Geol. Mag. **143** (*5*), *2006, pp. 679–697.* © *2006 Cambridge University Press*
doi:10.1017/S0016756806002494 *Printed in the United Kingdom*

Multistage growth and reworking of the Palaeoproterozoic crust in the Bergslagen area, southern Sweden: evidence from U–Pb geochronology

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(*Received* 25 April 2005; *accepted* 13 March 2006)

Abstract – The Svecofennian Domain of the Fennoscandian Shield constitutes a considerable volume of Palaeoproterozoic crustal growth, 2.1–1.86 Ga ago, in between the Archaean craton in the NE and the 1.85–1.65 Ga Transscandinavian Igneous Belt (TIB) in the south and west. The Bergslagen area is a classical ore province located in the southwestern part of the Svecofennian Domain of south-central Sweden. Its northern part is dominated by volcanic and plutonic rocks of a magmatic arc with continental affinity, while the SE part is made up by a sedimentary basin. The Bergslagen area shows a metamorphic zonation from lower to middle amphibolite facies in the north to upper amphibolite facies and locally granulite facies in the south; a small greenschist area exists in the west. Identifying the age spectra of inherited components, magmatic crystallization, as well as metamorphic episodes, provide important constraints on the geodynamic evolution of this centrally located piece of the Shield.

U–Pb zircon SIMS data presented in this paper complement the previous, regionally scattered TIMS data from this area. Magmatic zircons from two felsic metavolcanic rocks and two amphibolites (metagabbros) yield 1888 ± 12, 1892 ± 7 and 1887 ± 5, 1895 ± 5 Ma, respectively; i.e. within the 1.91–1.86 Ga range previously obtained for Early Svecofennian magmatism in Bergslagen. An augen gneiss from southern Bergslagen, assigned to the earliest TIB generation, yield an intrusive age of 1855 ± 6 Ma. Metamorphic monazites from the same rock indicate that deformation and elevated thermal activity prevailed 1.83–1.82 Ga ago (TIMS). Metamorphic zircons in high-grade metasedimentary rocks from the south and west yield ages of 1793 ± 5 and 1804 ± 10 Ma, in accordance with ages for regional peak metamorphism and migmatite formation found elsewhere in the southern Svecofennian province of Sweden. More importantly, a few zircon crystals and overgrowths in rocks from the north indicate an early metamorphic episode at *c.* 1.87 Ga, indicating that Bergslagen has experienced two major metamorphic events. Detrital and inherited zircons span the range 2.78–1.90 Ga, with an apparent gap at 2.45–2.1 Ga, which further emphasize previous observations of a major juvenile (< 2.1 Ga) and a minor Archaean provenance. This, and in particular the 1.94–1.91 Ga crystals present in the *c.* 1.89 Ga amphibolites, support the suggestion of a former Palaeoproterozoic pre-1.91 Ga crust in the Bergslagen area.

Keywords: geochronology, zircon, SIMS, TIMS, Bergslagen, Svecofennian, magmatic, inherited, metamorphic.

1. Introduction

The Fennoscandian Shield comprises an Archaean nucleus in the NE to which Proterozoic terranes have successively been accreted along the southern and western flanks (e.g. Gaál & Gorbatschev, 1987). Post-Archaean development started with rifting of the craton interior and margin at 2.45–2.0 Ga (e.g. Park *et al.* 1984; Gaál & Gorbatschev, 1987; Nironen, 1997). This was followed at 2.1–1.93 Ga by juvenile crust formation in several arc systems outboard the craton in the SW resulting in 'microcontinents', including the Bergslagen arc (e.g. Nironen, 1997; Nironen, Lahtinen & Korja, 2002; Lahtinen, Korja & Nironen, 2004). Sm–Nd results indicate that some of these microcontinents contain unexposed Archaean basement (Andersson, Neymark & Billström, 2002). This earliest 'proto-Svecofennian' development is documented in only a few preserved outcrops of magmatic rocks (e.g. Wasström, 1993; 1996; Skiöld *et al.* 1993; Lahtinen & Huhma, 1997; and references therein), but detrital zircons of this age are frequent in the associated metasedimentary rocks (Claesson *et al.* 1993; Lahtinen, Huhma & Kousa, 2002; Andersson *et al.* 2004a; Sultan, Claesson & Plink-Björklund, 2005).

The main Early Svecofennian rock-forming episode (1.91–1.86 Ga) resulted in thorough reworking of these early arc systems, partly including rift-related

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volcanism, and in formation of juvenile crust in areas between the microcontinents. The final accretion of this composite collage to the craton margin occurred at *c.* 1.86 Ga (e.g. Vivallo & Claesson, 1987; Baker, Hellingwerf & Oen, 1988; Lagerblad, 1988; Vivallo & Willdén, 1988; Gaál, 1990; Allen *et al.* 1996; Nironen, 1997). Subsequent long-term north(east)ward subduction resulted in pervasive reworking of the newly formed crust during the Late Svecofennian (*c.* 1.85–1.75 Ga). This reworking formed the voluminous Transscandinavian Igneous Belt (TIB) along the SW margin, as well as penecontemporaneous, mainly granitoid magmatism and regional metamorphism further continentward (e.g. Patchett, Todt & Gorbatschev, 1987; Andersson, 1991; Andersson *et al.* 2004b). Apart from a hiatus in magmatism between *c.* 1.75 and 1.71 Ga, continental reworking continued until *c.* 1.65 Ga, partly coeval with juvenile crustal growth further to the south and west, recorded in the Southwest Scandinavian Domain (Gorbatschev, 1980; 2004; Berglund & Larson, 1997; Åhäll & Larson, 2000).

In this paper we use the term 'Svecofennian' to cover processes of major crust formation and reworking in the time period *c.* 2.1–1.75 Ga, subdivided into: (1) 'proto-Svecofennian' earliest oceanic-arc crust formation (2.1–1.91 Ga), (2) Early Svecofennian period of reworking and main juvenile crust formation (*c.* 1.91–1.86 Ga), and (3) Late Svecofennian period of mainly underplating and reworking of the juvenile (and cratonic) crust (*c.* 1.85–1.75 Ga).

The Bergslagen area constitutes the SW part of the southern Svecofennian province (Gaál & Gorbatschev, 1987) in the Svecofennian Domain, and comprises a volcano-metallogenetic district rich in ores that probably extends eastwards into the Orijärvi area in southern Finland (e.g. Gaál, 1990). The northern and western areas of Bergslagen are dominated by volcanic formations, while the sedimentary successions of the Sörmland basin dominate the bedrock in the SE (e.g. Stålhös, 1991; Allen *et al.* 1996; Romer & Öhlander 1995), recently interpreted as an accretionary prism (Korja & Heikkinen, 2005). In the NE large areas are made up by Early Svecofennian intrusive rocks, referred to as the *c.* 1.89 Ga Uppland batholith (Fig. 1). The Bergslagen volcano-plutonic complexes have been interpreted as formed in a rifted mature (continental) arc setting (e.g. Oen *et al.* 1982; Vivallo & Claesson, 1987; Baker, Hellingwerf & Oen, 1988; Lagerblad, 1988; Vivallo & Willdén, 1988; Gaál, 1990; Allen *et al.* 1996). The metamorphic grade is high in southern Bergslagen, including the Sörmland basin (Rickard, 1988; Stålhös, 1991). Paragneisses in upper amphibolite facies are characteristically sillimanite-bearing, and locally, in the vicinity of Transscandinavian Igneous Belt intrusions, granulite facies conditions prevail (600–900 °C and 3–6 kbar) (Andersson, Larsson & Wikström, 1992; Wiktröm & Larsson, 1993; Andersson, 1997b). There is a gradual decrease in grade into lower–middle amphibolite facies conditions towards northern Bergslagen, where pelitic rocks are characteristically andalusite-bearing (*c.* < 600 °C and 2–4 kbar; Stålhös, 1991; Ripa, 1994; Sjöström & Bergman, 1998), with a minor greenschist facies area in the NW (Fig. 1; e.g. Lundström, 1995). A relatively abrupt change in metamorphic conditions to higher (upper amphibolite facies) grade occurs in northernmost Bergslagen and in the transition zone to the central Svecofennian province further north (Stålhös, 1991; Sjöström & Bergman, 1998). Bergslagen is truncated by several major shear zones, notably the Singö Shear Zone (SSZ) and the Ornö Banded Series (OBS) in the east that may have formed a conjugate set during N–S convergence. The main shearing took place > 1.85 Ga (Singö Shear Zone) and > 1.82 Ga (Ornö Banded Series) (Fig. 1; Persson & Sjöström, 2002; 2003; Hermansson *et al.* 2004; and references therein). These zones also coincide with breaks in metamorphic grade: higher grade NE of the Singö Shear Zone and lower grade SE of the Ornö Banded Series (Fig. 1).

Crystallization ages of rocks formed during the Early Svecofennian magmatic phase in Bergslagen are generally between 1.91 and 1.86 Ga (Åberg, 1978; Welin, Wiklander & Kähr, 1980; Åberg & Strömberg, 1984; Åberg, Levi & Fredrikson, 1984; Welin, 1987; Öhlander & Billström, 1989; Persson, 1993; Dobbe, Oen & Verdurmen, 1995; Kumpulainen *et al.* 1996; Ripa & Persson, 1997; Lundström *et al.* 1998; Persson & Persson, 1999). Two rocks in this area have yielded anomalously low U–Pb zircon ages: a granitic rock from western Bergslagen (*c.* 1850 Ma; Åberg *et al.* 1983), and a felsic metatuffite from the south (1836 ± 9 Ma; Dobbe, Oen & Verdurmen, 1995), respectively. However, the 1850 Ma age of the former is imprecise due to considerable scatter around the discordia. A recent study by Högdahl & Jonsson (2004) reinterpreted the granitic rock as a complex comprising mostly volcanic to subvolcanic rocks, and presented three new U–Pb zircon ages, all close to 1885 Ma, i.e. within the 'normal' age range. The age of the metatuffite has been reinterpreted as metamorphic, reflecting the medium-grade metamorphism in the area (Lundström *et al.* 1998).

Components older than 1.91 Ga have so far only been recorded from metasedimentary rocks. Detrital zircons from metasedimentary rocks in the southernmost parts of Bergslagen yielded ages ranging from 3.63 to 1.87 Ga and 2.12 to 1.87 Ga (Claesson *et al.* 1993; Sultan, Claesson & Plink-Björklund, 2005), where the Archaean proportion consititued about one fourth of the population. In central Bergslagen a metasedimentary rock hosts Archaean zircons in the age range 2.97–2.60 Ga and a dominating population of Palaeoproterozoic 2.04–1.90 Ga zircons (Claesson *et al.* 1993), while 2.70 and 1.96 Ga zircons have been reported from SW Bergslagen (Kumpulainen

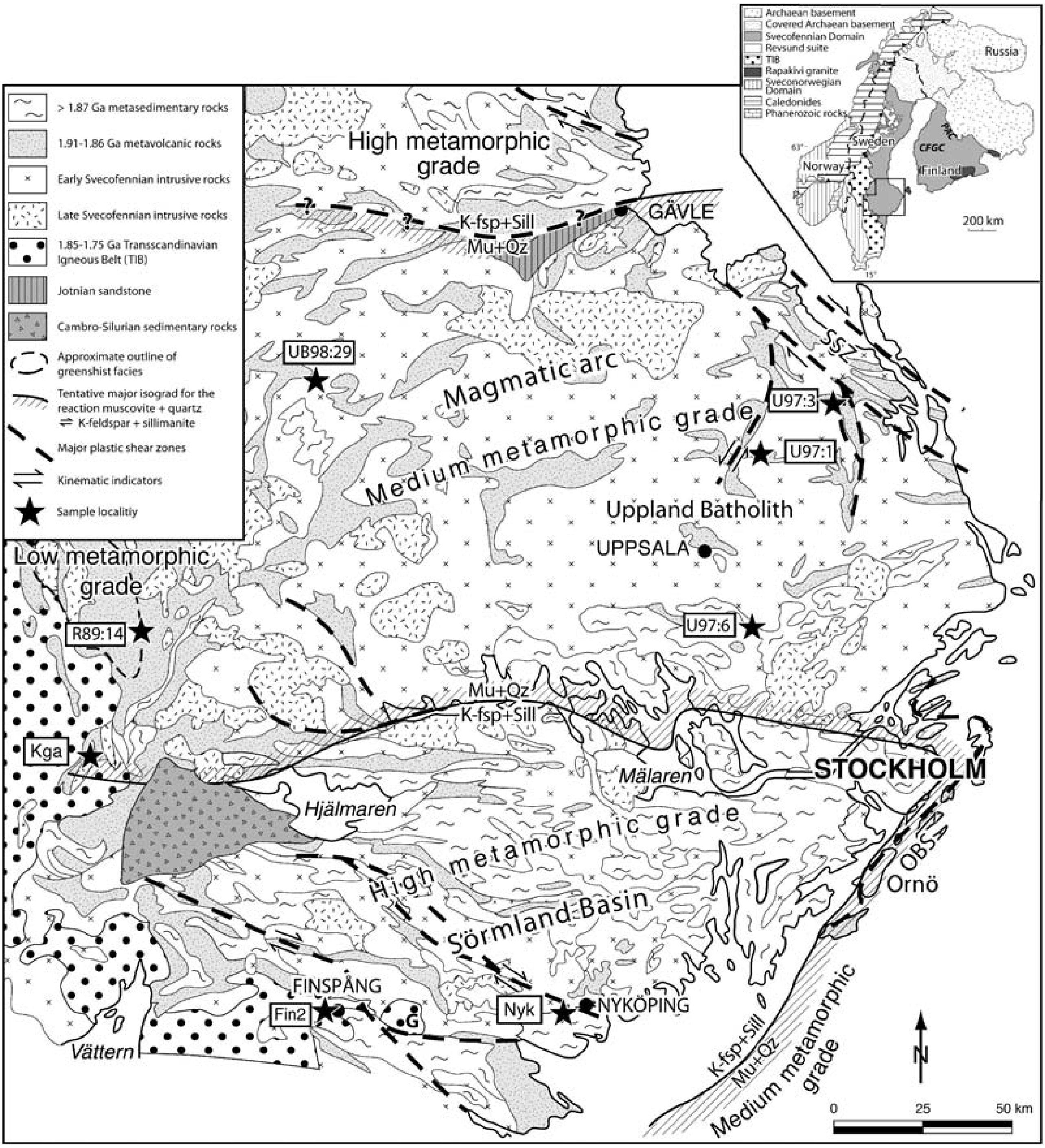

Figure 1. Geology of the Bergslagen area, southern Sweden (modified from Stålhös, 1991), showing sample sites (sample coordinates are found in Table 2). The Early Svecofennian formations (1.91–1.86 Ga) include the metasedimentary, metavolcanic and early orogenic intrusive rocks. The Late Svecofennian intrusions are 1.82–1.75 Ga old. The 'post-orogenic', Transscandinavian Igneous Belt (TIB) rocks are 1.85–1.78 Ga in this area, while the Jotnian sediments are 1.50–1.26 Ga. References in the text. G refers to the Graversfors intrusion (Early TIB). Shear zone kinematics from Sjöström & Bergman (1998), Person & Sjöström (2002, 2003). SSZ and OBS refer to the Singö Shear Zone and the Ornö Banded Series, respectively. Inset shows location of Bergslagen within the Fennoscandian Shield (frame). PAC is the Primitive Arc Complex of eastern Finland and CFGC is the Central Finland Granitoid Complex (e.g. Vaasjoki *et al.* 2003).

et al. 1996). The existence of old, pre-Svecofennian, components in metasedimentary rocks in Bergslagen is further substantiated by relatively low initial ε_{Nd} ratios of -2 to -3.7 (Miller *et al.* 1986; Patchett, Todt & Gorbatschev, 1987; Valbracht, 1991; Kumpulainen *et al.* 1996; Andersson, 1997a).

Timing of metamorphic events in the Bergslagen region is poorly constrained. Romer & Öhlander (1995)

Table 1. TIMS monazite U–Pb results, Finspång augen gneiss

Fraction	Weight (μg)	Number of crystals	U (ppm)	Pb_{tot} (ppm)	Pb_{com} (ppm)	$^{206}Pb/^{204}Pb$ (1)	rad. Pb atomic % $^{206-207-208}Pb$	$^{206}Pb/^{238}U$ (2)	$^{207}Pb/^{235}U$ (2)	$^{207}Pb/^{206}Pb$ (2)	$^{207}Pb–^{206}Pb$ (age)
Fin2 m1	6.0	1	8556	6062	5.71	21514	38.3–4.3–57.4	0.314 ± 20	4.82 ± 31	0.11146 ± 12	1823 ± 2
Fin2 m2	6.7	4		1584	3.63	5200	24.5–2.7–72.8			0.11181 ± 15	1829 ± 2.5
Fin2 m3	18.0	7	3290	4481	2.04	24331	20.6–2.3–77.1	0.3246 ± 22	5.008 ± 35	0.11190 ± 12	1831 ± 2

Pb_{tot} is total lead, and Pb_{com} common lead.
(1) $^{206}Pb/^{204}Pb$ only corrected for mass discrimination.
(2) Ratios corrected for mass discrimination, blank (0.01 ng U and 0.01 ng Pb), and common Pb after Stacey & Kramers (1975) at the crystallization age.
Errors are 2σ. U measurement failed for Fin2 m2.

reported a U–Pb monazite age of 1846 ± 1 Ma for a migmatite granite from central Bergslagen, as well as a U–Pb titanite age of 1789 ± 2 Ma for a skarn from the northwestern part of the region (Romer & Öhlander, 1994). Andersson (1997b) determined several U–Pb monazite ages from upper amphibolite grade regional metamorphic (1.82–1.80 Ga) and granulite grade contact metamorphic rocks close to intrusions of the Transscandinavian Igneous Belt (1.85–1.78 Ga) in southern Bergslagen.

Emplacement ages of a number of different pegmatites in Bergslagen, which may be closely related to the Svecofennian regional metamorphism, have been determined by columbite U–Pb geochronology to 1.82–1.78 Ga (Romer & Smeds, 1994; 1997; Romer, 1997). The studies so far bracket the timing of metamorphism in Bergslagen to the period 1.85–1.78 Ga, where only the titanite skarn age (1.79 Ga) of Romer & Öhlander (1994) is from the northern medium-grade (middle–lower amphibolite facies) region.

The metamorphic age range completely overlaps with that of Transscandinavian Igneous Belt intrusives along the margins in the south and west (1.85–1.78 Ga; Patchett, Todt & Gorbatschev, 1987; Jarl & Johansson, 1988; Persson & Wikström, 1993; Persson & Ripa, 1993, Wikström, 1996; Andersson, 1997a; Claeson & Andersson, 2000), while ages of the late orogenic granites within the Bergslagen area partly overlap and tend to be slightly younger (1.81–1.75; Åberg & Bjurstedt, 1986; Billström, Åberg & Öhlander, 1988; Stephens, Wahlgren & Annertz, 1993; Sundblad, Ahl & Schöberg, 1993; Ivarsson & Johansson, 1995; Bergman, Schöberg & Sundblad, 1995; Öhlander & Romer, 1996).

In order to add constraints on the Proterozoic development of the southern Svecofennian Province, SIMS and TIMS U–Pb analysis has been carried out on metaigneous and metasedimentary rocks from the Bergslagen area. In particular, important objectives were to identify inherited and metamorphic components in selected rocks from all over the Bergslagen region.

2. Analytical techniques

Mineral samples were separated using standard techniques. For U–Pb ID-TIMS analyses, between one and seven clear, inclusion-free monazite crystals (Table 1) were selected for each fraction. The chemical procedure for extracting U and Pb followed Krogh (1973) and Parrish *et al.* (1987), employed the micro capsule dissolution technique (Parrish, 1987), and was described in detail by Andersson (1997b).

In order to sample as wide a spectrum as possible of inherited, magmatic and metamorphic types, zircons separated for SIMS analysis from each sample were hand-sorted into groups of: (i) translucent crystals without visible inclusions or cores (generally smaller), and (ii) turbid crystals with inclusions and eventual cores. The zircons were mounted in epoxy together with chips of a zircon standard (Geostandards 91500), polished and coated with gold. U–Th–Pb analyses were performed using a Cameca IMS 1270 ion microprobe at the NordSIM laboratory of the Laboratory for Isotope Geology (LIG), Swedish Museum of Natural History in Stockholm, Sweden. Analytical procedures followed those described by Whitehouse *et al.* (1997) and Whitehouse, Kamber & Moorbath (1999). Analyses were conducted using a *c.* 4 nA O_2^- primary beam producing a spot size of *c.* 25 × 20 μm. Pb/U ratios were calibrated against a 1065 Ma reference (Geostandards 91500; Wiedenbeck *et al.* 1995), and error on this ratio includes error propagation of the calibration curve produced by repeated analyses on the zircon standard. Common lead correction was made for modern lead composition $^{206}Pb/^{204}Pb = 18.703$, $^{207}Pb/^{204}Pb = 15.629$, $^{208}Pb/^{204}Pb = 38.631$, and concordia diagrams were constructed using the Isoplot/Ex 2.49 program by Ludwig (2001).

Cathodoluminescence (CL) imaging was performed on all analysed crystals, and back-scattered electron (BSE) imaging on selected grains, with a Hitachi S-4300 FE-SEM (at LIG) and a Zeiss DSM962 (at GeoForschungsZentrum, Potsdam, Germany) scanning electron microscopes.

3. Results and interpretations

Eight samples were collected for geochronology: five from lower amphibolite facies areas of northern Bergslagen and three from higher grade areas of southern Bergslagen (see Appendix). All but one are metasupracrustal rocks and amphibolites; the last sample is a deformed megacrystic granite from southern Bergslagen. The sample localities are shown in Figure 1, and coordinates are found in Table 2. Zircons

Table 2. Ion-microprobe U–Th–Pb data for zircons from Bergslagen, southern Sweden

Sample/ spot no.	[U] ppm	[Th] ppm	[Pb] ppm	Th/U	$^{206}Pb/^{204}Pb$	Disc.% 2σ lim.	$^{207}Pb/^{206}Pb$	$\pm\sigma$ %	$^{207}Pb/^{235}U$	$\pm\sigma$ %	$^{206}Pb/^{238}U$	$\pm\sigma$ %	ρ	$^{207}Pb/^{206}Pb$ Age	$\pm\sigma$
U97:1 (coordinates, national grid: 666635/161850)															
n744-01a	149	85	60	0.57	5810	− 3.9	0.1148	0.49	4.897	2.2	0.3094	2.1	0.97	1876	8.8
n744-03a	305	129	130	0.42	2670		0.1140	2.1	5.259	3.1	0.3344	2.2	0.72	1865	38
n744-04a	892	294	280	0.33	6360	− 16.8	0.1107	0.22	3.939	2.1	0.2581	2.1	0.99	1810	3.9
n744-05a	556	307	241	0.55	28100		0.1143	0.27	5.364	2.1	0.3403	2.1	0.99	1869	4.9
n744-06a	1195	461	475	0.39	16200	− 0.8	0.1155	0.20	5.144	2.1	0.3231	2.1	1.00	1887	3.6
n744-07a	1228	452	438	0.37	32200	− 10.0	0.1142	0.21	4.557	2.1	0.2894	2.1	1.00	1867	3.8
n744-08a	176	52	62	0.30	11800	− 8.3	0.1147	0.40	4.667	2.2	0.2952	2.1	0.98	1874	7.1
n744-09a	1272	444	450	0.35	33400	− 10.7	0.1149	0.18	4.586	2.1	0.2895	2.1	1.00	1878	3.3
n744-10a	1952	1291	594	0.66	13200	− 18.2	0.1068	1.1	3.472	2.4	0.2359	2.1	0.90	1745	19
n744-11a	1488	444	517	0.30	17100	− 10.4	0.1139	0.24	4.514	2.1	0.2874	2.1	0.99	1863	4.3
n744-12a	1776	751	553	0.42	3540	− 13.4	0.1096	1.5	3.821	2.6	0.2529	2.1	0.82	1792	27
n744-13a	2022	738	464	0.36	5410	− 35.5	0.1045	0.26	2.682	2.2	0.1862	2.2	0.99	1705	4.9
n744-14a	439	153	171	0.35	21600	− 2.5	0.1162	0.25	5.123	2.1	0.3198	2.1	0.99	1898	4.4
n922-22c	749	569	124	0.76	1660	− 19.2	0.1068	8.0	1.895	8.4	0.1287	2.5	0.30	1746	140
n922-21a	175	53	44	0.31	6980	− 27.7	0.1091	1.1	3.183	2.4	0.2117	2.2	0.90	1784	20
n922-22a	345	149	91	0.43	3560	− 30.4	0.1111	0.57	3.277	2.2	0.2140	2.2	0.97	1817	10
n922-22b	1702	831	652	0.49	27300	− 0.3	0.1102	0.24	4.681	2.2	0.3080	2.1	0.99	1803	4.3
n922-20a	49	25	21	0.51	12700		0.1146	1.1	5.407	2.2	0.3421	1.9	0.88	1874	19
n922-19a	506	211	222	0.42	125000		0.1155	0.31	5.640	1.9	0.3542	1.9	0.99	1888	5.6
n922-18a	1885	786	810	0.42	23000		0.1146	0.16	5.521	1.9	0.3494	1.9	1.00	1874	2.9
n922-17a	1078	472	461	0.44	270000		0.1159	0.19	5.510	1.9	0.3449	1.9	1.00	1893	3.4
n922-16a	416	185	178	0.44	229000		0.1164	0.36	5.536	2.0	0.3448	1.9	0.98	1902	6.4
n922-15a	206	47	83	0.23	28800		0.1148	0.46	5.389	2.0	0.3403	1.9	0.97	1877	8.3
U97:3 (668180/163190)															
n748-07a	692	132	289	0.19	73900	0.2	0.1158	0.33	5.708	2.1	0.3574	2.0	0.99	1893	5.9
n748-08a	449	248	194	0.55	1370		0.1164	0.50	5.404	2.1	0.3367	2.0	0.97	1902	9.0
n748-1a	1720	585	754	0.34	26600	1.5	0.1156	0.17	5.743	2.0	0.3603	2.0	1.00	1889	3.0
n748-2a	551	177	219	0.32	35200		0.1152	0.25	5.255	2.0	0.3308	2.0	0.99	1883	4.5
n748-3a	1351	729	591	0.54	26200		0.1157	0.17	5.507	2.0	0.3452	2.0	1.00	1891	3.0
n748-4a	588	193	246	0.33	2870		0.1152	0.32	5.498	2.1	0.3461	2.0	0.99	1883	5.7
n748-5a	658	307	268	0.47	1540	− 0.4	0.1163	0.68	5.187	2.1	0.3236	2.0	0.95	1900	12
n748-5b	1663	546	720	0.33	14900		0.1166	0.34	5.655	2.3	0.3517	2.3	0.99	1905	6.2
n748-6a	737	136	296	0.19	52400		0.1164	0.30	5.516	2.1	0.3435	2.0	0.99	1902	5.4
U97:6 (661830/161545)															
n749-01a	2276	408	358	0.18	2160	− 47.2	0.09660	0.35	1.819	2.1	0.1366	2.1	0.99	1559	6.5
n749-01b	4382	1173	466	0.27	2640	− 58.1	0.08694	0.36	1.104	2.1	0.09212	2.0	0.99	1359	6.8
n918-02a	175	62	77	0.35	> 1e6		0.1207	0.92	5.908	2.3	0.3550	2.1	0.92	1967	16
n918-03a	195	62	68	0.32	4720	− 6.8	0.1138	1.3	4.535	2.6	0.2889	2.2	0.87	1862	23
n918-04a	186	100	78	0.54	8010	− 10.6	0.1282	0.69	5.686	2.2	0.3216	2.1	0.95	2074	12
n918-05a	106	82	56	0.77	51500		0.1286	0.54	6.969	2.2	0.3929	2.1	0.97	2080	9.6
n918-05b	107	64	51	0.60	8700		0.1279	0.75	6.458	2.3	0.3663	2.1	0.94	2069	13
n918-06a	991	760	455	0.77	61000		0.1175	0.21	5.556	2.2	0.3430	2.1	1.00	1918	3.8
n918-07a	3107	1885	718	0.61	2500	− 18.1	0.09828	2.7	2.482	7.8	0.1832	7.3	0.94	1592	50
n918-08a	376	266	184	0.71	27200		0.1233	0.29	6.289	2.2	0.3699	2.1	0.99	2005	5.1
n918-09a	286	82	119	0.29	53800		0.1167	0.37	5.588	2.2	0.3474	2.1	0.99	1906	6.6
n918-10a	414	107	152	0.26	3370	− 3.8	0.1146	0.42	4.894	2.2	0.3097	2.1	0.98	1874	7.6
n918-11a	195	97	84	0.50	42300		0.1164	0.44	5.496	2.2	0.3425	2.1	0.98	1901	7.8
n918-12a	2014	333	493	0.17	3100	− 30.1	0.1095	0.31	3.219	2.3	0.2132	2.3	0.99	1791	5.7
n918-13a	1142	213	411	0.19	28900	− 5.5	0.1161	0.28	4.957	2.2	0.3096	2.2	0.99	1898	5.0
n918-14a	248	36	93	0.14	22900	− 1.1	0.1175	0.53	5.290	2.2	0.3265	2.1	0.97	1919	9.5
n918-15a	1014	310	434	0.31	10000	− 22.1	0.1593	0.37	7.551	2.2	0.3437	2.1	0.99	2449	6.2
n918-16a	2243	838	844	0.37	1450	− 30.6	0.1602	0.86	6.653	2.4	0.3012	2.2	0.93	2458	15
n918-17a	1318	112	617	0.085	11700	− 14.2	0.1669	0.41	9.052	2.2	0.3933	2.2	0.98	2527	6.9
n918-18a	640	507	314	0.79	48500		0.1182	0.72	5.885	2.3	0.3610	2.1	0.95	1930	13
n918-19a	1094	120	380	0.11	1250	− 6.3	0.1157	0.66	4.829	2.2	0.3027	2.1	0.96	1891	12
n918-20a	2044	61	455	0.030	1800	− 40.8	0.1185	0.28	3.225	2.2	0.1975	2.1	0.99	1933	5.1
n918-9b	1480	73	580	0.050	121000		0.1141	0.22	5.469	2.1	0.3477	2.1	0.99	1866	4.0
R89:14 (663060/144590)															
n746-09a	136	106	63	0.78	151000		0.1157	0.53	5.486	2.0	0.3439	2.0	0.97	1891	9.5
n746-08a	105	75	48	0.71	33600		0.1152	0.64	5.523	2.0	0.3477	1.9	0.95	1883	12
n746-01a	294	263	126	0.90	613		0.09561	11	4.296	11	0.3259	2.6	0.23	1540	191
n746-02a	329	220	147	0.67	41000		0.1176	0.46	5.487	2.6	0.3385	2.6	0.98	1920	8.1
n746-03a	618	596	289	0.97	110000		0.1170	0.38	5.391	2.6	0.3343	2.5	0.99	1911	6.7
n746-04a	1048	911	478	0.87	49900		0.1154	0.23	5.356	2.6	0.3365	2.5	1.00	1887	4.1
n746-05a	729	683	354	0.94	92900		0.1151	0.32	5.547	2.6	0.3497	2.5	0.99	1881	5.8
n746-06a	314	366	153	1.2	40300		0.1159	0.34	5.377	2.6	0.3366	2.5	0.99	1893	6.1
n746-07a	571	444	247	0.78	4360		0.1130	1.9	5.022	3.2	0.3223	2.5	0.80	1848	34

Table 2. *(Contd.)*

Sample/ spot no.	[U] ppm	[Th] ppm	[Pb] ppm	Th/U	$^{206}Pb/^{204}Pb$	Disc.% 2σ lim.	$^{207}Pb/^{206}Pb$	$\pm\sigma$ %	$^{207}Pb/^{235}U$	$\pm\sigma$ %	$^{206}Pb/^{238}U$	$\pm\sigma$ %	ρ	$^{207}Pb/^{206}Pb$ Age	$\pm\sigma$
UB98:29 (668110/149230)															
n743-01a	239	92	99	0.39	61300		0.1161	0.34	5.369	2.1	0.3355	2.1	0.99	1897	6.1
n743-02a	414	202	169	0.49	344000	− 1.2	0.1160	0.21	5.172	2.1	0.3235	2.1	1.00	1895	3.8
n743-04a	165	116	65	0.70	64300	− 9.0	0.1192	0.77	4.971	2.3	0.3025	2.1	0.94	1944	14
n743-05a	444	408	196	0.92	96700	0.0	0.1157	0.36	5.194	2.1	0.3257	2.1	0.99	1890	6.5
n743-06a	721	847	328	1.2	152000	− 1.3	0.1155	0.48	5.099	2.2	0.3202	2.1	0.98	1888	8.7
n743-09a	144	63	60	0.44	102000		0.1164	0.66	5.426	2.0	0.3381	1.9	0.95	1901	12
n743-08a	248	171	112	0.69	122000		0.1166	0.41	5.548	2.0	0.3451	1.9	0.98	1905	7.3
n743-07b	100	54	44	0.53	47400		0.1140	0.99	5.483	2.2	0.3488	2.0	0.90	1864	18
n743-07a	84	37	36	0.45	66300		0.1145	0.67	5.476	2.0	0.3469	1.9	0.94	1872	12
Fin2 (650975/149720)															
n920-13a	413	165	168	0.40	71400		0.1132	0.35	5.149	2.0	0.3299	1.9	0.98	1852	6.3
n920-12a	276	102	119	0.37	8050		0.1207	0.53	5.835	2.0	0.3505	1.9	0.97	1967	9.4
n920-11a	84	47	37	0.55	> 1e6	− 1.1	0.1252	0.96	5.968	2.2	0.3457	1.9	0.90	2032	17
n920-10b	181	92	98	0.51	5400	− 5.3	0.1618	0.90	9.355	2.1	0.4194	1.9	0.91	2474	15
n920-10a	1016	34	396	0.033	48700		0.1135	0.29	5.435	1.9	0.3472	1.9	0.99	1857	5.2
n920-09c	802	34	306	0.043	2800		0.1132	0.39	5.299	2.0	0.3395	1.9	0.98	1851	7.0
n920-09b	87	52	42	0.60	9060		0.1244	0.84	6.381	2.1	0.3719	1.9	0.92	2021	15
n920-09a	564	25	214	0.044	20100		0.1135	0.34	5.286	2.0	0.3377	1.9	0.98	1857	6.2
n920-08a	260	51	85	0.19	418	− 9.7	0.1125	1.1	4.307	2.2	0.2778	1.9	0.86	1840	21
n920-07a	423	53	162	0.12	886		0.1142	0.64	5.254	2.0	0.3338	1.9	0.95	1867	11
n747-01a	754	132	276	0.17	2200	− 0.2	0.1127	0.37	4.858	2.6	0.3128	2.5	0.99	1843	6.7
n747-02a	355	54	130	0.15	2670		0.1129	0.56	4.942	2.6	0.3173	2.6	0.98	1847	10
n747-03a	686	499	285	0.73	2050	− 0.4	0.1128	0.43	4.858	2.6	0.3123	2.5	0.99	1845	7.8
n747-04a	965	52	275	0.054	5580	− 10.7	0.1037	0.29	3.639	2.6	0.2544	2.5	0.99	1692	5.4
n747-06a	425	61	159	0.14	24500		0.1134	0.31	5.082	2.1	0.3249	2.1	0.99	1855	5.6
n747-05a	211	115	121	0.55	51200	− 1.1	0.1592	0.33	9.593	2.1	0.4372	2.1	0.99	2447	5.6
Kga (657760/143135)															
n477-01a	647	28	243	0.044	5920		0.1097	0.23	4.852	0.57	0.3208	0.52	0.91	1795	4.3
n477-02a	551	29	214	0.053	7900		0.1114	0.25	5.078	0.57	0.3307	0.52	0.90	1822	4.5
n477-03a	1336	29	362	0.022	11400	− 18.9	0.1064	0.28	3.573	1.4	0.2435	1.3	0.98	1739	5.1
n477-04a	162	25	119	0.15	34400	− 1.8	0.1941	0.30	13.93	0.59	0.5207	0.50	0.86	2777	4.9
n477-05a	413	31	185	0.075	23400		0.1178	0.30	5.706	0.58	0.3512	0.50	0.86	1924	5.4
n477-6a	310	30	126	0.10	35000		0.1168	0.34	5.508	0.64	0.3420	0.54	0.85	1908	6.1
n477-07a	766	38	250	0.050	131000		0.1116	0.43	5.265	5.8	0.3422	5.7	1.00	1827	7.7
n477-08a	1898	130	563	0.068	162000		0.1095	0.20	4.687	5.8	0.3104	5.8	1.00	1791	3.6
n477-09a	788	46	232	0.058	2750		0.1099	0.56	4.674	5.8	0.3085	5.7	1.00	1797	10
Nyk (651450/157690)															
n479-01c	339	19	119	0.057	182000	− 0.8	0.1100	0.37	4.715	1.1	0.3108	1.0	0.94	1800	6.7
n479-01d	389	34	138	0.089	109000		0.1098	0.34	4.763	1.0	0.3147	0.93	0.94	1796	6.1
n479-02d	816	32	280	0.039	337000	− 3.2	0.1105	0.27	4.662	1.0	0.3061	0.94	0.96	1807	5.0
n479-2c	391	31	143	0.080	130000		0.1105	0.30	4.914	1.0	0.3225	0.93	0.95	1808	5.4

were extracted from all samples. Monazite was only found in the Finspång augen gneiss (Fin2; Table 1), and in the Kga and Nyk samples. Monazite from the latter two samples have previously been analysed by U–Pb TIMS geochronology (Andersson, 1997b). U–Pb ID-TIMS work (Table 1) was undertaken on monazite from the Finspång augen gneiss (Fin2). Zircons from all samples were analysed by SIMS. Results are reported in Table 2 and shown in Figure 2. Analyses with high common lead ($^{206}Pb/^{204}Pb < 5000$) have not been plotted and are not discussed, except for sample Fin2 (see below). Weighted averages discussed below are all ^{207}Pb–^{206}Pb ages.

Th/U ratios have sometimes been used to distinguish whether particular zircons grew from a melt or from a metamorphic fluid phase; those grown under subsolidus conditions usually have ratios lower than 0.2 (e.g. Williams & Claesson, 1987; Vavra *et al.* 1996; Schaltegger *et al.* 1999; Da Silva *et al.* 2000). Such systematics are, however, not always straightforward (e.g. Whitehouse *et al.* 1997), and depend on the chemical environment during (re)crystallization, i.e. whether another phase with high Th/U (e.g. monazite, allanite, etc.) is coprecipitating from the fluid (or melt), or not (e.g. Bea & Montero, 1999; Schaltegger *et al.* 1999). Coprecipitating monazite is very likely in peraluminous rocks, but less so in calc-alkaline and mafic rocks (e.g. Overstreet, 1967; Parrish, 1990), where allanite/epidote or titanite may act as Th-sinks instead. The youngest, metamorphic zircons in the high-grade metapelitic rocks of this study (Nyk and Kga) crystallized together with monazite (cf. Andersson, 1997b) and, consequently, have low Th/U ratios (< 0.1). A few of the old detrital zircons in sample U97:6 also have low Th/U ratios. These are rather high in U, metamict, and may originally have grown together with a high-Th/U phase, although the rock now lacks detrital or metamorphic monazite. In the augen gneiss sample (Fin2) some of the oscillatory zoned zircon crystals have Th/U ratios < 0.1, consitent with the rock´s peraluminous (Wikström & Aaro, 1986; Andersson & Wikström, 2001) and monazite-rich nature. Zircons in the remaining samples have medium to high Th/U ratios (> 0.1), and no associated monazite.

Twenty-three spots were analysed on grains from the felsic metavolcanic sample U97:1. The quality of the crystals is heterogeneous and consequently some have lost variable amounts of lead (Fig. 2a), and four have high common lead. A discordia fitted to 18 points (excluding one point which plots significantly to the left) gives an upper intercept of 1888 ± 12 Ma (MSWD = 5.4). Scatter in excess of analytical error is obvious around the discordia, and the concordant and near-concordant points do not all overlap within error (Fig. 2a). A weighted average of seven concordant points yields within error the same age as the discordia (1882 ± 11 Ma; MSWD = 6.2). One core of an oscillatory zoned crystal is near-concordant and considerably younger (1803 ± 4 Ma) than the rest, while two rim analyses of the same grain are strongly discordant (not plotted). U and Th contents vary strongly, as does the correlated brightness of the CL images (Fig. 3a), but Th/U ratios are relatively uniform, in the range 0.23–0.76. The crystals often show distinct oscillatory zoning (Fig. 3a).

The extrusive age of this metavolcanic rock is in the range 1880–1900 Ma, as approximated by the poorly defined upper intercept age 1888 ± 12 Ma (Fig. 2a). The age scatter among the individual crystals may reflect incorporation of xenocrysts from earlier magmatic pulses or protoliths. If so, the youngest concordant zircons at *c.* 1.87 Ga (05a 1869 ± 5 Ma) should be closest in age to the extrusion of the sample. However, no overgrowths or correlation between age, zoning, and CL-character have been identified. The single younger analysis at 1803 Ma in the core of an oscillatory zoned crystal, in which a marginal analysis falls close to the main discordia line, remains unexplained.

Nine spots were analysed in the metasupracrustal sample U97:3, of which three have high common Pb contents. The remaining analyses are all concordant and have ^{207}Pb–^{206}Pb ages in the range 1883–1905 Ma, overlapping within error (Table 2). A weighted average yields 1892 ± 7 Ma (MSWD = 2.5; Fig. 2b). Th/U ratios are in the range 0.18–0.56 and CL images are dark with some visible zonation; Fig. 3b).

This rock was earlier interpreted as a metasedimentary rock (Stålhös, 1989). However, we consider the dated rock to be metavolcanic in origin, based on the presence of amphibole, lack of muscovite, and relatively quartz-poor composition (cf. Stålhös, 1991). In spite of the limited analyses, the narrow age range accords with this interpretation. The rock may represent redeposited juvenile volcanic ejecta, where the 1892 ± 7 Ma age represents the time of eruption.

Twenty-three spots were analysed in the metasedimentary sample U97:6. Nine of these have high common lead contents (Table 2). The age scatter is significant (Fig. 2c). Three grains, close to 2.50 Ga, are considerably older than the rest. A group clusters tightly between 1898 and 1919 Ma, and six concordant to near-concordant points plot at widely scattered ages between 1930 and 2080 Ma. All these grains show concentric oscillatory zoning, with Th/U ratios varying from 0.03 up to 0.79, and U concentrations from *c.* 100 to 4400 ppm. CL and BSE images reveal that the discordant grains often are strongly metamict. One concordant point plots significantly to the left, at 1866 ± 4 Ma. This point is from the only observed dark unzoned overgrowth of a zoned core (Fig. 3c), and has a very low Th/U ratio (0.05).

All of the analysed crystals in this sample are detrital with varying provenance ages. Only one analysis in a CL-dark, unzoned, low Th/U rim, which is discordant to the oscillatory zoned (1906 ± 7 Ma) core, appears to record metamorphic growth (Fig. 3c). The 1866 ± 4 Ma age represents a maximum age for the overgrowth as an overlap with the core during analysis cannot be excluded. The range of detrital ages, from 1900 to 2080 Ma, followed by a hiatus to *c.* 2450 Ma is typical for Svecofennian metasedimentary rocks (e.g. Claesson *et al.* 1993; Lahtinen, Huhma & Kuosa, 2002; Sultan, Claesson & Plink-Björklund, 2005).

Two of the nine points measured in the amphibolite sample R89:14 have high common lead contents (Table 2), and seven analyses are concordant (Fig. 2d), with individual ^{207}Pb–^{206}Pb ages in the range 1881–1920 Ma. Two are notably older (1910 ± 7 and 1920 ± 8 Ma) than the the remining five. All these crystals are clear, and represent mainly fragments of bigger ones. CL images are generally relatively uniform and fairly dark, except one older grain with a brighter centre (Fig. 3f). Th/U ratios are high (0.68–1.15). The five youngest points overlap within error and a weighted average (Fig. 2d) gives 1887 ± 5 Ma (MSWD = 0.64).

The high Th/U ratios (up to 1.15) in this metabasite favour igneous crystallization, with no metamorphic influence. Even so there is a significant age variation among the measured crystals (Fig. 2d). This rock was interpreted by Lundström (1995) as intrusive into the felsic metavolcanic country rocks. One such felsic metavolcanic rock (ignimbrite) *c.* 10 km to the west was dated by U–Pb TIMS on zircon at 1891 ± 4 Ma (Lundström *et al.* 1998). The 1887 ± 5 Ma age, defined by the group of younger zircons, represents the time when the basic magma intruded into the volcanic pile, while the 1.91–1.92 Ga crystals are interpreted as xenocrysts.

Nine points were measured in the amphibolite sample UB98:29 (Fig. 2e; Table 2). A central group forms a linear array of six concordant and near-concordant analyses, yielding an upper intercept age of 1904 ±14 (MSWD = 0.14) and a weighted average of

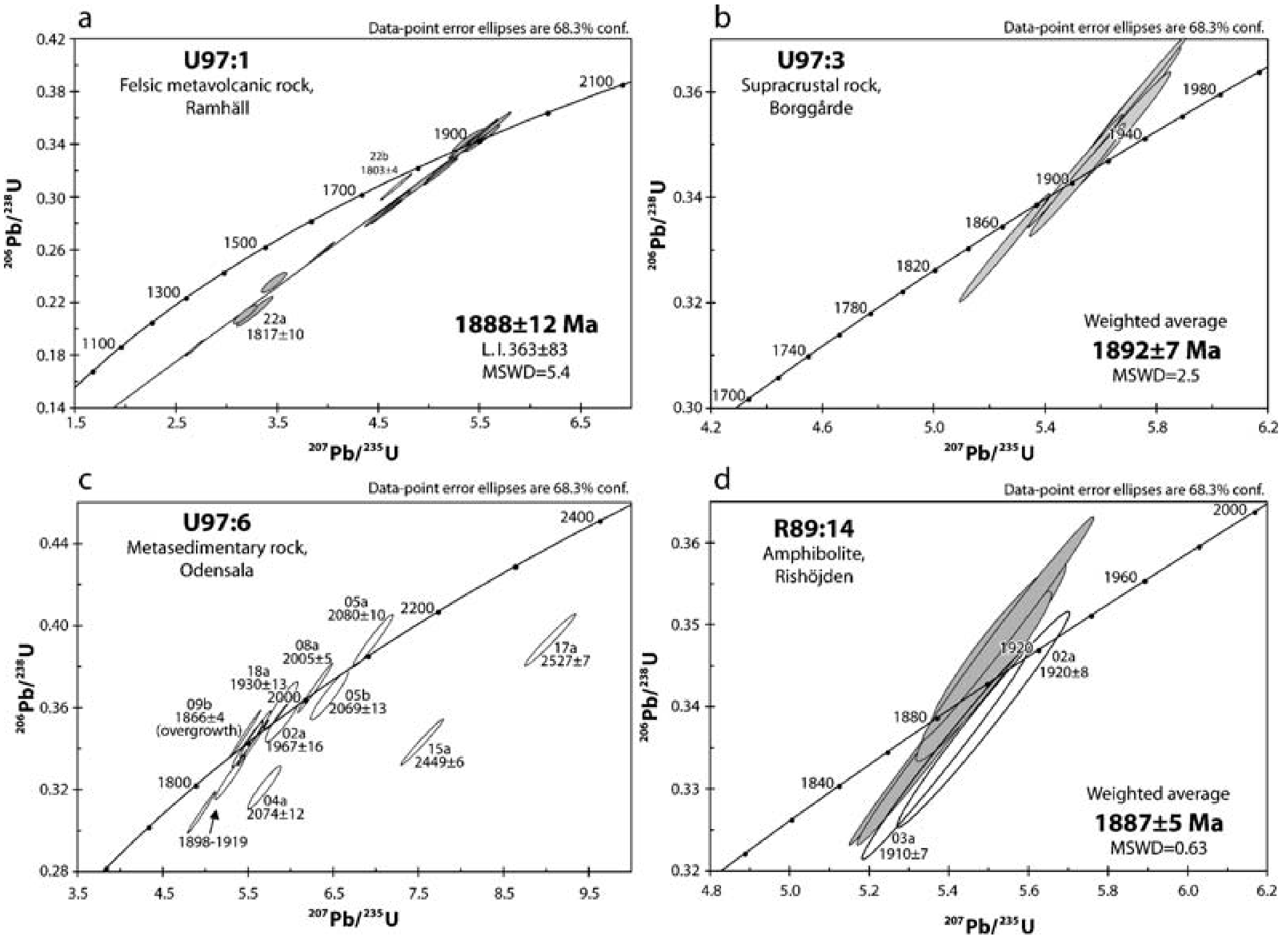

Figure 2. For legend see facing page.

1895 ± 5 Ma (Fig. 2e). One analysis lies discordantly (14 %) to the right of this group with a $^{207}Pb–^{206}Pb$ age of 1944 ± 14 Ma, while two others plot to the left with a $^{207}Pb–^{206}Pb$ average age of 1870 ± 20 Ma. Th/U ratios are all high (0.39–1.17), and the CL images are uniform and dark, except for the younger crystal which is brighter (Fig. 3h), and slightly lower in U and Th (Table 2).

The six overlapping, near-concordant points with the weighted average of 1895 ± 5 Ma (Fig. 2e) represent the intrusion age of the mafic magma. The one slightly discordant analysis plotting to the right of the others ($^{207}Pb–^{206}Pb$ age *c.* 1.94 Ga) is most probably from a grain picked up during emplacement, from pre-existing lithologies. Another crystal, with two analyses of *c.* 1.87 Ga, has lower contents of U and Th, and brighter diffuse CL-zoning than the others, but similarly high Th/U ratios (Table 2). The age is interpreted to reflect the maximum age of a metamorphic event.

Sixteen points were analysed in the augen gneiss from Finspång (Fin2), of which six have high common lead. Two spots yield Early Palaeoproterozoic ages of 2474 ± 15 and 2447 ± 6 Ma. Three others are concordant or near-concordant with $^{207}Pb–^{206}Pb$ ages in the range 2032–1967 Ma (Fig. 2f). Nine of the remaining 11 points plot together concordantly or near-concordantly within error, with a weighted average of 1852 ± 5 Ma (MSWD = 0.72), but five points have $^{206}Pb/^{204}Pb < 5000$. A discordia through these nine points and one slightly more discordant one yields an upper intercept of 1854 ± 5 Ma (MSWD = 0.35). One additional, more strongly discordant point, plots to the left of the discordia. Using only those four points with low common Pb results in a weighted average of 1855 ± 6 Ma (MSWD = 0.15; Fig. 2f). The *c.* 1.85 Ga crystals are elongate, with length/width ratios of *c.* 4 and CL images showing concentric oscillatory zoning (Fig. 3d). The analyses giving older ages are from brighter cores (lower U contents) with generally higher Th/U ratios (Table 2) in the range 0.36–0.58. The oscillatory zoned grains and margins are usually darker with low Th/U ratios (0.03–0.40), except one (0.71). Two TIMS monazite fractions yielded concordant $^{207}Pb–^{206}Pb$ ages of 1823 ± 2 and 1831 ± 2 Ma, respectively. In one additional fraction the U measurement failed, but the $^{207}Pb–^{206}Pb$ age was 1829 ± 3 Ma (Table 1; Fig. 2f).

The elongated, oscillatory zoned crystals represent magmatic crystallization at 1855 Ma. The variable, and sometimes low, Th/U ratios suggest local control

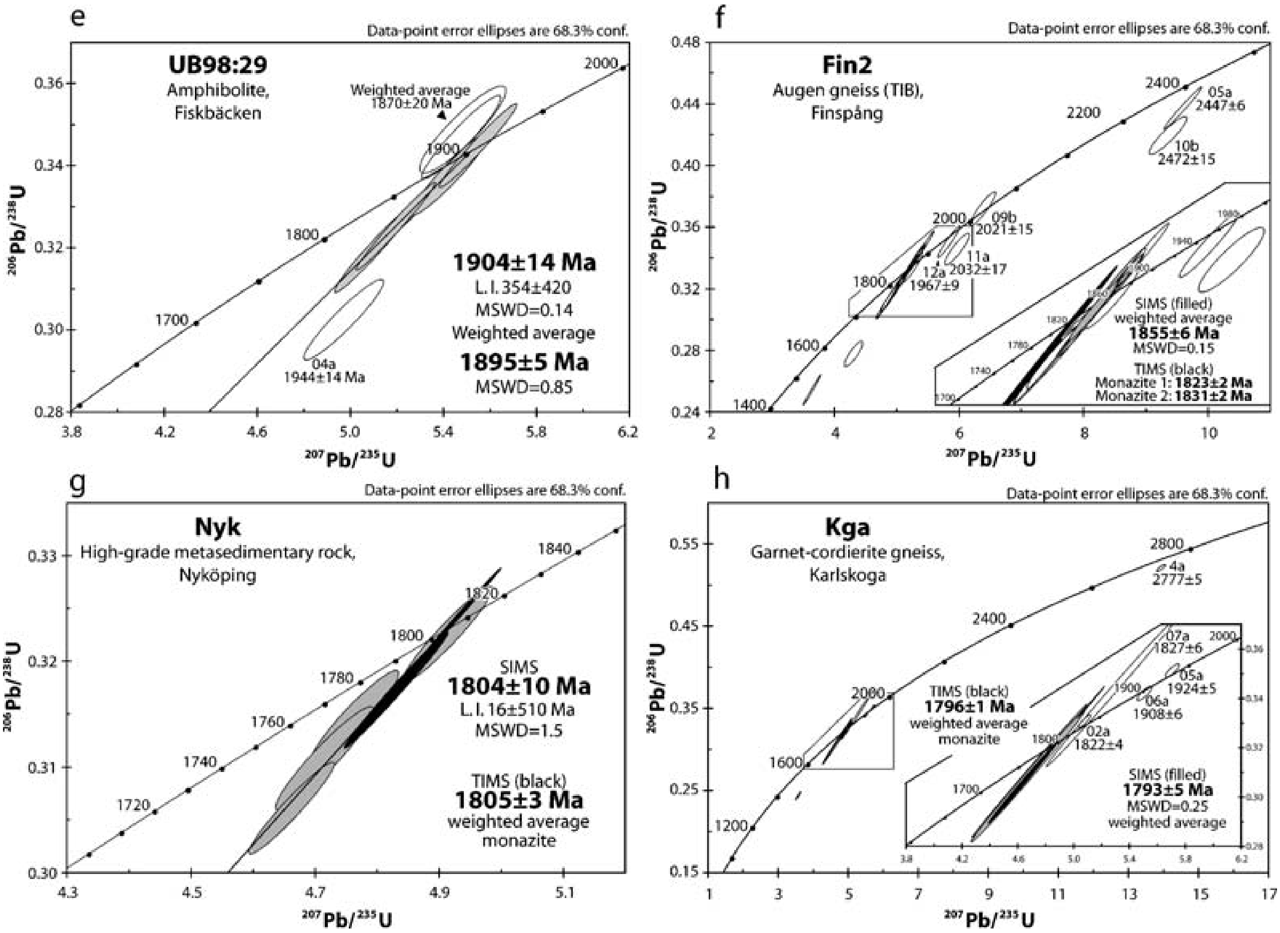

Figure 2. Concordia plots for the samples in this study. Sample locations are given in Figure 1. Analyses with high common lead ($^{206}Pb/^{204}Pb < 5000$) are not plotted or used in age calculations (except for sample Fin2; see text). In parts (g) and (h) monazite TIMS data from the same samples are plotted for comparison (Andersson, 1997b). Error ellipses and individual ages at 1σ level, except TIMS data (2σ). Discordia and averaged ages are given at 2σ. See text for discussion.

by inhomogeneous distribution of co-precipitating monazite. The age of metamorphism in this sample is not resolved, but the youngest ID-TIMS monazite age (1823 ± 2 Ma) may be close (Fig. 2f), although one monazite fraction from the adjacent high-grade metasedimentary rock yielded 1813 ± 1 Ma (Andersson, 1997b). Older monazite fractions around 1.83 Ga are thus likely to contain inherited components from igneous crystallization, or even xenocrystic material. The variable ages of the inherited zircon crystals have an age distribution similar to that in metasedimentary rocks in Bergslagen (e.g. sample U97:6; cf. also Claesson *et al.* 1993; Andersson *et al.* 2004a; Sultan, Claesson & Plink-Björklund, 2005), supporting a sedimentary contribution to the magma. This concurs with the strongly peraluminous composition of the Finspång augen gneiss (Wikström & Aaro, 1986; Andersson & Wikström, 2001) and its low initial ε_{Nd} value (*c.* -1.3; Wikström & Andersson, 2004).

Four points were measured in the metapelite from Nyköping (Nyk), two points each in the margins of two relatively large and clear grains. CL images are fairly dark with faint zoning. The ^{207}Pb–^{206}Pb ages of the almost concordant points (Fig. 2g) range between 1796 and 1808 Ma and a discordia calculated through these points yields an upper intercept of 1804 ± 10 Ma (MSWD = 1.5), while a weighted average of the ^{207}Pb–^{206}Pb ages results in 1803 ± 6 Ma. Th/U ratios are very low in these crystals (0.039–0.080), due to their low Th contents (Table 2).

The 1804 ± 10 Ma SIMS age is identical to the previously obtained monazite TIMS age of this metapelite sample (1805 ± 3 Ma; Andersson, 1997b). The consistent SIMS and TIMS results, and the fact that the zircons have very low Th/U, argues for metamorphic zircon and monazite crystallization at *c.* 1805 Ma.

Nine spots have been analysed in the centre of fairly small zircons from a granulite-grade garnet–cordierite gneiss from Karlskoga (Kga*)*. These crystals were gathered into two groups, one with somewhat larger grains (70–90 μm) deemed optically to have cores, and one with smaller clear grains (50–70 μm). The measured points in the small grains yielded the oldest ages. One grain is Archaean (^{207}Pb–^{206}Pb age of 2777 ± 5 Ma), and the other two are Palaeoproterozoic with concordant ^{207}Pb–^{206}Pb ages of 1923 ± 5 and

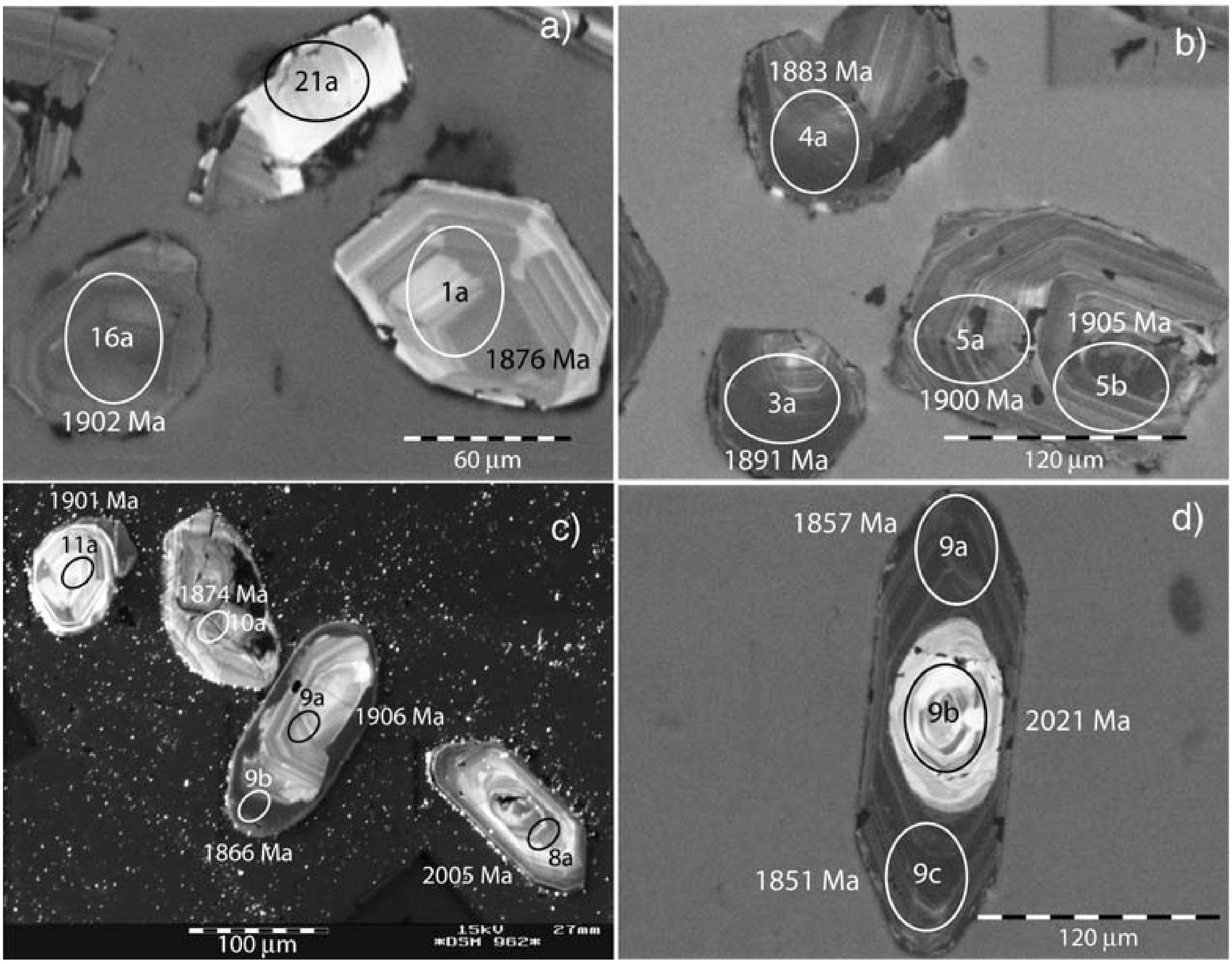

Figure 3. For legend see facing page.

1908 ± 6 Ma (Fig. 2h). Measured points in the larger grains are all younger than 1830 Ma. Three of these points have large uncertainties in their U/Pb ratios, due to high uncertainty in the standard measurements. The errors in their ^{207}Pb–^{206}Pb ages are, however, of the same magnitude as the others. The three youngest analyses have ages close to 1795 Ma with a weighted average of 1793 ± 5 Ma (MSWD = 0.25), while the two other grains have ages of 1822 ± 4 and 1827 ± 8 Ma (Fig. 2h). The oldest ages are found in CL bright crystals and cores, and CL images indicate that the *c.* 1825 Ma spots may represent mixed ages of core–margin overlaps (see below). Th/U ratios are low, below 0.1, except in the Archaean grain (0.154; Table 2).

The weighted average age of the three youngest crystals (1793 ± 5 Ma) is within error identical to the TIMS monazite age of the same rock (1796 ± 1 Ma; Andersson, 1997b). This, and the low Th/U ratios in the zircons, strongly argue in favour of metamorphic crystallization. The Archaean crystal and the two cores at 1924 and 1908 Ma (Fig. 2h) represent detritus in the sedimentary precursor to the garnet–cordierite gneiss, overgrown during the 1793 Ma gneiss-forming event. The 1924 and 1908 Ma detrital ages are considered minimum ages because some metamorphic components may have been included in the measurement of these small crystals. Similarly, the two grains giving ages of 1822 and 1827 Ma show some bright CL in their centres, material from which was probably included in the analyses. They are thus interpreted as representing mixed ages of a dominating metamorphic (low Th/U 1793 Ma component) and some older Palaeoproterozoic (probably 1900–1925 Ma) or Archaean component.

4. Discussion

The magmatic ages of the metabasic and metavolcanic rocks of this study fall in the range 1.90–1.88 Ga, i.e. within the 1.91–1.87 Ga range typical for Early Svecofennian magmatism in the Bergslagen area (cf. Lundström *et al.* 1998, and references therein).

The Finspång augen gneiss was interpreted by Wikström & Aaro (1986) and Wikström & Karis (1991) as a deformed part of the Transscandinavian Igneous Belt, and the intrusive age obtained here (1855 ± 6 Ma) verifies that this massif belongs to the

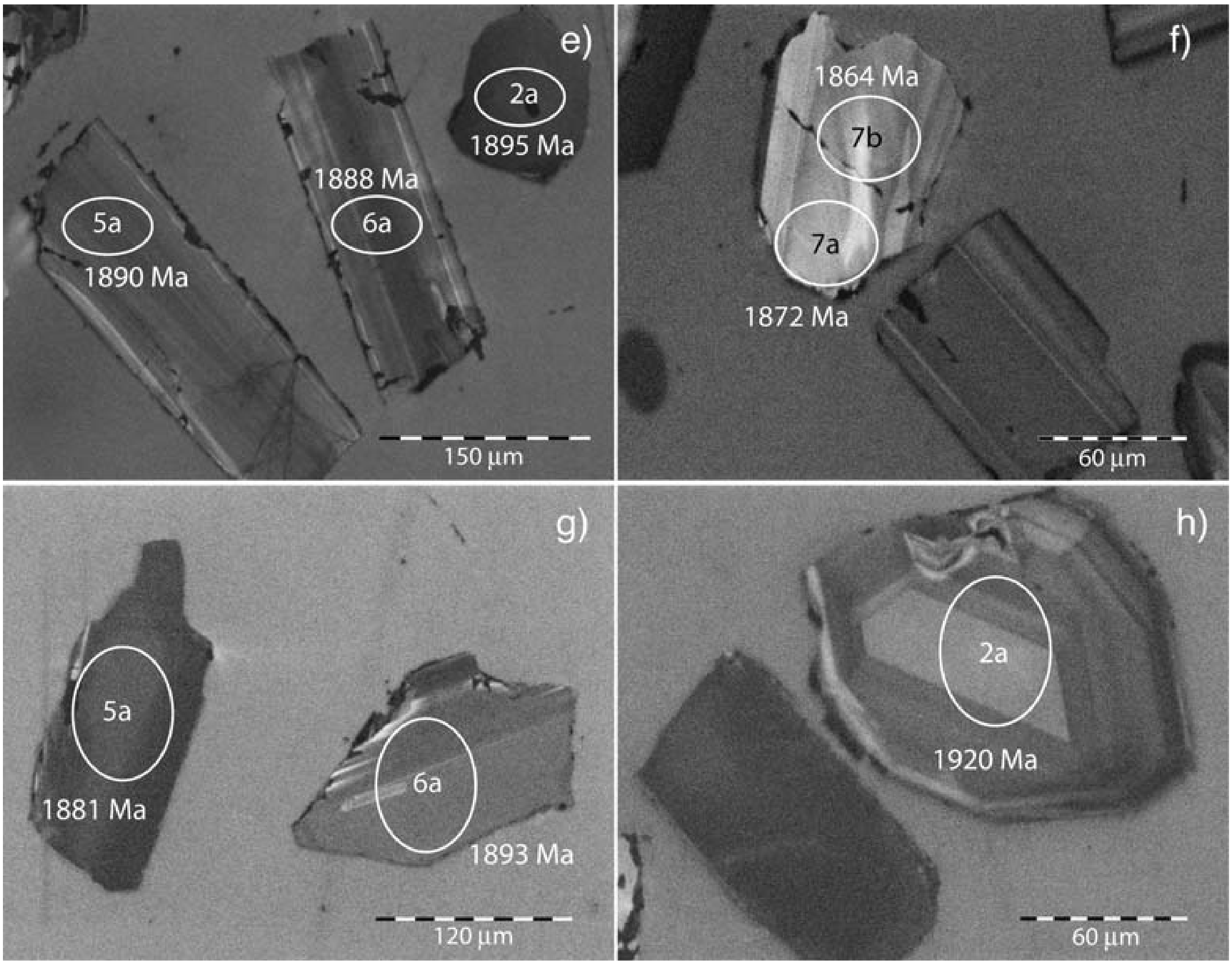

Figure 3. Cathodoluminescence pictures of selected crystals from the investigated samples. Ovals indicate analytical spots and ages are $^{207}Pb/^{206}Pb$ ages. (a) Sample U97:1: widely varying zoning pattern and U contents. (b) U97:3: typically oscillatory zoned crystals. (c) U97:6: note dark, unzoned, 1866 Ma overgrowth (9b) on bright inherited core (9a). (d) Fin2: *c.* 1.85 Ga, magmatic, oscillatory overgrowths on inherited core (poorer in U). (e, f) UB98:29: brighter crystal is younger, possibly metamorphic. (g, h) R89:14: faintly zoned crystals. One crystal with bright core zone is older.

earliest generation of Transscandinavian Igneous Belt magmatism ('TIB-0' at *c.* 1.85–1.83 Ga) in southwestern Bergslagen (Persson & Wikström, 1993; Wikström, 1996; Claeson & Andersson, 2000; Andersson & Wikström, 2004; and references therein). As a comparison, Andersson (1997b) reported an age of *c.* 1.83 Ga for the essentially undeformed adjacent Graversfors Transscandinavian Igneous Belt satellite body (Fig. 1). The Finspång gneiss is folded within the regional Late Svecokarelian structures (Wikström & Aaro, 1986; Wikström & Andersson, 2004) and high-temperature deformation was probably active as early as 1.83–1.82 Ga, as recorded by the monazites.

Two samples (U97:6 and UB98:29) indicate Early Svecofennian metamorphism in northern Bergslagen at *c.* 1.87 Ga, although these should be viewed as maximum ages. Evidence for such early metamorphism is now emerging from other localities in northern Bergslagen (Andersson *et al.* 2000, 2004a; Bergman *et al.* 2004; Hermansson *et al.* 2004, 2006).

A record of Early Svecofennian metamorphism (> 1.86 Ga) has been proposed to be present in the Skellefte district and neighbouring areas of northern Sweden (e.g. Wikström, Skiöld & Öhlander, 1996; Lundström, Persson & Bergström, 1999; Rutland, 2001; Rutland *et al.* 2001). The latter authors advocate episodes of deformation prior to 1.9 Ga, but this has been strongly contested by Weihed (2003) and Weihed *et al.* (2002). Metamorphic ages of 1860 Ma and younger have been reported for high-grade rocks from central Sweden (Sjöström & Bergman, 1998; Andersson *et al.* 2004a; Högdahl & Sjöström, 2006), and from 1845 to 1780 Ma in southern Bergslagen. The latter includes both regional metamorphism (Romer & Öhlander, 1995) and contact metamorphism related to Transscandinavian Igneous Belt rocks (Andersson, 1997b).

In Finland, however, a record of Early Svecofennian metamorphism (with the peak at *c.* 1885 Ma) exists around the Central Finland Granitoid Complex (CFGC;

Korsman *et al.* 1984, 1999; Vaasjoki & Sakko, 1988; Hölttä, 1988, 1995; Korja *et al.* 1994; Korsman, 1996; Kilpeläinen, 1998; Mouri, Korsman & Huhma, 1999). In these areas, characterized by tonalitic migmatites, there is no Late Svecofennian metamorphic overprint. A tectonometamorphic discordance has been proposed to separate the Central Finland Granitoid Complex area from the 'Potassium granite migmatite zone' (Korsman, 1996) in southern Finland, which was typically metamorphosed in Late Svecofennian times (1.85–1.81 Ga; Korsman *et al.* 1984, 1999; Vaasjoki & Sakko, 1988; van Duin, 1992; Väisänen, Mänttäri & Hölttä, 2002; Väisänen *et al.* 2004; Ehlers, Skiöld & Vaasjoki, 2004), contemporaneous with southern Bergslagen. Recently, however, Rutland, Williams & Korsman (2004) have, based on structural and chronological data, proposed that the metasediments of the Vammala migmatite belt of southern Finland were deposited prior to 1.92 Ga when they experienced a major episode of metamorphism, as evidenced by metamorphic zircon and monazite growth.

Inherited zircon crystals in both metasedimentary and metaigneous rocks span the age range 2.78 to 1.90 Ga. A majority of these crystals are 2.08–1.91 Ga old, with a minor, older group at 2.78–2.45 Ga. No magmatic rocks from either group have been found in the Bergslagen area. The lack of age data in the interval 2.5–2.1 Ga is a typical feature of detrital zircon populations in Svecofennian metasedimentary rocks (deposited $\geq$ 1.87 Ga) (Claesson *et al.* 1993; Lahtinen, Huhma & Kousa, 2002; Rutland, Williams & Korsman, 2004; Sultan, Claesson & Plink-Björklund, 2005; Bergman *et al.* 2006). Even though these studies do not fulfil the statistical and other criteria (cf. e.g. Vermeesch, 2004; Andersen, 2005) to reject the presence of protosources in the age range 2.5–2.1 Ga in the provenance areas, such sources in any case were of limited importance during erosion and Svecofennian sedimentation.

Magmatic rocks with ages of 1.95–1.92 Ga have been reported from the Bothnian basin and further north in Sweden (e.g. Wasström, 1993, 1996; Skiöld *et al.* 1993; Lundqvist, Vaasjoki & Persson, 1998). Welin, Christiansson & Kähr (1993) reported an age of 2.03 Ga for a garnet-bearing granite in central Sweden, but this age has been questioned based on the rock´s S-type chemistry and heterogenous, probably xenocrystic, zircon population (Lundqvist, Vaasjoki & Persson, 1998). In the 'Primitive Arc Complex' of central Finland 1.93–1.91 Ga old granitoids represent the oldest Svecofennian magmatism (Helovuori, 1979; Korsman *et al.* 1984; Vaasjoki & Sakko, 1988; Vaasjoki *et al.* 2003; and references therein), with no zircons older than 1.95 Ga (Lahtinen & Huhma, 1997; Vaasjoki *et al.* 2003). 1.95–1.91 Ga magmatic rocks were available for erosion during sedimentation at *c.* 1.89–1.87 Ga and at least part of the detrital material with such ages may be derived from source areas within the shield. In contrast, felsic igneous rocks in the age range 2.1–1.95 Ga are virtually unknown in the shield, and source areas for this detritus may have to be sought, for example to the SE in Sarmatia (Lahtinen, Huhma & Kousa, 2002).

The Archaean detritus common in metasedimentary rocks in Bergslagen and other areas of the Svecofennian Domain (Huhma *et al.* 1991; Claesson *et al.* 1993; Lahtinen, Huhma & Kousa, 2002; Andersson *et al.* 2004a; Rutland, Williams & Korsman, 2004; Sultan, Claesson & Plink-Björklund, 2005) is generally confirmed by numerous determinations of low ε_{Nd} values (e.g. Miller *et al.* 1986; Claesson, 1987; Huhma, 1987; Patchett, Todt & Gorbatschev, 1987; Valbracht, 1991; Claesson & Lundqvist, 1995; Kumpulainen *et al.* 1996; Lahtinen & Huhma, 1997; Andersson, 1997a). This combined evidence points to about 30–40 % Archaean material. Natural source areas for this detritus are found in the 3.2–2.5 Ga cratonic nucleus in the NE part of the shield (Koistinen *et al.* 2001), although the pre-3.2 Ga components have not yet been identified in outcrop. Continental areas in the west are possible suppliers of detritus to > 1.87 Ga Svecofennian sedimentation, e.g. the 3.9–2.5 Ga south Greenland craton (e.g. Nutman *et al.* 1996, 2004, and references therein), as indicated by recent Palaeoproterozoic supercontinent reconstructions (e.g. Karlstrom *et al.* 2001; Zhao *et al.* 2002, 2004, and references therein).

By contrast, in magmatic rocks of Bergslagen only minor older inherited material has so far been documented. This is supported by their relatively high ε_{Nd} ratios giving no evidence for Archaean basement material in the Bergslagen area (Patchett, Todt & Gorbatschev, 1987; Valbracht, 1991; Kumpulainen *et al.* 1996; Andersson, 1997a). The few Archaean zircons reported from magmatic rocks in Bergslagen may derive from assimilation of associated or buried sedimentary rocks (as in the case of the Finspång augen gneiss) that obtained this component by surface transport from other regions.

It is particularly significant that 1.94–1.91 Ga crystals occur in the *c.* 1.89 Ga metabasic rocks studied here. This indicates that the basic magmas, en route to their emplacement levels, passed through 1.94–1.91 Ga lithologies, suggesting that such rocks exist at depth below northern Bergslagen, although these are not yet recognized on the surface. A possible source is outlined by the recent interpretation of BABEL reflection seismic profiles, suggesting a Palaeoproterozoic lower crustal nucleus below Bergslagen (Korja & Heikkinen, 2005). Proto-Svecofennian zircons have been found in younger magmatic rocks elsewhere in the Svecofennian Domain. Kumpulainen *et al*. (1996) reported a 1.93 Ga old fraction in a *c.* 1.89 Ga rhyolite from Godegård in southern Bergslagen. Lundqvist, Vaasjoki & Persson (1998) found *c.* 1.93 and 1.89 Ga zircon populations from a granodiorite in the Bothnian basin, and Andersson *et al*. (2004a) reported pre- and

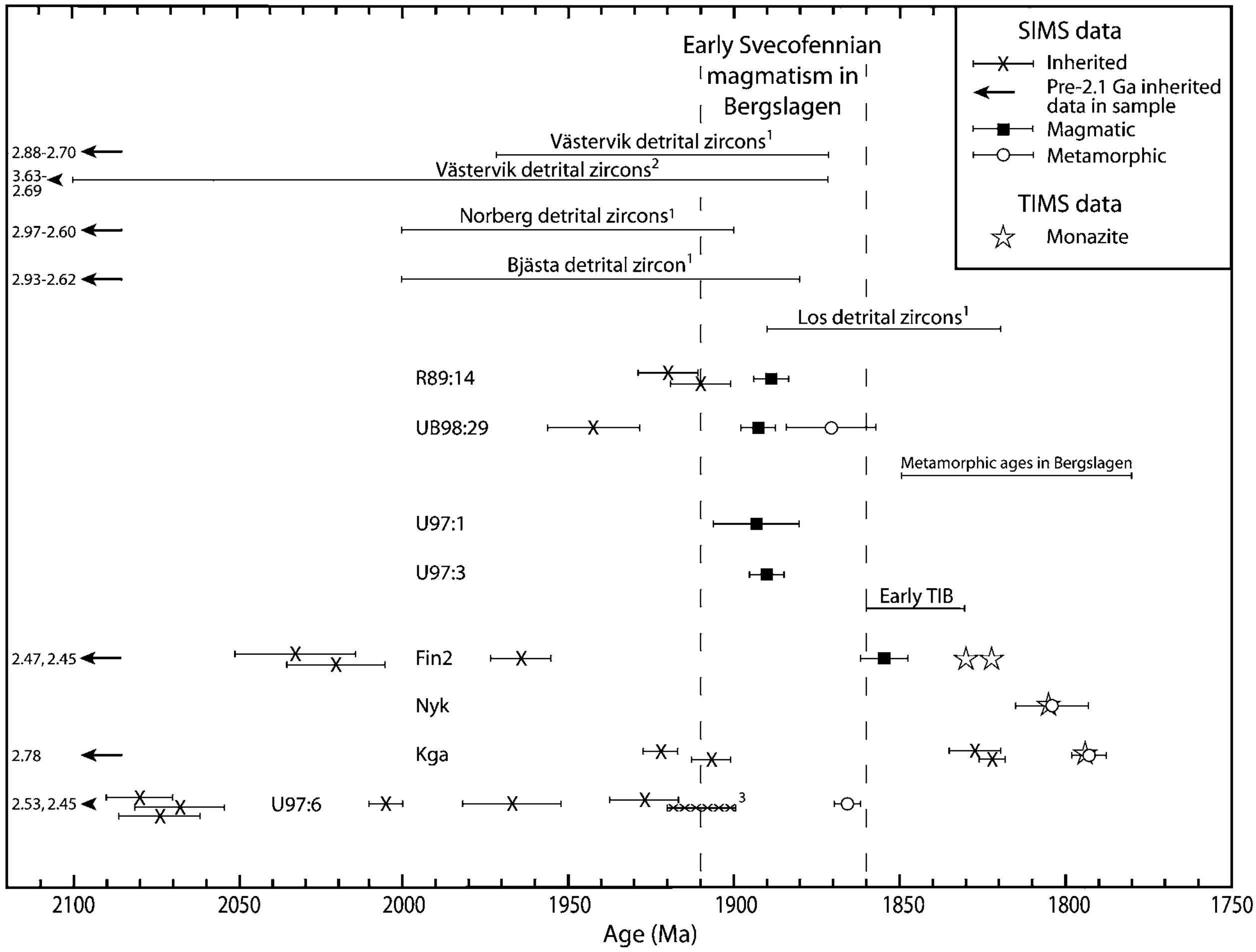

Figure 4. Summary of age data obtained in this study. For comparison, age ranges of Early Svecofennian magmatism and metamorphism in Bergslagen, and Early Transscandinavian Igneous Belt (TIB) plutonism along the southern and western Bergslagen margin are given (as summarized in Andersson *et al.* 2004b, and references in the text). In addition, age ranges from previously published studies on detrital zircons in other Svecofennian metasediments in Bergslagen and its nearest surroundings are shown (1: Claesson *et al.* 1993; 2: Sultan, Claesson & Plink-Björklund, 2005). 3: Age range of the group of youngest detrital zircons in U97:6 (1919–1898 Ma); cf. Fig. 2c).

post-1.90 Ga zircons in a charnockite from central Sweden. Likewise, Vaasjoki *et al.* (2003) report pre- and post-1.90 Ga zircon groups in Early Svecofennian granitoids of the Primitive Arc Complex of central Finland, while Ehlers, Skiöld & Vaasjoki (2004) report inherited zircons in the range 2.08–2.00 Ga in *c.* 1.88 Ga granodiorites of SW Finland.

The present data suggest a two-(multi?-)stage development of the Early Svecofennian crust in Bergslagen, and support earlier inferences about pre-1.90 Ga continental crust. These inferences suggest a pre-1.90 Ga formation of island arc–continental crust during oceanic subduction, followed by renewed subduction, accompanied by extension and rifting of the juvenile crust in the overriding plate during late Early Svecofennian times (1.90–1.86 Ga; e.g. Oen *et al.* 1982; Baker, Hellingwerf & Oen, 1988; Lagerblad, 1988; Gaál, 1990; Valbracht, 1991; Allen *et al.* 1996), accompanied by typically bimodal magmatism. Similarly, Lahtinen (1994) and Lahtinen & Huhma (1997) presented evidence and models for pre-1.90 (in fact 2.1–2.0) Ga continental crustal formation in both the central and southern Finland Svecofennian terranes (supported by the recent data of Ehlers, Skiöld & Vaasjoki, 2004). This was further developed into a model of Palaeoproterozoic microcontinent formation and accretion (2.1–1.88 Ga), followed by extension and collision in the framework of five orogenies for the whole Svecofennian Domain of the shield (Nironen, Lahtinen & Korja, 2002; Lahtinen, Korja & Nironen, 2004). However, to constrain crust formation, assembly, and orogenic evolution of the Svecofennian Domain in detail considerably more data are required, not least from the Bergslagen area.

5. Conclusions

Results from the present study are summarized in Figure 4 (excluding Archaean zircons).

(1) Felsic volcanic extrusions (1888 ± 12 and 1892 ± 7 Ma) are contemporaneous with

emplacement of mafic plutonic rocks (1887 ± 5, and 1895 ± 5 Ma) in northern Bergslagen, and fall in the same general range (1.91–1.87 Ga) as previously known magmatic ages from this province.

(2) Early Svecofennian metamorphism (at *c.* ≤ 1.87 Ga) is indicated by direct geochronology in northern Bergslagen for the first time.

(3) Inherited zircon crystals with ages of 2.78–1.90 Ga, and particularly crystals in metaigneous rocks in the age range 1.94–1.91 Ga, support earlier inferences of the existence of pre-1.91 Ga crust in the Bergslagen region. In contrast, no evidence for Archaean crust in this region was found. The previously observed age gap between *c.* 2.45 and 2.1 Ga for detrital zircons in the Svecofennian Domain is further emphasized by the present data.

(4) The Finspång augen gneiss in southern Bergslagen belongs to the earliest generation of Transscandinavian Igneous Belt intrusives present along the southern margin of the Svecofennian Domain; it intruded at 1855 ± 6 Ma, and experienced high-temperature deformation at least as late as 1.82 Ga.

Acknowledgements. Hans Schöberg, Per-Olof Persson, Paula Allart, and Marina Fischerström are thanked for technical and analytical advice during the ID-TIMS U–Pb work, while Kerstin Lindén, Torbjörn Sunde, Jessica Vestin, and Martin Whitehouse assisted skilfully during SIMS work. The NordSIMS facility in Stockholm is financed and operated under an agreement between the research councils of Denmark, Norway and Sweden, the Geological Survey of Finland, and the Swedish Museum of Natural History. Ursula Glenz, GeoForschungsZentrum, Potsdam, assisted with CL images of sample U97:6. Samples R89:14 and Fin2 were kindly supplied by Ingmar Lundström and Tore Qvarnström, respectively. This is NordSIMS contribution number 145.

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Appendix: sample descriptions

Sample U97:1

U97:1 is a felsic metavolcanic rock collected at Ramhäll, northeastern Bergslagen (Fig. 1), in a steep E–W-striking belt of volcanic rocks, located in Early Svecofennian granitoids (Stålhös, 1987). Close to the sampling site, the volcanic successions include a major marble lens and associated iron ores. The E–W belt shows a very consistent planar fabric and has recently been interpreted as a shear zone (Persson & Sjöström, 2003). Petrographically, the sample is fine-grained and contains the major phases quartz–K-feldspar–plagioclase–amphibole–biotite.

Sample U97:3

U97:3 was collected at Borggårde, eastern Bergslagen, from a small occurrence interpreted to be of metasedimentary origin, intercalated with felsic metavolcanic rocks and surrounded by Early Svecofennian intrusive rocks (Stålhös, 1989). It comprises quartz–plagioclase–biotite–amphibole–garnet, with zircon as a scarce accessory phase.

Sample U97.6

U97:6 is a lower amphibolite facies metasedimentary rock (Stålhös, 1972), sampled close to Odensala church, eastern Bergslagen. The major assemblage is quartz–plagioclase–biotite–amphibole–garnet. This area is cut by pegmatites including the rare-mineral-bearing lithium–caesium–tantalum type (Romer & Smeds, 1994; Öhlander & Romer, 1996).

Sample R89:14

Sample R89:14 is an amphibolite located at Rishöjden, in the low-grade area of western Bergslagen (Lundström, 1995). The sampled rock belongs to a group of basic intrusions which form horizontal sheet-like bodies cutting the vertical structures of surrounding metarhyolites (Lundström, 1995). The major assemblage is plagioclase (strongly altered) – amphibole (primary, brownish green, and secondary, pale green)–muscovite (secondary)–opaque.

Sample UB98:29

The amphibolite sample UB98:29 was collected at Fiskbäcken, northwestern Bergslagen. The relatively coarse-grained texture of this rock indicates an intrusive origin, and it is classified as a metagabbro (Strömberg, 1996). The two dominating minerals are amphibole and plagioclase, with the latter strongly altered to saussurite and sericite.

Sample Fin2

Sample Fin2 was collected in Finspång town, southern Bergslagen from an augen gneiss unit (Wikström & Karis, 1991) of the peraluminous Finspång massif (Wikström & Aaro, 1986). It is interpreted as a deformed Transscandinavian Igneous Belt granitoid (Wikström, 1984; Wikström & Karis, 1991; Andersson & Wikström, 2001; Wikström & Andersson, 2004). The major assemblage of the sample is K-feldspar (megacrysts and groundmass)–plagioclase–quartz–biotite. Both zircon and monazite are abundant in the rock.

Sample Nyk

Nyk is a pelitic gneiss regionally metamorphosed to upper amphibolite facies (garnet–cordierite–sillimanite–plagio-clase–K-feldspar–quartz–biotite) (Lundström, 1975). The sample was collected close to Nyköping in the Sörmland metasedimentary basin, southeastern Bergslagen.

Sample Kga

Kga is a granulite facies garnet–cordierite gneiss (garnet–cordierite–K-feldspar–plagioclase–biotite–quartz) (cf. Stephens, 1998) from Karlskoga, western Bergslagen, that was subjected to contact metamorphism due to Transscandinavian Igneous Belt intrusions (Andersson, Larsson & Wikström, 1992). Both Nyk and Kga samples have previously been analysed by U–Pb monazite TIMS geochronology (Andersson, 1997b).