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The Salittu Formation in southwestern Finland, part I: Structure, age and stratigraphy

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Abstract

Because of the relatively low metamorphic grade, low strain and well-preserved early structures in volcanic and volcaniclastic rocks, the Orijärvi triangle (in the center of the larger Orijärvi area) is one of the few locations in the Svecofennian orogen of southern Finland where lithostratigraphy has been determined. The geochemistry of the picritic-basaltic metavolcanic rocks of the Salittu Formation, located in the northeastern part of the Orijärvi area, has been characterized but otherwise the bedrock and structures have been barely touched. After remapping we give an interpretation of structural evolution, provide new age data to constrain duration of volcanism at Salittu, and refine the stratigraphy in the Orijärvi area.

The original stratigraphy is visible at Salittu: metabasalt overlies migmatitic gneisses, and metapicrite is on top. The rocks were folded during early Svecofennian D_1–D_2 deformations, and the large synformal structures developed as D_2–D_5 interference structures formed during late Svecofennian D_5 deformation. The structural pattern at Salittu is much the same as in the Orijärvi triangle.

The new age data, combined with earlier published data, constrains the Salittu volcanism at ca. 1875 Ma. The stratigraphy in the Orijärvi area consists of the early (1.90–1.89 Ga) volcanic Orijärvi Formation, overlain by the sedimentary Vetio Formation, the volcanic Kisko Formation, the volcanic-sedimentary Ahdisto Formation, the volcanic Toija Formation, and on top the Salittu Formation, all emplaced at 1.88–1.87 Ga. We propose a model in which the Orijärvi Formation represents magmatism at the margin of a microcontinent, and the overlying package represents sedimentation and magmatism above a subduction zone during an initial stage of microcontinental accretion. D_1 deformation occurred in an advanced stage of accretion, after emplacement of the volcanic rocks of the Salittu Formation.

Keywords: metavolcanic rocks, gabbros, volcanism, structural geology, stratigraphy, U–Pb, zircon, Paleoproterozoic, Orijärvi, Salittu, Finland

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1. Introduction

The Svecofennian bedrock in the Uusimaa belt in southernmost Finland is characterized by rather high strain and high grade metamorphism, reaching granulite grade e.g. in the West Uusimaa Complex (Fig. 1; Parras, 1958). There are also areas of lower amphibolite facies assemblages such as the Orijärvi area. Because Orijärvi is an old mine area, early geological studies were focused on ore mineralogy and alteration of the rocks around the mines. Eskola (1914) presented petrology of rocks in the larger Orijärvi area and described also an ultramafic rock at Salittu which he considered a plutonic peridotite.

The structural interpretation of the western Uusimaa belt by Tuominen (1957) included two shear zones, later called the Kisko and Jyly shear zones (Fig. 1). In the 1970's and 1980's the research projects by the Free University of Amsterdam led to a series of papers on metamorphism (Schreurs, 1984; Schreurs, 1985; Schreurs and Westra, 1986; Blom, 1988) and structures (Bleeker and Westra, 1987; Ploegsma and Westra, 1990; Ploegsma, 1991) in the central Uusimaa belt. Ploegsma and Westra (1990) called the triangular area bordered by Kisko and Jyly shear zones the Orijärvi triangle. They considered it a low-strain area that preserved early (D1) folds and was protected from regional deformation during later deformation events (D2, D3). More recently Skyttä et al. (2006) and Pajunen et al. (2008) described profoundly the structures in the Orijärvi triangle and presented slightly different interpretations of structural evolution. The rocks at Salittu, to the northeast of the Orijärvi triangle were studied as well: Schreurs et al. (1986) were the first to consider the ultramafic rocks at Salittu to be of volcanic origin and they presented basic geochemistry of these rocks. Since the 1980's tentative interpretations of structures at Salittu have been given by Skyttä and Mänttäri (2008) and Pajunen et al. (2008).

Because of the relatively low metamorphic grade, low strain and well-preserved early structures in volcanic and volcaniclastic rocks, the Orijärvi area is one of the few regions in southern Svecofennia of Finland where reliable structural evolution and lithostratigraphy can be determined. Väisänen and Mänttäri (2002) presented the stratigraphy of the supracrustal rocks in the Orijärvi area (including Salittu), with four formations: Orijärvi (oldest), Kisko, Toija, and Salittu (youngest), with an age span from 1.90 Ga to 1.88 Ga. The Salittu Formation (SFm) differs from the other ones by the characteristic metapicrites (metapicrite occurs in the Toija Formation as well). The geochemistry of the picritic-basaltic volcanism at Salittu is studied in a companion paper (Nironen, submitted).

Based on remapping of the area where the rocks of the SFm occur we give an interpretation of structural evolution. In addition, we provide additional age data to earlier published ages to constrain the duration of volcanism at Salittu, provide and interpretation of stratigraphy at Salittu, and refine the stratigraphy in the Orijärvi area.

2. Geological setting

The Paleoproterozoic tectonic evolution in Fennoscandia, including the Svecofennian orogeny, has been modeled basically in two ways. In the first semi-continuous, protracted accretion of island arcs to the Archean craton led to southwestward growth of the orogenic belt (Gorbatchev and Bogdanova 1993); in a refinement of this model continuous long-lived subduction occurred with repeated retreating and advancing stages (Hermansson et al., 2008; Stephens and Andersson, 2015). The second model involves accretion of microcontinents (Nironen, 1997), and the orogeny consisted of separate orogenic stages, including microcontinent accretion (1.92–1.87 Ga), continental extension (1.86–1.84 Ga), continental collision (1.84–1.79 Ga), and finally orogenic collapse and stabilization (1.79–1.77 Ga: Lahtinen et al., 2005). According to the second model the Svecofennian orogen in southern Finland may be divided into Central Svecofennia with a microcontinent in the core, and Southern Svecofennia with another microcontinent

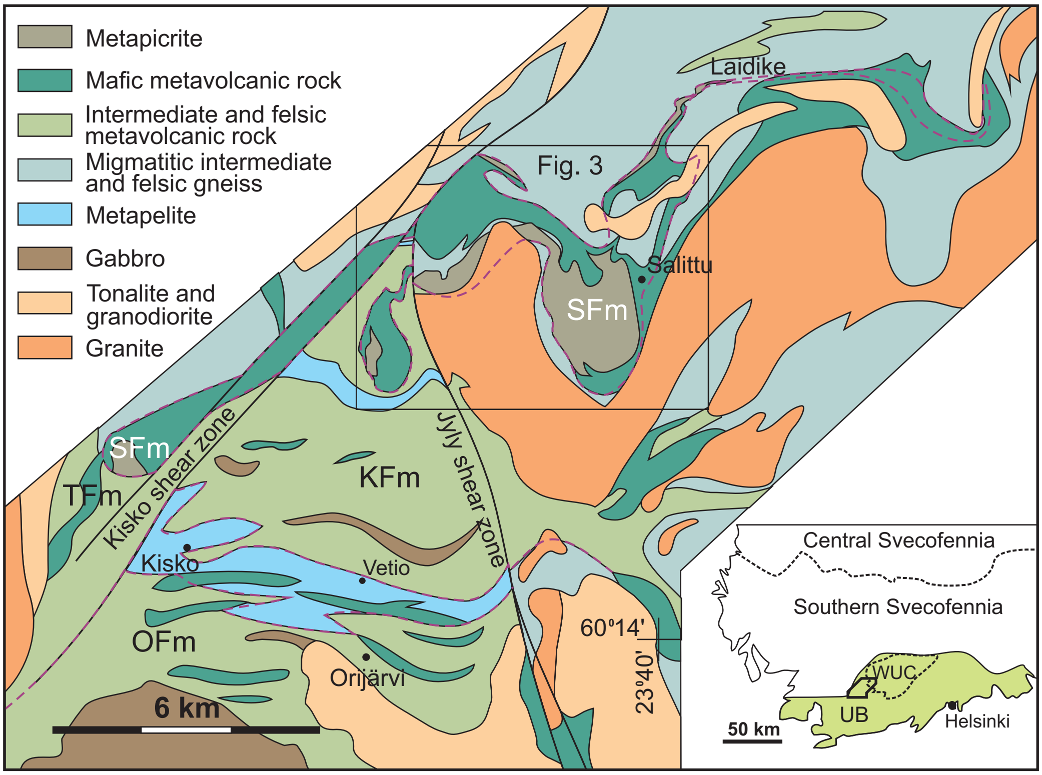

Figure 1. Geological map of the Orijärvi area, modified from Salli (1955) and Väisänen and Mänttäri (2002). OFm = Orijärvi Formation, KFm = Kisko Formation, TFm = Toija Formation, SFm = Salittu Formation. Inset shows location of the Orijärvi area (thick black line) within the Uusimaa belt (UB). WUC = West Uusimaa Complex.

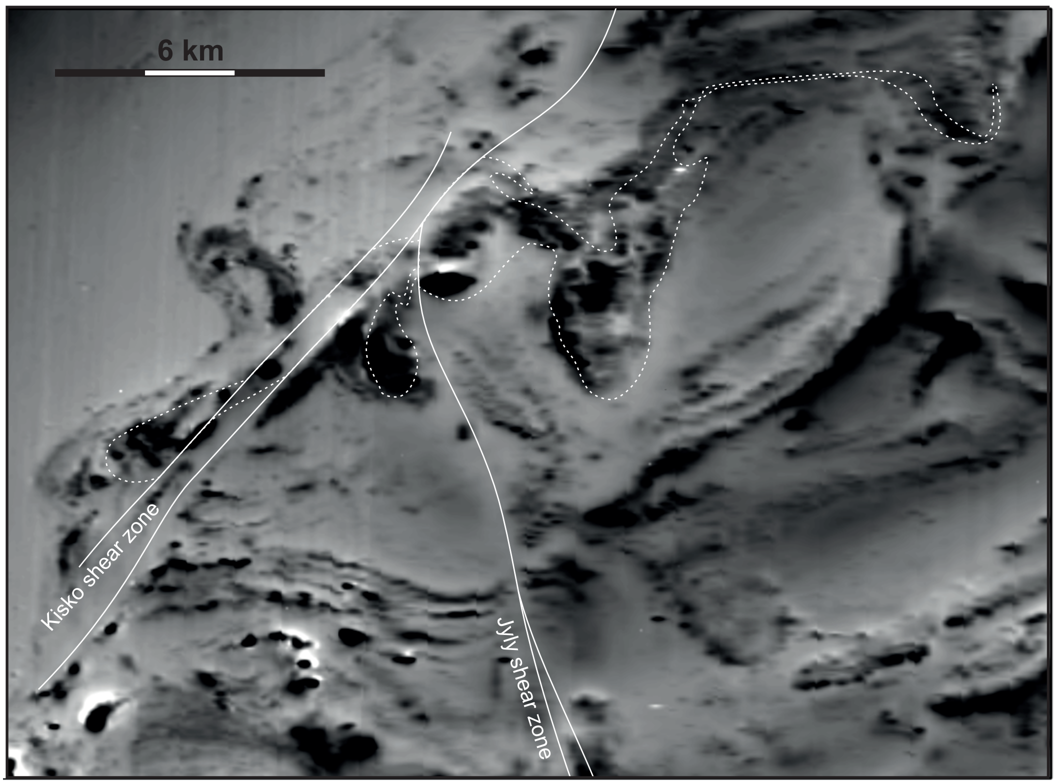

Figure 2. Aeromagnetic map of the Orijärvi area (GTK database). Major shear zones (solid line) and extent of the rocks of the Salittu Formation (broken line) are shown.

(Fig. 1, inset). The difference between these accretionary units is that whereas both display structural and metamorphic evolution attributed to microcontinental accretion, in Southern Svecofennia this evolution is overprinted by another one, attributed to continental collision.

Magmatism and metamorphism during the Svecofennian orogeny in Southern Svecofennia may be divided into the early Svecofennian (1.90–1.87 Ga) and the late Svecofennian (1.84–1.79 Ga) periods. The early Svecofennian period was followed by an intra-orogenic phase of minor magmatism (Bergman et al., 2008; Väisänen et al., 2012). The late Svecofennian granitic magmatism and associated high-temperature, low-pressure type metamorphism overprinted the early Svecofennian magmatism, metamorphism and deformation in a wide belt called the Late Svecofennian Granite-Migmatite zone (Ehlers et al., 1993) that covers southernmost Finland.

The supracrustal rocks in the Uusimaa belt are mainly metavolcanic and volcanic-derived metasedimentary rocks, with epiclastic schists and gneisses as minor constituents. They were metamorphosed during the early Svecofennian period at amphibolite facies conditions (Pajunen et al., 2008). The lower amphibolite facies rocks in the Orijärvi triangle include andalusite-cordierite schists; this area was largely preserved from the late Svecofennian deformation and metamorphism. Skyttä et al. (2006) explained the preservation of the early Svecofennian structures by strain partitioning into high-strain zones (the Jyly and Kisko shear zones; Figs. 1 and 2).

The metamorphic grade increases from the Orijärvi triangle, and most of the rocks at Salittu are migmatitic (Fig. 3). Peak regional metamorphic pressure of 3 kbar (Latvalahti, 1979) and temperature of 600° C (Schumacher and Czank, 1987) have been estimated for the Orijärvi triangle. Schreurs and Westra (1986) suggested an increase in peak metamorphic temperature from 550–650° C in the Orijärvi triangle to 700–825° C in the West Uusimaa Complex at 3–5 kbar pressure. Mouri et al. (2005) estimated 750–800° C and 4–5 kbar conditions in the West Uusimaa Complex during peak of the late Svecofennian metamorphism (1.83–1.81 Ga). According to Schreurs (1985), the increase in metamorphic grade and development of the West Uusimaa Complex was a thermal, virtually isobaric event, and therefore Schreurs and Westra (1986) suggested that the whole region, including the Orijärvi area, represents one crustal level. However, Skyttä et al. (2006) noted an abrupt change in metamorphic grade across the Jyly shear zone and interpreted that the Jyly and Kisko shear zones acted as block margins during the late Svecofennian period, with reverse (east-side-up) dip-slip movement in the Jyly shear zone.

3. Rock types

The SFm consists of pictitic and basaltic metavolcanic rocks. The extent of the SFm in the Orijärvi area is about 20 km (Fig. 1), and considering the metapicrite occurrences farther east the total length exceeds 30 km. In the aeromagnetic map the metavolcanic rocks of the SFm show up as positive anomalies (Fig. 2).

The metabasalt is generally compositionally banded (Fig. 4a) but breccia structures are rather common, and pillow structure occurs locally. The metapicrite varies from homogeneous to fragmental type that probably represents volcanic breccia. Inclusions of metabasalt in the metapicrite (Fig. 4b) show that picrite erupted after basalt and dragged fragments of the basalt into the lava. The boundary of the metapicrite and metabasalt is usually tectonized but sharp where a primary contact can be found (Fig. 4c).

In addition to the metavolcanic rocks, the supracrustal sequence at Salittu consists of felsic-intermediate gneisses and mafic schists (Fig. 3). The gneisses are migmatitic, with leucosome veins usually parallel to compositional layering. Quartz-feldspar gneiss is light-colored and faintly layered (Fig. 4d). As the gneiss does not contain aluminosilicate porphyroblasts, it is probably volcaniclastic in origin. Quartz-feldspar gneiss

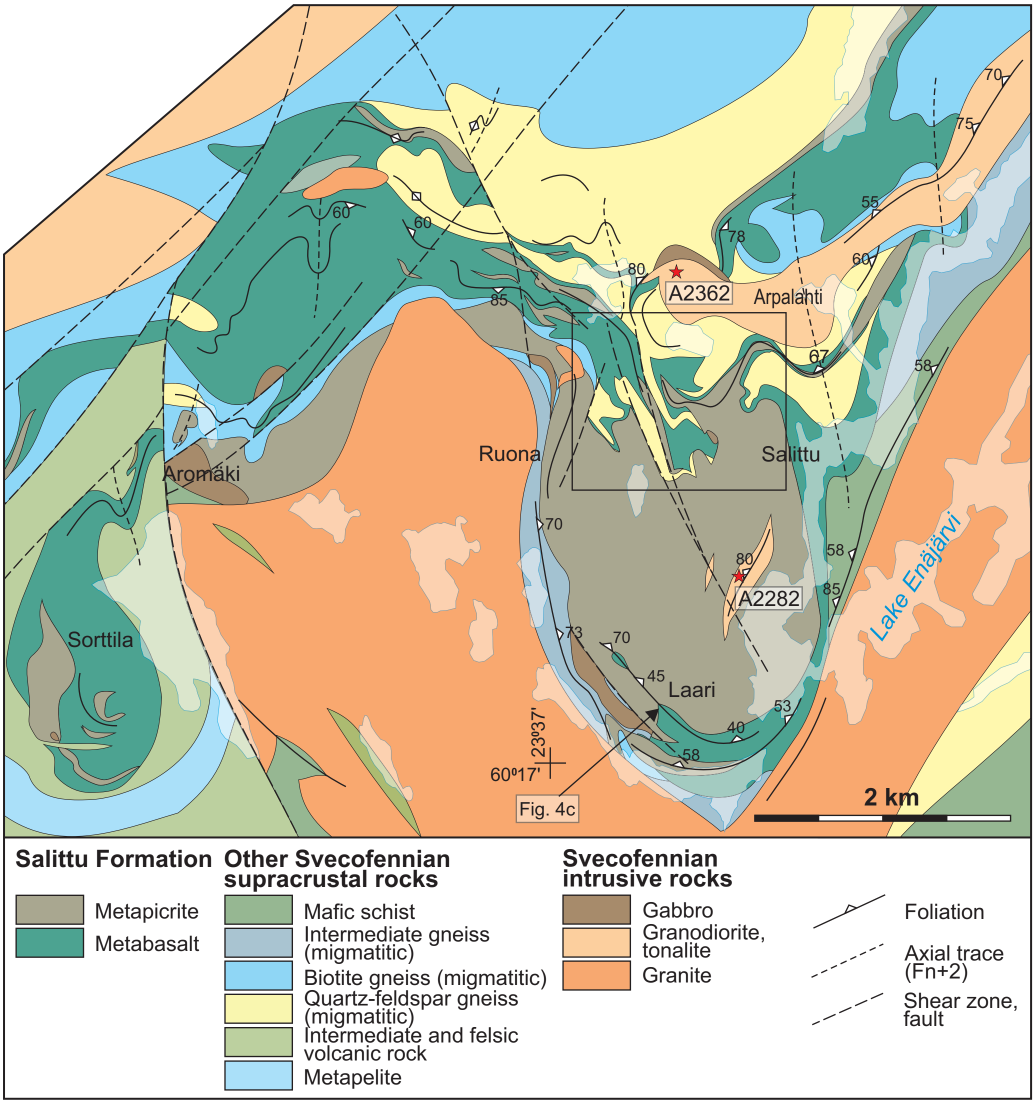

Figure 3. Geological map of the Salittu area, modified from Salli (1955), Schipper (1981), and Väisänen and Mänttäri (2002). Location of Figure 6 is shown by rectangle. Sample sites for age dating are shown by red stars. The foliation form lines show orientation of the prominent foliation (S_n or S_n–S_{n+1} in supracrustal rocks).

grades into cordierite-bearing biotite gneiss, indicating an increasing proportion of weathered material in the protolith. There are also gneisses of intermediate composition with some hornblende as well as diopside- and amphibole-bearing mafic schists; these rocks are likely of volcanic or volcanic-sedimentary origin. Garnet-bearing biotite paragneiss occurs scatteredly and represents epiclastic sedimentary rocks that are scarce at Salittu.

At Sorttila, within the Orijärvi triangle area (Fig. 5), the metavolcanic rocks are much similar to the metavolcanic rocks at Salittu. The rocks surrounding the metavolcanic rocks are intermediate, faintly layered schists, interpreted to have a volcanic or volcanic-sedimentary origin (tuffs or redeposited tuffs). Farther south the schists contain andalusite porphyroblasts indicating sedimentary origin.

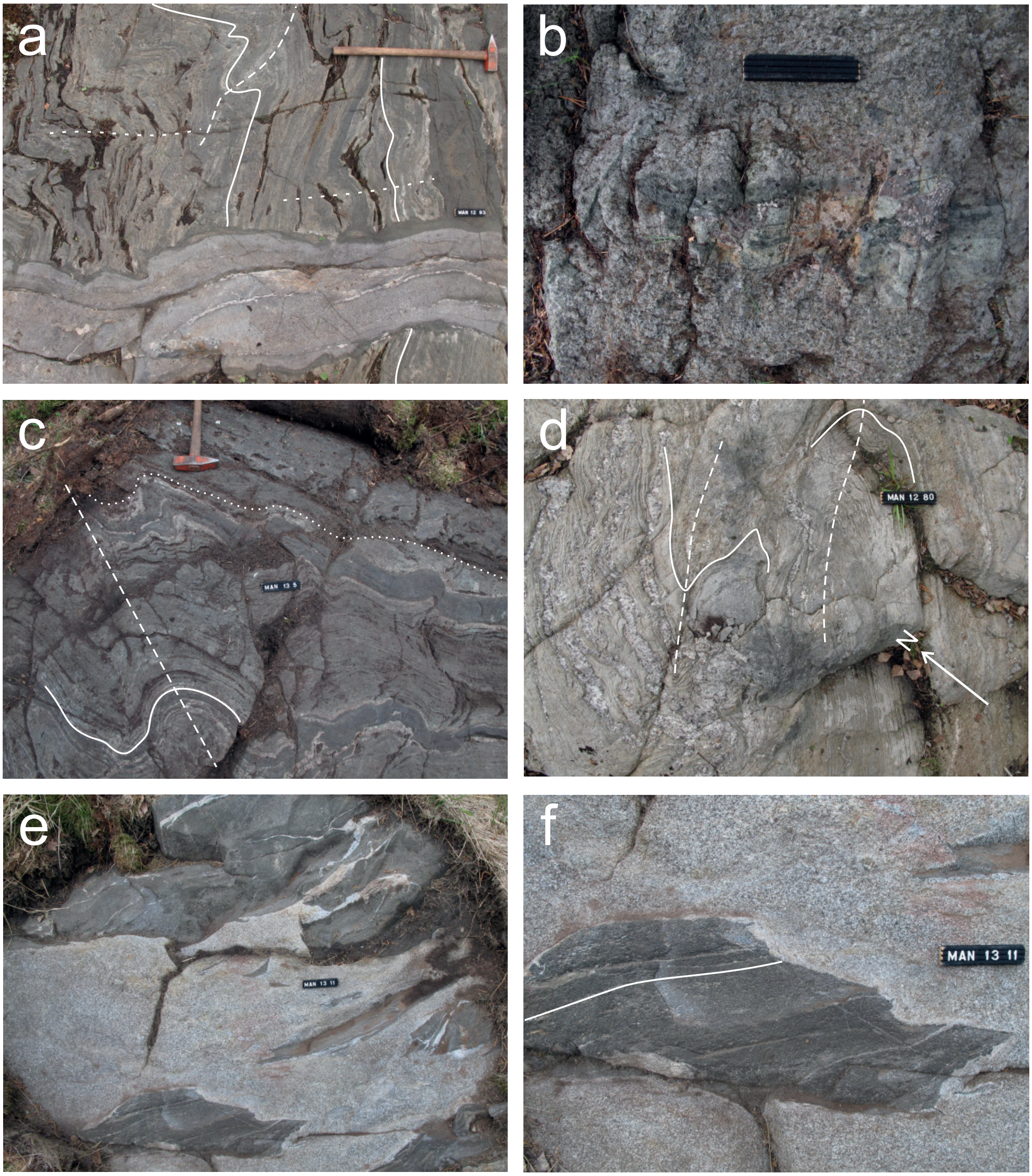

Figure 4. Outcrop photographs from the Salittu area. a. Foliation in metabasalt (S_n, solid white line), parallel to compositional layering, is deformed by tight, sinistral F_{n+1} folding which is folded by open folding (F_{n+2}). F_{n+1} and F_{n+2} axial traces are shown by broken and dotted white lines, respectively. A composite dike crosscuts all structures. Shaft of hammer (length 60 cm) points to north. b. Inclusion of metabasalt in metapicrite. Length of code bar 12 cm. c. Primary contact (dotted white line) between metabasalt of fragmental type (upper part of figure) and metapicrite. Dark alteration bands occur at the base of the metapicrite. Sinistral folding (F_{n+1}) of banding (S_n) is visible. Shaft of hammer points to north. d. Quartz-feldspar gneiss. Foliation parallel to compositional layering (S_n) and leucosome veins are tightly folded (F_{n+1}; axial trace shown). e. Intrusive contact of Arpalahti tonalite with metabasalt. f. Detail of the intrusive contact. Note how the almost undeformed tonalite crosscuts foliation (S_n) in the metabasalt.

Gabbroic bodies occur within and adjacent to the metavolcanic rocks. The largest body within the metavolcanic rocks at Aromäki (Fig. 3) is medium-grained and unfoliated. Gradational contacts to metapicrite and similar geochemistry (Nironen, submitted) suggest that the Aromäki gabbro is synvolcanic with respect to metapicrite. In contrast, the body at Laari consists of fine- to medium-grained, foliated intrusive rock that varies in composition from gabbro to diorite and is crosscut by the metapicrite.

An elongate body in the northern part of the Salittu area (Fig. 3), here named as the Arpalahti tonalite, crosscuts the metabasalt of the SFm (Figs. 4e, 4f). The tonalite is homogeneous, medium-grained, generally foliated but less foliated at the ends of the elongate body. A sample (A2362) was taken from the tonalite for dating. Another sample (A2282) was taken from a foliated, medium-grained granodiorite, here named as the Laari granodiorite, occurring within the metapicrite (Fig. 3).

A large, elongate granite body occupies the southern part of the study area and crosscuts all the supracrustal rocks (Fig. 1). It is medium- to coarse-grained and almost unfoliated except for a few shear zones. In addition, small granite bodies crosscut the ductile structures in the metavolcanic rocks but some of these granites are deformed by faults (Fig. 6).

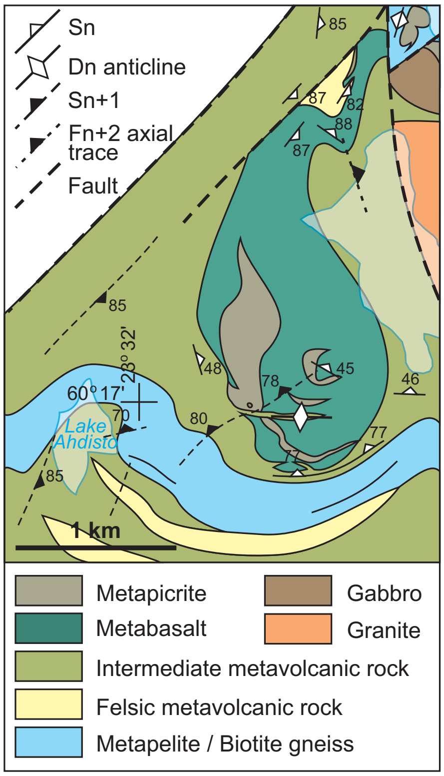

Figure 5. Geological map of the Sorttila area within the Orijärvi triangle, based on observations of this study and GTK bedrock database.

4. Structure

The Salittu Formation trends approximately ENE–WSW (Fig. 1). A small area in the center of the Salittu area, with good outcrop density and easily distinguishable rock types, was studied in detail to understand the structure (Fig. 6).The quartz-feldspar gneiss and biotite gneiss exhibit a weak compositional layering, and leucosome veins are (sub)parallel to the layering. Generally the amount of mica is small in the gneisses but a schistosity (S_n) can be seen in more mica-bearing layers (mainly in the biotite gneiss). Compositional layering in the metabasalt has the same orientation as layering in the adjacent quartz-feldspar gneiss, and is therefore considered to define D_n foliation.

S_n foliation and leucosome veins in the gneisses are tightly to isoclinally folded (F_{n+1}). A faint, steeply dipping to vertical S_{n+1} foliation can be identified in the axial plane in the mica-bearing layers (Fig. 4d). The F_{n+1} fold axis plunges vary from subhorizontal to steep, and a mineral lineation (L^n_{n+1}) is parallel to the fold axis. F_{n+1} folding is visible also in the metabasalt (Fig. 4a).

The F_{n+1} folding is deformed by upright, open to tight folding (F_{n+2}) with NNW–SSE to NNE–SSW striking subvertical axial planes (Figs. 4a and 6). Based on interpretation of the structures in the area of Figure 6, the synformal structure at Laari results from interference of two, almost orthogonal foldings F_{n+1} and F_{n+2}.

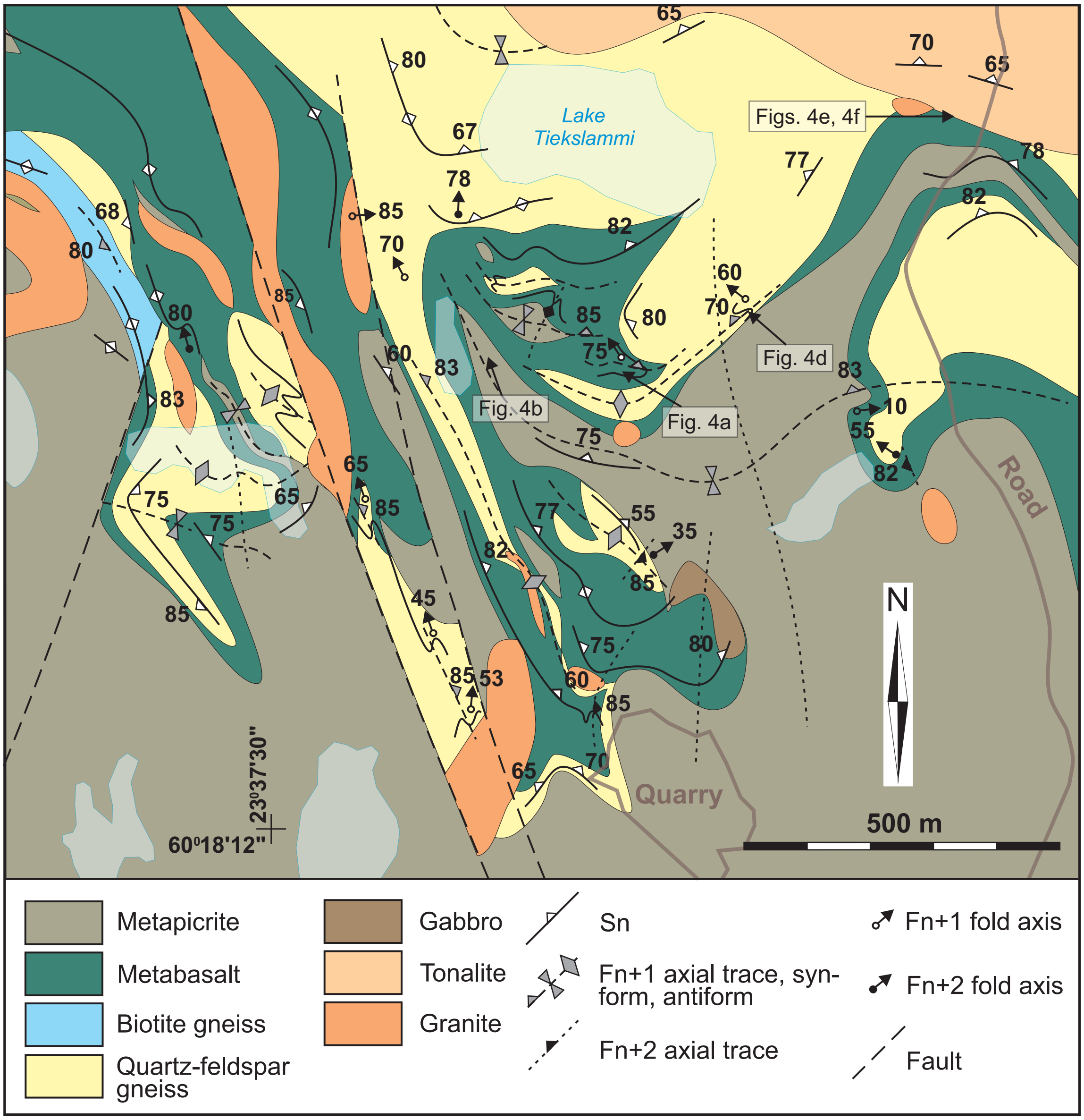

Figure 6. Structural interpretation of the Ruona-Arpalahti area.

At Sorttila, the structural sequence is correlative to that at Salittu. The intermediate metavolcanic schist displays a rather weak S_n foliation, and compositional layering in the metabasalt is parallel to S_n (Fig. 5). Rather open F_{n+1} folding deforms S_n in the schist and compositional layering in the metabasalt, with development of a schistosity in the axial plane. In the southern part of the area the intermediate schist crops out as an E–W directed narrow lense within the metabasalt. Since S_{n+1} foliation is visible within this lense, the structure is interpreted as a D_n anticline. In the northeastern part of Sorttila, S_n compositional layering in the metabasalt is tightly folded, with axial plane in NNE–SSW direction. This folding is interpreted as F_{n+2}.

The sharply crosscutting contact of the Arpalahti tonalite to the foliated (S_n) metabasalt

(Fig. 4f) indicates that the tonalite intruded during a late stage or after D_n. The tonalite is foliated (Figs. 2 and 6) but it is difficult to assess whether this foliation correlates with S_n or S_{n+1} in the supracrustal rocks because the two are parallel in many places. The form of the tonalite body suggests that it was deformed during D_{n+2} (Fig. 1).

An unfoliated composite (mafic) dike was found to sharply crosscut the foliation and folding in the metabasalt (Fig. 4a). The subvertical dike is oriented parallel to F_{n+2} fold axial plane, suggesting emplacement during D_{n+2}.

Numerous faults at Salittu appear as discontinuities in lithology (Fig. 6) and as linear depressions in topography. They are exposed on few outcrops where they occur as narrow mylonitic zones. Larger shear zones were interpreted on the basis of the aeromagnetic data (Fig. 2) and discontinuities in lithology.

To the north of Salittu, at Laidike (Fig. 1), there is tight folding with E-W axial trace. The relative age of this folding is open: it may be F_{n+1} folding, or it may be the youngest deformation in the area.

5. U–Pb zircon dating

5.1. Analytical methods

Zircon for LA-MC-ICPMS (Laser Ablation Multi-collector Inductively Coupled Plasma Mass Spectrometry) U–Pb dating were selected by hand-picking after heavy liquid and Frantz magnetic separation. The chosen grains were mounted in epoxy resin and sectioned approximately in half and polished. Back-scattered electron images (BSE) and cathodo luminescence (CL) images were taken using SEM (Scanning Electron Microscope) to target the spot analysis sites on mineral grains.

U–Pb dating analyses were performed using a Nu Plasma HR multicollector ICPMS at the Geological Survey of Finland in Espoo using a technique very similar to Rosa et al (2009) except that Analyte G2 193 nm laser laser microprobe was used. The analyses were made in static ablation mode with beam di- ameter of 20 μm, pulse frequency of 10 Hz, and beam energy density of 2.07 J/cm^2. A single U–Pb measurement included 30 s of on-mass background measurement, followed by 60 s of ablation with a stationary beam. Masses 204, 206 and 207 were measured in secondary electron multipliers and 238 in the extra high mass Faraday collector. Ion counts were converted and reported as volts by the Nu Plasma timeresolved analysis software. ^{235}U was calculated from the signal at mass 238 using a natural ^{238}U/^{235}U=137.88. Mass number 204 was used as a monitor for ^{204}Pb. Raw data were corrected for background, laser induced elemental fractionation, mass discrimination, and drift in ion counter gains and reduced to U–Pb isotope ratios by calibration to concordant reference zircon of known age, using protocols adapted from Andersen et al. (2004) and Jackson et al. (2004). Standard zircon GJ-01 (609 ± 1 Ma; Belousova et al., 2006) and an in-house standard zircon A1772 (2711 ± 3 Ma/TIMS; 2712 ± 1 Ma/SIMS; Huhma et al., 2012) were used for calibration. In addition, A382 (1877±2 Ma; Huhma et al., 2012) was measured as a quality control sample to check the calibration (1882 ± 6 Ma; MSWD of concordance = 4.8; Mean ^{207}Pb/^{206}Pb age = 1880 ± 5 Ma; MSWD = 0.95; n=25/25. MSWD = Mean Square Weighted Deviation). Age related common lead (Stacey and Kramers, 1975) correction was used when the analysis showed common lead contents above the detection limit. The calculations were done off-line, using an interactive spreadsheet program written in Microsoft Excel/ VBA by Tom Andersen (Rosa et al, 2009). To compensate for drift in instrument sensitivity and Faraday vs. electron multiplier gain during an analytical session, a correlation of signal vs. time was assumed for the reference zircons. A description of the algorithms used is provided in Rosa et al (2009).

The age calculations and plotting of the U–Pb isotope data were performed using the Isoplot/ Ex 3 program (Ludwig, 2003). All the ages were calculated with 2σ errors and without decay constant errors. Data-point error ellipses in the figures are at the 2 σ level.

Table 1. LA-MC-ICPMS zircon U-Pb data, Laari granodiorite and Arpalahti tonalite.

	Ages						*Ratios*											
Sample/ spot # 4)	$^{207}Pb/^{206}Pb$	±σ	$^{207}Pb/^{235}U$	±σ	$^{206}Pb/^{238}U$	±σ	$^{207}Pb/^{206}Pb$	±σ	$^{207}Pb/^{235}U$	±σ	$^{206}Pb/^{238}U$	±σ	ρ 1)	Disc. % 2)	U ppm	^{206}Pb ppm	$^{206}Pb/^{204}Pb$ measured	f_{206} % 3)
A2282 Laari granodiorite, Salittu																		
A2282_01a	1875	17	1897	30	1917	56	0,11467	0,00111	5,47436	0,19171	0,34626	0,01166	0,96	2,6	650	304	3,98E+05	0,00
A2282_02a	1857	17	1937	31	2014	59	0,11354	0,00111	5,74021	0,2047	0,36666	0,01257	0,96	9,8	606	300	8,33E+03	0,20
A2282_03a	1867	18	1882	33	1895	61	0,11416	0,00114	5,38019	0,20595	0,34182	0,01263	0,97	1,8	344	157	1,69E+04	0,10
A2282_04a	1873	18	1886	29	1898	54	0,11459	0,00112	5,40776	0,18575	0,34228	0,01127	0,96	1,5	488	226	5,58E+04	0,00
A2282_05a	1848	17	1914	34	1976	65	0,11301	0,00114	5,58777	0,22153	0,35861	0,01375	0,97	8	772	375	3,96E+04	0,10
A2282_06a	1878	17	1885	32	1891	59	0,1149	0,00117	5,40125	0,20304	0,34095	0,01234	0,96	0,8	446	203	2,76E+04	0,00
A2282_07a	1849	18	1908	34	1964	65	0,11303	0,00115	5,54948	0,21923	0,35608	0,01359	0,97	7,2	1011	478	1,19E+05	0,00
A2282_08a	1862	18	1913	30	1960	57	0,11387	0,00113	5,57782	0,19749	0,35527	0,01208	0,96	6,1	519	245	5,86E+04	0,00
A2282_09a	1852	17	1883	30	1911	55	0,11323	0,00112	5,38613	0,18778	0,345	0,01153	0,96	3,7	762	354	5,31E+04	0,00
A2282_10a	1843	17	1916	32	1985	60	0,11268	0,00113	5,60082	0,20558	0,36049	0,01273	0,96	8,9	756	362	5,84E+04	0,00
A2282_11a	1922	17	1915	33	1910	62	0,1177	0,00122	5,59545	0,21724	0,3448	0,0129	0,96	-0,7	491	227	2,52E+04	0,01
A2282_12a	1875	17	1888	33	1900	60	0,11468	0,00121	5,42167	0,20626	0,34287	0,01254	0,96	1,6	372	167	2,48E+04	0,00
A2282_13a	1862	18	1908	31	1950	58	0,11387	0,00116	5,5444	0,19832	0,35313	0,01211	0,96	5,4	819	378	6,82E+04	0,00
A2282_14a	1867	18	1895	30	1921	56	0,11417	0,00116	5,46497	0,19227	0,34717	0,01169	0,96	3,4	474	218	2,51E+04	0,00
A2282_15a	1866	18	1877	31	1886	58	0,11414	0,00118	5,34853	0,19652	0,33986	0,01199	0,96	1,2	464	204	3,99E+04	0,00
A2282_16a	1903	19	1896	37	1889	67	0,11646	0,00128	5,46795	0,23304	0,34052	0,01402	0,97	-0,8	447	201	7,53E+03	0,33
A2282_17a	1887	19	1909	34	1929	63	0,11545	0,00123	5,55415	0,21838	0,34892	0,01321	0,96	2,6	502	226	2,93E+04	0,00
A2282_18a	1878	19	1851	37	1826	67	0,1149	0,00123	5,1881	0,22457	0,32749	0,01374	0,97	-3,2	673	297	1,18E+05	0,00
A2282_19a	1900	18	1963	34	2023	65	0,1163	0,0012	5,91236	0,2297	0,3687	0,01381	0,96	7,6	328	155	8,79E+03	0,21
A2282_20a	1871	18	1859	32	1849	58	0,11443	0,00122	5,23955	0,19813	0,3321	0,01205	0,96	-1,4	567	250	2,79E+04	0,03
A2362 Arpalahti tonalite, Salittu																		
A2362-01a	1893	12	1904	32	1915	61	0,11581	0,00081	5,52316	0,20703	0,34589	0,01274	0,98	1,4	261	113	2,56E+04	0,00
A2362-01b	1898	13	1914	30	1929	56	0,11614	0,00085	5,58698	0,1928	0,34889	0,01177	0,98	1,9	301	131	4,31E+04	0,08
A2362-02a	1870	13	1874	32	1878	60	0,11434	0,00083	5,33294	0,19885	0,33827	0,01237	0,98	0,5	165	69	3,08E+04	0,18
A2362-03a	1916	12	1871	28	1831	50	0,11733	0,00087	5,31375	0,17172	0,32848	0,01033	0,97	-5,1	177	72	1,90E+04	0,00
A2362-03b	1867	12	1916	54	1961	106	0,11418	0,00075	5,59663	0,35236	0,3555	0,02226	1,00	5,8	120	36	1,52E+04	0,11
A2362-04a	1894	14	1891	33	1888	62	0,1159	0,00093	5,43847	0,21125	0,34032	0,01293	0,98	-0,3	141	58	1,32E+04	0,00
A2362-04b	1874	13	1892	29	1908	54	0,11464	0,00084	5,44572	0,18247	0,34452	0,01126	0,98	2,1	407	174	3,62E+04	0,00
A2362-05a	1898	13	1892	30	1887	55	0,11614	0,00092	5,44656	0,18814	0,34011	0,01144	0,97	-0,6	94	39	8,55E+03	0,00
A2362-05b	1890	14	1879	29	1869	54	0,11567	0,0009	5,36322	0,18343	0,33628	0,0112	0,97	-1,3	229	95	1,57E+04	0,00
A2362-06a	1877	14	1886	28	1894	52	0,1148	0,0009	5,40749	0,17571	0,34163	0,01077	0,97	1,1	362	153	2,93E+04	0,00
A2362-07a	1911	20	1896	34	1882	63	0,11699	0,0014	5,46874	0,21948	0,33902	0,01299	0,96	-1,7	127	54	8,21E+03	0,72
A2362-07b	1865	13	1859	46	1854	85	0,11405	0,00089	5,23994	0,28005	0,33321	0,01762	0,99	-0,7	135	38	1,04E+04	0,17
A2362-08a	1885	13	1880	28	1875	52	0,11534	0,00086	5,36721	0,17498	0,33748	0,01071	0,97	-0,7	331	137	2,70E+04	0,00
A2362-09a	1854	15	1969	30	2080	60	0,11339	0,00092	5,95307	0,20761	0,38076	0,01291	0,97	14,2	323	152	3,09E+04	0,01
A2362-10a	1882	11	1897	58	1910	112	0,11511	0,00071	5,47386	0,37255	0,34488	0,02337	1,00	1,7	104	33	1,66E+04	0,00
A2362-11a	1881	11	1882	57	1884	108	0,11509	0,00073	5,38483	0,35816	0,33935	0,02247	1,00	0,1	167	52	2,48E+04	0,00
A2362-12a	1878	11	1875	57	1872	108	0,11486	0,00071	5,33657	0,3557	0,33696	0,02236	1,00	-0,3	124	38	1,34E+04	0,03
A2362-13a	1885	11	1916	58	1946	113	0,1153	0,00074	5,60161	0,3785	0,35236	0,0237	1,00	3,8	100	32	1,52E+04	0,00
A2362-14a	1881	11	1892	57	1902	109	0,1151	0,00072	5,44621	0,36112	0,34319	0,02265	1,00	1,3	94	29	1,89E+04	0,00
A2362-15a	1882	11	1885	56	1887	107	0,11516	0,00072	5,39844	0,35597	0,33998	0,02232	1,00	0,3	133	41	1,75E+04	0,00
A2362-16a	1888	11	1884	56	1881	107	0,11552	0,00073	5,39769	0,35475	0,3389	0,02217	1,00	-0,4	99	30	1,25E+04	0,00
A2362-17a	1872	11	1900	57	1926	109	0,11453	0,00071	5,49734	0,36159	0,34812	0,0228	1,00	3,3	217	68	2,32E+04	0,03
A2362-18a	1883	11	1883	56	1884	106	0,11519	0,00072	5,3909	0,35155	0,33943	0,02203	1,00	0,1	129	39	1,89E+04	0,00
A2362-19a	1888	11	1892	56	1896	107	0,11553	0,00073	5,44605	0,35549	0,34189	0,02221	1,00	0,5	136	41	1,32E+04	0,01
A2362-19b	1875	11	1897	55	1917	106	0,11472	0,00075	5,47624	0,35165	0,34622	0,02212	1,00	2,5	351	108	5,61E+04	0,00
A2362-20a	1882	12	1858	53	1837	99	0,11514	0,00077	5,23398	0,32471	0,32968	0,02034	0,99	-2,8	148	42	1,52E+04	0,00
A2362-20b	1869	12	1883	54	1896	102	0,1143	0,00076	5,39062	0,33708	0,34204	0,02127	0,99	1,7	272	80	4,44E+04	0,03
A2362-21a	1876	11	1859	52	1845	97	0,11472	0,00076	5,24123	0,31922	0,33135	0,02006	0,99	-1,9	134	38	1,55E+04	0,00
A2362-21b	1874	11	1883	52	1891	100	0,1146	0,00075	5,38553	0,32937	0,34085	0,02072	0,99	1,1	369	107	1,14E+05	0,00
A2362-22a	1869	11	1925	55	1977	108	0,1143	0,00074	5,65604	0,35918	0,35888	0,02267	1,00	6,7	433	132	3,67E+08	0,00
A2362-23a	1834	11	1827	52	1821	97	0,11213	0,00073	5,04624	0,30958	0,32641	0,01991	0,99	-0,8	173	48	6,65E+03	0,27
A2362-24a	1877	11	1879	53	1882	100	0,11479	0,00075	5,36488	0,33209	0,33896	0,02087	0,99	0,3	262	76	3,08E+04	0,00
A2362-25a	1866	11	1880	53	1893	101	0,11411	0,00075	5,37167	0,33281	0,3414	0,02103	0,99	1,7	161	47	1,56E+04	0,03
A2362-26a	1850	12	1856	52	1861	97	0,1131	0,00077	5,21875	0,31636	0,33467	0,02016	0,99	0,7	178	50	9,46E+03	0,21
A2362-27a	1884	11	1926	54	1966	106	0,11526	0,00075	5,66653	0,35491	0,35656	0,02221	1,00	5	148	45	2,08E+04	0,00

All errors are in 1 sigma level. 1) Error correlation in conventional concordia space. 2) Age discordance at closest approach of 1 sigma error ellipse to concordia. 3) Percentage of common ^{206}Pb in measured ^{206}Pb, calculated from the ^{204}Pb signal assuming a present-day Stacey & Kramers (1975) model terrestrial Pb-isotope composition. 4) In BSE/Cl images, all the analysed zircon domains in sample A2282 are texturally similar and in sample A2362 the xxa refers BSE darker center and xxb the paler outer/rim domains.

5.2. Results

The zircon population from the A2282 Laari granodiorite is very homogeneous consisting mostly of prismatic (length : width = 2–6) and transparent grains. Magmatic zoning is visible especially in the rather dark Cl images (Fig. 7). Twenty zircon grains were dated using LA-MC-ICPMS (Table 1). On the concordia diagram (Fig. 7a), the U-Pb data plot in a tight cluster defining a 1876 ± 8 Ma age with a high MSWD of concordance. Therefore the mean Pb–Pb age of 1871 ± 9 Ma (Fig. 7b) is considered the best age estimate for the timing of granodiorite crystallization.

Zircon in the A2362 Arpalahti tonalite is transparent to translucent and stubby. The BSE and CL images usually show BSE-darker/CL-paler center and BSE-paler/CL-darker outer/rim domains. In the CL images, mixed type (oscillatory and sector zoning) growth zoning is visible (Fig. 8). Altogether 35 zircon domains from 27 grains were U-Pb dated (Table 1). Ages from the center and outer/rim zircon domains are coeval within the error limits. On the concordia diagram (Fig. 9), all the U-Pb data plot in a tight cluster determining an age of 1878 ± 5 Ma for the tonalite crystallization.

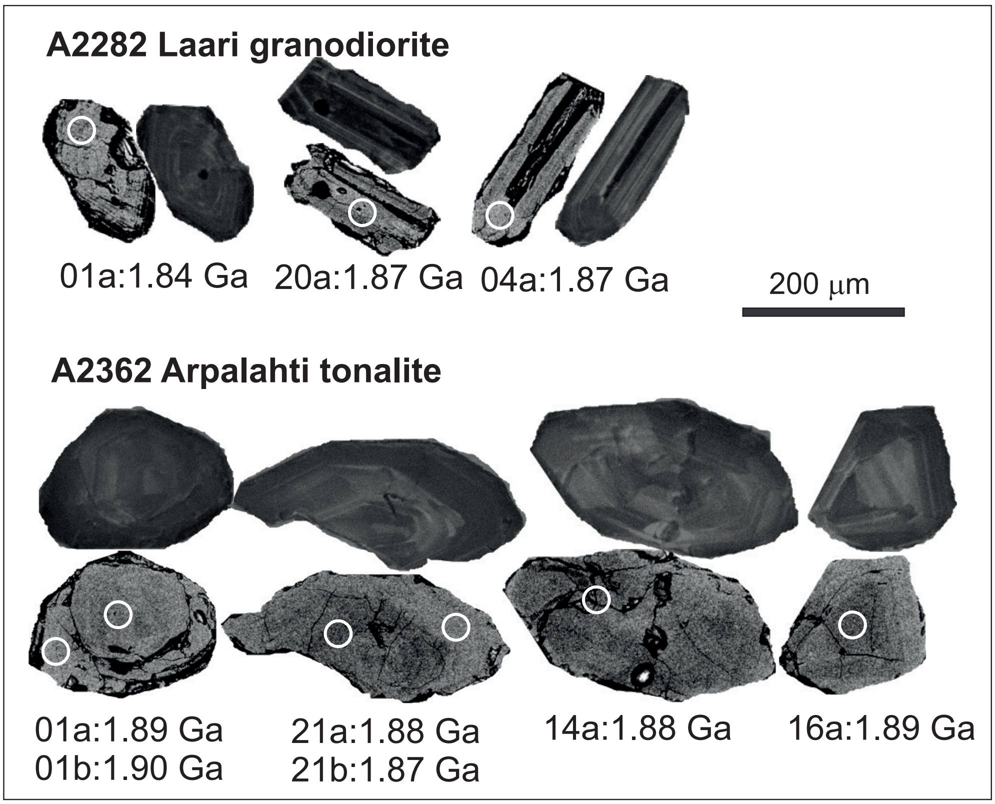

Figure 7. Representative BSE and CL images of U–Pb dated zircon grains from Salittu. The analysis spot sites on the BSE images, corresponding analysis numbers and resulted $^{207}Pb/^{206}Pb$ ages are indicated. See Table 1.

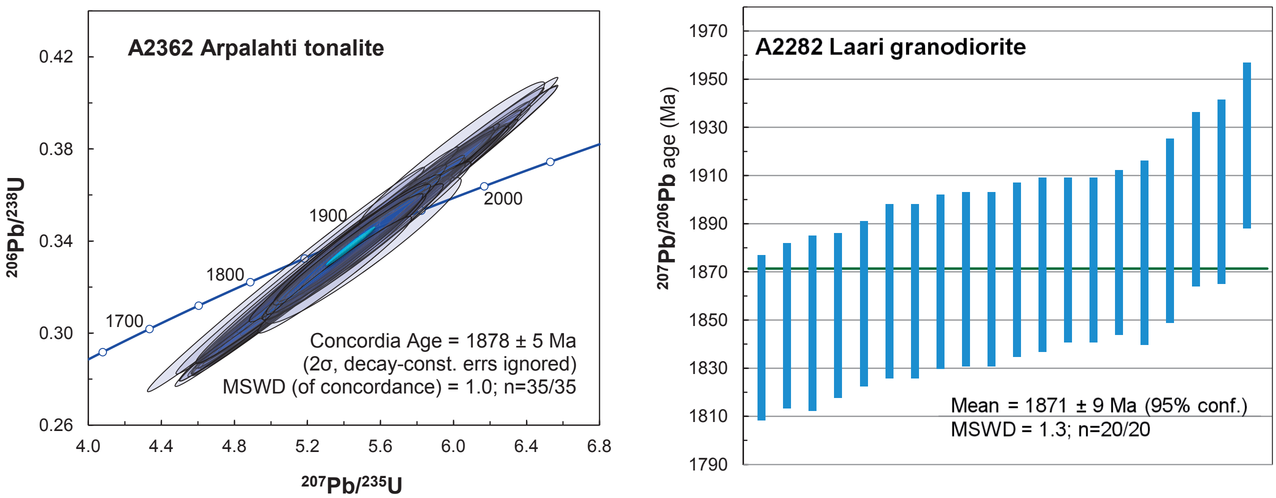

Figure 8. U-Pb age data of the Laari granodiorite. a) Concordia plot. b) $^{207}Pb/^{206}Pb$ age distribution and mean age.

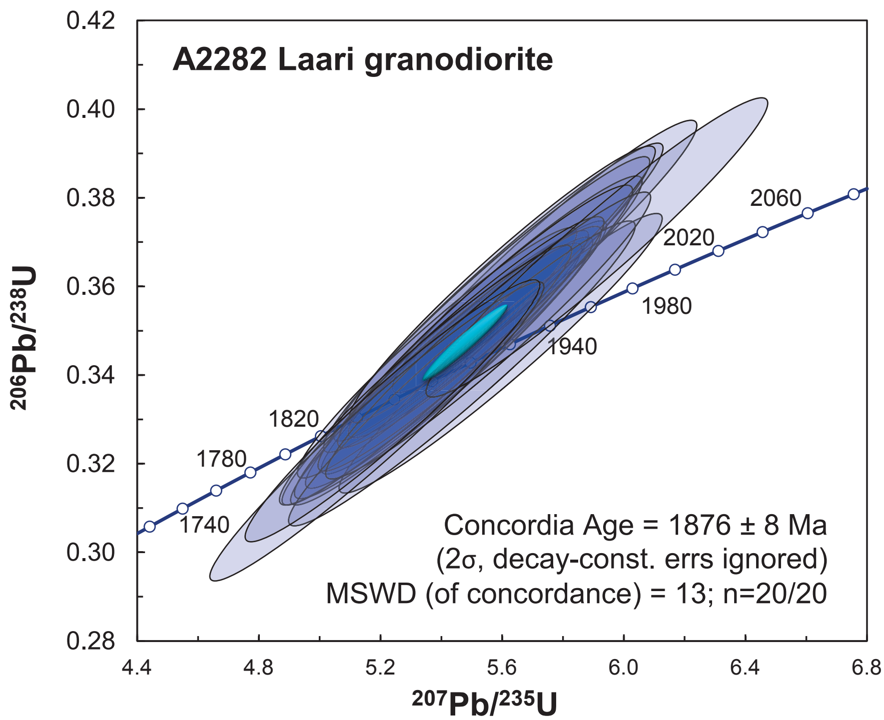

Figure 9. Concordia plot of the Arpalahti tonalite.

Table 2. Interpretations of deformation events, their styles and orientations of foliations and axial traces in the Orijärvi area.

Ploegsma & Westra 1990	Skyttä et al. 2006	Skyttä & Mänttäri 2008	Pajunen et al. 2008	Present study
Early Svecofennian (1.90–1.87 Ga)				
D1 (subhorizontal, thrusting)	D1 (subhorizontal, thrusting)	D1 (subhorizontal, thrusting)	D_A (horizontal, thrusting)	Dn (subhorizontal, thrusting? - upright)
D2	D2 (upright, ENE-WSW)	D2	D_B-D_C-D_D (extension - compression - transpression)	Dn+1 (upright)
Late Svecofennian (1.84–1.79 Ga)				
		D3 (subhorizontal, extension)	D_G-D_H (upright, E-W - horizontal, extension)	
	D3 (upright, E-W)	D4 (upright, ENE-WSW)	D_H (transpression, upright, NE-SW)	
	D4 (upright N-S & NE, NW shear zones)	D5 (NE, NW shear zones)	D_I (upright, N-S)	Dn+2 (upright, N-S)
Post-Svecofennian (1.53 Ga)				
D3 (upright, N-S)				

6. Discussion

6.1. Structural correlation

Skyttä et al. (2006) presented a structural sequence for the Orijärvi area with two early Svecofennian and two late Svecofennian deformation events D_1–D_2 and D_3–D_4, respectively (Table 2). In their sequence S_1 is a weak schistosity subparallel to bedding and possibly associated with thrusting during D_1; D_2 deformation occurred at 1.88–1.87 Ga Ma and caused the penetrative S_2 foliation with ENE–WSW trend and development of a D_2 synform in the northern part of the Orijärvi triangle (Orijärvi sub-area in their terminology, including Sorttila). Subhorizontal D_2 structures were refolded into upright folds in the southern part of the Orijärvi triangle during D_3 deformation. The Kisko and Jyly shear zones and the different crustal sections in the present erosion surface are the result of E–W contraction during retrograde D_4 deformation that continued in brittle regime. Later Skyttä and Mänttäri (2008) refined the deformational sequence over a larger area by adding a crustal extension stage D_3 to the beginning of late Svecofennian deformation, at 1835–1825 Ma. According to Skyttä and Mänttäri (2008), the area east of the Jyly shear zone (including the Salittu area) is dominated by upright D_4 structures with ENE–WSW axial planes. D_4 deformation started by N–S contraction, and counterclockwise rotation of contraction caused strain localization into the subvertical D_5 Kisko and Jyly shear zones that were formed after 1820 Ma.

In the structural scheme of southern Finland by Pajunen et al. (2008) the western part of the Uusimaa belt is characterized by late Svecofennian dome-and-basin interference structure ($D_{G+H\pm I,}$ see Table 2).

Väisänen and Skyttä (2007) found evidence for contrasting movement senses e.g. in the Jyly shear zone. They proposed that the late Svecofennian oblique-slip reverse faulting (sinistral, east-side-up) caused a positive flower structure in the Jyly shear zone, followed by extensional displacement (east-side-down) in brittle conditions. Väisänen and Skyttä (2007) tentatively suggested such an extension at 1.79–1.77 Ga.

The occurrence of metabasalt inclusions in the metapicrite (Fig. 4b) and the rock type distribution in Figures 5 and 6 imply that the original stratigraphic succession is essentially preserved at Salittu and Sorttila: metapicrite overlies metabasalt, and metavolcanic schists and gneisses are below these rocks. The oldest structure at Salittu is the penetrative composite layering S_n that can be identified in the gneisses and in the metabasalts. Preservation of the original stratigraphy suggests that the volcanic sequence remained flat-lying during D_n deformation. However, the style of D_n is unclear; it may have been thrusting (see Table 2). D_n anticline at Sorttila may result from upright, rather open F_n folding during a late stage of progressive D_n deformation. S_n and S_{n+1} at Sorttila are the same as S_1 and S_2 of Skyttä et al. (2006). We interpret that S_n and S_{n+1} at Salittu may be correlated with S_1 and S_2 as well. The leucosome veins in the gneisses probably developed along S_0/S_1 surfaces during early stages of D_2 deformation, and were folded during progressive D_2.

We did not find evidence of D_3/D_H crustal flattening that was suggested to have initiated late Svecofennian deformation (Skyttä and Mänttäri 2008). Neither did we find evidence of large-scale, open, upright D_4 folding with ENE–WSW trending axial planes which was supposed to characterize the Salittu area; however the tight folding at Laidike, with E–W trending axial planes, is possibly D_4 deformation. F_{n+2} folding with N–S trending axial planes corresponds to F_5 and D_I of Skyttä and Mänttäri (2008) and Pajunen et al. (2008), respectively.

Structural interpretation of the Orijärvi area is shown in Figure 10. The structure at Sorttila was interpreted as D_2 synform by Skyttä et al. (2006) but the ovoidal form of exposed metavolcanic rocks suggest the effect of F_5 folding. We interpret that the synforms at Laari and Sorttila are the result of two almost orthogonal folding phases, early Svecofennian (D_2) and late Svecofennian (D_5). If this interpretation is correct, the Orijärvi triangle was not totally preserved from late Svecofennian deformation (cf. Ploegsma and Westra, 1990).

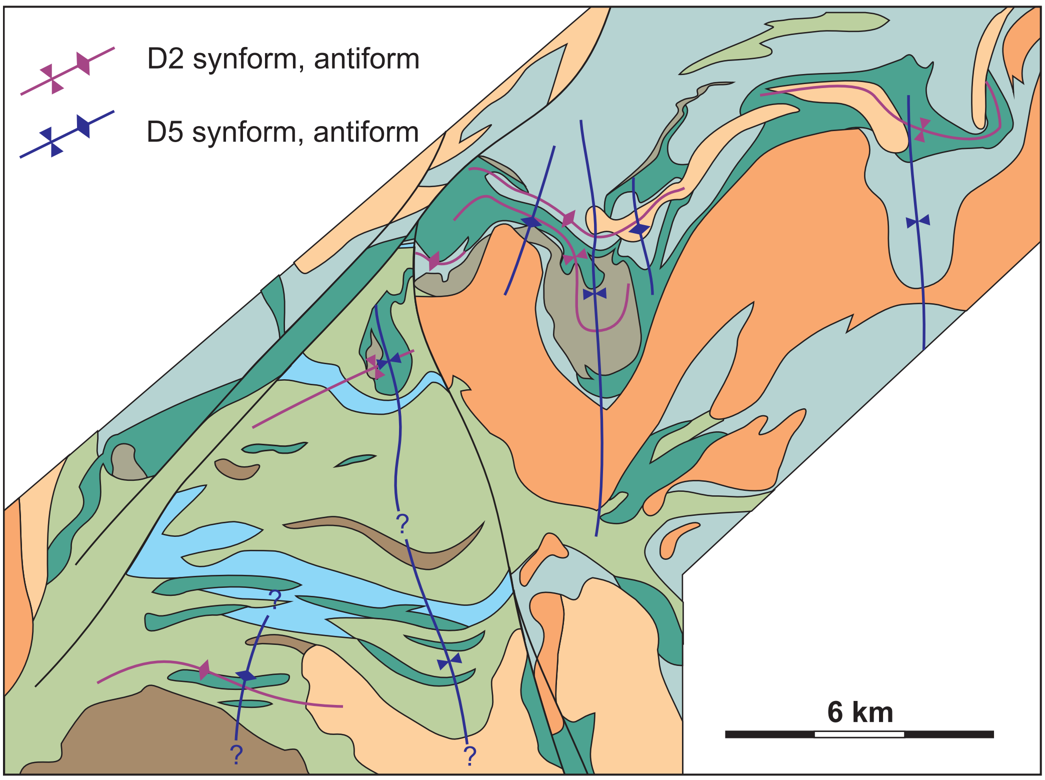

Figure 10. Structural interpretation of the Orijärvi area. Rock types as in Figure 1.

6.2. Metamorphism and structural evolution

As noted by Blom (1988), the S_1-S_2 foliation is in places orthogonal to the metamorphic isogrades, showing that the late Svecofennian thermal pulse overprinted the early Svecofennian structural grain. At Salittu the segregation of melt parallel to S_1 foliation is interpreted to be an early Svecofennian feature. The implication is that the gneisses represent a deeper section of the crust than rocks in the Orijärvi triangle; the rise in metamorphic grade is not related to the late Svecofennian thermal event.

The Orijärvi area is within the Late Svecofennian Granite-Migmatite zone (Ehlers et al., 1993) but we could not find migmatization at Salittu that was explicitly late Svecofennian. The large granite body around Salittu (Fig. 1) is late Svecofennian because it crosscuts in places early Svecofennian structures (Fig. 3). The small, undeformed granite bodies that occur along faults (Fig. 6) are similar to granites in D_5 shear zones (cf. Skyttä and Mänttäri, 2008, Fig. 2). The composite mafic dike at Salittu shows that magmatism during D_5 deformation was not only granitic.

Ploegsma and Westra (1990) interpreted that folding (F_3) with N–S trending axial planes governs the large-scale structure at Salittu, and Pajunen et al. (2008) came to a similar conclusion (their F_1). These interpetations appear to be valid because the form of the large granite body is curved and the granite contains deformation zones, suggesting large-scale D_5 deformation (Fig. 10).

6.3. Stratigraphy and age correlation

The oldest ages in the Southern Svecofennia are 1898 ± 9 Ma for the Orijärvi granodiorite (Väisänen et al., 2002) and 1895.3 ± 2.4 Ma for a synplutonic rhyolite at Orijärvi (Väisänen and Mänttäri, 2002). Moreover, Väisänen and Mänttäri (2002) obtained 1878.2 ± 3.4 Ma age from a dacite from the upper part of the Kisko formation. Subsequently Väisänen and Kirkland (2008) attained 1878 ± 4 Ma concordia age from core domains of zircons from a felsic metavolcanic rock of the Toija Formation, and interpreted it to represent the crystallization of the rock. They attempted to date the SFm and although they did not get a direct age they concluded that the formation was deposited between 1878 and 1875 Ma.

The Arpalahti tonalite gives the minimum age of for the Salittu metabasalt, and the Laari granodiorite most probably crosscuts the metapicrite. Considering the error limits, the 1874 ± 8 Ma and 1878 ± 5 Ma ages for the Laari granodiorite and Arpalahti tonalite, respectively, are the same as the ages obtained from the Kisko and Toija formations. As the dacite from the upper part of the Kisko Formation yielded an 1878.2 ± 3.4 Ma age, the timing of picritic-basaltic volcanism may be estimated to ca. 1875 Ma.

A belt of metapelite (metagreywacke) occurs between the Orijärvi and the Kisko Formations, in the southern limb of the Orijärvi D_2 synform (Fig. 1; Väisänen and Mänttäri, 2002; Skyttä et al., 2006). The Vetio site within this belt was studied in detail by Skyttä et al. (2006), Pajunen et al. (2008), and Sayab et al. (2015). Detrital zircons in the metapelite are older (2.72, 2.12–1.93 Ga; Claesson et al., 1993) than zircons in the volcanic rocks in the Orijärvi and Kisko Formations. We interpret that the metapelite represents a sedimentary basin that developed after the earlier magmatic event at 1.90–1.89 Ga, and sediments from a large area were mixed in the basin. Since the stratigraphic position of the belt is known within the Orijärvi triangle, we here name it Vetio Formation.

The metasedimentary rocks as well as the overlying intermediate metavolcanic rocks at Sorttila (Fig. 5) were previously considered to be the uppermost part of the Kisko Formation (Väisänen and Mänttäri, 2002). Since the lower part of the Kisko Formation consists solely of metavolcanic rocks, we consider that these rocks form a separate formation within the Orijärvi triangle, here named as Ahdisto Formation. The stratigraphic position of the Ahdisto Formation is correlative to the quartz-feldspar gneisses and biotite gneisses at Salittu. Field evidence suggests that the Toija Formation

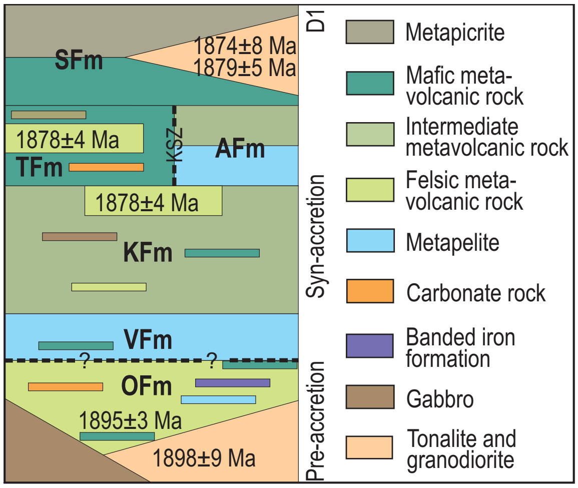

Figure 11. Stratigraphy of the Orijärvi area (not in scale), and related tectonics. Ages from Väisänen and Mänttäri (2002), Väisänen et al. (2002), Väisänen and Kirkland (2008) and this study. Broken line represents a tectonic discordance. OFm = Orijärvi Formation, VFm = Vetio Formation, KFm = Kisko Formation, TFm = Toija Formation, AFm = Ahdisto Formation, SFm = Salittu Formation, KSZ = Kisko shear zone. See text for explanation.

conformably underlies the Salittu Formation (Väisänen and Mänttäri, 2002). Therefore both the Ahdisto and Toija Formations are stratigraphically below the Salittu Formation but as they are separated by the Kisko shear zone, their mutual position is unknown.

The refined stratigraphy of the Orijärvi area is shown in Figure 11. The Orijärvi Formation is lowermost, with felsic and mafic metavolcanic rocks and interlayers of marble, metapelite and iron formation; stratigraphic interpretation by Latvalahti (1979) differs from that of Colley & Westra (1987). The synvolcanic granodioritic and gabbroic intrusions are shown. The overlying metasedimentary unit is the new Vetio Formation, with metapelite containing interlayers of mafic metavolcanic rock. The Kisko Formation consists of mafic to felsic metavolcanic rocks. The new Ahdisto Formation between the Kisko and Salittu Formations consists of metasedimentary rocks at the base and intermediate metavolcanic rocks on top. The Toija Formation consists of mafic and felsic metavolcanic rocks, with interlayers of marble and a unit of ultramafic metavolcanic rock in the upper part. Uppermost are the ultramafic-mafic metavolcanic rocks of the SFm, crosscut by the Arpalahti tonalite and the Laari granodiorite.

In the stratigraphic column (Fig. 11) rocks of 1890–1880 Ma age are missing which is somewhat strange considering that this was a period of voluminous crust formation during the Svecofennian orogeny. The ca. 20 Ma age gap between the Orijärvi and Kisko Formations is somewhat loosely constrained because no age data exists from the bottom of the Kisko Formation but at present we take the age gap to be real.

The ca. 1875 Ma picritic-basaltic volcanism of SFm gives the maximum age of D_1 deformation. If the presented stratigraphy and structural correlation are correct, deformation of the stratigraphic package overlying the Orijärvi Formation experienced D_1 deformation at an advanced stage of accretion.

Isotopic evidence suggests an evolved crust in southernmost Finland with an age of ≥ 2.0 Ma (Lahtinen and Huhma 1997), and therefore a ≥ 2.0 Ma microcontinental core is assumed to have existed to the south of the Orijärvi area during accretion in the models of Nironen (1997) and Lahtinen et al. (2005). In line with these models we propose the following evolution: the Orijärvi Formation represents magmatism and sedimentation at the margin of this older block that was approaching the subduction zone in the overriding plate; and the overlying volcanic package was formed above the subduction zone when the block started to accrete to another microcontinental block (i.e. accretion of Southern Svecofennia and Central Svecofennia). Because of the proposed duality (pre-accretion vs. accretion sequences) a tectonic discordance is inferred in Figure 11 between the Orijärvi and Vetio Formations. The metapelites of the Vetio Formation represent sediments of the forearc basin and the metavolcanic rocks of the Kisko and Toija Formations are volcanic rocks of the growing magmatic arc. The metavolcanic rocks of the Salittu Formation represent a rifting episode in the magmatic arc (Väisänen and Mänttäri, 2002; Nironen, submitted).

7. Conclusions

The Salittu formation consists of pictitic and basaltic metavolcanic rocks. The original stratigraphy is visible at Salittu although slightly obscured by deformation: metabasalt overlies migmatitic gneisses and metapicrite is on top. The early Svecofennian D_1 foliation is visible both in the metavolcanic and metasedimentary rocks. The rocks were folded during the early Svecofennian D_2 deformation, and the large synformal structures developed as D_2–D_5 interference structures during late Svecofennian D_5 deformation. The structural pattern at Salittu is much the same as in the Orijärvi triangle although the metamorphic grade is higher.

The rise in metamorphic grade from the Orijärvi triangle to Salittu is the result of early Svecofennian reverse faulting in the Jyly shear zone, not related to the late Svecofennian thermal event that caused the development of the West Uusimaa Complex.

The new age data, together with earlier published data constrains the Salittu volcanism at ca. 1875 Ma. The stratigraphy in the Orijärvi area is refined to consist of the early (1.90–1.89 Ga) volcanic unit (Orijärvi Formation), overlain by a mainly sedimentary unit (Vetio Formation), the Kisko Formation, a volcanic-sedimentary unit (Ahdisto Formation), the Toija Formation, and on top the Salittu Formation; these were all emplaced at 1.88–1.87 Ga. We propose a model in which the Orijärvi Formation represents magmatism at the margin of a microcontinent, and the overlying package sedimentation and magmatism above a subduction zone during an initial stage of microcontinental accretion. First identified deformation occurred in an advanced stage of accretion after the emplacement of the volcanic rocks of the Salittu Formation.

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