian arc (Kemnitz, 2007). With reference to Gondwana, such characteristics link the Lusatian–Izera basement with the West Africa part of the Gondwana supercontinent (Linnemann *et al.*, 2007).

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