Bulletin of the Geological Society of Finland, Vol. 80, 2008, pp. 73–87

U-Th-Pb zircon geochronology on igneous rocks in the Toija and Salittu Formations, Orijärvi area, southwestern Finland: Constraints on the age of volcanism and metamorphism

Markku Väisänen[1)]* and Christopher L. Kirkland[2)]
[1)] *Department of Geology, FI-20014 University of Turku, Finland*
[2)] *Swedish Museum of Natural History, SE-104 05 Stockholm, Sweden*

Abstract
Zircons from a felsic volcanic rock in the Toija Formation and a synvolcanic gabbro intrusion in the Salittu Formation within the Orijärvi area were dated by U-Th-Pb SIMS in order to provide depositional constraints on these formations. Zircon crystals from the felsic rock preserve a two-stage crystallisation history with zoned core domains and homogeneous rim domains. Inner domains yield a 1878 ± 4 Ma concordia age, interpreted to determine the crystallisation of this rock. Rims yield a 1815 ± 3 Ma concordia age interpreted to determine the regional metamorphism. Small rounded zircon grains from the Salittu gabbro, located within the Jyly shear zone, yield a concordia age of 1792 ± 5 Ma. We interpret the grain textures to suggest that they recrystallised from inherited zircon seeds during the heat and fluid flow into the shear zone. Although no direct ages for the Salittu Formation have been recovered, field relationships imply that it was deposited between 1878 – 1875 Ma.

Key words: metavolcanic rocks, gabbros, volcanism, volcanic arcs, metamorphism, absolute age, U/Th/Pb, zircon, Proterozoic, Paleoproterozoic, Toija, Orijärvi, Salittu, Suomusjärvi, Finland

*Corresponding author, email: markku.vaisanen@utu.fi

1. Introduction

The Orijärvi area (Fig. 1) can be regarded as a key study area for understanding the crustal growth of the Svecofennian Orogen in southern Finland because the area consists of well-preserved volcanic rocks metamorphosed at lower temperatures compared to its migmatitic surroundings (Eskola, 1914; Ploegsma & Westra, 1990; Skyttä et al., 2005; Skyttä et al., 2006). Väisänen & Mänttäri (2002) performed geological mapping of the area and divided it into four formations, which from oldest to youngest were named, the Orijärvi, Kisko, Toija and Salittu Formations. According to their model, the Orijärvi and Kisko Formations represent growth of the volcanic arc whereas the Toija and Salittu Formations represent rifting of the arc. This scenario was based on field relationships and geochemical compositions of the volcanic rocks and has only partly been verified by absolute ages since only volcanic rocks of the Orijärvi and Kisko Formations have been dated by U-Pb on zircon (1895.3 ± 2.4 Ma and 1878.2 ± 3.4 Ma, respec-

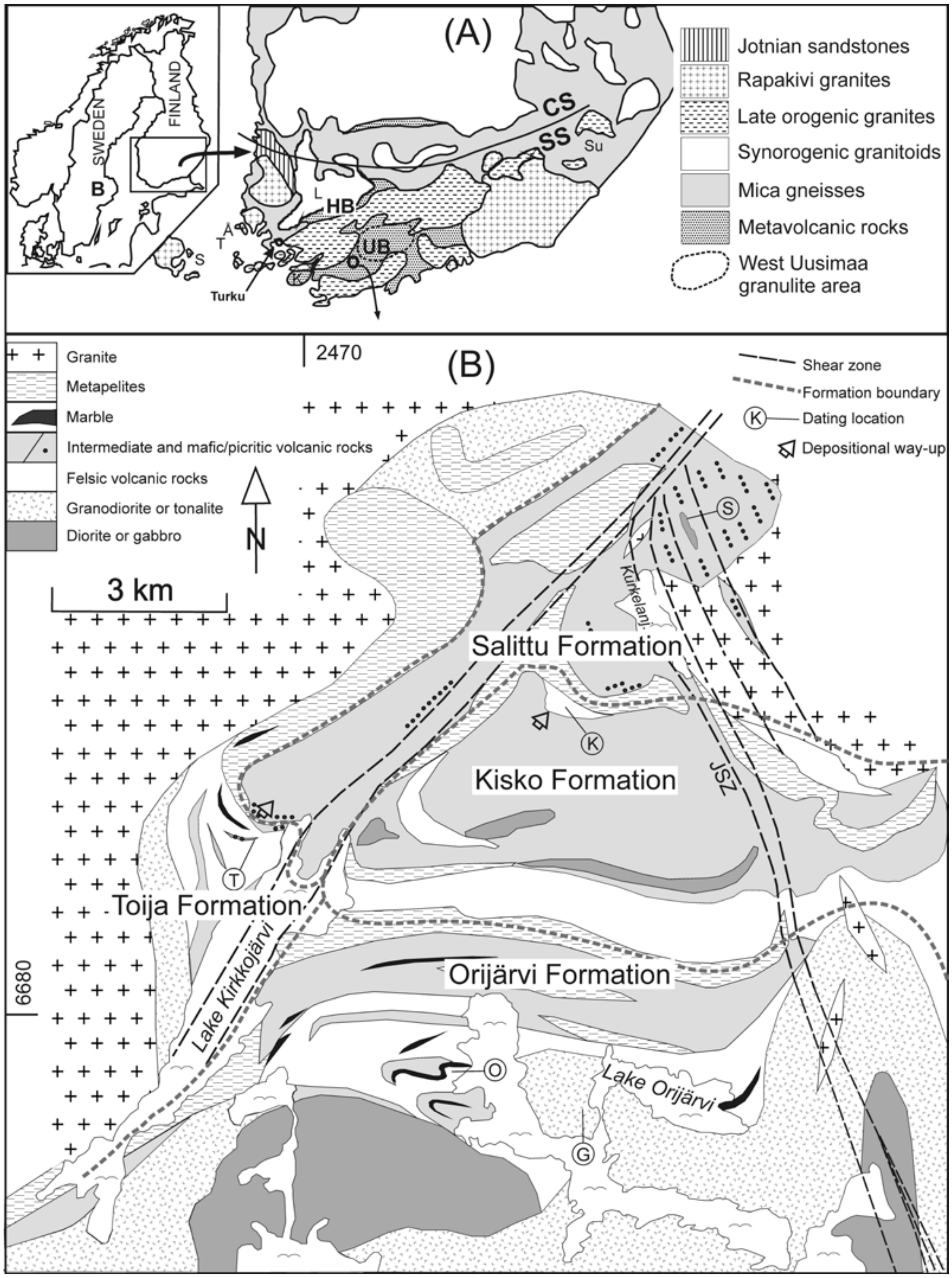

Fig. 1. (A) Main geological units, modified after Korsman et al. (1997). B = Bergslagen; CS = central Svecofennia; HB = Häme belt; K = Kemiö; L = Loimaa; O = Orijärvi; S = Sottunga; SS = southern Svecofennia; Su = Sulkava; T = Torsholma; UB = Uusimaa belt; Å = Åva. (B) Geological map of the Orijärvi area, modified after Skyttä et al. (2006). JSZ = Jyly shear zone; Kurkelanj. = Kurkelanjärvi. Previous and present dating locations: G = Orijärvi granodiorite (Huhma, 1986; Väisänen et al., 2002); K = Kisko dacite (Väisänen & Mänttäri, 2002); O = Orijärvi rhyolite (Väisänen & Mänttäri, 2002); S = Salittu gabbro (this study); T = Toija felsic volcanite (this study).

tively). In addition, zircon crystals from the synvolcanic Orijärvi granodiorite yielded an age of c. 1900 Ma (1891 ± 13 Ma in Huhma, 1986; 1898 ± 9 Ma in Väisänen et al., 2002).

The Toija and Salittu Formations remain, however, undated and therefore their stratigraphic positions are unverified. Of particular interest is the age of the Salittu Formation as it consists of EMORB-type tholeiitic basalts and picrites. Kähkönen (2005) pointed out that the Salittu Formation could be in tectonic contact with the other members of the stratigraphy and therefore predate them. This is possible as the area is known to contain several generations of shear zones that may have juxtaposed rocks of different ages (Ploegsma, 1989; Skyttä et al., 2006; Väisänen & Skyttä, 2007). The tectonostratigraphic position of the Salittu Formation is evidently of crucial importance in the tectonic modelling of the Svecofennian Orogen in this region. The interested reader may find additional details on the local geology in Väisänen & Mänttäri (2002).

In order to test the proposed tectonostratigraphy of Väisänen & Mänttäri (2002), we sampled extrusive volcanic and intrusive synvolcanic rocks from the Toija and Salittu Formations and separated zircons from them. The zircons were dated by the ion microprobe (SIMS) at the Nordsim laboratory, Swedish Museum of Natural History.

2. Geological setting

The bedrock in the Orijärvi area belongs to the Uusimaa belt, a volcanic belt in southern Finland that mainly consists of bimodal volcanics and sediments. Plutonic rocks of synvolcanic, synorogenic, late-orogenic, post-orogenic and anorogenic stages are common (e.g. Nironen, 2005). The Uusimaa belt and the Häme belt, which is another volcano-sedimentary sequence to the north of the Uusimaa belt, together form a terrane called southern Svecofennia (Kähkönen, 2005), although Korja et al. (2006) regard the Uusimaa and Häme belts as separate (suspect) terranes. All existing age determinations of volcanic and synvolcanic plutonic rocks within southern Svecofennia are bracketed between ~ 1.90 – 1.88 Ga (Patchett & Kouvo, 1986; Vaasjoki, 1994; Reinikainen, 2001; Väisänen & Mänttäri, 2002; Väisänen et al., 2002; Ehlers et al., 2004; Skyttä et al., 2005, Torvela et al., 2008). It is proposed that southern Svecofennia is separated from the central Svecofennia terrane, lying to the north, by a suture zone (Lahtinen, 1996; Rämö et al., 2001; Fig. 1).

The tectonic setting of volcanism in southern Svecofennia and its proposed extension in the Bergslagen area in south-central Sweden is still controversial. In Sweden, the tectonic setting is well-constrained to a back-arc basin (e.g. Allen et al., 1996; Kumpulainen et al., 1996) but in Finland, volcanic arc, back-arc rifting and rifting of older continental margin settings have been proposed (e.g. Lindroos & Ehlers, 1994; Kähkönen, 2005; Weihed et al., 2005). A combination of these settings was proposed by Väisänen & Mänttäri (2002) and Väisänen & Westerlund (2007). Older underlying continental crust below these units is supported by initial ε_{Nd} values close to zero in some magmatic rocks both in Sweden (Valbracht et al., 1994) and in Finland (Huhma, 1986; Lahtinen & Huhma, 1997). In addition, recent ion microprobe zircon results include Archaean and 2.1 – 1.92 Ga inherited zircons in younger intrusive rocks, indicating crustal sources of these ages (Väisänen et al., 2002; Ehlers et al., 2004; Andersson et al., 2006b).

After volcanism, plate tectonic movements caused the collision of southern Svecofennia with central Svecofennia at 1.88 – 1.86 Ga followed by an extensional period at 1.86 – 1.84 Ga. A new collisional event took place after c. 1.84 Ga followed again by several extensional periods (Lahtinen et al., 2005). Many of these events were associated with magmatism of both mantle and crustal origin. The original volcanic and volcano-sedimentary formations were subject to multistage deformation and metamorphism, resulting in a complex mosaic of the Svecofennian Orogen (e.g. Ehlers et al., 1993; Nironen, 1997; Korsman et al., 1999; Korja & Heikkinen, 2005; Lahtinen et al., 2005). Because of deformation partitioning, part of the Orijärvi area escaped the most severe overprinting tectonometamorphism and, therefore, makes an excellent locality to study the early Svecofennian crustal growth in southern Finland (Ploegsma & Westra, 1990; Skyttä et al., 2006). However, due to the complexity of the orogenic processes, different parts of the area are now in dissimilar tectonostratigraphic positions. The Orijärvi and Kisko Formations are non-migmatitic with co-existing andalusite and cordierite in pelitic rocks (Eskola, 1914; Ploegsma & Westra, 1990). The Toija Formation is located on the western side of the prominent Kisko shear zone. K-feldspar and sillimanite co-exist in pelites and leucosomes are also present, implying that metamorphic temperatures were high enough for melting in this formation. The rocks of the Salittu Formation occur principally to the north of the Toija Formation, on the western side of the Kisko shear zone and on the eastern side of the Jyly shear zone. The Salittu Formation contains pelitic rocks which are strongly migmatitic granulite facies garnet-cordierite gneisses. A small piece of the Salittu Formation also exists above the Kisko Formation, within a low-grade domain (Fig. 1b).

3. Samples

Two samples were collected for dating purposes, one from the Toija Formation and another from the Salittu Formation.

3.1. Toija Formation (felsic volcanic rock: 26MJV06)

The Toija Formation consists of bimodal volcanic rocks, generally rhyolites and pillow-basalts with intercalations of marbles and pelitic sedimentary rocks. At the inferred upper part of the formation, a few hundred metres from the contact with the overlying Salittu Formation, picritic rocks occur together with felsic volcanic rocks. The sample for geochronology is collected from a layered felsic volcanic rock. Geochemical analysis of a similar sample collected from the same outcrop is published in Väisänen & Mänttäri (2002, sample TOS-2). Proximal to the sample location a picritic rock outcrops (sample TOM-1 in Väisänen & Mänttäri, 2002). The contact between the felsic and ultramafic rock is unexposed.

3.2. Salittu Formation (gabbro: 19MJV05)

The Salittu Formation consists mainly of tholeiitic pillow-basalts and picritic volcanic rocks. Sedimentary units include pelites, some of which are graphite-bearing and intercalated marbles. The formation also contains gabbroic rocks and occasional mafic dykes crosscutting the picritic lavas (Väisänen & Mänttäri, 2002). There is a complete absence of felsic volcanic rocks within the Salittu Formation. A gabbro which has a chemical composition very similar to that of the volcanic rocks from the Salittu Formation (data in Väisänen & Mänttäri, 2002) was selected for geochronology. Using the Th-Hf-Ta diagram of Wood (1980), the gabbro plots in the EMORB-field. Basaltic and picritic volcanic rocks also from the same formation plot in the same field. N-MORB normalised multielement diagram display no negative Nb-Ta anomalies as would be expected for more evolved magmas. These features, among with other geochemical similarities, suggest that the gabbro is a synvolcanic intrusion. Synvolcanic plutonism is common in southern Svecofennia in Finland (Väisänen et al., 2002; Ehlers et al., 2004; Peltonen, 2005) and in Bergslagen in Sweden (Allen et al., 1996; Andersson et al., 2006b). The sampled gabbro is a deformed and elongated, sill-like intrusion that is surrounded by sulphide- and graphite-bearing migmatitic mica gneisses, picrites and basalts. They are folded but not intensively sheared. Granites and pegmatites are also common. The locality is, in broad sense, within the N-S oriented Jyly shear zone system and, in detail, between the Jyly II and III fault zone strands of Väisänen & Skyttä (2007). Geochemical analysis of the gabbro is presented in the Appendix.

4. Analytical methods

Zircons were extracted from the samples by crushing, sieving, shaking table, hand magnet, Franz isodynamic separator, heavy liquid (methylene iodide) and hand picking. About 100 grains per sample were mounted in epoxy resin and polished to approximately half grain thickness to reveal their interiors. Grains were imaged by BSE with a Cambridge Instruments Stereoscan 360 electron microscope at Åbo Akademi University, Turku, Finland to determine their internal structure. Measurement of Pb/U, Pb/Pb isotope ratios and U, Th, and Pb concentrations were performed on a Cameca IMS 1270 ion microprobe at the Swedish Museum of Natural History following methods described by Whitehouse et al. (1999) and Whitehouse & Kamber (2005). A flat-bottomed spot with homogeneous primary beam density and a diameter less than 20 µm was achieved using a 150-µm aperture inserted into the defocused primary beam. The resulting primary beam ion current density (O_2^-) was typically c. 4 nA. Sample changing effects were minimised by optimising the energy offset to maximum transmission in the ±15 eV energy window at the start of each automated analysis. Automated beam centring in the field aperture and mass calibration, of the isotopic peaks, was also performed prior to data collection. Measurements were taken over 12 cycles through the respective mass stations measured in peak skipping mode. U/Pb ratio calibration was based on analyses of the standard zircon 91500 (Wiedenbeck et al., 1995) and used a best-fit power law relationship between U_2O/U versus U/Pb. Common Pb correction is assumed to represent present day surface contamination and is effectively corrected using the terrestrial average Pb-isotopic composition of Stacey and Kramers (1975) i.e. $^{207}Pb/^{206}Pb$=0.83, which has been shown as appropriate for this laboratory (Kirkland et al., 2008).

5. Results

5.1. Toija Formation (felsic volcanic rock: 26MJV06)

The zircon crystals separated from sample 26MJV06 are generally euhedral but heterogeneous in size and colour. Small stubby crystals (40 – 100 µm long) and elongate prismatic (100 – 300 µm long) grains, which vary between transparent to turbid, occur (Fig. 2a-2d). A few small rounded zircons are also present. BSE-images indicate that the zircons experienced two growth stages. The core domains usually reveal well-developed (growth) zoning (Fig 2a), which is especially prevalent among the stubby crystals. Many zircons reveal a thin, homogeneous, light coloured rims. Some rims were of sufficient size to site an ion microprobe analysis (Fig. 2b-2c). In a few cases, the rim domains penetrate into the interior of the crystals along transgressive zones that crosscut the oscillatory zoning in the cores thus leaving only patches of the original zircon (Fig. 2b). The rim domains are often metamict, in contrast to the inner domains that only show alteration features along cracks. Fractures are widespread in most crystals making it difficult to find a suitable homogeneous area for analysis.

Forty-one spots on thirty-three zircon crystals were analysed (Table 1). The majority of spots were aimed at zoned prismatic zircons and the inner domains of apparently multiple growth stage crystals. A small number of analyses were carried out on the rim regions. The analysed data show a wide scatter of ages, many of which are discordant and/or with large errors (Fig. 3a). This is due to the prevalence of fractures: many spots overlapped cracks, metamict domains and some are partly on the mounting epoxy, while trying to analyse the narrow rim regions. Furthermore, some analyses are inadvertent mixtures. Those analytical positions on fractures, metamict domains and touching epoxy have increased common Pb ($f^{206}Pb$%, see Table 1) resulting in larger common Pb corrected $^{207}Pb/^{206}Pb$ age errors. Some indication of greater Pb loss is also suggested from the analytical locations touching fractures, though the timing of this Pb loss is not constrained. However, using only concordant data which forms clusters at the 2σ level, two distinctly different populations can be readily distinguished. Zoned crystals and inner domains yield a well defined concordant age of 1878.2 ± 3.5 Ma (Fig. 3b, hereafter referred to as 1878 ± 4 Ma). This is interpreted to reflect the crystallisation age of this felsic volcanic rock. Analyses on rims yield a concordant age of 1814.8 ± 2.9 Ma (Fig. 3c, hereafter referred to as 1815 ± 3 Ma) which is inter-

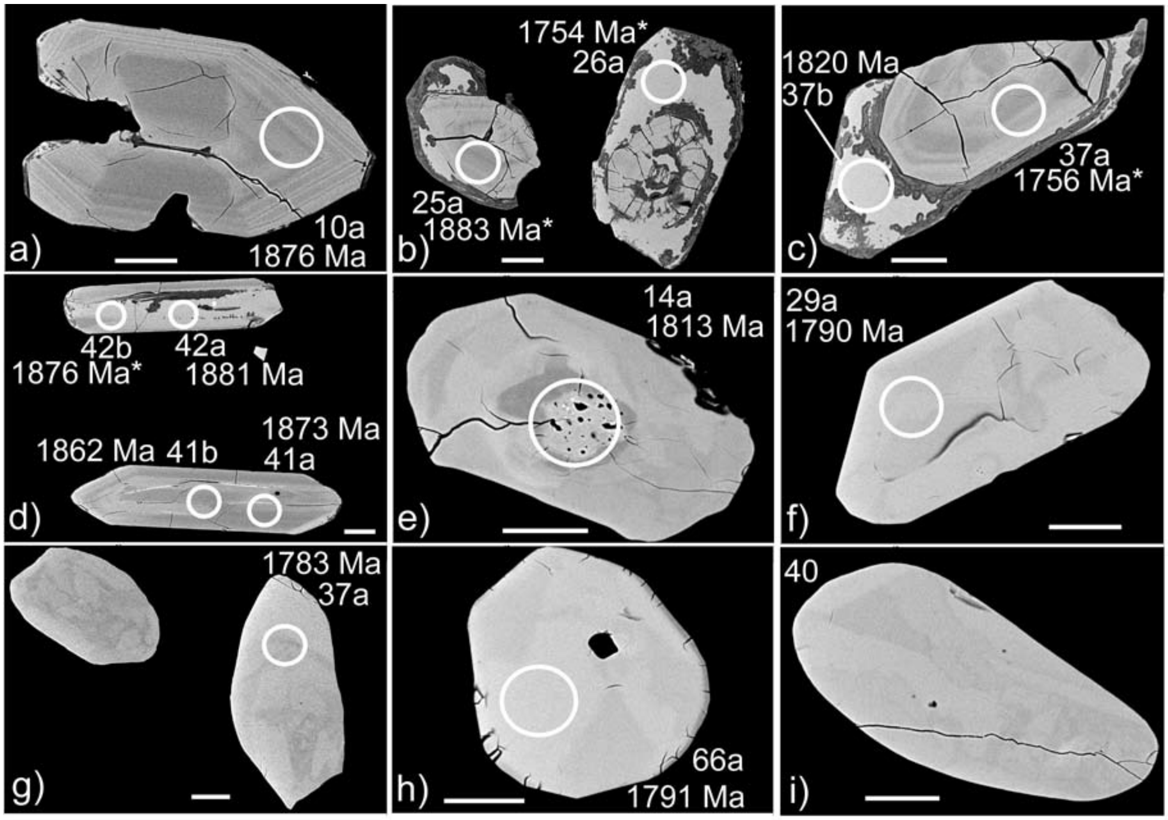

Fig. 2. SEM backscatter electrons (BSE) images of selected zircons demonstrating the morphological varieties in the dated crystals. 2a–2d are from the Toija felsic volcanic rock and 2e–2i are from the Salittu gabbro. Ellipses indicate analysed regions and the text refers to the $^{207}Pb/^{206}Pb$ age. For errors and other analytical results, see Table 1. Not all the shown dates were used in the concordia age calculations because of discordance promoted by Pb loss along fractures or grain margin. These are indicated by a *. Zircon 40 in 2i was not analysed but is shown to demonstrate the rounded, ovoid shape. Scale bar (white line) in each figure is 20 μm.

preted to correspond to the timing of a metamorphic overprint. The Th/U ratios of those analyses used for calculations in the BSE homogenous rims are lower, generally 0.01 – 0.1, compared to the Th/U ratios in the older domains which average 0.2 – 0.5 (Table 1). Low Th/U ratios are typical for metamorphic zircon (e.g. Hoskin & Black, 2000) although exceptions have been reported (see discussion in Whitehouse & Kamber, 2005).

5.2. Salittu Formation (gabbro:19MJV05)

Only a small population of zircon grains were recovered from the gabbro (c. 100 crystals). The zircons are all small, 50 – 100 μm in length and subhedral or anhedral in shape (Fig. 2e-2i). Most crystals have rounded shapes and some are completely ovoid or spherical (Figs 2h and 2i). In BSE-images the crystals are dominantly homogeneous and reveal no zoning but a weak patch-like texture can be observed in some crystals. A small sub-population of crystals contains small turbid inner cores. In general these domains were too small to position an analytical spot on, although one larger region of this texture was of sufficient size and was analysed (spot 14a in Table 1; Fig. 2e).

Nine spots on nine different zircons were analysed (Table 1). All nine analyses fall on concordia and cluster with a concordia age of 1794.7 ± 4.4 Ma. The old-

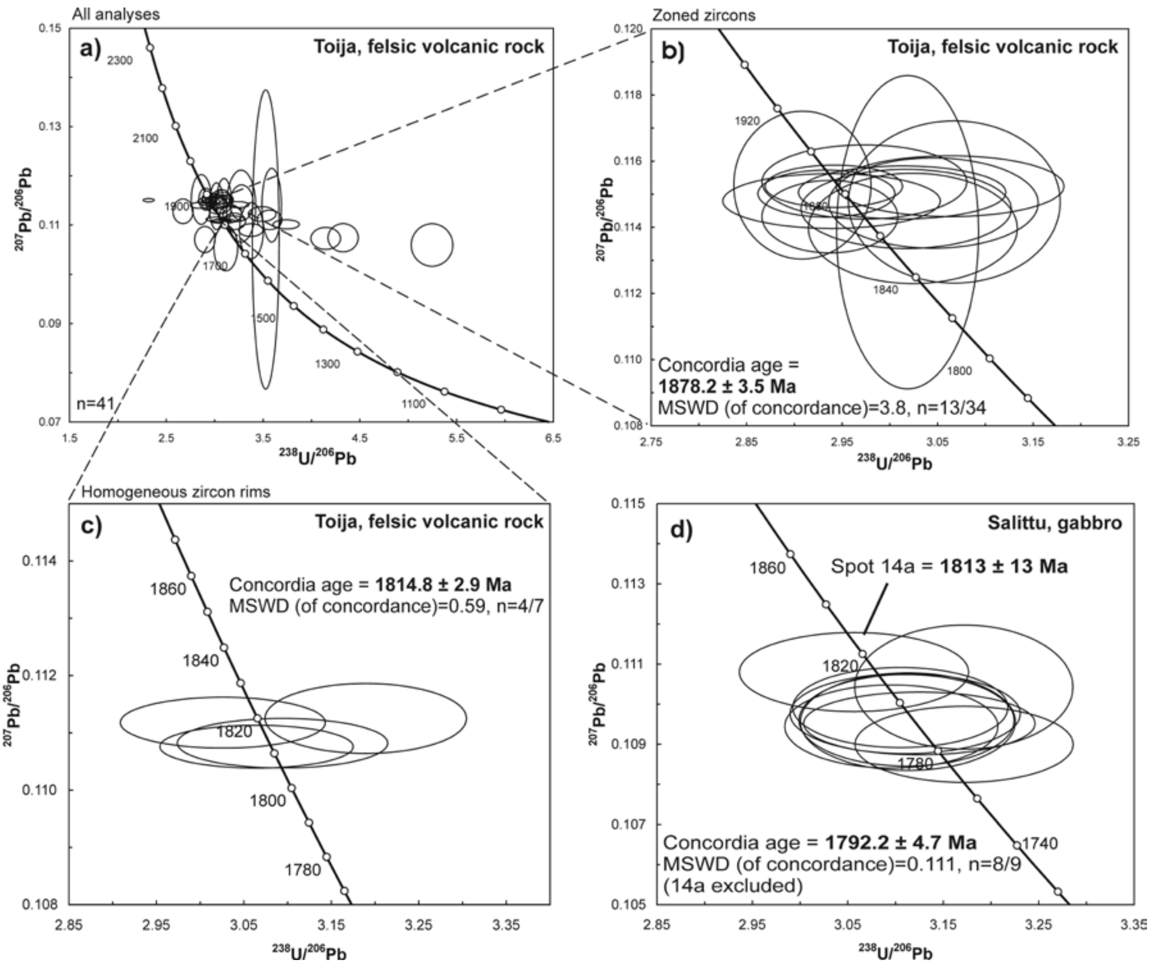

Fig. 3. Tera-Wasserburg concordia diagrams for SIMS analyses from the Orijärvi area with 2σ error ellipses. a), b) and c) are from the Toija felsic volcanic rock (26MJV06) and d) is from the Salittu gabbro (19MJV05). Ages are presented at 2σ level. Analyses/spots used in calculations in diagrams b) and c) are shown in bold and italics, respectively, in Table 1.

est of the spots was positioned on an inner domain of grain 14a. It yields a concordant age of 1813 ± 13 Ma. This inner domain has somewhat higher Th concentration and consequently, higher Th/U ratio compared to the other analyses (Table 1). Because there is only one analysis on the inner domains, the significance of the 1813 Ma age is ambiguous. It may be a mixture of older inherited age and younger metamorphic age. Therefore, we do not use this age in calculations. The eight other analyses yield a concordant age of 1792.2 ± 4.7 Ma (Fig. 3d, hereafter referred to as 1792 ± 5 Ma). On the basis of the reasons discussed below, we interpret this to reflect the timing of metamorphic event.

6. Discussion and conclusions

6.1. Age of volcanism

Väisänen & Mänttäri (2002) interpreted the Toija Formation to reflect the beginning of arc rifting and opening of a back-arc or intra-arc basin. Consequently, the Toija Formation was inferred to be younger than the Orijärvi and Kisko Formations that belong to the volcanic arc. The 1878 ± 4 Ma crystallisation

Table 1. U-Th-Pb SIMS isotopic results from rocks of the Orijärvi area

Sample 26MJV06: Felsic volcanic rock from the Toija Formation (National Coordinates: N=6688265, E=3302963)													Ages			
Zircon/spot	Comment	(U) ppm	(Th) ppm	(Pb) ppm	Th/U	f_{206}%	$^{238}U/^{206}Pb$	±σ (%)	$^{207}Pb/^{206}Pb$	±σ (%)	Disc. % conv.	Disc. % 2σ lim.	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ
n2395-58a		429	99	166	0.23	0.29	3.063	1.59	0.115	0.33	–3.81	–0.40	1884	6	1821	25
n2395-55a		844	234	343	0.28	0.02	2.939	1.58	0.115	0.29	0.69		1877	5	1888	26
n2395-48a		765	286	312	0.37	1.08	3.024	1.48	0.114	0.60	–1.35		1864	11	1842	24
n2395-42a		855	388	356	0.45	0.31	3.013	1.47	0.115	0.28	–2.03		1881	5	1848	24
n2395-41a		403	153	169	0.38	1.18	3.017	1.49	0.115	0.43	–1.69		1873	8	1845	24
n2395-30a		678	188	271	0.28	0.06	2.973	1.47	0.115	0.40	–1.02		1886	7	1869	24
n2395-19a		481	105	186	0.22	0.88	3.068	1.48	0.115	0.87	–3.50		1876	16	1819	23
n2395-10a		359	62	139	0.17	1.60	3.046	1.47	0.115	0.49	–2.78		1876	9	1830	24
n2809-41b		228	86	94	0.38	1.31	3.019	1.01	0.114	1.70	–1.06		1862	30	1845	16
n2809-65b		267	53	107	0.20	0.07	2.937	0.99	0.114	0.39	1.21		1869	7	1889	16
n2809-67a		169	44	71	0.26	2.46	2.908	1.01	0.115	0.79	1.28		1884	14	1905	17
n2809-113a		719	221	295	0.31	0.01	2.934	1.00	0.115	0.25	0.66		1880	5	1891	16
n2809-115a		1188	1091	561	0.92	0.01	2.941	1.00	0.115	0.23	0.18		1884	4	1887	16
n2395-83a	Hom. rim	1870	33	690	0.02	0.01	3.026	1.58	0.111	0.16	1.38		1819	3	1841	25
n2395-78a	Hom. rim	1564	18	563	0.01	0.01	3.094	1.59	0.111	0.16	–0.48		1813	3	1805	25
n2395-50a	Hom. rim	2178	23	792	0.01	0.01	3.065	1.47	0.111	0.14	0.54		1812	2	1820	23
n2395-37b	Hom. rim	2117	51	745	0.02	0.39	3.189	1.48	0.111	0.22	–3.88	–0.86	1820	4	1758	23
n2395-46a		1641	493	619	0.30	0.71	3.238	1.49	0.114	0.22	–8.03	–5.12	1867	4	1735	23
n2395-44a		238	56	93	0.23	5.21	3.278	1.49	0.115	1.29	–9.51	–3.04	1873	23	1716	22
n2395-65a		590	114	261	0.19	1.59	2.673	1.59	0.113	0.91	12.90	7.25	1845	16	2049	28
n2395-62a		537	149	203	0.28	19.16	3.527	1.62	0.107	11.59	–9.11		1750	198	1609	23
n2395-39a		311	103	121	0.33	1.76	3.165	1.47	0.112	0.59	–4.10	–0.19	1836	11	1770	23
n2395-38a	Hom. rim	1064	110	385	0.10	1.51	3.204	1.48	0.111	0.61	–3.83		1812	11	1751	23
n2395-37a	Core	694	134	204	0.19	6.15	4.327	1.48	0.107	1.09	–26.21	–20.64	1756	20	1340	18
n2395-35a		440	131	177	0.30	1.37	3.081	1.48	0.115	0.56	–4.21	–0.40	1881	10	1812	23
n2395-33b	Hom. rim	1082	68	328	0.06	0.66	3.740	1.49	0.110	0.35	–17.08	–14.14	1802	6	1528	20
n2395-33a	Core	524	37	120	0.07	5.51	5.246	1.67	0.106	1.67	–38.11	–29.82	1731	30	1125	17
n2395-28a		319	104	117	0.33	6.41	3.380	1.48	0.109	2.13	–6.99		1781	38	1671	22
n2395-27a		440	91	144	0.21	0.68	3.567	1.47	0.111	0.73	–13.90	–9.69	1817	13	1593	21
n2395-26a	Hom. rim	2408	68	680	0.03	4.57	4.147	1.51	0.107	0.85	–22.89	–18.26	1754	15	1392	19
n2395-25a		641	725	305	1.13	8.49	2.862	1.49	0.115	1.79	3.01		1883	32	1932	25
n2395-13a		209	37	78	0.18	6.75	3.115	1.61	0.105	1.80	4.95		1720	33	1795	25
n2395-6a		574	55	196	0.10	1.75	3.374	1.52	0.109	0.48	–6.94	–3.32	1782	9	1673	22
n2395-5b	Zoned rim	535	205	211	0.38	3.23	2.899	1.48	0.107	1.01	10.60	4.68	1750	18	1911	25
n2395-5a	Zoned core	817	324	302	0.40	2.24	3.273	1.89	0.114	2.63	–8.75		1862	47	1719	29
n2395-2b		837	291	358	0.35	0.34	2.852	1.48	0.114	0.22	4.54	1.28	1864	4	1937	25
n2395-2a	Overgrowth?	1099	596	402	0.54	4.62	3.491	1.67	0.112	0.54	–13.06	–9.23	1836	10	1624	24
n2809-42b		1365	703	568	0.52	1.20	3.097	1.01	0.115	1.69	–4.40		1876	30	1804	16
n2809-48b		349	106	123	0.30	5.35	3.589	1.15	0.114	2.68	–16.95	–4.67	1865	48	1584	16
n2809-105a		147	40	57	0.28	0.23	3.081	0.99	0.115	0.59	–4.57	–1.27	1887	11	1812	16
n2809-112a		7815	3172	4128	0.41	0.00	2.318	1.01	0.115	0.12	27.31	24.67	1881	2	2312	20

Table 1. continued

Sample 19MJV05: Gabbro from the Salittu Formation (National Coordinates: N=6693698, E=3312949)													Ages			
Zircon/spot	Comment	(U) ppm	(Th) ppm	(Pb) ppm	Th/U	f_{206}%	$^{238}U/^{206}Pb$	±σ (%)	$^{207}Pb/^{206}Pb$	±σ (%)	Disc. % conv.	Disc. % 2σ lim.	$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ
n2396-46a		231	67	89	0.29	0.04	3.113	1.48	0.110	0.45	0.23		1792	8	1796	23
n2396-73a		289	138	116	0.48	0.03	3.112	1.48	0.110	0.43	0.24		1793	8	1796	23
n2396-64a		330	128	130	0.39	0.03	3.105	1.48	0.110	0.37	0.11		1798	7	1800	23
n2396-37a		377	108	142	0.29	0.03	3.172	1.47	0.109	0.36	−1.04		1783	6	1767	23
n2396-29a		319	116	125	0.36	0.06	3.096	1.47	0.109	0.39	0.92		1790	7	1805	23
n2396-25a		189	57	72	0.30	0.58	3.172	1.47	0.110	0.57	−2.51		1806	10	1767	23
n2396-14a	Core	322	730	190	2.27	0.08	3.057	1.61	0.111	0.36	0.76		1813	7	1825	26
n2396-66a		486	236	194	0.49	0.02	3.125	1.58	0.110	0.29	−0.11		1791	5	1790	25
n2396-70a		286	121	113	0.42	0.02	3.110	1.58	0.110	0.37	0.11		1795	7	1797	25

Ion microprobe U-Th-Pb data. Toija Formation; analyses used for calculating the age of the zoned zircons are marked in **bold**, those used for the rim age are marked in *italics*. Salittu Formation; all analyses are used. Hom. rim is homogeneous rim. f_{206}% is percentage of common ^{206}Pb estimated from the measured ^{204}Pb, assuming a present-day Stacey & Kramers (1975) model. Disc. % conv. is the discordance in conventional concordia space. Disc. % 2σ is the degree of discordance of the $^{207}Pb/^{206}Pb$ and $^{206}Pb/^{238}U$ ages at the 2σ level estimated from the measured ^{204}Pb. Age calculations use the routines of Ludwig (2003) and follow the decay constant recommendations of Steiger & Jäger (1977).

age (detailed above) from the upper part of the Toija Formation is consistent with this hypothesis as the age is within errors, identical, to the uppermost part of the Kisko Formation dated at 1878.2 ± 3.4 Ma (Väisänen & Mänttäri, 2002). This verifies that (i) the previously dated rock from the Kisko Formation represented the end of arc growth and (ii) the bimodal extension-related Toija Formation is of approximately the same age as the cessation of arc growth. This is consistent with plate tectonic models of subduction-related magmatism where the arc ceases to grow approximately at the same time as the back-arc opening starts (e.g. Wilson, 1989).

The Salittu Formation with its tholeiitic basalts and picrites is unfavourable for U-Pb zircon geochronology. However, metamorphic zircon was recovered (see below). Several lines of evidence indicate that the 1878 ± 4 Ma age for the upper part of the Toija Formation also applies to the lower part of the Salittu Formation. The most important arguments for this are: (i) picritic rocks already started to occur before deposition of the Salittu Formation which itself is dominated by this rock type. In the upper part of the Toija Formation the onset of ultramafic magmatism took place simultaneously with felsic magmatism, producing a similar rock assemblage to that found in the overlying Salittu Formation. Ultramafic and felsic volcanic rocks occurring together have also been found elsewhere e.g. locality 245MV95 (Väisänen & Mänttäri, 2002) and Suomusjärvi (Salli, 1955). (ii) The Salittu Formation has a primary depositional contact with the underlying Toija Formation (Figs 2d and 2g in Väisänen & Mänttäri, 2002). This would be impossible if the Salittu Formation were an older unit. (iii) A thin picritic layer has also been mapped in the upper part of the Kisko Formation, stratigraphically just below the c. 1878 Ma dated dacitic rock (272MV95 in Väisänen & Mänttäri, 2002). This indicates that the ultramafic magmatism, i.e. the Salittu Formation proper, started after the main arc volcanism.

Skyttä et al. (2006) dated two synorogenic dykes, located c. 15 km to the south from the present area, that intrude into deformed rocks during D2 defor-

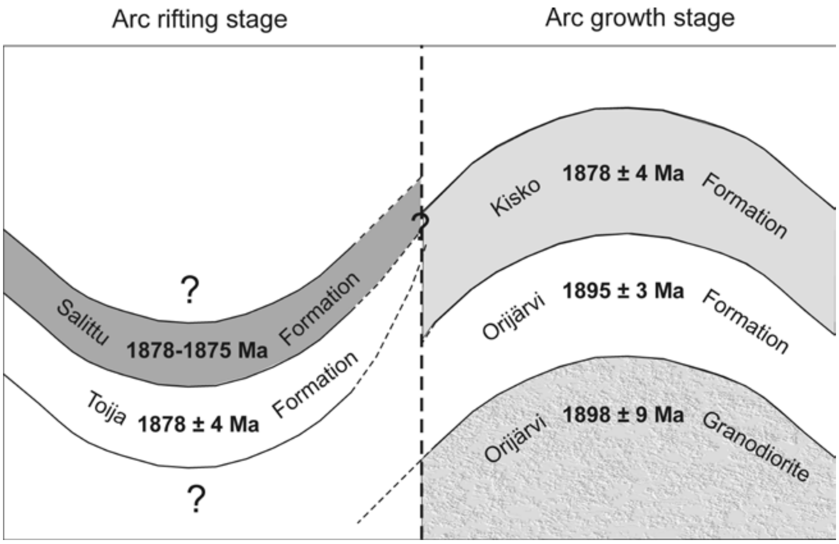

Fig. 4. Schematic stratigraphic profile of the different formations in the Orijärvi area, shown prior to deformation events.

mation at 1877 ± 3 Ma (granodiorite) and 1876 ± 4 Ma (intermediate dyke). These ages, referred to as 1875 Ma magmatism in Skyttä et al. (2006), give the minimum age for volcanism in the area.

In summary, the bimodal volcanism in the Orijärvi area started at c. 1895 Ma. This volcanic arc grew until c. 1878 Ma when rifting commenced. During the rifting stage, bimodal volcanic rocks and marbles were deposited. The rifting intensified and opened channels to dominantly ultramafic and mafic EMORB-type volcanism at c. 1878 – 1875 Ma (Väisänen & Mänttäri, 2002 and this study). A schematic cartoon of the stratigraphy is shown in Fig. 4.

6.2. Age of metamorphism

The 1815 ± 3 Ma rims on the magmatic zircons from the Toija Formation are morphologically different to the zoned inner domains that they crosscut. The c. 1815 Ma domains are irregular, inward-penetrating, and have patchy reaction zones; their inner contact follows primary growth zones within core regions of the zircon. In addition, Th/U ratios in the rims are in most cases lower compared to those of the inner domains. There is no apparent record of the parent isotope compositions in the recrystallised rim domains, which implies complete re-equilibration of the isotope systems by a coupled dissolution–reprecipitation process (Geisler et al., 2007). This is consistent with descriptions of solid-state metamorphic recrystallisation of zircon (Hoskin & Black, 2000; Rayner et al., 2005). This is compatible with previous investigations on late Svecofennian metamorphism. 20 km to the east of the present location, within the West Uusimaa granulite area, Mouri et al. (2005) dated monazites from high-grade gneisses by U-Pb TIMS. The leucosome monazites yielded concordant ages of 1819 ± 2 Ma, 1818 ± 2 Ma and 1816 ± 2 Ma, while the mesosome monazites yielded an age of 1832 ± 2 Ma. The results were interpreted to indicate that continuous high temperature metamorphism prevailed between 1830 – 1815 Ma. The leucosome monazite ages are identical within error to the metamorphic zircon ages obtained in this study.

The 1792 ± 5 Ma age obtained from the Salittu gabbro is not straightforward to interpret. We interpret that this age is related to metamorphism. There are several pieces of evidence that support this argument: (i) The field relationships indicate that the gabbro was intruded in the very early stages of the Svecofennian Orogeny, as it was deformed in all deformation events that affected the supracrustal rocks that it intrudes. This suggests an intrusion age older than c. 1875 Ma for the gabbro. Typically rocks within the 1.81 – 1.79 Ga age bracket are classified as post-collisional or post-orogenic in the Svecofennian Orogeny and invariably show crosscutting field relationships to their host rocks (e.g. Hopgood et al., 1983, Ehlers et al., 2004; Eklund & Shebanov, 2005; Nironen, 2005). (ii) Geochemical compositions of post-orogenic intrusions in Finland show mantle characteristics that are highly enriched in Sr, Ba, P, LILEs and LREEs and have a strong negative Nb-Ta anomaly (Eklund et al., 1998; Andersson et al., 2006a and references therein). The composition of the Salittu gabbro is different having an EMORB-type chemistry, similar to the volcanic rocks in the Salittu Formation (Väisänen & Mänttäri, 2002), which prompts a synvolcanic interpretation for the intrusion. (iii) BSE images of the zircons show rather homogenous textures similar to those described for metamorphic zircons (e.g. Hoskin & Black, 2000). Metamorphically recrystallised zircon grains typically contain featureless (homogeneous) domains and, in other areas, fading of internal structures similar to some of the textures we observe (Fig. 2e-2i; e.g. Geisler et al., 2007).

The textural features of the zircon grains, within the Salittu gabbro, indicate a metamorphic origin. However, it is difficult to differentiate between subsolidus nucleation of neocrystals or metamorphic recrystallisation of older grains. The rounded external morphology of some of the crystals resembles mechanically abraded detrital zircons, perhaps indicating that the zircons originally were xenocrysts entrained into the gabbro from the surrounding sedimentary rocks. On the other hand, rounded zircon shapes can be produced during metamorphism through volume reduction of crystals if in contact with Zr-undersaturated intergranular fluid or melt (Hoskin & Black, 2000). In either case the age of such zircon reflects metamorphic processes.

High temperature conditions prevailed in the West Uusimaa area at 1815 Ma whereas Sm-Nd garnet ages indicate that metamorphic temperatures had cooled to around 650°C at c. 1.80 Ga (Mouri et al., 2005). This indicates that either the 1792 Ma zircons crystallised at rather low temperatures or high heat flow/fluid flow was localised. The fact that the gabbro is situated in the Jyly shear zone likely had a pronounced influence on the zircon grains. The Jyly shear zone is known to have grown sillimanite on its fault planes (Ploegsma, 1989; Väisänen & Skyttä, 2007), therefore this zone was at high temperature and a suitable conduit for fluid flow, which could have promoted localised zircon growth or recrystallisation.

In summary, the peak metamorphism in the Orijärvi area took place at 1815 ± 3 Ma. The younger localised metamorphism at 1792 ± 5 Ma probably took place as a result of later hot fluid flow along the Jyly shear zone. This implies that the shear zone was active at this time.

6.3. Regional implications

The new age results from the Orijärvi area show that volcanism and associated plutonism lasted about twenty million years, and even longer if Bergslagen in south-central Sweden is considered (Allen et al., 1996). The earliest stage of arc volcanism at 1895 Ma is bimodal with marbles and iron formations. Although controversial, some older hidden crust may have been involved with the formation of the early volcanism (Lahtinen & Huhma, 1997). The possible hidden crust has been referred to as "proto-Svecofennian" by Andersson et al. (2006b). The volcanic arcs grew until about 1878 Ma when the tectonic regime turned to rifting (Väisänen & Mänttäri, 2002). Considering the short time span involved, the rifting stage must have been intensive as manifested by nearly juvenile ultramafic volcanism that was erupted in the Salittu Formation (initial ε_{Nd} value +3.1 at 1.89 Ga; Huhma, 1986). The majority of supracrustal rocks between the Uusimaa and Häme belts likely belong to this stage, e.g. volcanic and sedimentary rocks in the Turku area (Väisänen & Westerlund, 2007).

The extensional period was, however, very short-lived and swapped into contraction when southern Svecofennia collided with central Svecofennia (arc-accretion, e.g. Lahtinen, 1994; Korja & Heikkinen, 2005). This effectively prevented the Salittu Formation from widening into oceanic crust *sensu stricto* (i.e. an ophiolite). The collision triggered new synorogenic magmatism which in our study area has been dated at c. 1875 Ma (Skyttä et al., 2006). This marks the minimum age of the synvolcanic stage and the commencement of the synorogenic stage. In the Turku area, the latter magmatism is slightly younger, at around 1870 Ma (Nironen, 1999; Väisänen et al., 2002).

Our 1815 ± 3 Ma and 1792 ± 5 Ma age results from the Orijärvi area are the first U-Pb zircon ages performed on the late Svecofennian metamorphism in the West Uusimaa and adjacent areas. Previous c. 1830-1815 Ma U-Pb ages were obtained from monazites (Mouri et al., 2005). These monazite ages coincide with the 1824 ± 5 Ma monazite age of metamorphism from Kemiö, 50 km west of Orijärvi (Levin et al., 2005). Ages of crustally derived granites can also be used to deduce the timing of heat flow. Nine ion-microprobe ages on granites and pegmatites from the West Uusimaa area (Skyttä & Mänttäri, 2008) apparently form three age groups. Five of the granites are coeval, within error, with the 1815 ± 3 Ma metamorphism reported here, three of the granites are coeval with the 1832 ± 2 Ma mesosome monazites reported by Mouri et al. (2005) and one of the granites is older at 1843 ± 3 Ma. The oldest granite has so far no proved metamorphic counterpart in the West Uusimaa area although it was obviously derived from crustal melting.

The 1792 ± 5 Ma metamorphic age from the gabbro within the Jyly shear zone is the youngest zircon age for metamorphism so far reported from southern Finland. 1793 ± 5 Ma zircon age has been obtained from high-grade metasediment in Berslagen, south-central Sweden (Andersson et al., 2006b). However, similar U-Pb monazite ages for metamorphism from mica gneisses in Sulkava, southeastern Finland, have been reported by Vaasjoki & Sakko (1988) and Baltybaev et al. (2006), who obtained ages of 1796 ± 5 Ma and 1794.7 ± 4.6 Ma, respectively. Ages close to 1795 Ma are quite common in crustally derived plutonic rocks in SW Finland. In addition to several post-orogenic intrusions dated in Suominen (1991), monazite from the Oripää Granite in the Loimaa area has been dated at 1794 ± 10 Ma (Nironen, 1999), monazite in the Torsholma pegmatite yielded an age of 1795 ± 4 Ma and zircon in the Sottunga Granite, within the South Finland shear zone, yielded an age of 1790 ± 6 Ma (Ehlers et al., 2004). The 1790.6 ± 7.5 Ma zircon cores in the Åva granite are also interpreted to indicate the age of magma formation, although it was emplaced later at c. 1762 Ma (Eklund & Shebanov, 2005). All these results imply an active tectonothermal regime at c. 1.80 – 1.79 Ga within the Svecofennian Orogen in southern Finland.

It has previously been assumed that metamorphism was younger in southeastern Finland than elsewhere in southern Finland (Korsman et al. 1999). As the geochronology dataset for this large region of Finland has developed it now seems more likely that the whole belt, including Bergslagen in Sweden, preserves evidence of long-lasting late Svecofennian heat flow. These include 1.83 – 1.79 Ga metamorphic ages in Bergslagen in Sweden (Andersson et al., 2006b and references therein). In southwestern Finland there is zircon data on 1.85 Ga and 1.83 Ga metamorphism (Torvela et al., 2008), 1824 ± 5 Ma and 1804 ± 14 Ma leucosomes (Väisänen et al. 2002) and c. 1830 Ma and younger monazite data from granites (Suominen, 1991; Nironen, 1999; Ehlers et al., 2004). From the West Uusimaa area there are from c. 1.85 Ga to 1813 ± 4 Ma granites (Kurhila et al., 2005; Skyttä & Mänttäri, 2008) and from 1832 ± 2 Ma to 1792 ± 5 Ma metamorphic ages (Mouri et al., 2005 and this study). From southeastern Finland ages on metamorphism and crustally derived granites ranges from c. 1840 Ma to 1795 Ma (Korsman et al. 1984; Nykänen, 1988; Vaasjoki & Sakko, 1988). Theses data, with apparent clusters at c. 1850 – 1840, 1830 – 1820, 1815 and 1795 Ma, suggest that high heat flow events may have been episodic rather than continuous, in accordance with the view of Kurhila et al. (2005) who stated that the heat flow took place in separate pulses. Successive intrusion of mafic magmas into mid-crustal levels could account for the episodic heat pulses (Schreurs & Westra, 1986). Metamorphism at c. 1795 Ma was evidently more widespread in the deeper crust and can now only be sporadically detected at the present erosion level.

Acknowledgements

This study was funded by the Academy of Finland (project 117311) and the Turku University Foundation. Karin Högdahl, Hannu Huhma and Editor Petri Peltonen reviewed the manuscript and suggested many improvements. Sören Fröjdö at Åbo Akademi University helped tuning the shaking table and Pietari Skyttä helped drafting of the map. Martin Whitehouse, Kerstin Lindén and Lev Ilyinsky are thanked for help during the analytical session at the Nordsim laboratory. The Nordsim facility 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. This is Nordsim publication 208.

Appendix 1. *Geochemical analysis of the sample 19MJV05 from the Salittu gabbro.*

Element abundances of the Salittu gabbro are listed below. The analysis was performed at Acmelab Ltd., Canada. Major elements were analysed by ICP and trace elements by ICP-MS methods. Additional analytical information is available at www.acmelab.com.

Major elements: SiO_2 = 48.07; TiO_2 = 1.99; Al_2O_3 = 13.51; Fe_2O_3 = 15.29; MnO = 0.22; MgO = 6.56; CaO = 9.98; Na_2O = 3.22; K_2O = 0.4; P_2O_5 = 0.21;

Trace elements: Ba = 82.5; Rb = 7.4; Sr = 271.6; Cs = 0.7; Zr = 122.3; Hf = 3.8; Y = 25; Nb = 10.4; Ta = 0.7; U = 0.4; Cr = 41; Ni = 47; Th = 0.9; La = 12.2; Ce = 30.1; Nd = 19.3; Sm = 5.1; Eu = 1.78; Tb = 0.93; Lu = 0.27; Er = 2.58; Yb = 2.34; Y = 25;

References

Allen, R.L., Lundström, I., Ripa, M., Simeonov, A. & Christofferson, H., 1996. Facies analysis of a 1.9 Ga, continental margin, back-arc, felsic caldera province with diverse Zn-Pb-Ag-(Cu-Au) sulfide and Fe oxide deposits, Bergslagen region, Sweden. Economic Geology 91, 979–1008.

Andersson, U. B., Eklund, O., Fröjdö, S. & Konopelko, D., 2006a. 1.8 Ga magmatism in the Fennoscandian Shield: lateral variations in subcontinental mantle enrichment. Lithos 86, 110–136.

Andersson, U.B., Högdahl, K., Sjöström, H. & Bergman, S., 2006b. Multistage growth and reworking of the Palaeoproterozoic crust in the Bergslagen area, southern Sweden; evidence from U-Pb geochronology. Geological Magazine 143, 679–697.

Baltybaev, S.K., Levchenkov, O.A., Levsky, L. K., Eklund, O. & Kilpeläinen, T., 2006. Two metamorphic stages in the Svecofennian domain: Evidence from the isotopic geochronological study of the Ladoga and Sulkava metamorphic complexes. Petrology 14, 247–261.

Ehlers, C., Lindroos, A. & Selonen, O., 1993. The late Svecofennian granite-migmatite zone of southern Finland – a belt of transpressive deformation and granite emplacement. Precambrian Research 64, 295–309.

Ehlers, C., Skiöld, T. & Vaasjoki, M., 2004. Timing of Svecofennian crustal growth and collisional tectonics in Åland, SW Finland. Bulletin of the Geological Society of Finland 76, 63–91.

Eklund, O., Konopelko, D., Rutanen, H., Fröjdö, S. & Shebanov, A. D., 1998. 1. 8 Ga Svecofennian post-collisional shoshonitic magmatism in the Fennoscandian Shield. Lithos 45, 87–108.

Eklund, O. & Shebanov, A., 2005. Prolonged postcollisional shoshonitic magmatism in the southern Svecofennian domain – a case study of the Åva granite-lamprophyre ring complex. Lithos 80, 229–247.

Eskola, P., 1914. On the petrology of the Orijärvi region in southwestern Finland. Bulletin de la Commission géologique de Finlande 40, 277 p.

Geisler T., Schaltegger, U. & Tomaschek, F., 2007. Re-equilibration of zircon in aqueous fluids and melts. Elements 3, 43–50.

Hopgood, A.M., Bowes, D.R., Kouvo, O. & Halliday, A.D., 1983. U-Pb and Rb-Sr isotopic study of polyphase deformed migmatites in the Svecokarelides, southern Finland. In: Atherton, M.P. & Gribble, C.D. (eds.) Migmatites, Melting and Metamorphism. Nantwich: Shiva, 80–92.

Hoskin, P.W.O. & Black, L.P., 2000. Metamorphic zircon formation by solid-state recrystallization of protolith igneous zircon. Journal of Metamorphic Geology 18, 423–439.

Huhma, H., 1986. Sm-Nd, U-Pb and Pb-Pb isotopic evidence for the origin of the early Proterozoic Svecokarelian crust in Finland. Geological Survey of Finland, Bulletin 337, 48 p.

Kähkönen, Y., 2005. Svecofennian supracrustal rocks. In: Lehtinen, M., Nurmi, P. A. & Rämö, O. T. (eds.) Precambrian Geology of Finland – Key to the Evolution of the Fennoscandian Shield. Developments in Precambrian Geology 14. Elsevier, Amsterdam, 343–405.

Kirkland, C.L., Daly, J.S. & Whitehouse, M.J., 2008. Basement-cover relationships of the Kalak Nappe Complex, Arctic Norwegian Caledonides and constraints on Neoproterozoic terrane assembly in the North Atlantic region. Precambrian Research 160, 245–276.

Korja, A. & Heikkinen, P., 2005. The accretionary Svecofennian orogen - insight from the BABEL profiles. Precambrian Research 136, 241–268.

Korja, A., Lahtinen, R. & Nironen, M. 2006. The Svecofennian orogen: a collage of microcontinents and island arcs. Geological Society, London, Memoirs 32, 561–578.

Korsman, K., Hölttä, P., Hautala, T. & Wasenius, P., 1984. Metamorphism as an indicator of evolution and structure of crust in eastern Finland. Geological Survey of Finland, Bulletin 328, 40 p.

Korsman, K., Koistinen, T., Kohonen, J., Wennerström, M., Ekdahl, E., Honkamo, M., Idman, H. & Pekkala, Y., (eds.) 1997. Bedrock map of Finland 1:1 000 000. Geological Survey of Finland, Espoo, Finland.

Korsman K., Korja T., Pajunen M., Virransalo P. & GGT/SVEKA Working Group, 1999. The GGT/SVEKA transect: structure and evolution of the continental crust in the Paleoproterozoic Svecofennian orogen in Finland. International Geology Review 41, 287–333.

Kumpulainen, R., Mansfeld, J., Sundblad, K., Neymark, L. & Bergman, T., 1996. Stratigraphy, age and Sm-Nd isotope systematics of the country rocks to Zn-Pb sulfide deposits, Åmmeberg district, Sweden. Economic Geology 91, 1009–1021.

Kurhila, M., Vaasjoki, M., Mänttäri, I., Rämö, T. & Nironen, M., 2005. U-Pb ages and Nd isotope characteristics of the lateorogenic, migmatizing microcline granites in southwestern Finland. Bulletin of the Geological Society of Finland 77, 105–128.

Lahtinen, R. 1994. Crustal evolution of the Svecofennian and Karelian domains during 2.1 – 1.79 Ga, with special emphasis on the geochemistry and origin of 1.93-1.91 Ga gneissic tonalites and associated supracrustal rocks in the Rautalampi area, central Finland. Geological Survey of Finland, Bulletin 378, 128 p.

Lahtinen, R., 1996. Geochemistry of Palaeoproterozoic supracrustal and plutonic rocks in the Tampere–Hämeenlinna area, southern Finland. Geological Survey of Finland, Bulletin 389, 113 p.

Lahtinen, R. & Huhma, H., 1997. Isotopic and geochemical constraints on the evolution of the 1.93 – 1.79 Ga Svecofennian crust and mantle in Finland. Precambrian Research 82, 13–34.

Lahtinen, R., Korja, A. & Nironen, M., 2005. Paleoproterozoic tectonic evolution. In: Lehtinen, M., Nurmi, P. A. & Rämö, O. T. (eds.) Precambrian Geology of Finland – Key to the Evolution of the Fennoscandian Shield. Developments in Precambrian Geology 14, Elsevier, Amsterdam, 481–531.

Levin, T., Engström, J., Lindroos, A., Baltybaev, S. & Levchenkov, O., 2005. Late-Svecofennian transpressive deformation in SW Finland – evidence from late-stage D3 structures. GFF 127, 129–137.

Lindroos, A. & Ehlers, C., 1994. Emplacement, deformation and geochemistry of bimodal volcanics in Vestlax, SW Finland. Geological Survey of Finland, Special Paper 19, 173–184.

Ludwig, K.R., 2003. User's Manual for Isoplot 3.00. A Geochronological Toolkit for Microsoft Excel. Berkley Geochronology Center, Special Publication No. 4.

Mouri H., Väisänen M., Huhma H. & Korsman K., 2005. Sm-Nd garnet and U-Pb monazite dating of high-grade metamorphism and crustal melting in the West Uusimaa area, southern Finland. GFF 127, 123–128.

Nironen, M. 1997. The Svecofennian Orogen: a tectonic model. Precambrian Research 86, 21–44.

Nironen, M., 1999. Structural and magmatic evolution in the Loimaa area, southwestern Finland. Bulletin of the Geological Society of Finland 71, 57–71.

Nironen, M. 2005. Proterozoic orogenic granitoid rocks. In: Lehtinen, M., Nurmi, P. A. & Rämö, O. T. (eds.) Precambrian Geology of Finland – Key to the Evolution of the Fennoscandian Shield. Developments in Precambrian Geology 14, Amsterdam, Elsevier, 443–479.

Nykänen, O., 1988. Pre-Quaternary rocks of the Virmutjoki map-sheet area, 1:100 000. Explanation to the map sheet 4121. Geological Survey of Finland, Espoo, 64 p.

Patchett, J. & Kouvo, O., 1986. Origin of continental crust of 1.9-1.7 Ga age: Nd isotopes and U-Pb zircon ages in the Svecokarelian terrain of South Finland. Contributions to Mineralogy and Petrology 92, 1–12.

Peltonen, P. 2005. Svecofennian mafic-ultramafic intrusions. In: Lehtinen, M., Nurmi, P. A. & Rämö, O. T. (eds.) Precambrian Geology of Finland – Key to the Evolution of the Fennoscandian Shield. Developments in Precambrian Geology 14. Elsevier, Amsterdam, 407–441.

Ploegsma, M., 1989. Shear zones in the West Uusimaa area, SW Finland. Ph.D. thesis, Vrije Universiteit te Amsterdam, 134 p.

Ploegsma, M. & Westra, L., 1990. The Early Proterozoic Orijärvi triangle (southwest Finland): a key area on the tectonic evolution of the Svecofennides. Precambrian Research 47, 51–69.

Rämö, O.T., Vaasjoki, M., Mänttäri, I., Elliott, B.A. & Nironen, M., 2001. Petrogenesis of the post-kinematic magmatism of the Central Finland Granitoid Complex I; radiogenic isotope constraints and implications for crustal evolution. Journal of Petrology 42, 1971–1993.

Rayner, N., Stern, R.A. & Carr, D., 2005. Grainscale variations in trace element composition of fluid-altered zircon, Acasta Gneiss Complex, northwestern Canada. Contributions to Mineralogy and Petrology 148, 721–734.

Reinikainen, J., 2001. Petrogenesis of Paleoproterozoic marbles in the Svecofennian Domain, Finland. Geological Survey of Finland, Report of Investigations 154, 84 p.

Salli, I. 1955. Suomusjärvi. Geological map of Finland 1:100 000: pre-Quaternary rocks, sheet 2023, Geological Survey of Finland, Espoo.

Schreurs, J. & Westra, L., 1986. The thermotectonic evolution of a Proterozoic, low pressure, granulite dome, West Uusimaa, SW Finland. Contributions to Mineralogy and Petrology 93, 236–250.

Skyttä, P., Käpyaho, A. & Mänttäri, I., 2005. Supracrustal rocks in the Kuovila area, southern Finland: structural evolution, geochemical characteristics and the age of volcanism. Bulletin of the Geological Society of Finland 77, 129–150.

Skyttä, P. & Mänttäri, I., 2008. Structural setting of late Svecofennian granites and pegmatites in Uusimaa Belt, SW Finland: Age constraints and implications for crustal evolution. Precambrian Research 164, 86–109, doi:10.1016/j.precamres.2008.04.001

Skyttä, P., Väisänen, M. & Mänttäri, I., 2006. Preservation of Palaeoproterozoic early Svecofennian structures in the Orijärvi area, SW Finland – Evidence for polyphase strain partitioning. Precambrian Research 150, 153–172.

Stacey, J.S. & Kramers, J.D., 1975. Approximation of terrestrial lead isotope evolution by a two-stage model. Earth and Planetary Science Letters 26, 207–221.

Steiger, R.H. & Jäger, E., 1977. Subcomission on geochronology: convention on the use of decay constants in geo- and cosmochronology. Earth and Planetary Science Letters 36, 359–362.

Suominen, V., 1991. The chronostratigraphy of southwestern Finland with special reference to Postjotnian and Subjotnian diabases. Geological Survey of Finland, Bulletin 356, 100 p.

Torvela, T., Mänttäri, I. & Hermansson, T., 2008. Timing of deformation phases within the South Finland shear zone, SW Finland. Precambrian Research 160, 277–298.

Vaasjoki, M., 1994. Valijärven hapan vulkaniitti: minimi Hämeen liuskejakson iäksi. Summary: Radiometric age of a meta-andesite at Valijärvi, Häme schist zone, southern Finland. Geologi 46, 91–92.

Vaasjoki, M. & Sakko, M., 1988. The evolution of the Raahe-Ladoga zone in Finland: isotopic constraints. Geological Survey of Finland, Bulletin 343, 7–32.

Väisänen, M. & Mänttäri, I., 2002. 1.90-1.88 Ga arc and back-arc basin in the Orijärvi area, SW Finland. Bulletin of the Geological Society of Finland 74, 185–214.

Väisänen, M., Mänttäri, I. & Hölttä, P., 2002. Svecofennian magmatic and metamorphic evolution in southwestern Finland as revealed by U-Pb zircon SIMS geochronology. Precambrian Research 116, 111–127.

Väisänen, M. & Skyttä P., 2007. Late Svecofennian shear zones in southwestern Finland. GFF 129, 55–64.

Väisänen, M. & Westerlund G., 2007. Palaeoproterozoic mafic and intermediate metavolcanic rocks in the Turku area, SW Finland. Bulletin of the Geological Society of Finland 79, 127–141.

Valbracht, J., Oen I. S. & Beunk, F. F., 1994. Sm-Nd isotope systematics of 1.9-1.8 Ga granites from western Bergslagen, Sweden: inferences on a 2.1-2.0 Ga crustal precursor. Chemical Geology 112, 21–37.

Weihed, P., Arndt, N., Billström, K., Duchesne, J-C., Eilu, P., Martinsson, O., Papunen, H. & Lahtinen, R., 2005. Precambrian geodynamics and ore formation: the Fennoscandian Shield. Ore Geology Reviews 27, 273–322.

Whitehouse, M.J. & Kamber, B.S. 2005. Assigning dates to thin gneissic veins in high-grade metamorphic terranes: a cautionary tale from Akilia, southwest Greenland. Journal of Petrology 46, 291–318.

Whitehouse, M.J., Kamber, B. & 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. Chemical Geology 160, 201–224.

Wiedenbeck, M., Allé, P., Corfu, F., Griffin, W.L., Meier, M., Oberli, F., von Quadt, A., Roddick, J.C. & Spiegeln, W., 1995. Three natural zircon standards for U-Th-Pb, Lu-Hf, trace element and REE analyses. Geostandards Newsletter 19, 1–23.

Wilson, M., 1989. Igneous Petrogenesis. Chapman and Hall, London, 466 p.

Wood, D.A., 1980. The application of a Th-Hf-Ta diagram to problems of tectonomagmatic classification and to establish the nature of crustal contamination of basaltic lavas of the British Tertiary Volcanic Province. Earth and Planetary Science Letters 50, 11–30.