Geological Quarterly, 2009, **53** (3): 255–272

Late Svecofennian sedimentary basins in the crystalline basement of NE Poland and adjacent area of Lithuania: ages, major sources of detritus, and correlations

Ewa KRZEMIŃSKA, Janina WISZNIEWSKA, Grazina SKRIDLAITE and Ian S. WILLIAMS

Krzemińska E., Wiszniewska J., Skridlaite G. and Williams I. S. (2009) — Late Svecofennian sedimentary basins in the crystalline basement of NE Poland and adjacent area of Lithuania: ages, major sources of detritus, and correlations. Geol. Quart., **53** (3): 255–272. Warszawa.

The Precambrian basement of Lithuania and NE Poland, much of which is metasedimentary paragneiss, can be accessed only by cores from deep boreholes. Ion microprobe dating of detrital zircons from samples of these metasediments and the geochemical signatures of the rocks provide new insights into their age and provenance. Detrital zircons from metasediments from Jastrzębna and Mońki have Pb isotopic ages in the ranges 3.14–1.83 and 3.53–1.82 Ga, respectively. Similar results have been obtained from the Lithuanian Lazdijai and Bliudziai paragneisses. About 30% of the analysed Polish detrital zircon cores are of Late Archaean age (2.90–2.60 Ga) and about 70% are Palaeoproterozoic (2.10–1.90 Ga), similar to the age distributions of detrital zircon from Svecofennian metasediments exposed in Central Sweden and Southern Finland. The youngest detrital zircon sub-groups indicate maximum deposition ages of about 1.86 Ga, similar to the ages of exposed Svecofennian sedimentary basins. Possible source rocks of comparable ages and affinities can be found within Fennoscandia, Greenland and Sarmatia.

Ewa Krzemińska, Janina Wiszniewska, Polish Geological Institute-National Research Institute, Rakowiecka 4, PL-00-975 Warszawa, Poland; e-mails: ewa.krzemińska@pgi.gov.pl, janina.wiszniewska@pgi.gov.pl; Grazina Skridlaite, Institute of Geology and Geography, T. Sevcenkos 13, LT-03223 Vilnius, Lithuania; e-mail: skridlaite@geo.lt; Ian S. Williams, Research School of Earth Sciences, The Australian National University, Canberra ACT 0200, Australia; e-mail: ian.williams@anu.edu.au *(received: July 31, 2008; accepted: August 17, 2009).*

Key words: Fennoscandia, Sarmatia, Greenland, Svecofennian basin, detrital zircon, geochronology.

INTRODUCTION

Unlike the well-exposed Precambrian bedrock of Finland and Sweden, the Fennoscandian basement rocks of the East European Craton (EEC) in NE Poland, Lithuania and Belarus are completely covered by Phanerozoic sediments that are up to 5 km thick in the west. The main structures in the buried basement have been delineated, however, primarily by geophysics (Bogdanova *et al*., 2006). Imaging based on the EUROBRIDGE' 95 seismic profile (Eurobridge'95 Working Group, 2001) and magnetic and gravimetric mapping (Wybraniec, 1999) shows clear structural continuities between the deep crustal domains of Lithuania and Poland. Moreover, Nd depleted mantle model ages for basement metasediments from NE Poland and Lithuania (2.09–1.99 Ga) are similar to those obtained from Svecofennian rocks in the Baltic Shield (Claesson and Ryka, 1999). It appears that a major phase of crust formation in the Polish and Lithuanian parts of the EEC was related to the Svecofennian Orogeny (Claesson *et al*., 2001; Skridlaite and Motuza, 2001; Krzemińska *et al*., 2005, 2007; Wiszniewska *et al*., 2007; Krzemińska and Wiszniewska, 2007).

The host rocks to the 1.83–1.80 Ga igneous suites in the basement of central NE Poland range from low-grade metasediments to paragneisses (Krzemińska *et al*., 2005, 2007; Wiszniewska *et al*., 2007). They have been intersected in several drill holes, for example at Mońki, Jastrzębna, Wigry and Czyże (Fig. 1A, B).

Sediments deposited in a basin preserve a record of tectonic processes and the geological history of their source terrains. Using a combination of petrographic and geochemical studies, together with geochronological data, it is possible to infer not only the nature of the source region of sediments, but also the principal ages of the source rocks and of pre- and post-depositional thermal events. Zircon is most widely used

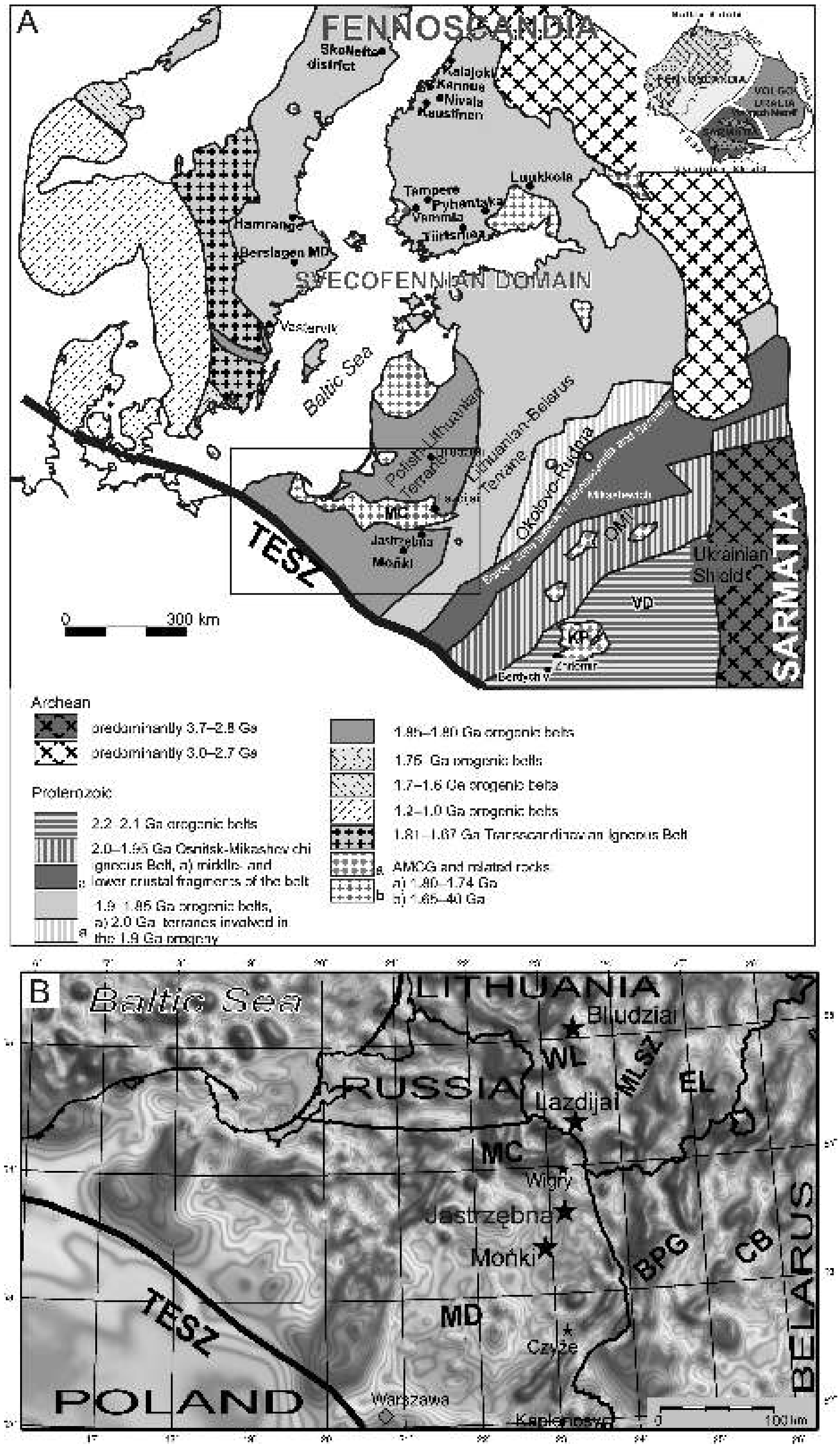

Fig. 1. General location map showing the crystalline basement of NE Poland and adjacent area of Lithuania

A — major tectonic subdivisions and age of the crust within the western part of the East European Craton (modified after Bogdanova *et al*., 2006) showing the location of the key drill holes as well as previous detrital zircon studies of Svecofennian metasediments. Some structural units are marked by the letters: KP — Korosten Pluton, MC — Mazury Complex, OMI — Osnitsk–Mikashevichi Igneous Belt, TESZ — Trans-European Suture Zone, VD — Volyn domain; **B** — magnetic anomaly map of the crystalline basement of NE Poland and adjoining areas of Lithuania and Belarus (Wybraniec, 1999), showing the area of the Mazowsze domain (MD) and the location of the sampling sites: Mońki and Jastrzębna and Bliudziai and Lazdijai boreholes. The locations of the deep drill holes at Czyże, Wigry and Kaplonosy are also shown; major structural units are marked by: BPG — Belarus–Podlasie Granulite Belt, CB — Central Belarus Belt, EL — East Lithuanian domain, WL — West Lithuanian Granulite domain, MC — Mazury Complex, MD — Mazowsze domain, MLSZ — Middle Lithuanian Suture Zone for dating detritus, primarily because it crystallises from most silica-saturated magmas, is very resistant both chemically and physically to most crustal geological processes, and is isotopically robust. U-Pb dating of detrital zircon grains from a sediment is the best method for determining the ages of the silica-saturated igneous rocks and metamorphic events in its source region, thus helping to identify terrains from which the sediment might have been derived. This in turn can be used to infer tectonic processes and to make regional palaeogeographic reconstructions. Such studies of sediment provenance, however, require the accurate dating of a large number of detrital zircon grains (Dodson *et al*., 1988; Vermeesh, 2004) and a careful distinction between gneous and metamorphic zircon growth.

The presence of detrital zircon has been noted in several geochronological studies of the basement rocks in Latvia, Lithuania and Belarus (e.g., Mansfeld, 2001; Claesson *et al*., 2001), but the basic characteristics of terrigenous sediments (e.g., the age spectrum of the detrital material and the maximum age of deposition) that can be obtained from a large number of zircon age measurements have rarely been determined on the unexposed basement rocks of Fennoscandia. Metasediments from only two drill holes, at Mońki and Jastrzębna in NE Poland, have been well documented by both whole-rock geochemistry and detrital zircon geochronology (Williams *et al*., 2009).

Here we present the first regional review of all current results from detrital zircon U-Pb studies in NE Poland and adjacent areas of Lithuania, and compare these data with those from Palaeoproterozoic structural units in the Fennoscandian, Laurentian (Greenland) and Sarmatian megablocks.

GEOLOGICAL CONTEXT

The Fennoscandian crustal segment consists of several domains: the Archaean, Svecofennian and south-west Scandinavian domains, and the Transscandinavian Igneous Belt (Fig. 1A). The age of these domains decreases from the north-east to the south-west. Archaean (3.2–2.5 Ga) granitoid-gneiss complexes are preserved in the Karelian Province (Sorjonen-Ward and Lukkonen, 2005). The oldest known Palaeoarchaean rocks are the *ca*. 3.4 Ga gneisses (Sm-Nd isochron age) from the Vodlozero Block, SE Karelian Province (Puchtel *et al*., 1991) and trondhjemite gneisses up to *ca*. 3.5 Ga (U-Pb zircon) from the Pudasjärvi Granulite Belt, NE Finland (Mutanen and Huhma, 2003).

The younger part of the Baltic Shield, the Svecofennian domain, was formed during the multiphase Svecofennian Orogeny (1.92–1.79 Ga), probably by progressive accretion of several island arcs (terranes) and continent–continent collision (Lahtinen *et al*., 2005, 2008; Korja *et al*., 2006). It was preceded by rifting of the Archaean Karelian protocraton and formation of oceanic crust (Korja *et al*., 2006) and large marginal basins (Rutland *et al*., 2004; Williams *et al*., 2008). The Svecofennian domain forms a large area of Palaeoproterozoic crust within Fennoscandia (10^6 km^2), and extends southwards as far as the Trans European Suture Zone (TESZ) beneath the Phanerozoic sedimentary cover in Lithuania and Poland (Williams *et al*., 2009; Fig. 1A).

The domain is dominated by granitoids and subordinate volcanic rocks that were intruded over a relatively short period of time (*ca*. 1.90–1.87 Ga: Huhma, 1986; Gaal and Gorbatschev, 1987; Gorbatschev and Bogdanova, 1993) into metasediments deposited only shortly beforehand (Lahtinen 1996; Claesson *et al*., 2001; Rutland *et al*., 2004; Williams *et al*., 2008). The bulk of the detritus was derived from Palaeoproterozoic (2.1–1.9 Ga), and to a lesser extent Archaean (3.0–2.6 Ga), sources (Claesson *et al*., 1993; Lahtinen *et al*., 2002; Rutland *et al*., 2004; Sultan *et al*., 2005; Bergman *et al*., 2008; Williams *et al*., 2008).

The Svecofennian domain consists of northern and southern volcanic belts and a central marine basin, the Bothnian Basin (e.g., Gaal and Gorbatschev, 1987; Claesson *et al*., 1993). These regions have been distinguished as the northern, central and southern subprovinces (e.g., Lahtinen *et al*., 2002). The northern subprovince is composed of 1.95–1.85 Ga volcanic and plutonic rocks (Lahtinen *et al*., 2005; Korja *et al*., 2006), plus metasedimentary rocks. The central subprovince consists mainly of metagreywacke and metapelite of the Bothnian Basin (*ca*. 10 km thick) with intercalations of mafic volcanic rocks (e.g., Nironen, 1997). The southern subprovince is made up of metapelite, quartzite, migmatite and S-type granite, associated with quartz-feldspar gneisses, carbonate rocks and metavolcanics (Lahtinen *et al*., 2002). The volcanic rocks of the Southern Subprovince have ages of 1.90–1.88 Ga; the granitoids are slightly younger (1.87–1.85 Ga; Lahtinen *et al*., 2005 and references therein). Both the Central and Southern Subprovinces are assumed to be the remnants of island arcs. The rocks of the whole Svecofennian domain were subject to late low-P, high-T metamorphism that peaked at *ca*. 1.88 Ga in Finland and 1.85–1.75 Ga in Sweden (Mouri *et al*., 1999; Lahtinen *et al*., 2002; Rutland *et al*., 2004; Skiöld and Rutland, 2006; Högdahl *et al*., 2008).

The Svecofennian Orogeny (1.92–1.79 Ga) has been divided into the Lapland-Savo, Fennian Svecobaltic and Nordic crustal orogenic stages (Lahtinen *et al*., 2005; Korja *el al*., 2006; Lahtinen *et al*., 2008). Well preserved and exposed Svecofennian sequences such as in the Tampere Schist Belt (TSB) and Vammala Migmatite Belt (VMB) in Finland, and in the Bergslagen mining district (BMD), Skellefte district (SD) and Västervik Basin (VB) in Sweden, belong to the Bothnian Basin. They are key areas for studying the evolution of the Svecofennian Orogen (e.g., Kähkönen and Leveinen, 1994; Lahtinen, 2000; Lahtinen *et al*., 2002; Rutland *et al*., 2004; Sultan *et al*., 2005; Bergman *et al*., 2008). During that protracted orogeny, the Bothnian Basin or several similar basins extended from the area of present-day Central Sweden south-east into Southern Finland, and to the south into the Baltic–Belarus region (Lithuania, NE Poland and Belarus), where the Precambrian basement is now covered by a thick layer of Phanerozoic sediments (Claesson *et al*., 1993).

The principal basement crustal domains in Lithuania, Poland and Belarus have all been defined using magnetic and gravimetric mapping (Fig. 1A, B) and geophysics (e.g., Wybraniec, 1999; Eurobridge'95 Working Group, 2001;

Bogdanova, 2005; Bogdanova *et al*., 2006). The southeastern margin of Fennoscandia and a major lithospheric discontinuity that separates Fennoscandia from Sarmatia (e.g., the Mińsk Fault) have been recognized in a few tectonically discrete complexes. The oldest one, close to the Central Belarus Suture Zone (CBsz), is the Okolovo subterrane, which contains komatiitic and tholeiitic rocks of oceanic island-arc affinity. Intercalations of metasediments include black shales and ferruginous volcanogenic deposits, including juvenile metavolcanics dated at *ca*. 1.98 Ga (Bibikova *et al*., 1995).

Strong linear NNE-trending magnetic and gravimetric anomalies to the west of the Okolovo subterrane and CBsz delineate the Belarus–Podlasie Granulite Belt (BPG; Taran and Bogdanova, 2003). This belt consists mostly of Palaeoproterozoic (1.90–1.87 Ga) meta-igneous, mature-arc-related granulites (Bibikova *et al*., 1995; Skridlaite and Motuza, 2001; Bogdanova *et al*., 2006) and subordinate high grade metasediments (Bogdanova *et al*., 2001; Taran and Bogdanova, 2003). Crust of similar age, but formed in a back-arc palaeoenvironment and metamorphosed to amphibolite facies, comprises the adjacent East Lithuanian (EL) domain. These two domains form the Lithuanian–Belarus terrane (1.90–1.85 Ga).

Further west again, younger (1.85–1.80 Ga) crust with a different gravity and magnetic signature and thickness forms the Late Palaeoproterozoic Polish-Lithuanian terrane (Bogdanova *et al*., 2006). The main component of this terrane is the West Lithuanian Granulite domain (WL), consisting of high-grade, mostly marine pelites and intermediate to felsic and calc-alkaline island-arc igneous rocks (Skridlaite and Motuza, 2001; Motuza, 2005). Separating the Lithuanian-Belarus and Polish-Lithuanian terranes is a transition border zone, the Middle Lithuanian Suture Zone, which has been interpreted as the product of Late Palaeoproterozoic subduction processes and final amalgamation of the two terranes. Mafic to felsic metavolcanic rocks of arc affinity in the MLSZ have ages of 1.86–1.84 Ga (Motuza, 2005; Motuza *et al*., 2006).

At the southern extension of the WL and/or MLSZ lies the Mazowsze domain (MD). This area in Poland was initially defined as the Mazovian or Mazowsze Archaean massif (Depciuch *et al*., 1975; Kubicki and Ryka, 1982; Ryka, 1984), but following recognition that it consists predominantly of rocks of metasedimentary origin intruded by 1.83–1.80 Ga magmatic arc-related granitoids (Wiszniewska *et al*., 2006; Krzemińska *et al*., 2007), the domain is now interpreted as representing a Late Palaeoproterozoic arc palaeoenvironment (Bogdanova, 2005; Bogdanova *et al*., 2006; Krzemińska and Wiszniewska, 2007).

METASEDIMENTS IN THE LITHUANIAN BASEMENT

Metasedimentary and metaigneous rocks recovered from several drill holes in the western, central and southern WL and southern Middle Lithuanian Suture Zone (MLSZ) have been described by Skridlaite and Motuza (2001). The predominant supracrustal rocks in the WL are metapelitic and felsic (metapsammitic) gneisses. The protoliths of the EL metasediments were mostly greywacke and arkose, in places with an admixture of volcaniclastic material. They have been metamorphosed to biotite–plagioclase–quartz gneiss, in places with garnet and sillimanite (Motuza, 2008).

Metasediments from the central WL (Bliudziai drill hole Bl-150, Fig. 1) contain large amounts of volcanic material and alternate with metavolcanic and volcaniclastic rocks. The Bl-150 rocks include peraluminous gneiss (metapelite) with lenses and veins of anatectic cordierite-bearing granite, alternating with meta-andesitic porphyritic hypersthene and biotite gneiss. The metapelite contains quartz, plagioclase, K-feldspar, biotite, garnet, cordierite, sillimanite, zircon and opaque minerals. Restitic (melanocratic) areas are surrounded by lenses and irregularly shaped patches that are remnants of cordierite- and K-feldspar-bearing granitic partial melts. The hypersthene biotite gneiss consists of porphyritic plagioclase in a fine-grained matrix of hypersthene, biotite and plagioclase, with minor quartz, opaque minerals and apatite. All have been metamorphosed under upper amphibolite to lower granulite facies conditions. Numerous P-T estimates from the Bl-150 metapelite provide evidence for a stepwise P-T path from *ca*. 650°C at 600 MPa down to *ca*. 450°C at 300 MPa, indicating tectonic exhumation.

Similar metasedimentary and metaigneous complexes have been identified in Southern Lithuania, along the MLSZ at Lazdijai (drill holes Lz-13 and 32, Fig. 1). The basement core from Lz-13 consists of banded rocks ranging from marble and chlorite- and biotite-schist (i.e. metasediment) to metavolcanic biotite gneiss and clinopyroxene-bearing amphibolite. The gneisse have a wide range of compositions from fine-grained biotite gneiss to garnet-and-staurolite bearing gneiss. Minor marble layers are former limestones, containing hornfels strips at the boundaries with silicic rocks, mainly amphibolite. The sediments have been metamorphosed at lower- to upper-amphibolite facies. Most have undergone ductile deformation, transforming some rocks into mylonite and blastomylonite that have later been reworked under more brittle conditions. The schists consist of quartz and mica (commonly chlorite), with fine garnet and biotite. They are enriched in opaque minerals, occasionally magnetite, haematite and covellite intergrowths. The last might originate from submarine hydrothermal vents associated with volcanic activity. The complex has been intruded by a *ca*. 1.5 Ga anorthosite–mangerite–charnockite–granite (AMCG) suite (Skridlaite *et al*., 2003*a*).

Basement rocks from the Lz-32 drill hole consist of fine- to medium-grained anatectic granitoids and coarse-grained I-type granites interspersed with melanocratic strips and lenses of garnet-biotite bearing gneisses or amphibolites. The gneisses, which consist of biotite, garnet, plagioclase, quartz, K-feldspar, opaque minerals and zircon, are metamorphosed sedimentary rocks (Fig. 2A, B). The amphibolites, alternating with pyroxene gneisses, are interpreted as metamorphosed dacites, andesites and basalts (Skridlaite and Motuza, 2001). Veins and lenses of homogeneous medium-grained garnet-bearing granite that formed as a result of anatectic melting of the gneisses are also present.

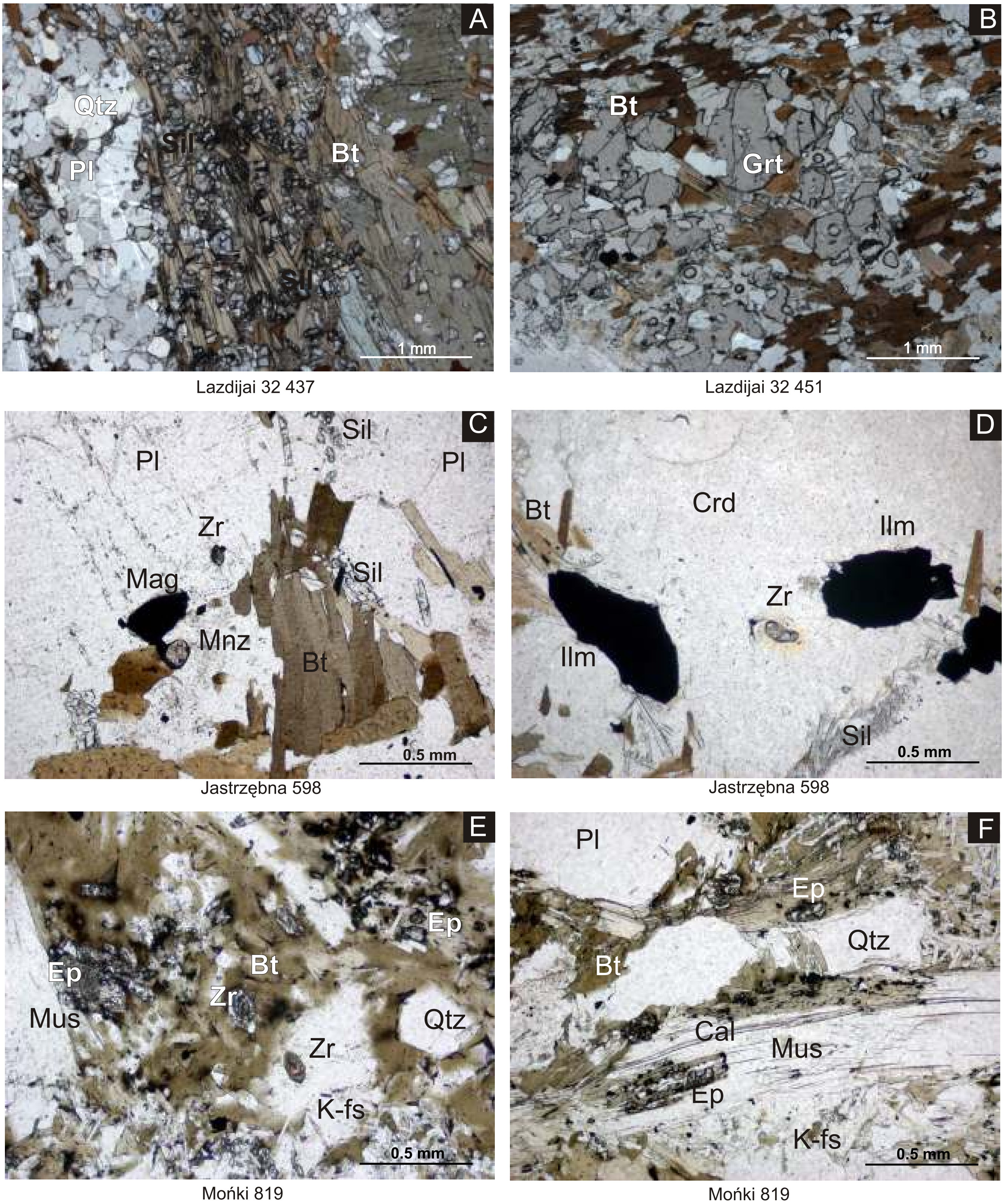

Fig. 2. Photomicrographs of paragneisses from drill holes at: A–B — Lazdijai 32, depth 437 and 451 m; C–D — Mońki, depth 819 m; E–F — Jastrzębna depth 598 m

Qtz — quartz, Pl — plagioclase, K-fs — K-feldspar, Mus — muscovite, Bt — biotite, Ep — epidote, Sil — sillimanite, Crd — cordierite, Grt — garnet, Zr — zircon, Mnz — monazite, Ilm — ilmenite, Mag — magnetite, Cal — calcite

METASEDIMENTS FROM THE BASEMENT OF CENTRAL NE POLAND

Metasedimentary rocks received a little attention in the early documentation of drill core from the Polish basement. The most important mentions were those of metasediment in drill cores from Jastrzębna and Mońki (Fig. 1B) in the Mazowsze domain. According to Ryka (1976), the upper part of the basement core from the Jastrzębna 1 drill hole (depth 512–570 m) was dominated by pegmatites and amphibolites, but from 570 to 717 m the main rock types were intercalated sillimanite-cordierite gneiss (*ca*. 70 m) and biotite gneiss (*ca*. 60 m). The biotite and sillimanite-cordierite gneisses have low levels of Cr, Ni, V and Co, but high Ba and Sr, consistent with a sedimentary protolith, possibly shale or mudstone (Dziedzic, 1976).

Williams *et al*. (2009) included a sample from the Jastrzębna drill hole in their study of the geochemistry and detrital zircon from Polish basement paragneisses. The selected sample, J-598 (depth interval 589.0–598.7 m), was typical of the main homogeneous, but partially migmatised sequence of cordierite-sillimanite gneiss. The mineral assemblage included plagioclase (An30-42), K-feldspar, quartz, biotite, cordierite and sillimanite (Fig. 2C, D), and a high abundance of ilmenite-magnetite, which probably accounts for the presence of a local magnetic anomaly. The ratio of plagioclase to K-feldspar was about 3:1. Biotite Fe-Mg-Ti geothermometry (Henry *et al*., 2005) indicated metamorphic temperatures of 540 to 625°C, reflecting metamorphism at amphibolite facies.

The upper part of Mońki IG 2 drill hole, (depth 626 to 745 m) traversed metasediments ranging from quartzite to quartz-schist. From 745 to 1539 m it intersected monotonous grey gneisse alternating with a few veins of pegmatite and some intermediate mafic metavolcanic rocks. The age and stratigraphic relationships of the Mońki rocks have not been closely investigated. The quartzite has been interpreted as Jotnian metasediment (Kubicki and Ryka, 1996) and the gneiss as an enigmatic terrigenous sequence (Wołkowicz, 1996). Metamorphic temperatures estimated using biotite geothermometry (Henry *et al*., 2005) ranged from greenschist facies at 475–520°C (819 m depth) to amphibolite facies at 595–620°C at deeper levels (1240–1460 m).

The Mońki sample analysed by Williams *et al*. (2009), M-819, was been taken from the upper part (depth 819 m) of the grey gneiss sequence. It was more quartz rich, and had less biotite relative to muscovite than the lower gneiss sequence. The main minerals in decreasing abundance were quartz, K-feldspar, plagioclase, biotite, muscovite and epidote, with accessory apatite, zircon, opaque oxides and chlorite (Fig. 2E, F). Laminae of biotite, muscovite and epidote were intercalated with laminae dominated by quartz and feldspars on a thin-section scale. The rare plagioclase was almost completely sericitised.

Sillimanite- and garnet-biotite-sillimanite gneiss sequences up to 300 m thick have also been intersected by drill holes at Czyże, Wigry and Sokółka within the MD and at Kaplonosy, within the BPG (Fig. 1). These gneisses also have sedimentary protoliths.

The metasediments from all drill holes are rather similar to one another in major element composition. Plotted on the classification diagram of Herron (1988), the compositions of the Jastrzębna metasediments analysed by Williams *et al*. (2009) straddle the boundary between the shale and Fe-shale fields (Fig. 3A). The Mońki gneiss compositions, with slightly lower Fe, lie mostly within the shale field. The same SiO_2/Al_2O_3 and Fe_2O_3/K_2O relationships are characteristic of the Czyże, Wigry, and Kaplonosy paragneisses (Ryka, 1989, 1996; Jackowicz, 2000). They are probably metamorphosed Fe-shale, shale and greywacke. Similarly, various gneisses in mostly the lower sedimentary units from both the central and southern subprovinces of the Svecofennian domain in Sweden and Finland (Lahtinen *et al*., 2002) also have the major element compositions of shale and greywacke.

The major element contents of the metasediments, when used to calculate functions F1–F2 (Roser and Korsch, 1988), discriminate four main provenance groups. The metasediments from the Polish and Lithuanian drill holes plot in the felsic P3 field (Fig. 3B), but they scatter into the intermediate P2 and mafic P1 fields, in which nearly all the immature and more mature equivalents from the exposed Svecofennian (central and southern subprovinces) lie. The trend from P1 (Jastrzębna samples) to P3 (Mońki samples) suggests a shift towards a more felsic bulk composition of the source materials. This trend continues in the samples from the exposed Svecofennian, with individual samples falling within fields P1, P2 and mostly P3 (Fig. 3B).

ANALYTICAL METHODS FOR DETRITAL ZIRCON GEOCHRONOLOGY

The zircon grains were studied in thin section using an electron microscope fitted with back-scattered electron (BSE) and cathodoluminescence (CL) detectors before ion microprobe analysis. All samples were also examined by the *LEO* electron microprobe and by *CL VIS-View 900* instrument at the Polish Geological Institute-National Research Institute. The BSE and CL digital images were used to help diagnose the sedimentary origin of the zircon and, from the zoning textures, the possible rock types in the region from which the protolith sediments were derived.

Studies of detrital zircon age populations require the dating of large numbers of grains. If a suite of analyses is to have a 95% chance of sampling all detrital components with a relative abundance of 5% or more, then it is necessary to date at least 60 (Dodson *et al*., 1988; Fedo *et al*., 2003) and possibly 120 (Vermeesh, 2004) randomly selected crystals. Data sets with less analyses risk missing minor components and misrepresenting the relative abundances of even the major components (Andersen, 2005).

Zircons from the two Polish paragneiss samples (Mońki and Jastrzębna) were dated by U-Th-Pb using the *SHRIMP II* at the Australian National University (Williams *et al*., 2009). U-Pb compositions of detrital zircons from the Lithuanian metasediments (drill cores Bl-150 and Lz-32) were measured using the *NORDSIM Cameca ims1270* at the Swedish Museum

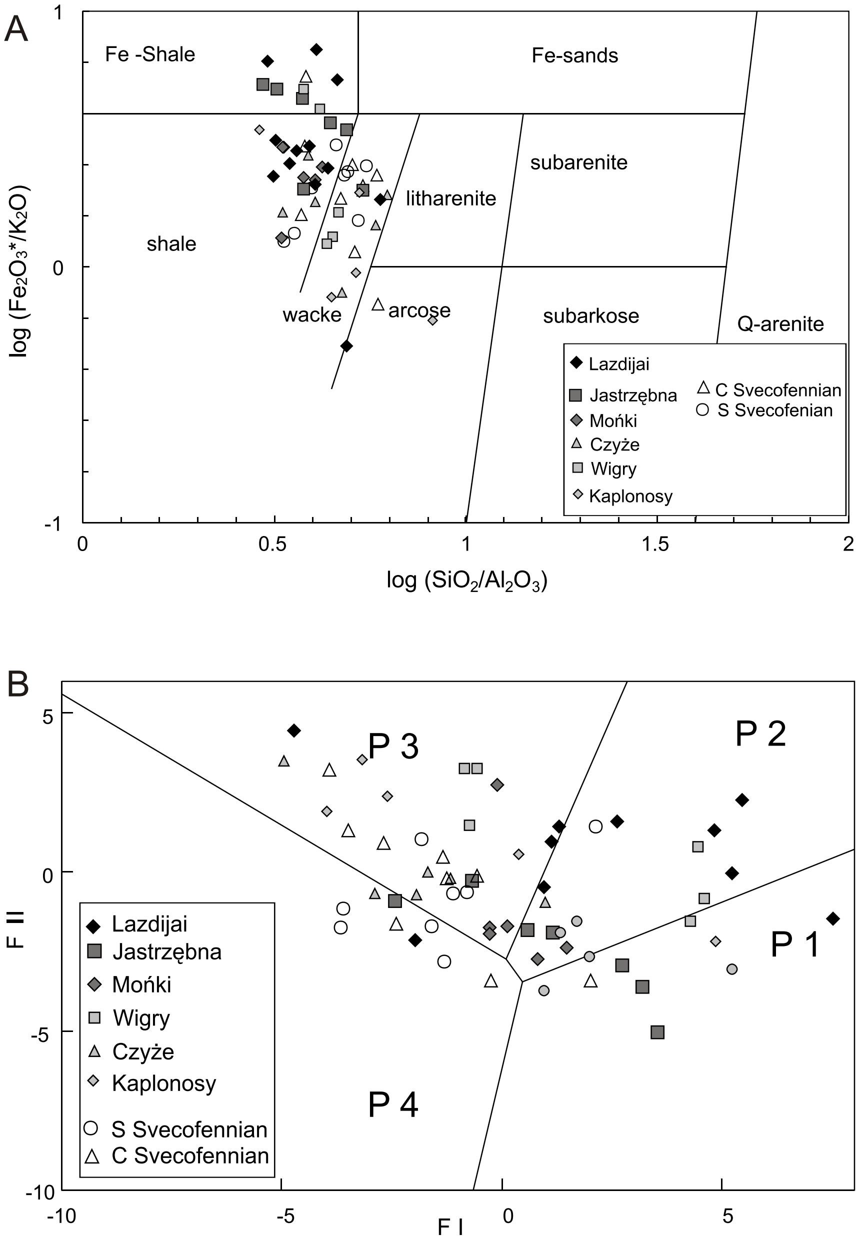

Fig. 3. Comparisons of Polish and Lithuanian paragneisses by A — classification diagram (Herron, 1988) and B — provenance discrimination diagram (Roser and Korsch, 1988) for fields of rocks: P1 — mafic, P2 — intermediate, P3 — felsic, P4 — quartzose

Analytical data from drill holes: Jastrzębna, Mońki (Williams *et al*., 2009), Wigry (Jackowicz, 2000), Czyże (Ryka, 1996), Kaplonosy (Ryka, 1989), Lazdijai and Bliudziai (Skridlaite, unpubl. data) and for SS — south Svecofennian metasediments and CS — central Svecofennian metasediments after Lahtinen *et al*. (2002)

of Natural History, Stockholm. Analytical procedures closely followed those described by Whitehouse *et al*., (1999). Pb/U ratios were calibrated against zircon standard 91500 (1065 Ma: Wiedenbeck *et al*., 1995). Common Pb was corrected using the present day Pb composition calculated using the model of Stacey and Kramers (1975), assuming that the Pb was laboratory-derived surface contamination. Age calculation methods followed the recommendations of Ludwig (1991) with uncertainties given at the 95% confidence level.

ZIRCON TYPES AND MORPHOLOGY

Zircon grains from Bl-150 were either included in biotite or plagioclase. They ranged in shape, but with two dominant types. The first type was short (<150 μm diameter) and prismatic, the second was elongated (up to 250 μm diameter) with rounded terminations (Fig. 4A). Both types showed oscillatory zoning in CL images, indicating mainly an igneous origin. Some grains had distinct cores and overgrowths.

Zircon grains from Lz-32 occurred mainly as inclusions in biotite. The elongated, rounded, oval-shaped zircons ranged in size from 100 to 300 μm diameter. In some grains, an oscillatory zoned core was surrounded by a distinct broad rim (Fig. 4B).

The zircon grains from the Mońki drill core were described by Williams *et al*. (2009) as mostly large (100–200 μm diameter), pale to dark brown, finely fractured, clear to turbid, subhedral prismatic crystals. Every zircon grain consisted of a large core surrounded by a very weakly luminescent and relatively thin overgrowth (Fig. 4C). The zoning in the cores ranged from simple oscillatory zoning to sector zoning or banded zoning. Some cores showed evidence of extensive recrystallisation; in others there was no zoning at all.

Zircons from the Jastrzębna sample were described as medium to large (80–200 μm diameter), pale grey-brown, mostly stubby subhedral to euhedral grains with fractured, inclusion-rich cores and clear, inclusion-free overgrowths. The cores ranged from euhedral crystals to crystal fragments and clearly rounded grains (Fig. 4D). Zoning in the cores was not as varied as in the Mońki sample, but ranged from simple oscillatory zoning to chaotic and unzoned. The overgrowths were thick, probably due to a higher grade of metamorphism.

U-Pb DETRITAL ZIRCON GEOCHRONOLOGY

Fifteen zircon grains (14 cores and 5 overgrowths) were analysed from Bliudziai (Bl-150) depth 1510 m and six grains (6 cores and 1 overgrowth) from Lazdijai (Lz-32) depth 475 m. Several core age groups can be distinguished, ranging from

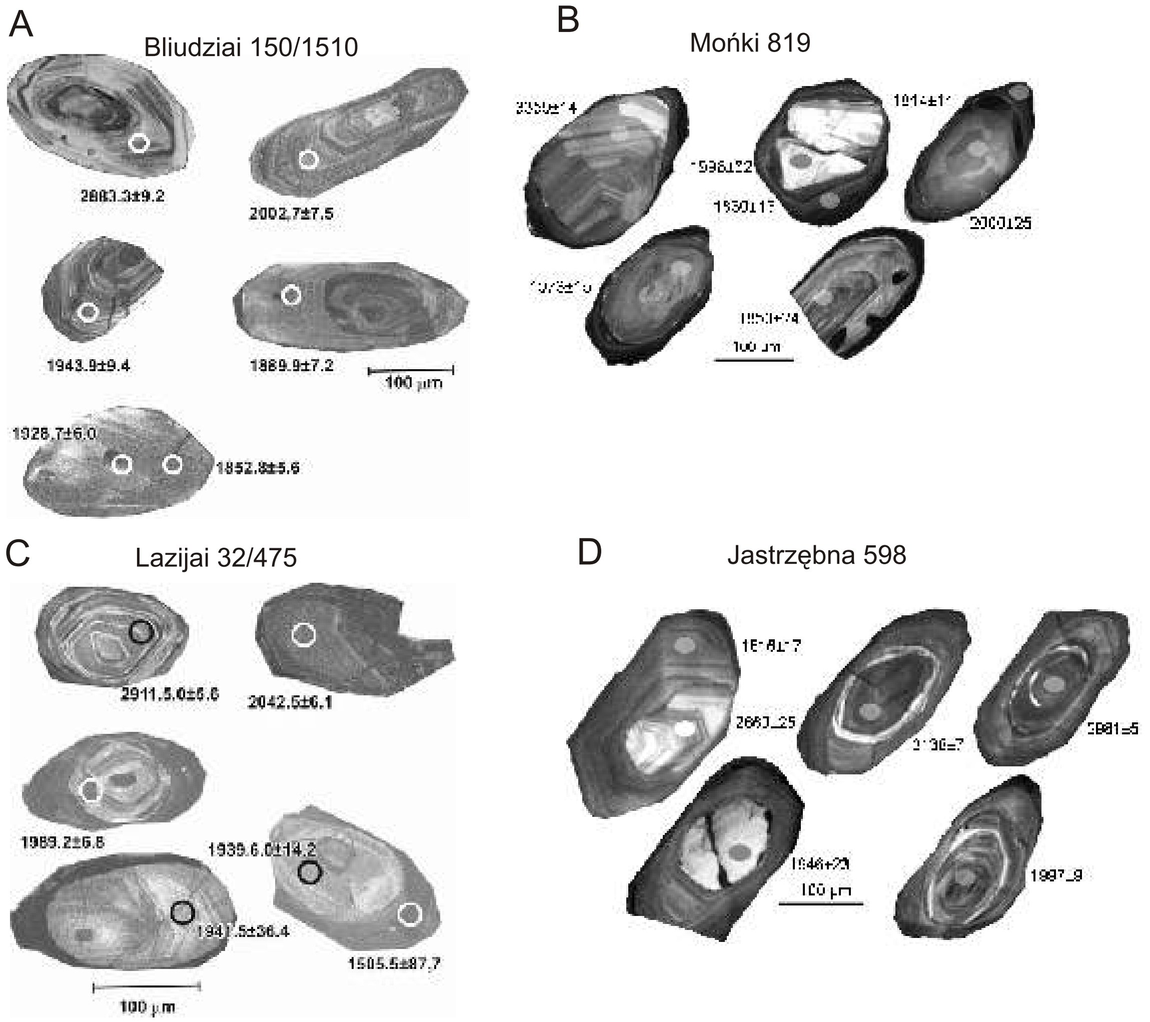

Fig. 4. Cathodoluminescence images of zircon grains selected from paragneisses

A — Bliudziai 150 depth 1510 m; B — Lazdijai 32 depth 475 m; C — Mońki depth 819 m; D — Jastrzębna depth 598 m; circles show the approximate location and size of *SHRIMP* and *NORDSIM* analytical spots respectively; ages in Ma and uncertainties 1σ

ca. 3.0 to *ca*. 1.85 Ga (Figs. 5 and 6), consistent with the rocks being of sedimentary origin.

The two oldest oscillatory-zoned cores gave $^{207}Pb/^{206}Pb$ ages of *ca*. 3.08 and *ca*. 2.88 Ga (Fig. 5). Several zircon grains gave ages of *ca*. 2.00, 1.94 and 1.90–1.89 Ga. One grain had a 2.00 Ga core and 1.89 Ga overgrowth, another a 1.94 Ga core and 1.85 Ga overgrowth.

The six rounded zircons from the Lz-32 anatectite yielded ages in the range *ca*. 2.90–1.50 Ga. One grain with pronounced oscillatory zoning gave 2.91 Ga, and a group of weakly zoned grains gave ages of *ca*. 2.00 and 1.94 Ga. One 1.94 Ga core was surrounded by a wide, weakly luminescent *ca*. 1.50 Ga overgrowth. Discordance in the Bl-150 and Lz-32 zircons (Fig. 5) reflects radiogenic Pb loss.

Many more zircon grains were dated from the Mońki and Jastrzębna samples (Williams *et al*., 2009). Consistent with the diverse sources of the Mońki rocks, the 54 zircon cores analysed had a wide range of U-Th-Pb isotopic compositions. Most analyses were concordant or nearly so, but the Pb isotopic ages ranged from 3.53 to 1.82 Ga (Fig. 5). However, the distribution of ages was not uniform, with most ages in the range 2.1–1.9 Ga, and minor clusters at 2.7 and *ca*. 1.85 Ga.

Similarly, 49 zircon cores analysed from Jastrzębna 598 also had a wide range of isotopic compositions (Williams *et al*., 2009). With a few exceptions, the U-Pb ages were concordant or nearly so, but the $^{207}Pb/^{206}Pb$ apparent ages ranged from *ca*. 3.14 to 1.83 Ga (Fig. 5). Most of the analyses (60%) had $^{207}Pb/^{206}Pb$ ages in the range 2.1–1.9 Ga. There was a much smaller cluster of ages in the range 2.8–2.6 Ga.

When the $^{207}Pb/^{206}Pb$ ages of the Polish and Lithuanian samples are plotted as relative probability histograms (Fig. 6), it is evident that the large cluster at 2.1–1.9 Ga is composed of several subgroups. The most objective way to estimate the ages of those subgroups is by mixture modelling (Sambridge and Compston, 1994), a procedure whereby groups of zircons with similar ages can be recognized within a mixed zircon population, and the mean age of, and fraction of the population in, each group estimated.

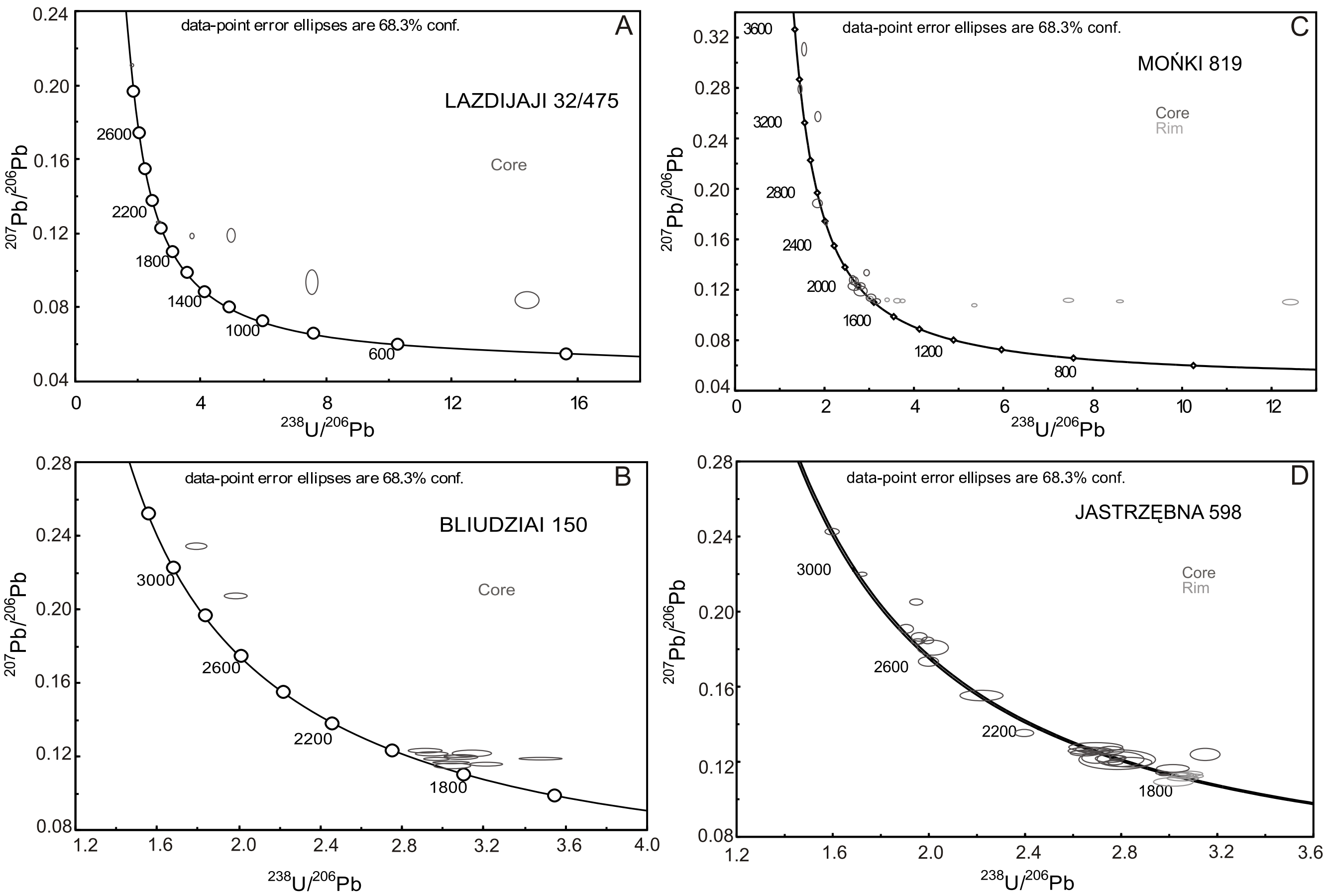

Fig. 5. Concordia diagrams for detrital zircons from samples: A — Lazdijai 32, B — Bliudziai 150, C — Mońki 819, D — Jastrzębna 598

The U/Pb isotopic data from Mońki and Jastrzębna are given in the table published by Williams *et al*. (2009); the isotopic ratios are plotted with 1σ uncertainties

Three main age groups accounted for about 85% of the analyses in the 2.1–1.9 Ga cluster from the Mońki sample; 2050 ±15, 1990 ±25 and 1965 ±15 Ma. There was a minor group at 1915 ±25. A cluster of 8 analyses at *ca*. 1.85 Ga possibly consisted of two age groups, 1870 ±20 and 1830 ±25 Ma. Of the 8 analyses in the *ca*. 2.7 Ga group, only 4 gave the same age within analytical uncertainty 2715 ±20 Ma (Williams *et al*., 2009).

Mixture modelling of the Jastrzębna zircon analyses identified three possible age groups; 2035 ±12, 1986 ±6 and 1927 ±15 Ma. There was a minor group at *ca*. 1.86 Ga. Of 10 analyses in the older group, five had a similar ^{207}Pb/^{206}Pb age within analytical uncertainty, 2690 ±12 Ma. There were no detrital grains aged between 2.2 and 2.4 Ga (Williams *et al*., 2009).

There are insufficient analyses from the Lithuanian samples to apply mixture modelling, but many of the zircon grains have apparent ages similar to those from the Polish samples.

MAXIMUM AGE OF DEPOSITION

The deposition ages of sedimentary rocks can be constrained using the ages of interlayered volcanic or crosscutting igneous rocks. Maximum deposition ages can also be constrained using the ages of the youngest detrital zircon. Sediments from active orogens commonly contain a zircon population contributed by contemporaneous volcanism, the age of which is effectively the deposition age. Care must be taken, however, zircon is well known to be susceptible to radiogenic Pb loss, reducing its Pb/U apparent age, and in metamorphic rocks it is vital to distinguish between detrital and post-depositional metamorphic overgrowths (Mezger and Krogstad, 1997; Whitehouse *et al*., 1999; Rutland *et al*., 2004; Williams *et al*., 2008).

Records of widespread Late Palaeoproterozoic igneous activity in Eastern Europe are preserved both in basement drill cores from Lithuania and Poland, and in igneous rocks exposed at the surface. For example, igneous rocks from Kuršiai-65, within the WL, and Graužai-105 (1844 ±5 and 1837 ±6 Ma, respectively: Motuza *et al*., 2006), granodiorite from Žeimiai-347, MLSZ (*ca*. 1.84 Ga: Rimsa *et al*., 2001), meta-andesite from Virbaliskis (1842 ±6 Ma: Motuza *et al*., 2006), the Mońki metavolcanics (1836 ±8 Ma: our unpubl. data), the Jastrzębna pegmatite and Bargłów metavolcanics (1826 ±12 and 1835 ±28 Ma, respectively: Krzemińska *et al*., 2006), plus arc-related granites from Rajsk and Pietkowo (1826 ±6 and 1818 ±15 Ma, respectively: Krzemińska *et al*., 2007).

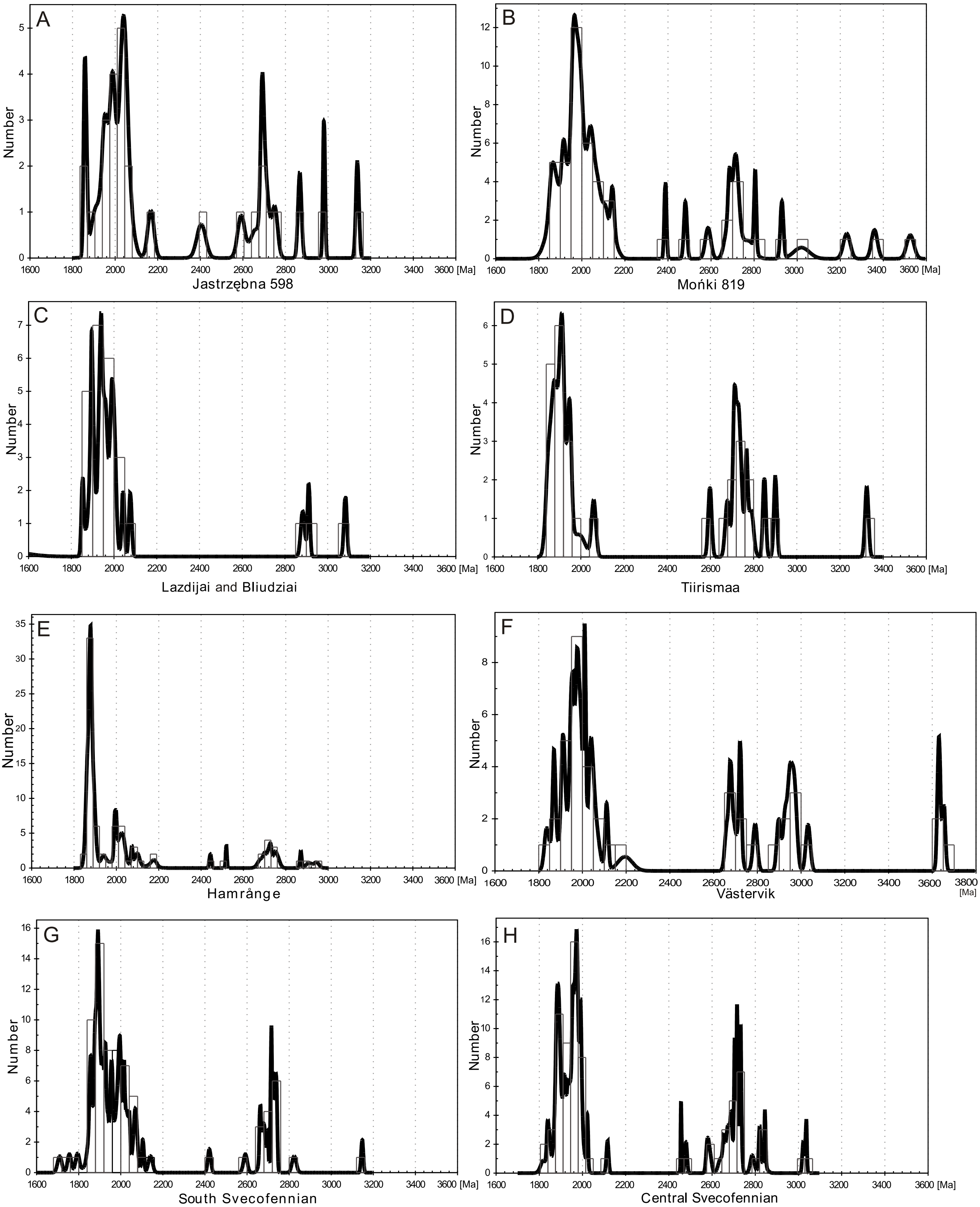

Fig. 6. Relative probability plots of $^{207}Pb/^{206}Pb$ detrital zircon ages for: A — Mońki sample 819; B — Jastrzębna sample 598, C — Bliudziai and Lazdijai samples (jointly) and selected Svecofennian metasediments

Data from: Bergman *et al*., 2008 (Tiirismaa and Hamrange); Sultan *et al*., 2005 (Västevik); Lahtinen *et al*., 2002 (South and Central Svecofennian); Williams *et al*., 2009 (Mońki and Jastrzębna)

Two of the youngest Jastrzębna detrital cores have a mean age of *ca*. 1.86 Ga, only slightly higher than the ages of the metamorphic zircon overgrowths and monazite, 1841 ±14 and 1827 ±20 Ma, respectively. Using the same criterion, the more mature Mońki sample has a maximum deposition age of 1830 ±35 Ma (Williams *et al*., 2009). The ages of the youngest detrital zircon cores from the Jastrzębna and Mońki samples are nearly coincident with that of nearby igneous activity, therefore the protolith sediments must have been deposited at or soon after *ca*. 1.86 Ga.

The maximum deposition ages measured on metasediments from the northern part of the Svecofennian domain are mostly similar. For example, mature quartzites from upper stratigraphic levels (southern Svecofennian) have maximum deposition ages of *ca*. 1.87–1.86 Ga (Lahtinen *et al*., 2002). Clastic sediments in the Västervik Basin, SE Sweden, were deposited after 1.85 Ga (Sultan *et al*., 2005). The maximum age of sedimentation from four localities: Luukkola (1842 ±10 Ma), Pyhäntaka (1865 ±11 Ma), Tiirsimaa (1848 ±13 Ma), Hamrånge (1855 ±10 Ma), as given by the youngest detrital zircon cores, has been constrained to time interval 1.88–1.83 Ga (Bergman *et al*., 2008). These data suggest that during the time interval 1.87–1.83 Ga there were several contemporaneous sedimentary basins in southern Fennoscandia. This period corresponds to an hiatus between two major phases of the Svecofennian Orogeny in Sweden and Finland, 1.89–1.86 Ga and 1.83–1.79 Ga, the Fennian and Svecobaltic episodes, respectively (Korja *et al*., 2006; Lahtinen *et al*., 2008), a period referred to as the intra-orogenic phase in the Fennoscandian Shield (Bergman *et al*., 2008).

Sedimentation in the Jastrzębna, Mońki, Lazdijai and Bliudziai areas was probably initiated at a late stage in the Svecofennian Orogeny, possibly reflecting deposition of detritus in advance of a continent-continent collision (Korja *et al*., 2006; Lahtinen *et al*., 2008).

COMPARISON OF DETRITAL ZIRCON AGES WITHIN THE SVECOFENNIAN DOMAIN

Several studies in the last decade have confirmed that the pattern of detrital zircon ages first found by Huhma *et al.* (1991) is a feature of the Svecofennian metasediments throughout Finland and Sweden (e.g., Claesson *et al*., 1993; Lahtinen *et al*., 2002; Rutland *et al*., 2004; Sultan *et al*., 2005; Bergman *et al*., 2008; Williams *et al*., 2008). The relative abundance of the various components differs from place to place (Fig. 6), but all analysed metasediments, either immature and mature, or from the upper or lower Svecofennian, are characterized by a major population of 2.1–1.9 Ga zircon, a scarcity of zircon with ages of 2.6–2.1 Ga, and a cluster of ages at about 2.9–2.7 Ga. Some have rare grains up to 3.5 Ga (e.g., up to 3.44 Ga at TSB, Southern Finland, and 3.32 Ga at Västervik, Central Sweden). Claesson *et al*. (1993) concluded that most of the Archaean zircon ages were consistent with sediment derivation from the Archaean craton in the northeastern part of the Baltic Shield, but that some of detritus must have been derived from unidentified Archaean sources elsewhere.

Common applications of detrital zircon geochronology include the comparison or/and correlation of detrital ages between sedimentary basins, or between a basin and its potential source regions. In many cases, however, such comparisons are compromised because insufficient grains from any one sample have been dated and there is a high probability that either some age components have been missed or their relative abundances are not accurately represented. Ideally, the characterization of a detrital zircon population should be based on the dating of at least 60 grains (Dodson *et al*., 1988; Fedo *et al*., 2003). Recent detrital zircon studies of metasediments from Central Sweden and Southern Finland (e.g., Sultan *et al*., 2005; Bergman *et al*., 2008) have been based on a large number of age measurements, sufficient to quantify the relative abundances of the major detrital components.

The detrital zircon age measurements from the Jastrzębna (n = 49) and Mońki (n = 54) metasediments (Fig. 6A, B) are sufficient in combination for quantitative comparisons (Williams *et al*., 2009). In contrast, the smaller data sets from Lazdijai and Bliudziai reported here (n = 6 and 20, respectively) are indicative only and risk missing minor components and misrepresenting the relative abundances of the major age groups. In summarizing the data from the Lithuanian metasediments (Fig. 6C), therefore, we have been able to document only a major Palaeoproterozoic population at 2.1–1.9 Ga, and a common Archaean group at 3.1–2.9 Ga.

The Pb isotopic ages measured by Bergman *et al*. (2008) on detrital zircon from quartz-rich metasandstones from the upper parts of Svecofennian sequences in southeastern Finland (Luukkola, Pyhäntaka and Tiirismaa) and east-central Sweden (Hamrånge) had a multimodal distribution very similar to that previously reported from that part of Fennoscandia, namely a high abundance of 2.1–1.9 Ga Palaeoproterozoic grains and a subsidiary population at 2.9–2.7 Ga. A sample from Tiirismaa (Arc Complex of Southern Finland), for example, had two main age groups, 2.1–1.85 and 2.95–2.60 Ga. The oldest grain was 3.3 Ga (Fig. 6D).

On the other hand, the detritus from Hamrånge in Central Sweden (Berslagen area) was somewhat different. The typical Palaeoproterozoic group at 2.1–1.85 Ga was dominant in those quartz-rich metasediments, but Archaean zircon was very rare. It was proposed that the youngest detritus from Hamrånge was derived from the nearby, extensive 1.86–1.84 Ga Ljusdal Batholith in Sweden (Bergman *et al*., 2008).

The southernmost outcrops of late Svecofennian metasedimentary rocks in Sweden are near Västervik, close to the Baltic Sea coast. Detrital zircons from three samples from a variety of sedimentary environments and lithofacies towards the bottom of the Västervik Basin have been dated by Sultan *et al*. (2005). The metasediments contained a mixture of Palaeoproterozoic (75%) and Archaean (25%) zircon. The main Palaeoproterozoic group ranged in age from 2.12 to 1.87 Ga, there were two Archaean groups (2.72–2.69 Ga and 3.03–2.95 Ga), and the oldest detrital grain was 3.64 Ga (Fig. 6F).

The detrital zircon age spectra from the central and southern Svecofennian sequences are all rather similar (Fig. 6G, H). Although the number of grains analysed per sample is relatively low, the combined results (about 158 analyses from 9 samples) show the typical mix of about 70% Palaeoproterozoic and 30% Archaean grains (Lahtinen *et al*., 2002).

The Mońki, Jastrzębna and combined Lazdijai and Bliudziai zircon age spectra (Fig. 6) are generally consistent with those from metasediments from different structural segments of the Svecofennian domain. The predominance of Palaeoproterozoic zircon with ages between *ca*. 2.10 and 1.90 Ga, plus small amounts of Archaean zircon (2.90–2.60 Ga) is evident on all diagrams, as well as the gap in the spectra between *ca*. 2.6 and 2.1 Ga.

The small number of detrital zircon ages reported from other units in the Baltic–Belarus region show a similar time span. Five dated grains from a metapelitic granulite from the BPG had ages of *ca*. 1.93 to 1.84 Ga (Claesson *et al*., 2001). Mansfeld (2001) measured ages of 2.24–2.11 Ga on a few detrital zircons from a metagreywacke within the East Lithuanian domain (EL).

Thus, all reported ages of detrital zircon from metasediments in the Polish–Lithuanian (e.g., WL and MD) terrane (*sensu* Bogdanova *et al*., 2006) are very similar to those of zircon from metasediments in the conventionally-defined Svecofennian domain.

POSSIBLE SOURCE AREAS

Accurate reconstructions of the locations of continents during the Early Palaeoproterozoic are essential in the quest to identify potential source areas for the Svecofennian sediments. The latitudes of several microcontinents established using palaeomagnetism suggest that, during the Palaeoproterozoic, there was a large landmass assembled along global scale 2.1–1.8 Ga collisional orogens (Zhao *et al*., 2002). The configuration of that supercontinent (Hudsonland: Pesonen *et al*., 2003, or Columbia: Rogers and Santosh, 2002; Zhao *et al*., 2004) remains controversial. However, in most plate tectonic reconstructions, the NW part of Baltica (*sensu* Fennoscandia) was attached to Greenland (part of Laurentia) and Western Amazonia. The former position of Amazonia remains problematic — only one reconstruction has placed that continent in close proximity to Baltica (at 1.65 Ga: Pesonen *et al*., 2003 and references therein). The major Palaeoproterozoic crust-forming event in Amazonia, the Trans-Amazonian Orogeny, occurred at 2.2–2.0 Ga (Tassinari and Macambira, 1999). On current geological evidence from the Amazonian craton, it appears that the continent was not assembled before 1.65 Ga (Pesonen *et al*., 2003).

Other possible microcontinents, namely Baltica and Greenland, drifted independently between 2.15 and 1.93 Ga, until collision in the period 1.9–1.8 Ga (Pesonen *et al.*, 2003), forming coeval orogens in Baltica and Laurentia (the Nagssugtoqidian Orogen). The most important crust forming events from Greenland are illustrated in Figure 7A. Palaeoproterozoic zircon ages of *ca*. 1.97–1.80 Ga have been measured on several igneous suites from Eastern Greenland, for example the Ammasslik arc-related Intrusive Complex (Nutman *et al*., 2008). Gneisses in Eastern Greenland have an Archaean (3.04–2.73 Ga) igneous protolith. Palaeo/Eoarchaean rocks (3.85–3.57 Ga) are found in the Itsaq Gneiss Complex. They are cut by *ca*. 3.50 Ga Ameralik dykes. The Amitsoq Gneisses in Western Greenland have ages of *ca*. 3.85–3.50 Ga (Nutman *et al*., 2000; Cates and Mojzsis, 2006; Hölttä *et al*., 2008). Zircon geochronology on both orthogneiss complexes confirms a minimum age for the igneous protoliths of *ca*. 3.75 Ga (Fig. 7A).

The zircon ages from Greenland bear some resemblance to the detrital zircon age spectra characteristic of the Svecofennian metasediments. For example, the oldest zircon found in sediments from the Västervik Basin has ages in the range 3.65–3.63 Ga (Sultan *et al*., 2005). It has been proposed, however, that such zircon, which is not well represented in the rocks of the Baltic Shield, is derived from the Sarmatia megablock (*op cit*.).

The reconstructions of Palaeoproterozoic Fennoscandia have not considered amalgamation with the Sarmatia Block at 2.0–1.8 Ga because of palaeomagnetic data (Pesonen *et al*., 2003 and references therein) that imply that Baltica (*sensu* Fennoscandia) and Ukraine (*sensu* Sarmatia) remained separated until sometime between 1.77 and 1.70 Ga (Elming *et al*., 2001; Bogdanova *et al*., 2001; Lubnina *et al*., 2009) when Ukraine docked with Baltica from the south. There are, however, no reliable palaeomagnetic data from Baltica at 2.0 Ga or Ukraine at 1.88 Ga (Pesonen *et al*., 2003) that would indicate the configuration and drift history of those microcontinents while the Svecofennian sediments were being deposited. The palaeopoles for Ukraine at 2.0 Ga indicate moderate northerly latitudes in the same hemisphere as Laurentia. At 1.88 Ga, Baltica was positioned at low to intermediate latitudes, so the two microcontinents in all likelihood were some distance apart at the time.

Disregarding the configuration of the continents at the time of sediment deposition (*ca*. 1.86 Ga), several previous workers (e.g., Lahtinen *et al*., 2002; Sultan *et al*., 2005) have speculated on the provenance of the Svecofennian detritus, preferring the Sarmatian megablock as the source area for the oldest zircon grains. However, magmatism as early as 3.5 Ga has also been documented in Central Finland (Mutanen and Huhma, 2003), and trondhjemite gneisses from the Pudasjärvi Granulite Belt, at up to *ca*. 3.4 Ga, are the oldest rocks so far identified in Fennoscandia (Fig.7A) (Slabunov *et al*., 2006).

Sarmatia contrasts with Fennoscandia in terms of the timing of early crustal formation, for example in the higher relative abundance of Eo/Palaeoarchaean (3.65–3.00 Ga) crust. The Sarmatian megablock is composed of Archaean continental crust formed between 3.7 and 2.7 Ga, with intervening Palaeoproterozoic belts accreted to the Archaean cores between 2.2–2.1 and 2.0–1.9 Ga (Claesson *et al*., 2006). Igneous rocks formed at 2.10–1.90 Ga are therefore common within several continental blocks: Fennoscandia, Sarmatia and Laurentia (Greenland).

The 2.10–1.90 Ga detrius in Fennoscandia was derived mainly from mafic and felsic volcanic rocks. The southern Svecofennian metasediments, however, are characterized by a predominance of 2.1–2.0 Ga zircon thought to be derived from sources with alkaline affinity (Lahtinen *et al*., 2002). Considering the age span and observed compositional relationships, (*op*. *cit*.) the Sarmatia Block has been similarly

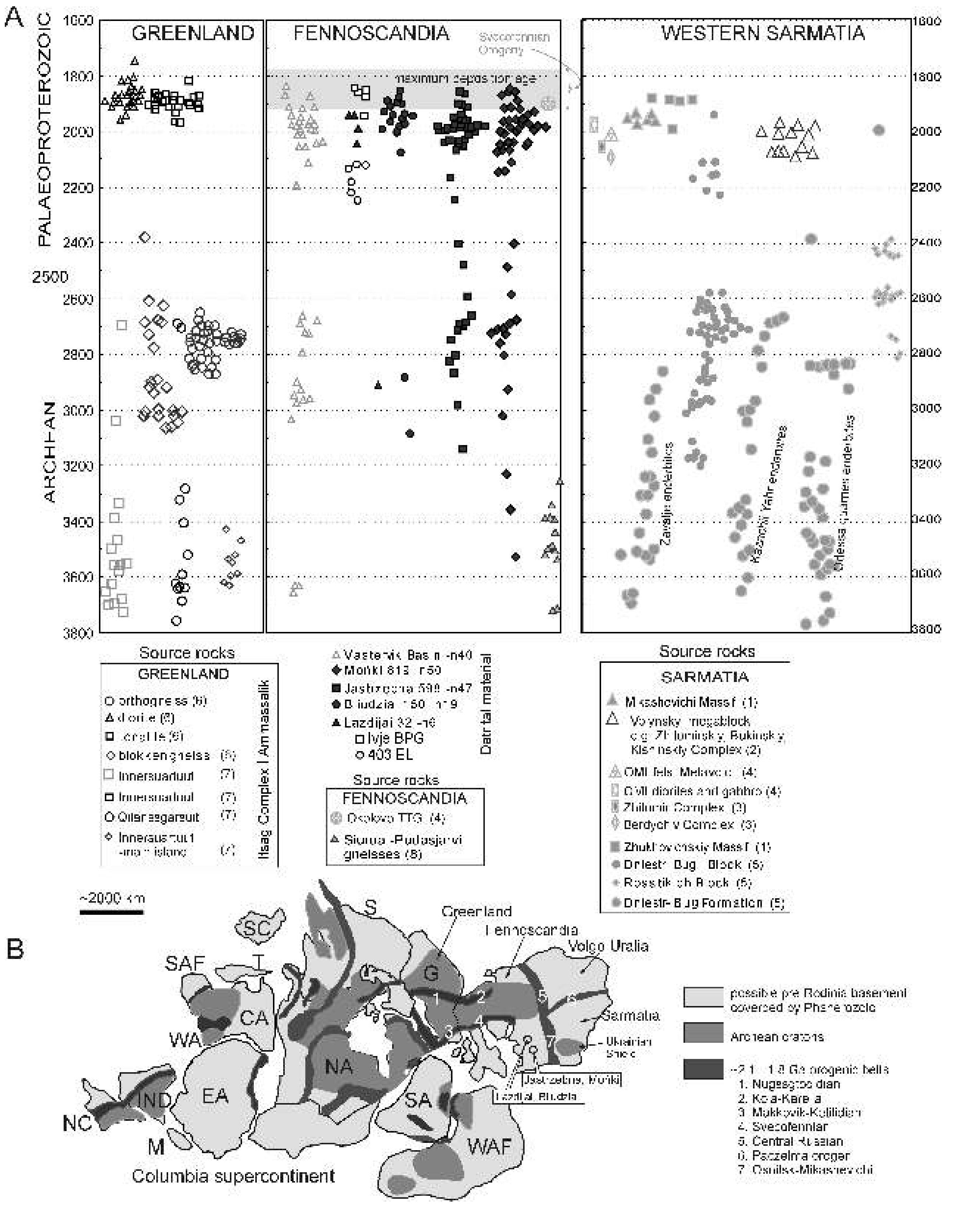

Fig. 7. The major source — detritus correlations

A — combined detrital zircons $^{207}Pb/^{206}Pb$ ages of metasedimentary rocks from Mońki, Jastrzębna, Lazdijai 32, Bliudziai 150 drill hole and from the Västervik Basin with igneous events of the Fennoscandian Okolovo Terrane and Greenland (a) and Sarmatia (b) discussed in the text. Data from: (1) Shcherbak *et al*., 1990; (2) Shcherbak *et al*., 2008, (3) Shcherbak *et al*., 2005; (4) Bogdanova *et al*., 2006; (5) Claesson *et al*., 2006; (6) Nutman *et al*., 2008; (7) Cates and Mojzsis, 2006; (8) Mutanen and Huhma, 2003; the major stages of Svecofennian Orogeny: 1.92–1.77 Ga (Korja *et al*., 2006) are marked by grey shading; **B** — possible configuration of continents of Palaeo-Mezoproterozoic Columbia supercontinent redrawn from Zhao *et al*., 2002; CA — Central Australia, EA — East Antarctica, G — Greenland, IND — India, M — Madagascar, NA — North America, NC — North China, S — Siberia, SAF — South Africa, SC — South China, T — Tarim, WA — West Australia, WAF — West Africa nominated as the possible source area for the southern Svecofennian metasediments. The metasediments from the Polish and Lithuanian boreholes have chemical compositions indicative of mostly felsic, but ranging to intermediate and mafic sources (Fig. 3B).

The chemical evidence for the provenance of the Mońki and Jastrzębna paragneisses has been discussed in detail by Williams *et al*. (2009). Trace element signatures of the metasediments (La-Th-Sc, Th-Sc-Zr/10) suggested that the source of their protolith was an active continental margin or continental (mature) island arc. The REE and Y contents of the Jastrzębna paragneiss indicated a relatively high proportion of source rocks of tonalite–trondhjemite–granodiorite (TTG) affinity and unweathered components with oceanic arc signatures (Williams *et al*., 2009). The nearest known occurrence of TTG rocks is the Okolovo–Rudma Fennoscandian subterrane (Taran, 2001; Bogdanova *et al*., 2006). That area is dominated by *ca*. 2.0 Ga metavolcanic rocks, including 1982 ±26 Ma metadacites (Bogdanova *et al*., 1994). A range of juvenile TTG magmas was emplaced in the northwestern part of the CBsz (including the Okolovo terrane) at *ca*. 1.9 Ga (Claesson *et al*., 2001; Taran and Bogdanova, 2003).

The relatively immaturity of the metasediments recovered from the Polish boreholes suggests that the source of the detritus was not far distant from the depositional basin. Most of the crust in the adjacent orogenic belts (e.g., the Lithuanian terrane) formed at *ca*. 1.85 Ga, however, so is too young to be a potential source. There is no doubt that in the Palaeoproterozoic, at 2.0–1.8 Ga, the distance between the Jastrzębna–Mońki area and the 1.9 Ga TTG bodies within the Central Belarus domain (*sensu* Bogdanova *et al*., 2001) and the Okolovo terrane (*sensu* Bogdanova *et al*., 2006) was different from that at the present day, but it is unlikely to have exceeded 300–400 km (Skridlaite *et al*., 2003*b*). Further, the BPG (and other units of the Lithuanian–Belarus terrane) and the Fennoscandian affinity Okolovo subterrane had a close relationship and shared a similar geodynamic evolution at that time (Bogdanova *et al*., 2006), so might well have been the proximal source areas.

The global-scale network of Palaeoproterozoic (*ca*. 2.1–1.8 Ga) collisional orogens led to the formation of a supercontinent Columbia/Hudsonia/Nuna at *ca*. 1.8 Ga (Fig. 7B). Fennoscandia, in the middle, was bounded by Laurentia in the NE, Volgo–Uralia in the E, Sarmatia in the SE, an unknown microcontinent in the SW, and possibly Amazonia in the W (e.g., Rodgers and Santosh, 2002; Zhao *et al*., 2002, 2004).

There are some marked similarities between the Svecofennian detrital zircon age spectra and the ages of important zircon-forming events in the more distal area of western Sarmatia (Fig. 7B). The Osnitsk–Mikashevichi Igneous Belt (OMI: 2.00–1.95 Ga), the youngest orogenic belt at the northwestern margin of Sarmatia (Bogdanova *et al*., 2006), contains granitoid massifs of Palaeoproterozoic age, for example the Mikashevichi, Zhukhovichskiy (1970 ±15 and 1900 ±45 Ma, respectively: Shcherbak *et al*., 1990) and Osnytsya complexes (2.02–1.97 Ga: Shcherbak *et al*., 2002). Furthermore, the Volyn domain on the western edge of Sarmatia is dominated by 2.1–2.0 Ga mafic and felsic metavolcanic and hypabyssal rocks, for example the Berdychiv enderbites with ages of *ca*. 2140, *ca*. 2166 and 2058 ±6 Ma (Shcherbak *et al*., 2005) and the Zhitomir Complex (2.08–2.0 Ga: Shcherbak *et al*., 2002, 2008).

Late Archaean granites with ages in the range 2.7–2.6 Ga, and slightly older high-grade metamorphic rocks, are present in the western Ukrainian Shield (Claesson *et al*., 2006). Eo/Palaeoarchaean rocks up to 3.6 Ga have also been found in the same region. There are *ca*. 3.75–3.65 Ga high-grade enderbites (Shcherbak *et al*., 2005; Claesson *et al*., 2006) and 3.4–3.1 Ga mafic enderbites (Lesnaya *et al*., 1995) in the Dniestr–Bug region of western Sarmatia and early Archaean two-pyroxene mafic granulites in the Tivriv Block (Bibikova *et al*., 2000).

The similarities between the Svecofennian detrital zircon age spectra and the zircon ages from the western edge of the Sarmatian Block are significant. They might suggest a closer spatial relationship between Baltica and Sarmatia, before final docking at 1.77–1.70 Ga, than has previously been considered. Further testing of that concept awaits the availability of palaeomagnetic data that reliably indicate the location of the two microcontinents at the time the Svecofennian sediments were being deposited (*ca*. 2.00–1.86 Ga).

As illustrated in Figures 7A and B, the ages of potential source rocks from proximal Svecofennian terranes of Fennoscandia, as well as igneous suites from distal Greenland (eastern Laurentia) and probably more distal western Sarmatia, are closely consistent with the detrital zircon age spectra measured on the Svecofennian metasediments throughout Finland, Sweden, Lithuania, and Poland.

CONCLUSIONS

There is now convincing evidence for the widespread presence of Palaeoproterozoic metasedimentary rocks in the basement of the Polish-Lithuanian and Lithuanian-Belarus terranes (*sensu* Bogdanova *et al*., 2006), and that these terranes are the buried southernmost part of the Svecofennian domain. Detrital zircon geochronology provides proof of a direct genetic link between those buried basement metasediments and the Svecofennian metasediments exposed in Sweden and Finland. The combined patterns of detrital zircon ages from Mońki, Jastrzębna, Lazdijai and Bliudziai show the same clusters of ages as commonly observed at different locations within the classical Svecofennian domain:

1. A predominance of Palaeoproterozoic ages between *ca*. 2.10 and 1.90 Ga;
2. Small numbers of Archaean ages between 2.90 and 2.60 Ga;
3. A marked scarcity of ages between *ca*. 2.4 and 2.2 Ga.

The youngest detrital zircon groups from the Jastrzębna and Mońki metasediments indicate significantly differenta maximum deposition ages of 1856 ±6 and 1829 ±9 Ma respectively .

These data from deep boreholes in Southern Lithuania and northeastern Poland, located very close to the present day boundary between the Fennoscandian and Sarmatian megablocks, suggest that at *ca*. 1.86 Ga (Late Svecofennian), in the southern part of Fennoscandia, several continental margin sedimentary basins existed more or less coevally.

The detrital zircon ages compiled here demonstrate the widespread occurrence of early Archaean (>3.2 Ga) detritus within the Svecofennian metasediments that is not a typical age of the Fennoscandian crust. On the other hand, detrital zircon populations with ages of *ca*. 2.1–1.9 and 2.8–2.6 Ga, as well as 3.1 and 3.5 Ga from the Jastrzębna, Mońki, Lazdijai and Bliudziai boreholes closely match the ages of most episodes of crust formation in the Fennoscandian Block, as well as in Greenland (eastern Laurentia): 3.75–3.06, 2.85–2.64 and 1.96–1.88 Ga, and the Sarmatia microcontinent: 3.69–3.24, 3.15–3.03, 2.79–2.59, 2.1–2.0 and 1.97–1.88 Ga.

The relative immaturity of the basement metasediments discovered by drilling in Poland and Lithuania suggests that their source area was quite close to the depositional basins. The close similarity between the detrital zircon age spectra and the crust formation ages in eastern Laurentia and Sarmatia is consistent with these microcontinents having been in close proximity to Fennoscandia for at least 100 m.y. before they finally docked at 1.77–1.70 Ga.

Acknowledgements. We thank Prof. S. Bogdanova for her constructive criticism of, and comments on, the first version of this manuscript, which helped us to clarify our arguments. The geochronological study of the basement of Poland is being carried out under the auspices of a collaborative research agreement between the Polish Geological Institute-National Research Institute and The Australian National University. G. Skridlaite thanks SYNTHESYS for support made available by the European Community Research Infrastructure Action under the FP6 Structuring the European Research Area Programme (project SE-TAF-1535), M. Whitehouse and the NORDSIM facility at the Swedish Museum of Natural History, the Lithuanian State Science and Study Foundation (LSSSF) and the VISBY-Programme administered by the Swedish Institute.

REFERENCES

ANDERSEN T. (2005) — Detrital zircons as tracers of sedimentary provenance: limiting conditions from statistics and numerical simulation. Chem. Geol., **216** (3–4): 249–270.

BERGMAN S., HÖGDAHL K., NITRONEN M., OGENHALL E., SJÖSTROM H., LUNDQVIST L. and LAHTINEN R. (2008) — Timing and Palaeoproterozoic intra-orogenic sedimentation in the central Fennoscandian Shleld: evidence from detrital zircons in matasandstone. Precambrian Res., **161** (3–4): 231–249.

BIBIKOVA E. V., BOGDANOVA S. V., GORBATSCHEV R., CLAESSON S. and KIRNOZOVA T. I. (1995) — Isotopic Age, Nature, and Structure of the Precambrian Crust of Belarus. Stratigraph. Geol. Correl., **3** (6): 591–601.

BIBIKOVA E. V., CLAESSONS S. and STEPHANYUK L. M. (2000) — The oldest granitoids of the Southwestern of Ukrainian Shield (the Bug region): U-Pb and Sm-Nd isotope ages (in Russian). XIX seminal “Geology of magmatic rocks”, April, 6–7, 2000: 25–26.

BOGDANOVA S. V. (2005) — The East European Craton: some aspects of the Proterozoic evolution in its south-west. Spec. Pap. Pol. Miner. Soc., **26**: 18–24.

BOGDANOVA S. V., BIBIKOVA E. and GORBATSCHEV R. (1994) — Paleoproteroterozoic U–Pb zircon ages from Belorussia: new tectonic implication for the East European Craton. Precambrian Res., **68**: 231–240.

BOGDANOVA S. V., GORBATSCHEV R. and STEPHENSON R. A. (2001) — EUROBRIDGE: Palaeoproterozoic accretion of Fennoscandia and Sarmatia. Tectonophysics, **339** (1–2): 39–66.

BOGDANOVA S., GORBATCHEV R., GRAD M., JANIK T., GUTERCH A., KOZLOWSKAYA E., MOTUZA G., SKRIDLAITE G., STAROSTENKO I., TARAN L. and EUROBRIDGE and POLONAISE WORKING GROUP (2006) — EUROBRIDGE: new inslight into the geodynamic evolution of East European craton. In: European Lithosphere Dynamics (eds. D. G. Gee and R. A. Stephenson). Geol. Soc., London, Mem., **32**: 599–625.

CATES N. L. and MOJZSIS S. J. (2006) — Chemical and isotopic evidence for widespread Eoarchean metasedimentary enclaves in southern west Greenland. Geochim. Cosmochim. Acta, **70**: 4229–4257.

CLAESSON S., BIBIKOVA E., BOGDANOVA S. and SKOBELEV V. (2006) — Archaean terranes, Palaeoproterozoic reworking and accretion in the Ukrainian Shield. East European Craton Geol. Soc., London, Mem., **32**: 645–654.

CLAESSON S., BOGDANOVA S. V., BIBIKOVA E. V. and GORBATSCHEV R. (2001) — Isotopic evidence for Palaeoproterozoic accretion in the basement of the East European Craton. Tectonophysics, **339** (1–2): 1–18.

CLAESSON S., HUHMA H., KINNY P. D. and WILLIAMS I. S. (1993) — Svecofennian detrital zircon ages — implications for the Precambrian evolution of the Baltic Shield. Precambrian Res., **64**: 109–130.

CLAESSON S. and RYKA W. (1999) — Nd model ages of the Precambrian crystalline basement of the NE Poland between EUROBRIDGE and TESZ. Abstact of Seventh Eurobridge Workshop: May 26–30, 1999. Suwalki–Szelment Poland: 17–18.

DEPCIUCH T., LI J. and SYLWESTRZAK H. (1975) — K-Ar age of the rocks of the crystalline basement of north-east Poland (in Polish with English sumary). Kwart. Geol., **19** (4): 759–779.

DODSON M. H., COMPSTON W., WILLIAMS I. S. and WILSOM J. F. (1988) — A search for ancien detrital zircons in Zimbabwean sediments. J. Geol. Soc., **145** (6): 977–983.

DZIEDZIC A. (1976) — Wyniki badan geochemicznych. In: Jastrzebna IG 1. Profile głębokich otworów wiertniczych Inst. Geol., **36**: 68–75.

EMLING S-A., PESONEN N. P. and KRAVCHENKO S. (2001) — Paleomagnetism of Proterozoic rocks From Ukrainian Shields: new tectonic reconstructions of the Ukrainian and Fennoscandian Shields. Tectonophysics, **339** (1–2): 19–38.

EUROBRIDGE’95 WORKING GROUP (2001) — Deep seismic profiling within the East European Craton. Tectonophysics, **339**: 153–175.

FEDO C. M., SIRCOMBE K. N. and RAINBIRD R. H. (2003) — Detrital zircon analysis of sedimentary record. In: Zircon (eds. J. M. Hanchar and P. Hoskin). Rev. Miner. Geochem., **53**: 277–303.

GAAL G. and GORBATSCHEV R. (1987) — An outline of Precambrian evolution of the Baltic Shield. Precambr. Res., **35**: 15–52.

GORBATSCHEV R. and BOGDANOVA S. (1993) — Frontiers in the Baltic Shield. Precambrian Res., **64**: 3–21.

HENRY D. J., GUIDOTTI C. V. and THOMSON J. A. (2005) — The Ti-saturation surface for low-to-medium pressure metapelitic biotite: implications for Geothermometry and Ti-substitution Mechanisms. Am. Miner., **90**: 316–328.

HERRON M. M. (1988) — Geochemical classification of terrigenous sands and shales from core or log data. J. Sediment. Petrol., **58**: 820–829.

HÖGDAHL K., SJÖSTRÖM H., ANDERSSON U. B. and AHL M. (2008) — Continental margin magmatism and migmatisation in the west-central Fennoscandian Shield. Lithos, **102**: 435–459.

HÖLTTÄ P., BALAGANSKYY V., GARDE A. A., MERTANEN S., PELTONEN P., SLABUNOV A., SORJONEN-WARD P. and WHITEHOUSE M. (2008) — Archean of Greenland and Fennoscandia. Episodes, **31**: 13–19.

HUHMA H. (1986) — Sm-Nd, U-Pb, Pb-Pb isotopic evidence for the origin of the Early Proterozoic Svecokarelian crust in Finland. Geol. Surv. Finland Bull., **337**: 1–48.

HUHMA H., CLAESSON S., KINNY P. D. and WILLIAMS I. S. (1991) — The growth of Early Proterozoic crust: new evidence from Svecofennian detrital zircons. Terra Nova, **3**: 175–178.

JACKOWICZ E. (2000) — Wyniki badań chemicznych. In: Wigry IG 1. Profile głębokich otworów wiertniczych. Państw. Inst. Geol., **95**: 38–42.

KÄHKÖNEN Y. and LEVEINEN J. (1994) — Geochemistry of metasedimnetary rocks of the Palaeoproterozoic Tampere schist belt, southern Finland. In: Geochemistry of Proterozoic Supracrustal Rocks in Finland (eds. M. Nironen and Y. Kähkönen). Geol. Surv. Finland Spec. Pap., **19**: 117–136.

KORJA A., LAHTINEN R. and NIRONEN M. (2006) — The Svecofennian orogen: a collage of microcontinents and island arcs. In: European Lithosphere Dynamics (eds. D. G. Gee and R. A. Stephenson). Geol. Soc. London, Mem., **32**: 561–578.

KRZEMIŃSKA E., WILLIAMS I. and WISZNIEWSKA J. (2005) — A Late Paleoproterozoic (1.80 Ga) subduction-related mafic igneous suite from Łomża, NE Poland. Terra Nova, **17**: 442–449.

KRZEMIŃSKA E. and WISZNIEWSKA J. (2007) — Swekofeńskie korzenie “masywu” mazowieckiego. Prz. Geol., **55** (4): 294–295.

KRZEMIŃSKA E., WISZNIEWSKA J. and WILLIAMS I. S. (2006) — Revision of the “oldest” rocks age in the crystalline basement of NE Poland (in Polish with English summary). Prz. Geol., **54** (11): 967–973.

KRZEMIŃSKA E., WISZNIEWSKA J., WILLIAMS I. S. and DÖRR W. (2007) — Late Paleoproterozoic arc-related granites from the Mazowsze domain, NE Poland. In: Granitoids in Poland (eds. W. Kozłowski and J. Wiszniewska): 41–56. AM Monograph.

KUBICKI S. and RYKA W. (1982) — Geological Atlas of crystalline basement in Polish part of the East European Platform (in Polish with English summary). Wyd. Geol., Warszawa.

KUBICKI S. and RYKA W. (1996) — Wstęp W. Ryka In: Mońki IG 1, IG 2. Profile głębokich otworów wiertniczych Państw. Inst. Geol., **84**: 5–9.

LAHTINEN R. (1996) — Geochemistry of Palaeoproterozoic supracrustal and plutonic rocks in the Tampere-Hämeenlinna area, southern Finland. Geol. Surv. Finland Bull., **389**: 1–113.

LAHTINEN R. (2000) — Arechean-Proterozoic transition: geochemistry, provenance and tectonic setting of metasedimentary rocks in central Fennoscandian Shield, Finnland. Precambrian Res., **104**: 147–174.

LAHTINEN R., HUHMA H. and KOUSA J. (2002) — Contrasting source components of the Paleoproterozoic Svecofennian metasediments: detrial zircon U-Pb, Sm-Nd and geochemical data. Precambrian Res., **116**: 81–109.

LAHTINEN R., KORJA A. and NIRONEN M. (2005) — Paleoproterozoic tectonic evolution. In: Precambrian Geology of Finland — Key to the Evolution of the Fennoscandian Shield (eds. M. Lehtinen, P. A. Nurmi and O.T. Rämö): 481–532. Elsevier, Amsterdam.

LAHTINEN R., GARDE A. A. and MELEZHIK V. A. (2008) — Paleoproterozoic evolution of Fennoscandia and Greenland. Episodes **31**: 20–28.

LESNAYA I. M., PLOTKINA T. E., STEPANIUK V. M. and BARTNICKY E. N. (1995) — Age stages formation of the mafite-enderbitic association of the Bug River area (in Russian). Geochemistry and ore formation, **21**: 56–68.

LUBNINA N., BOGDANOVA S. and SHUMLYANSKYY L. (2009) — The collision between Volgo-Sarmatia and Fennoscandia at *ca*. 1.76 Ga: new palaeomagnetic data from the Ukrainian Shield. Geophys. Res. Abstracts, 11, EGU2009-13229-3, EGU General Assembly.

LUDWIG K. R. (1991) — ISOPLOT, a plotting and regression program for radiogenic data. USGS Open-file Report: 91–445.

MANSFELD J. (2001) — Age and ε_{Nd} constraints on the Palaeoproterozoic tectonic evolution in the Baltic-Sea region. Tectonophysics, **339**: 135–151.

MEZGER K. and KROGSTAD E. J. (1997) — Interpretation of discordant U-Pb zircon ages: an evaluation. J. Metamorph. Petrol., **15**: 127–140.

MOTUZA G. (2005) — Structure and formation of the crystalline crust in Lithuania. Miner. Soc. Poland, Spec. Pap., **26**: 70–80.

MOTUZA G. (2008) — Southern Fennoscandia II (Lithuania). Geology of Europe, vol. 1: Precambrian and Palaeozoic (ed. T. McCann): 38–43. Geol. Soc., London.

MOTUZA G., MOTUZA V., SALNIKOVA E. and KOTOV A. (2006) — Orogenic magmatism in Western Lithuania. Geophys. Res. Abstracts., **8**: 06050, 2006.

MOURI H., KORSMAN K. and HUHMA H. (1999) — Tectono-metamorphic evolution and timing of the melting processes in the Swecofennian tonalite-Trondhjemite migmatite belt: example from Luopiinen, Tampere area, S Finland. Bull. Geol. Soc., Finland, **71**: 31–56.

MUTANEN T. and HUHMA H. (2003) — The 3.5 Ga Siurua trondhjemite gneiss in the Archean Pudasjärvi Granulite Belt, northern Finland. Bull. Geol. Soc., Finland, **75**: 51–68.

NUTMAN A. P., BENNETT V. C., FRIEND C. R. L and NORMAN M. D. (2000) — Metaigneous (non-gneissic) tonalites and quartz-diorites from an extensive *ca*. 3800 Ma terraine south of the Isua supracrustal belt, southern West Greenland; constrains on early crust formation. Contribut. Mineral. Petrol., **137**: 364–388.

NUTMAN A. P., KALSBEEK F. and FRIEND C. R. L. (2008) — The Nagssugtoqidian orogen in Southerneast greenland: evidence for Paleoproterozoic collision and plate assembly. Am. J. Sc., **308**: 529–572.

PESONEN L. J., ELMING S-A., MERTANEN S., PISAREVSKY S., D’AGRELLA-FILHO M. S., MEERT J. G. and SCHMIDT P. W. (2003) — Palaeomagnetic configuration of continents during the Proterozoic. Tectonophysics, **375**: 289–324.

PUCHTEL I. S., ZHURAVLEV D. Z. and KULIKOVA V. V. (1991) — Komatiites of Vodlozero Block in the Baltic Shield: a window into the Early Archean Mantle. Dokl. Akad. Nauk SSSR, **317**: 197–200.

RIMSA A., BOGDANOVA S. V., SKRIDLAITE G. and BIBIKOVA E. (2001) — The Randamonys TTG-intrusion In southern Lithuania: evidence of a 1.84 Ga island arc. J. Conf. Abstrakt, **6**: 368–359.

ROGERS J. J. W. and SANTOSH M. (2002) — Configuration of Columbia, a Mesoproterozoic supercontinent. Gondwana Res., **5**: 5–22.

ROSER B. P. and KORSH R. J. (1986) — Determination of tectonic setting of sandstone-mudstone suites using SiO_2 content and K_2O/Na_2O ratio. J. Geol., **94**: 635–650.

ROSER B. P. and KORSCH R. J. (1988) — Provenance signatures of sandstone- mudstone suite determined using discriminant function analysis of major-element data. Chem. Geol., **67**: 119–139.

RUTLAND R. W. R., WILLIAMS I. S. and KORSMAN K. (2004) — Pre-1.91 Ga deformation and metamorphism in the Palaeoproterozoic Vammala Migmatite Belt, southern Finland, and implications for Svecofennian tectonics. Bull Geol. Soc. Finland, **76**: 93–140.

RYKA W. (1976) — Wyniki badan petrograficznych skał podloza krystalicznego. In: Jastrzębna IG 1. Profile głębokich otworów wiertniczych Inst. Geol., **36**: 21–46.

RYKA W. (1984) — Precambrian evolution of the East European Platform in Poland. Biul. Inst. Geol., **347**: 17–28.

RYKA W. (1989) — Wyniki badań geochemicznych skał metamorficznych. In: Kaplonosy IG 1. Profile głębokich otworów wiertniczych. Państw. Inst. Geol., **68**: 116–120.

RYKA W. (1996) — Wyniki badań geochemicznych. In: Czyże IG 1. Profile głębokich otworów wiertniczych Państw. Inst. Geol., **85**: 45–50.

SAMBRIDGE M. S. and COMPSTON W. (1994) — Mixture modelling of multi-component data sets with application to ion-probe zircon ages. Earth Planet. Sc. Letters, **128** (3–4): 373–390.

SHCHERBAK N. P., ARTEMENKO G. V., SKOBELEV V. M., STEPANYUK L. M. and PONOMARENKO O. M., (2002) — Geochronology of the Precambrian of the Ukrainian Shield on the basis of the “Reference” isotope data. Miner. J., **24**: 87–110.

SHCHERBAK N. P., ARTEMENKO G. V., LESNYA I. M. and PONOMARENKO A. N. (2005) — Geochronology of the Early Precambrian of the Ukrainian Shield (in Russian). The Archean. Naukova Dumka, Kiev.

SHCHERBAK N. P., ARTEMENKO G. V., LESNYA I. M. and PONOMARENKO A. N. (2008) — Geochronology of the Early Precambrian of the Ukrainian Shield. The Proterozoic. Naukova Dumka, Kiev.

SHCHERBAK N. P., PAP A. M., BARTNITSKY E. N. and ZAYATS A. P. (1990) — The uranium-lead isotopicage of granitoids in Belarussia (in Russian). Doklady Akad. Nauk Belorusskoy SSR, **34**: 740–743.

SKIÖLD T. and RUTLAND R. W. R. (2006) — Successive ~1.94 Ga plutonism and ~1.92 Ga deformation and metamorphism south of the Skellefte district, northern Sweden. Substantation of the marginal basin accretion hypothesis of Swecofennian evolution. Precambrian Res., **148**: 181–204.

SKRIDLAITE G. and MOTUZA G. (2001) — Precambrian domains in Lithuania: evidence of terrane tectonics. Tectonophysics, **339**: 113–133.

SKRIDLAITE G., WILLINGSHOFER E. and STEPHENSON R. (2003*b*) — P-T-t modelling of Proterozoic terranes in Lithuania: geodynamic implications for accretion of SW fennoscandia. GFF, **125**: 201–211.

SKRIDLAITE G., WISZNIEWSKA J. and DUCHESNE J. C. (2003*a*) — Ferro-potassic A-type granites and related rocks in NE Poland and S Lithuania: west of the East European Craton. Precambrian Res., **124**: 305–326.

SLABUNOV A. I., LOBACH-ZHUCHENKO S. B., BIBIKOVA E. V., SORJONEN-WARD P., BALAGANSKY V. V., VOLODICHEV O. I., SHCHIPANSKY A. A., SVETOV S. A., CHERKULAEV V. P., ARESTOVA N. A. and STEPANOV V. S. (2006) — The Archean nucleus of Fennoscandian (Baltic) Shield. In: European Lithosphere Dynamics (eds. D. G. Gee and R. A. Stephenson). Geol. Soc., London, Mem., **32**: 627–644.

SORJONEN-WARD P. and LUKKONNEN E. (2005) — Archean rocks. In: The Precambrian Geology in Finland — Key to Evolution of the Fennoscandian Shield (ed. Lahtinen, Nurmi Rämö): 19–99. Elsevier, Amsterdam.

STACEY J. S. and KRAMERS J. D. (1975) — Approximation of terrestrial lead isotope evolution by a two-stage model. Earth Planet. Sc. Letters, **2**: 207–221.

SULTAN L., CLAESSON S. and PLINK-BJÖRKLUND P. (2005) — Proterozoic and Archean ages of detrital zircon from the Palaeoproterozoic Västrvik Basin, SE Sweden: Implications for provenance and timing of deposition. GFF, **127**: 17–24.

TARAN L. (2001) — Distribution of the Okolovo unit in the crystalline basement of Belarus (in Russian). Litasfera, **1** (14): 72–81.

TARAN L. N. and BOGDANOVA S. V. (2003) — Metamorphism of Paleoproterozoic Paragneisses in the Belarus-Podlyasie Granulite Belt: Prograde-Retrograde Evolution. Petrology C/C of Petrologiia, **11** (4): 383–399.

TASSINARI C. C. G. and MACAMBIRA M. J. B. (1999) — Geochronological provinces of the Amazonian Craton. Episodes, **22**: 174–182.

VERMEESH P. (2004) — How many grains are needed for a provenance study? Earth Planet. Sc. Letters, **224**: 441–451.

WHITEHOUSE M. J., KAMBER B. and MOORBATH S. (1999) — Age significance of U-Th-Pb zircon data from early Archaean rocks of west Greenland — a reassessment based on combined ion-microprobe and imaging studies. Chem. Geol., **160**: 201–224.

WIEDENBECK M., ALLE P., CORFU F., GRIFFIN W. L., MEIER M., OBERLI F., Von QUADT A., RODDICK J. C. and SPIEGEL W. (1995) — Three natural zircon standards for U-Th-Pb, Lu-Hf trace element and REE analysis. Geostand. Newslet., **19**: 201–224.

WILLIAMS I. S., RUTLAND R. W. R. and KOUSA J. (2008) — A regional 1.92 Ga Tectonothermal episode in Ostrobothnia, Finland: Implications for models of Svecofennian Accretion. Precambrian Res. (2008), doi: 10.1016/j.precamres. 2008. 05.004

WILLIAMS I. S., KRZEMIŃSKA E. and WISZNIEWSKA J. (2009) — An extension of the Svecofennian orogenic province into NE Poland: evidence from geochemistry and detrital zircon in Paleoproterozoic paragneisses. Precambrian Res., **17**: 234–254.

WISZNIEWSKA J., KRZEMIŃSKA E. and DÖRR W. (2007) — Evidence of arc-related Svecofennian magmatic activity in the southwestern margin of the East European Craton in Poland. Gondwana Res., **12**: 268–278.

WISZNIEWSKA J., KRZEMIŃSKA E. and WILLIAMS I. S. (2006) — Geochemical and geochronological correlation of Svecofennian metasedimentary complexes from NE Poland, central Finland and Sweden. AGU General Assembly. Vienna, April 2006.

WOŁKOWICZ K. (1996) — Wyniki badań geochemicznych. In: Mońki IG 1, IG 2. Profile głębokich otworów wiertniczych Państw. Inst. Geol., **84**: 37–56.

WYBRANIEC S. (1999) — Transformation and visualization of potential field data. Państw. Inst. Geol., Spec. Pap., **1**: 1–88.

ZHAO G. C., CAWOOD P. A., WILDE S. A. and SUN M. (2002) — Review of global 2.1–1.8 Ga orogens: implications for a pre-Rodinia supercontinent. Earth Sc. Rev., **59**: 125–162.

ZHAO G., SUN M., WILDE S. A. and LI S. Z. (2004) — A Paleo-Mesoproterozoic supercontinent; assembly, growth and breakup. Earth Sc. Rev., **67**: 91–123.