STRUCTURAL AND MAGMATIC EVOLUTION IN THE LOIMAA AREA, SOUTHWESTERN FINLAND

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Within the Loimaa area there is a junction of the general E-W structural trend of southern Finland and a NW-N-NE curving trend. The structure of the area is dominated by ductile D_3 and D_4 deformations with E-W and N-S axial traces, respectively. The typical semicircular structures in the study area are interpreted as F_3–F_4 fold interference structures.

The predominant plutonic rocks in the Loimaa area are penetratively foliated tonalites and granodiorites which probably intruded during D_2 deformation. Peak regional metamorphism at upper amphibolite facies and emplacement of the Pöytyä Granodiorite ca. 1870 Ma ago occurred during D_3 deformation. The ductile style of D_4 deformation in the Loimaa area is probably related to the high-grade metamorphism at 1850–1810 Ma in the late Svecofennian granite-migmatite (LSGM) zone immediately south of the study area. The Oripää Granite was emplaced during D_4 deformation. The structural evolution in the Loimaa area may be correlated with the evolution further to the northwest (Pori area) and north (Tampere–Vammala area) whereas correlation to the south and west is problematic. A transpressional model presented for the LSGM zone is not applicable to the Loimaa area.

Key words: structural geology, structural analysis, granites, absolute age, U/Pb, Svecofennian, Proterozoic, Loimaa, Finland

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INTRODUCTION

The Loimaa area, although rather poorly exposed, is geologically interesting because the general E-W structural trend of southern Finland shifts into a curving trend (Fig. 1). The late Svecofennian granite-migmatite (LSGM) zone of southern Finland extends to the southern part of the study area. Moreover, one of the highest positive magnetic anomalies in southern Finland exists within the study area. The Loimaa area, formerly mapped in the late 1940's and early 1950's (Salli 1953a, 1953b; Huhma 1957, 1959), was remapped by the author as a reconnaissance work for the Global Geoscience Transects Project. Within the limited time (2½ months) emphasis was set on structures, especially on structural setting of certain intrusions.

LITHOLOGY

Supracrustal rocks

The mineral assemblages in the Loimaa area indicate generally upper amphibolite facies with slight differences in metamorphic grade. Salli (1953b) divided the metavolcanic rocks of the area into two groups according to lithological differences and preservation of primary structures. This distinction still appears to hold. The rather well-preserved metavolcanic rocks in the southern part of the study area (Fig. 1) continue further east as part of the Forssa group of the Häme Belt (see Hakkarainen 1994). In the southeastern part mafic-intermediate pyroclastic rocks containing uralite or less commonly plagioclase clasts predominate. Agglomerate and flow breccia (Fig. 2a) are common. Plagioclase porphyry (Fig. 2b) occurs in an E-W trending, dike-like setting in the southeastern part of the study area. Layered intermediate to felsic tuffites become predominant towards the south.

The metavolcanic rocks of the other group consist mainly of hornblende gneisses. The origin of these rocks is not as obvious as for the first group, but lithological layering with alternating amphibolitic and feldspar-rich layers suggests a volcaniclastic origin. Massive uralite porphyry occurs in places as interlayers within the hornblende gneisses. The faintly layered quartz-feldspar-biotite gneisses devoid of Al-rich porphyroblasts are interpreted as tuffite interlayers. In many places, especially in the western part of the study area, migmatization and assimilation have been so intense that it is difficult to assess whether the rock is a paragneiss or an orthogneiss (Fig. 2c). Moreover, the supracrustal rocks are cross-cut by ubiquitous pegmatite and granite dikes, therefore the area denoted as "hornblende gneiss" in Fig. 1 is very heterogeneous in character.

The metasedimentary rocks are in the northern part of the study area migmatitic, mostly K-feldspar-cordierite gneisses with some garnet-cordierite gneisses, and in the southwestern part migmatitic garnet-cordierite gneisses. The northern margin of the LSGM zone near Aura, marked as granite in Fig. 1, consists of garnet-bearing pegmatite and granite with relics of supracrustal rocks (mostly garnet-cordierite gneiss). The composition of garnets within the gneiss xenoliths indicates granulite facies conditions (Pentti Hölttä, personal communication 1996).

Plutonic rocks

Equigranular, medium-grained tonalites and granodiorites are the most voluminous and probably the oldest plutonic rocks in the Loimaa area. They typically contain mafic magmatic enclaves, and a penetrative foliation can be seen in these rocks (Fig. 2d). The quartz dioritic body in the northern part of the study area grades into tonalite and may thus be coeval with the latter. The E-W trending coarse-porphyritic granite cross-cuts tonalite at some outcrops.

The two oblate bodies in the central part of the study area (NW and SW of Loimaa) consist of granodiorite and granite. At the margins of these bodies migmatitic gneisses grade into a moderately foliated plutonic rock containing wall-rock fragments. This rock is cross-cut by a slightly foliated homogeneous granite.

The large area southwest of Alastaro, mapped as granite, is lithologically heterogeneous. A magnetite-bearing granite apparently causes the high positive magnetic anomaly in Fig. 3. The granite is deformed and variable in grain size, ranging from a fine-grained type to pegmatite. It contains abundant gneissic and tonalitic xenoliths, and the vague appearance of some xenoliths suggests assimilation by the granite. Numerous pegmatitic and aplitic dikes cross-cut both the xenoliths and the host granite.

The southwestern part of the study area is dominated by the large (160 km^2) Pöytyä Granodiorite. The main part of the pluton is equigranular, medium-grained and granodioritic to tonalitic in composition. A granitic, porphyritic phase with K-feldspar phenocrysts (up to 15 mm in length) makes up the central part of the pluton. The gra-

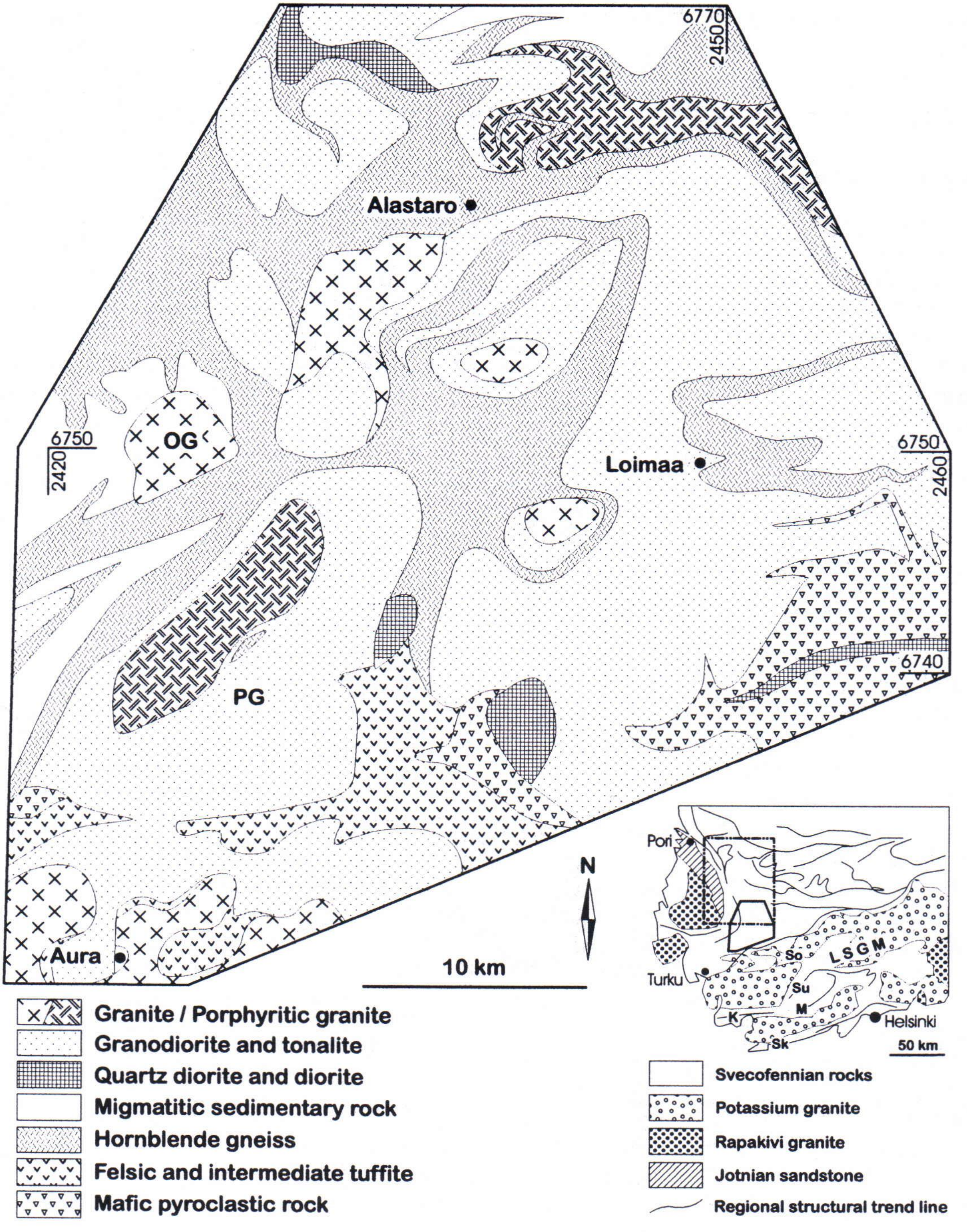
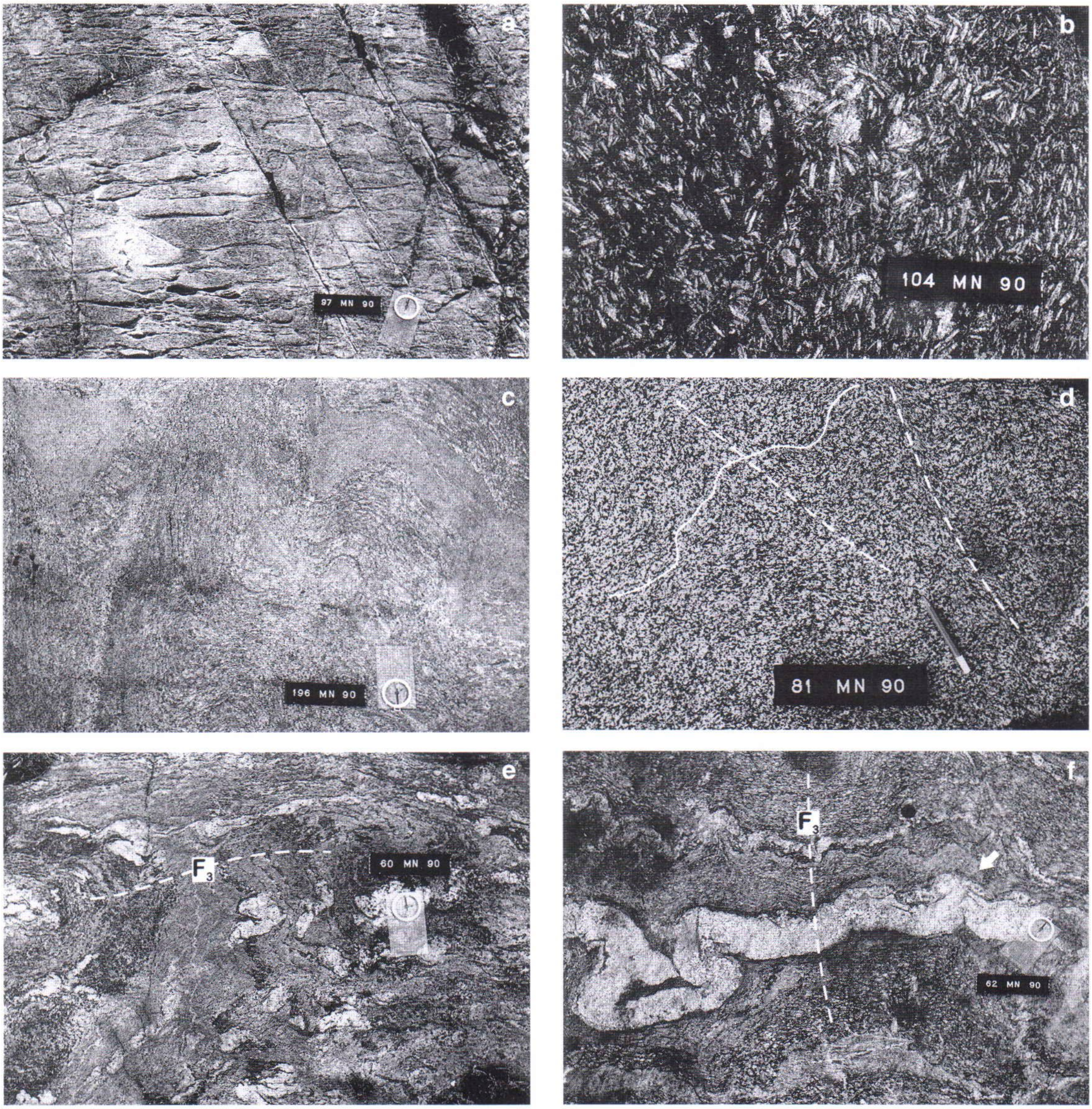


Fig. 1. Geological map of the Loimaa area. PG = Pöytyä Granodiorite, OG = Oripää Granite. The small map shows the general structural trends in southern Finland, location of the study area (thick solid line) and the area of Fig. 7 (thick dotted line). LSGM = late Svecofennian granite-migmatite zone, Sk = Skåldö, K = Kemiö, M = Mustio, Su = Suomusjärvi, So = Somero.



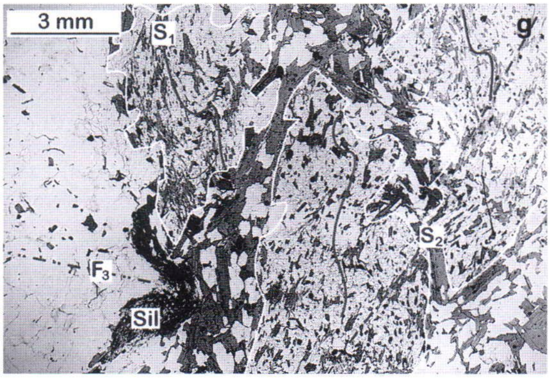


Fig. 2. a) Flow breccia in amphibolite. x = 6744.66, y = 2455.83. Length of code bar 15 cm. b) Plagioclase-phyric rock, possibly a dike rock. x = 6741.14, y = 2456.66. c) A felsic gneiss of uncertain origin cross-cut by felsic dikes. Both rock types are folded by F_4. x= 6753.85, y = 2425.42. d) Granodiorite with a penetrative foliation (marked by solid white line) that is folded (fold axial trace is marked by broken line). x = 6752.10, y = 2449.62. e) K-feldspar-cordierite gneiss. The penetrative foliation (S_1–S_2) parallel to the lithological layering is folded by F_3 (axial trace is marked by broken line). Migmatite neosome occurs as a vein along the axial plane and as irregular patches in the

dational contact between the rock units suggests a close temporal relationship in their emplacement. Mafic magmatic enclaves and pegmatitic dikes are fairly common in both phases.

The Oripää Granite in the western part of the area is a pinkish rock with highly variable grain size (usually pegmatitic). The granite contains abundant fragments of foliated garnet-bearing biotite gneiss and hornblende gneiss. In many places the rock shows schlieren migmatitic structures consisting of lensoid fragments of biotite gneiss in pegmatitic granite. In the central part of the pluton granite brecciates a medium-grained, foliated tonalite.

STRUCTURES

Low-altitude aeromagnetic maps were used to support the structural and lithological interpretation. The structural pattern in the study area is characterized by semicircular structures (Fig. 3). Another notable feature is the E-W trending structural pattern in the northern and central part, NE trending pattern in the southwestern part, and N trending pattern in the northwestern part.

The northern part of the study area, occupied by K-feldspar-cordierite gneisses, is the key area for the structural interpretation. The predominant foliation (preferred orientation of mica, and differentiation into mica-rich and quartz-feldspar layers in micaceous lithologies) is subparallel to the lithological layering (Figs. 2e and 2f). K-feldspar porphyroblasts contain an internal foliation defined by inclusion trails of small biotite flakes (Fig. 2g). This foliation is microfolded, and axial traces of these folds are parallel to the orientation of coarser biotite in the matrix. Because no microfabric could be identified parallel to the lithological layering and the inclusion trails define the oldest observed microstructure, lithological layering is interpreted as bedding (S_0), internal foliation as S_1, and matrix foliation as S_2. Presumably K-feldspar grew at an early stage of D_2, preserving the early microstructure (S_1) while subsequent coarsening of the matrix and the development of S_2 obliterated S_1 outside the porphyroblasts (cf. Passchier & Trouw 1996, Fig. 7.4).

Lithological layering in the metasedimentary rocks in the northern part is folded by open to isoclinal F_3 folds with steep, approximately E-W striking axial surfaces. Leucosome in K-feldspar-cordierite gneisses occurs as veins subparallel to S_2 (Fig. 2f), as veins parallel to F_3 axial plane, and as irregular patches at the hinges of mesoscopic F_3 folds (Fig. 2e). These features indicate migmatization during D_3. In microscale S_3 is expressed as differentiated crenulation cleavage, associated with reorientation and recrystallization of coarse (syn-D_2) biotite into small grains, minor growth of biotite parallel to F_3 axial plane, pinitization of cordierite, and sparse formation of muscovite. Commonly cordierite has overgrown coarse biotite that has been reoriented during F_3 folding. In areas of intense S_3 crenulation cleavage, clusters of fibrous sillimanite within cordierite express two generations of microfolding, the older with axial plane parallel to S_2 in the matrix (not shown) and the younger parallel to S_3 (Fig. 2g). Provided that sillimanite microfolds are related to regional folding episodes, the microstructures described above suggest that cordierite grew during syn-D_3 peak of metamorphism when most of the leucosome was formed, preserved F_2 and F_3 microfolded sillimanite as relics, and continued to grow after F_3 folding. Relics of sillimanite in cordierite are quite common in pelitic rocks of southern Finland, and microfolding in sillimanite seems to be related to regional folding at least in some areas (e.g. Nironen 1995).

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fold hinge area. x = 6767.68, y = 2445.48. f) K-feldspar-cordierite gneiss with leucosome veins subparallel to the composite layering (S_1–S_2). The hinge of isoclinally folded bedding between the two leucosome veins is shown by the white arrow. The composite layering and leucosome veins are folded openly by F_3. The drill hole in upper right part is the sample site for the thin section shown in Fig. 2g. x = 6769.19, y = 2444.35. g) F_2 microfolding in K-feldspar porphyroblasts (marked by white line) and leucosome folded by F_3. The cluster of fibrous sillimanite (Sil), preserved in undeformed cordierite, is microfolded with an axial trace parallel to that of mesoscopic F_3 folding.

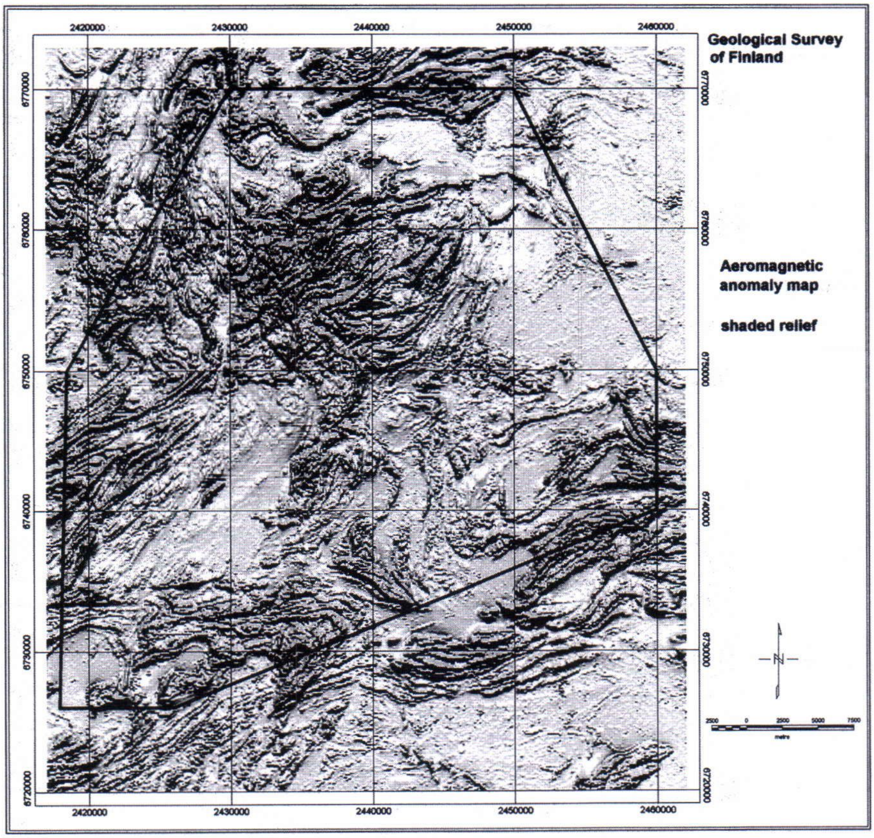

Fig. 3. Low-altitude aeromagnetic map of the Loimaa area.

Regional D_2 deformation was complex in southern Finland (Koistinen et al. 1996) but on the basis of the structures in the northern part the penetrative foliation elsewhere in the study area, marked by coarse biotite and hornblende, is probably S_2; S_0 and S_1 have largely been transposed to be subparallel to S_2. Upright F_3 folding transposed D_2 and earlier structures to generally steep orientations (Fig. 4). F_4 folding exists in the study area as open to tight, mesoscopic to macroscopic folds with approximately N-S trending axial traces. In microscale older fabrics are crenulated in mica-rich layers showing minor growth of biotite in the axial plane. The large semicircular structures, including the oval-shaped domal structure in the center of the Loimaa area, are interpeted as type 1 (dome-and-basin) F_3–F_4 fold interference structures.

Where the tonalites and granodiorites are in contact with supracrustal gneisses the penetrative foliation in the plutonic rock is parallel to the composite (S_0–S_2) layering in the gneisses, hence it is interpreted as contemporaneous with S_2. In areas where the penetrative foliation in the tonalites and granodiorites is folded (Fig. 2d) the fold axial surface is parallel to S_3 in supracrustal rocks. These features imply that the tonalites and granodiorites were emplaced before D_2 or, more probably, at the early stage of D_2 deformation. Also the porphyritic granite in the north as well as the magnetite-bearing granite in the central part of the study area contain a D_3 fabric.

The NE trending foliation in the southwestern part of the study area is a composite foliation: migmatitic garnet-cordierite gneisses exhibit differentiated layering in the mesosome that in places is tightly folded (Fig. 5a). The fold style and the occurrence of leucosome are similar to those in the northern part of the study area (cf. Figs. 2e and 5a). Leucosome garnet is devoid of inclusions

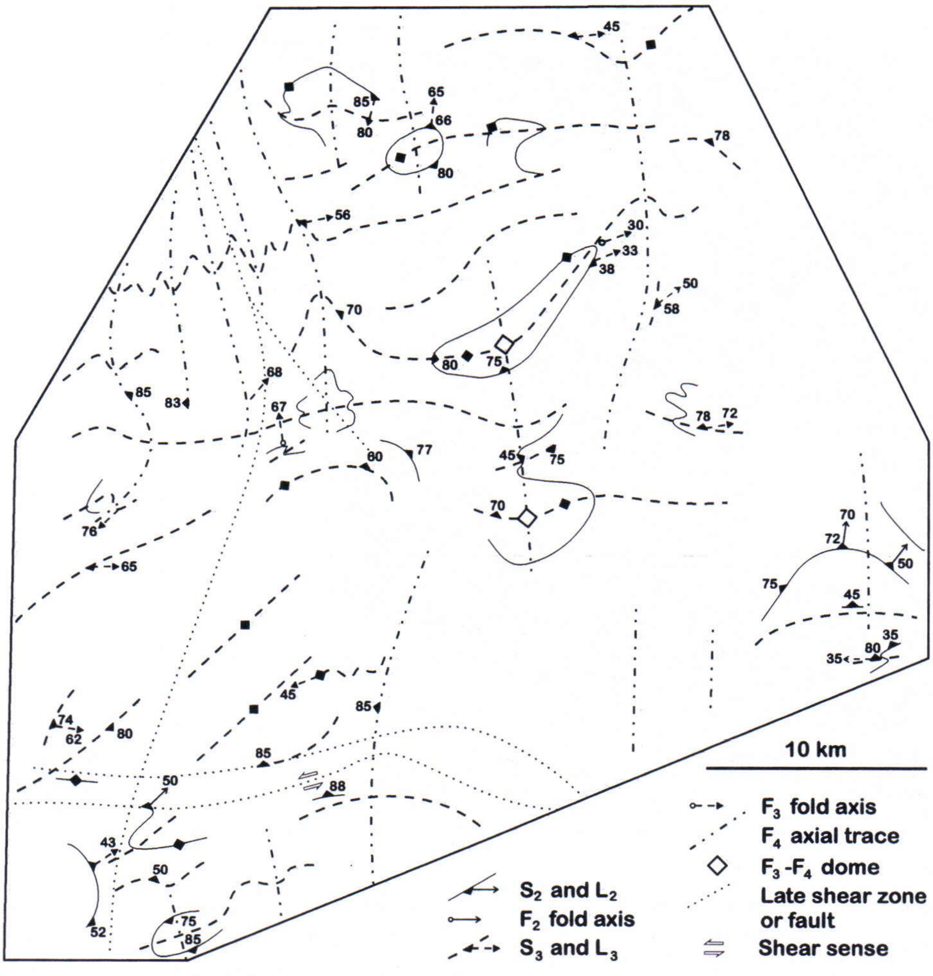

Fig. 4. Structural interpretation of the Loimaa area, based on geological data and aeromagnetic signatures.

whereas garnet in the mesosome is riddled with inclusions. Inclusion trails in the mesosome garnet curve and continue as a crenulated foliation in the matrix (Fig. 5b). Microfolding in the matrix is generally tighter than the curvature of the inclusion trails. The matrix foliation is interpreted as S_2, and the overprinting folding is F_3. The microstructures indicate that garnet in the mesosome grew when F_3 crenulation had initiated and that the tightening of microfolds in the matrix is the result of continued D_3 deformation after garnet growth.

The Pöytyä Granodiorite is pervasively deformed, and also the cross-cutting pegmatite dikes are deformed. In the northern part of the pluton deformation increases in intensity outwards from the center and the foliation is parallel to the margins. In places the penetrative foliation is deformed by ductile shear zones that are commonly filled with granitic material. In a sample from the granodioritic-tonalitic main phase close to the boundary of the porphyritic variety, feldspars are dynamically recrystallized whereas strain-free quartz is an evidence of post-deformation static recrystallization (Fig. 5c). Penetrative foliation as well as ductile deformation of feldspars indicate deformation at temperatures above ca. 500°C (Gapais 1989). In contrast, feldspars in the por-

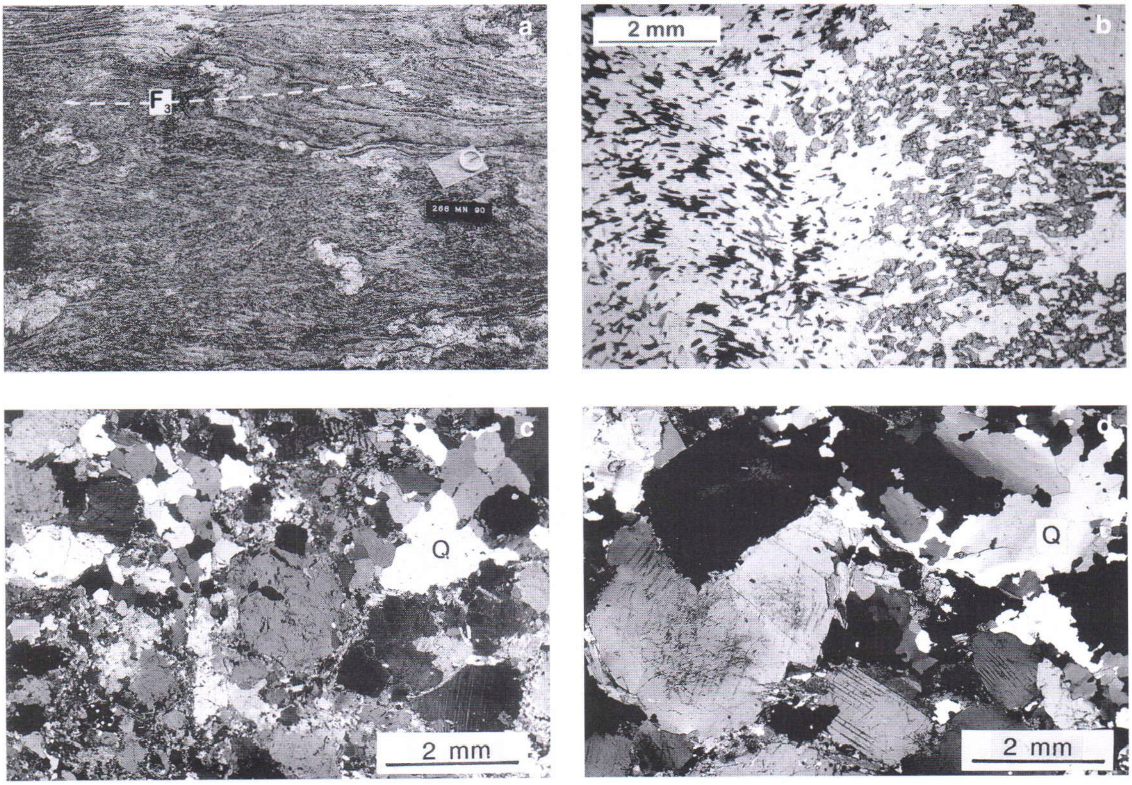

Fig. 5. a) Migmatitic garnet-cordierite gneiss (F_3 axial trace is marked by broken line). Length of code bar 15 cm. x = 6742.80, y = 2421.00. b) Microfolding in garnet-cordierite gneiss. Note curvature of inclusion trails in garnet (right) and plagioclase-quartz rim around garnet. x = 6742.56, y = 2421.25. c) Microstructure of the equigranular granodiorite of the Pöytyä batholith. Feldspars are largely recrystallized whereas quartz (Q) is undeformed and has straight or gently curving quartz/quartz grain boundaries (e.g. at the upper right corner). x = 6742.00, y = 2431.46. d) Microstructure of the porphyritic granite of the Pöytyä batholith. Feldspars are subhedral and rather undeformed. The high-angle boundaries within strained quartz (Q) indicate subgrain rotation recrystallization, and irregular margins indicate grain boundary migration recrystallization. x = 6740.50, y = 2424.04.

phyritic phase are subhedral and weakly deformed (Fig. 5d). Strain concentrated in quartz which deformed by combined subgrain rotation recrystallization and grain boundary migration recrystallization, indicating deformation at high temperatures (cf. Passchier & Trouw 1996, p. 48). These microstructural differences suggest that the granodioritic-tonalitic phase was emplaced earlier and deformed during cooling while the porphyritic phase – emplaced later – was still close to melting temperature during coeval deformation. Such conditions would also result in a thermal pulse causing the observed static recrystallization of quartz in the surrounding granodiorite-tonalite.

The composite foliation in the supracrustal gneisses is parallel to the contact of the Pöytyä Granodiorite. Close to the northern contact S_2 is folded in the gneisses, and the fold axial planes are parallel to the contact. Hence the granodiorite deformed D_2 structures during emplacement, i.e. the Pöytyä Granodiorite was emplaced after D_2 deformation. At the southern margin S_3 continues across the contact into the pluton – this is also evident from the aeromagnetic map (Fig. 3). At the southwestern margin of the Pöytyä Granodiorite S_3 cuts across a weak contact-parallel fabric. Since the microfabrics in the two plutonic phases indicate deformation at high temperatures, and since syntectonic pluton emplacement commonly involves geometric continuity between the country rock foliation and pluton fabric (e.g. Brun & Pons 1981), the emplacement of the Pöytyä Granodiorite probably took place during D_3 deformation.

F_4 folding becomes regionally dominant towards the northwest. The fairly symmetric style in the zone of intense F_4 folding suggests that the zone is not a shear zone but a zone of shortening. This is consistent with the large variation in plunge of mineral lineations in the zone. However, in many places it was impossible to assess the foliation-lineation relationships in the composite foliation planes.

The pegmatites of the Oripää Granite are in places slightly deformed, especially in the eastern part, but a medium-grained granitic type is undeformed. Undeformed magnetite-bearing granite/pegmatite dikes, similar to the undeformed Oripää Granite, are emplaced along axial surfaces of mesoscopic F_4 folds in the western part of the study area.

Late shear zones and faults are clearly visible in low-altitude aeromagnetic maps as distinct negative anomalies cross-cutting the lithological layering. An E-W trending anomaly in the southern part of the study area is a subvertical mylonite zone, 100–300 m in width. In places an S-C fabric is developed, indicating predominantly horizontal (sinistral) sense of movement (Fig. 4). Intense retrogressive alteration (e.g. epidotization) characterizes the mylonite zone.

The NNW trending anomaly in the northwestern part of the study area (Fig. 4), diverging into two anomalies towards the south, has been studied in detail by geophysical methods (Geologian tutkimuskeskus et al. 1993). It is a brittle fault that is younger than the mylonite zone and locates circulation of ground water.

U-Pb ISOTOPIC DATA

U-Pb isotopic analyses were made on zircon, titanite and monazite from the Pöytyä Granodiorite and the Oripää Granite (Fig. 1 and Table 1). The ages were calculated using the decay constants of Jaffey et al. (1971). The precision is given at the 2-sigma level.

A total of 15 zircon fractions from the porphyritic phase of the Pöytyä Granodiorite yielded an age of 1869 ± 8 Ma (Fig. 6a). An age of 1870 ± 5 Ma was obtained from eight zircon fractions of the granodioritic-tonalitic main phase, i.e. the same as from the porphyritic phase within error limits (Fig. 6b). There is no indication of an older zircon population except xenoliths that yield distinct isotopic ratios (Fig. 6a, fraction L). The 1784 ± 8 Ma and 1780 ± 12 Ma titanite ages from the porphyritic phase and granodioritic-tonalitic phase, respectively, date cooling through the ca. 650°C closure temperature of titanite (Scott & St-Onge 1995, Pidgeon et al. 1996).

The zircons of the Oripää Granite consist of red and pale brown varieties. There are several differences between the two populations (Table 1, Fig. 6c): 1) uranium content is higher in the red variety in the same density fraction; 2) the red zircons are less discordant than the pale brown ones; and 3) the two populations form separate regression lines, 1850 ± 27 Ma (red) and 1860 ± 41 Ma (pale brown). The large error limits show that there is a considerable dispersal in the two zircon populations suggesting mixing between them. The 1794 ± 10 Ma age of monazite dates cooling below ca. 700°C (Parrish 1990, Mezger et al. 1991).

STRUCTURAL AND MAGMATIC EVOLUTION

In addition to the age data of the Pöytyä and Oripää granitoids referred above, a few U-Pb datings of rocks in southwestern Finland help constraining the timing of the structural and magmatic evolution in the Loimaa area. A ca. 1890 Ma age has been obtained from a felsic metavolcanic rock of the Häme Belt (Vaasjoki 1994). This age defines the upper limit for the earliest deformation in the Loimaa area. Gabbroic pegmatoids with a ca. 1885 Ma pooled zircon age record the earliest deformational episodes in the Skåldö area (Fig. 1) whereas a ca. 1875 Ma monazite age from the

Table 1. U-Pb isotopic data on zircon, titanite and monazite from the Loimaa area. Error values for each ratio point are available on request.

Sample, fraction [1]		Concentration (ppm) ^{238}U	^{206}Pb	$^{206}Pb/^{204}Pb$ measured	Isotopic composition of lead $^{206}Pb = 100$ ^{204}Pb	^{207}Pb	^{208}Pb	Atomic ratios and radiometric ages, Ma $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A121 Pöytyä Granodiorite, porphyritic phase										
A	d>4.5	458.1	119.10	1968	.04650	11.936	9.670	.3005	4.684	.11306
								1693	1764	1849
B	titanite; 3.5<d<3.6;	83.8	23.27	1117	.08654	12.088	42.355	.3211	4.829	.10909
	Ø>160							1794	1789	1784
C	4.3<d<4.5	503.7	136.19	3794	.02372	11.651	8.234	.3125	4.881	.11329
								1752	1798	1853
D	4.3<d<4.5; abr	511.1	132.28	2159	.04536	11.914	9.877	.2991	4.660	.11299
								1686	1760	1848
E	4.3<d<4.5; Ø>70;	506.5	136.09	5535	.01713	11.566	8.223	.3106	4.853	.11334
	abr							1743	1794	1853
F	4.3<d<4.5; Ø>70	497.7	130.03	4020	.02298	11.614	8.501	.3020	4.706	.11303
								1701	1768	1848
G	4.2<d<4.3; Ø>130;	605.1	161.28	8095	.01088	11.430	6.915	.3080	4.792	.11283
	abr							1731	1783	1845
H	4.0<d<4.2; Ø>110;	777.8	194.24	3778	.02501	11.542	7.067	.2886	4.458	.11203
	abr; turbid							1634	1723	1832
I	4.0<d<4.2; abr;	916.1	231.87	4599	.02057	11.492	7.511	.2925	4.522	.11213
	long							1654	1735	1834
J	4.3<d<4.5; Ø<70;	487.3	129.41	5239	.01793	11.555	8.867	.3069	4.787	.11312
	abr							1725	1782	1850
K	d>4.5; abr	454.9	123.15	4759	.01949	11.522	8.337	.3129	4.856	.11257
								1754	1794	1841
L	3.6<d<4.2; abr;	1254	285.62	2084	.04712	11.892	15.278	.2633	4.085	.11253
	short (xenolith)							1506	1651	1840
M	4.3<d<4.5; Ø>70;	518.7	137.52	6576	.01326	11.500	7.713	.3064	4.782	.11320
	abr							1723	1781	1851
N	4.2<d<4.5; abr;	535.6	141.95	7657	.01033	11.433	6.771	.3063	4.769	.11293
	large							1722	1779	1847
O	4.2<d<4.5; Ø>70;	594.1	163.73	7648	.01103	11.476	7.218	.3185	4.974	.11326
	abr; red							1782	1814	1852
P	4.2<d<4.5; Ø>110;	483.0	127.88	3965	.02406	11.645	8.863	.3060	4.775	.11319
	abr							1720	1780	1851
A296 Pöytyä Granodiorite, main phase										
A	4.2<d<4.3; Ø>70;	603.2	122.29	5727	.01203	11.273	8.944	.2343	3.589	.11109
	abr; turbid							1357	1547	1817
B	4.3<d<4.5; abr;	421.4	103.27	12587	.006606	11.368	8.716	.2833	4.405	.11279
	clear							1607	1713	1844
C	4.2<d<4.3; Ø>70;	681.1	135.62	7317	.01246	11.210	8.600	.2301	3.503	.11040
	abr; clear							1335	1527	1806
D	4.0<d<4.2; Ø>70;	833.7	150.21	5953	.01568	11.144	8.893	.2082	3.138	.10930
	abr; clear, short							1219	1442	1787
E	4.0<d<4.2; Ø>110;	642.1	135.82	7576	.007777	11.257	8.447	.2445	3.758	.11151
	abr; turbid, short							1409	1583	1824
F	abr; long	658.8	138.15	6269	.01259	11.287	9.600	.2424	3.714	.11116
								1398	1574	1818
G	4.3<d<4.5; Ø>70;	421.3	102.50	11664	.006343	11.369	8.798	.2812	4.374	.11283
	abr; clear, short							1597	1707	1845
H	4.3<d<4.5; abr;	411.4	96.97	9549	.004830	11.309	8.236	.2725	4.223	.11243
	large, short, turbid							1553	1678	1839
J	titanite; 3.5<d<3.6;	59.2	15.59	509	.19188	13.474	29.382	.3043	4.555	.10857
	Ø>160; abr							1712	1741	1775

Table 1 (continued)

A1357 Oripää Granite										
A	4.2<d<4.3; red	874.9	296.01	4064	.02374	11.589	11.735	.3176	4.935	.1127
								1778	1808	1843
B	4.2<d<4.3; pale	772.6	232.42	2139	.04434	11.883	6.681	.2929	4.554	.1128
								1655	1740	1844
C	4.0<d<4.2; red	1217.1	354.97	2878	.03416	11.682	5.792	.2871	4.439	.1122
								1626	1719	1834
D	4.0<d<4.2; pale	1104.1	281.97	1910	.05096	11.870	6.347	.2491	3.838	.1117
								1433	1600	1828
E	3.6<d<4.0; pale	1490.9	292.16	913	.10872	12.337	8.167	.1857	2.780	.1086
								1097	1350	1775
F	3.6<d<4.0; red	1519.2	325.59	1285	.07701	11.964	6.779	.2072	3.119	.1092
								1213	1437	1785
M	monazite	1633.8	4825.9	7163	.01277	11.140	940.8	.3234	4.890	.1097
								1806	1800	1794

(1) d = density (g cm^{-3}), Ø = grain size (µm), abr = grains abraded. All data corrected for blank (0.88 ng Pb, 0.37 ng U).

same area has been interpreted as the age of regional metamorphism (Hopgood et al. 1983).

D_1 and D_2 deformations and syn-D_2 emplacement of the oldest tonalites and granodiorites in the Loimaa area probably took place during 1890–1880 Ma. Since the magnetite-bearing granite contains tonalitic xenoliths it is younger than the tonalites and granodiorites, but the relationship between emplacement of the granite and D_2 is unknown. The Pöytyä Granodiorite was emplaced at ca. 1870 Ma during D_3 deformation. Because the structural setting of the E-W trending coarse-por-

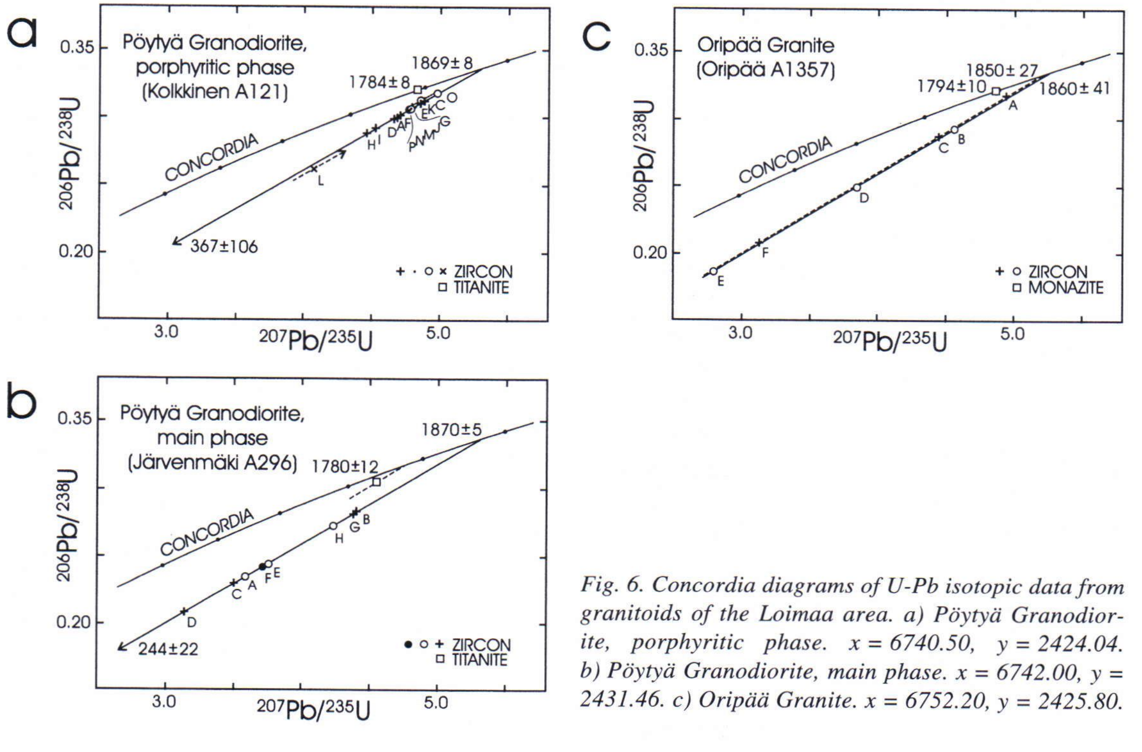

Fig. 6. Concordia diagrams of U-Pb isotopic data from granitoids of the Loimaa area. a) Pöytyä Granodiorite, porphyritic phase. x = 6740.50, y = 2424.04. b) Pöytyä Granodiorite, main phase. x = 6742.00, y = 2431.46. c) Oripää Granite. x = 6752.20, y = 2425.80.

phyritic granite in the northern part of the area is apparently controlled by D_3, the granite may be coeval with the Pöytyä Granodiorite.

Ductile F_4 folding has influenced the structural pattern in the Loimaa area. In the north, E-W striking F_3 axial surfaces are in places deformed by large-scale F_4 folding. The strike of S_3 planes shift into NE towards the southwestern part of the study area. There is few data on the orientation of foliation within the weakly deformed granites in the two domal F_3–F_4 fold interference structures, therefore the relationship between emplacement and folding episodes is unclear. The structure occupied by the Oripää Granite is also interpreted as an F_3–F_4 interference structure. The structures within the granite suggest dominantly in situ melt segregation. The medium-grained granite may represent a melt fraction that segregated from the host rock. The lack of deformation in the granite and the occurrence of the granitic dikes at F_4 axial surfaces suggest emplacement at a late stage of D_4. Unfortunately, zircons from the Oripää Granite yield minor information of magmatism and deformation, but the monazite dating indicates that the Oripää Granite was still hot ca. 1795 Ma ago.

DISCUSSION

Domal structures have been described in several areas of southwestern Finland. Van Staal and Williams (1983) interpreted such a structure in the Kemiö area (Fig. 1) as an F_3–F_4 interference structure in which F_3 axial traces generally trend NNE or NE (see also Verhoef & Dietvorst 1980) whereas the F_4 axial plane is E-W trending. The domal structure in the Mustio area having a granitic core was interpreted by Bleeker and Westra (1987) as an F_2–F_3 interference structure (F_2 and F_3 axial surfaces striking E-W and NNE, respectively) whereas Veenhof and Stel (1991) considered the structure as the result of a single-phase progressive event (D_1). Stel et al. (1989) interpreted the two domal structures in the Somero area (both with granitic cores) as the result of superposition of F_3 folding (NW to WNW striking axial surfaces) on earlier structures. These examples show that dome-and-basin structures similar to those in the Loimaa area are common in southwestern Finland, but interpretations of the folding episodes related to their formation and of the orientation of fold axial surfaces vary considerably.

Approximately N-S striking late Svecofennian foliation is visible all over southern Finland, especially in the eastern part (Gaál 1982 and references therein). Hopgood (1984) revealed the structural sequence in the migmatites of the Skåldö area (Fig. 1), including N-S striking planar structures. Schreurs and Westra (1986) described mesoscopic and macroscopic folding (their F_3) with subvertical N-S striking axial surfaces in the Suomusjärvi–Mustio area (Fig. 1) and concluded that granulite grade metamorphism preceded F_3 folding.

Pietikäinen (1994) studied the Pori Shear Zone northwest of the Loimaa area and concluded that dilatational brecciation of tonalites to form veined gneisses took place during regional D_3. He further inferred that mylonitization at the northeastern margin of the zone (within the steeply NE dipping Main Fault Zone; Fig. 7) occurred during late D_3 and that the zone is extensional with normal offsets down to the north and a minor sinistral horizontal component. The S_1 compositional banding in the tonalites of the Pori area is the predominant foliation and therefore similar to the penetrative S_2 foliation in the tonalites and granodiorites of the Loimaa area. An L_2 lineation is the most prominent structural element in the Pori area, while L_3 is the predominant lineation in the Loimaa area. Hence it seems that D_1 structures in the Pori area may be correlated with D_2 structures in the Loimaa area; possibly S_1, visible as inclusion trails in porphyroblasts, did not develop in the Pori area. Accordingly, D_3 structures in the Pori area are correlated with D_4 structures in the Loimaa area (Fig. 7). However, this correlation is ambiguous because in the Pori area late D_3 deformation was shearing whereas in the Loimaa area F_4 folding was probably the result of shortening.

Kilpeläinen (1998) concluded that the D_2 and D_3 (possibly also D_1) structures in the Tampere–Vammala area, north of the Loimaa area, developed due to N-S compression. The D_3 structures in the Loimaa area are compatible with N-S compression but the original attitude of the D_2 structures remains unrevealed because they have largely been transposed by younger deformation. Kilpeläinen (1998, Fig. 49) interpreted also the NNW-SSE and N-S trending structures as D_3 structures; these are interpreted as D_4 structures in the present study (Fig. 7).

The structural pattern in the Turku–Uusikaupunki area, southwest and west of the Loimaa area, is characterized by folds with vergence towards the northwest (Selonen & Ehlers 1998, Väisänen & Hölttä 1999). The granites within the the LSGM zone occur as sheets in generally gently dipping planar structures (Ehlers et al. 1993, Selonen et al. 1996). Ehlers et al. (1993) presented a model for the granites of the LSGM zone including emplacement 1840–1830 Ma ago along pre-existing subhorizontal structures and simultaneous westward translation; i.e. emplacement within a transpressional zone with vergence to the NW. The granites were subsequently refolded in open folds with E-W trending fold axes. Lindroos et al. (1996) dated a pegmatite in the Kemiö area and concluded that transpressional deformation lasted until 1803 Ma. As the overall structural pattern in the LSGM zone differs from that in the Loimaa area, where the E-W to NE-SW trending structures were developing already 1870 Ma ago, it is difficult to apply this model to the Loimaa area.

Metamorphism culminated between 1850 Ma and 1810 Ma in the LSGM zone (Korsman et al. 1984). U-Pb ages between 1860 Ma and 1820 Ma have been obtained from pyroxene-bearing granitoids around Turku, southwest of the Loimaa area (Suominen 1991, van Duin 1992; see inset in Fig. 1). The granitoids are located within a granulite facies area and contain zoned zircons with secondary overgrowths. A study of zircon from a pyroxene granitoid suggests that the 1862 Ma age is a mixed age of two populations, one giving primary crystallization ca. 1880 Ma ago and the other giving the high-grade metamorphism at ca. 1843 Ma (Suominen 1991, Appendix 5). Moreover, a tonalite outside the granulite facies area, 30 km SW of the Pöytyä Granodiorite, yielded a 1869 ± 5 Ma age (van Duin 1992). These data strongly suggest that the 1860–1820 Ma ages of the pyroxene granitoids are the result of high-grade metamorphism, as concluded by Väisänen et al. (1994), and that the true age of these plutons is ca. 1870 Ma. The high-grade metamorphism in the LSGM zone presumably caused a thermal event with temperature exceeding the closure temperature of the U-Pb system for titanite in the adjacent Loimaa area. Therefore the 1780 Ma age of titanite in the Pöytyä Granodiorite records the decrease in the temperature of regional metamorphism below ca. 650°C. This late metamorphic peak in the LSGM zone is probably also the reason for the ductile style of D_4 deformation.

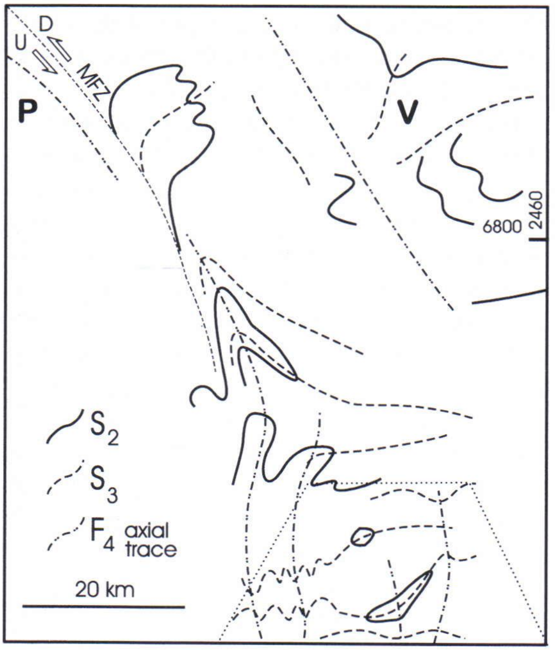

Fig. 7. A model to correlate the structures in the Loimaa, Pori (P), and Vammala (V) areas. Interpretation is based on low-altitude aeromagnetic map and field observations by the author in the southern part, Pietikäinen (1994) in the Pori area, and Kilpeläinen (1998) in the Vammala area. The relative sense of movement in the Main Fault Zone (MFZ) is shown by arrows and block movement directions (U = up, D = down). The study area is shown by a dotted line.

SUMMARY

The oldest tonalites and granodiorites in the Loimaa area were emplaced 1890–1880 Ma ago, before the end of regional D_2 deformation. The Pöytyä Granodiorite intruded 1870 Ma ago during D_3 deformation. The Oripää Granite was emplaced in an F_3–F_4 fold interference structure during D_4. Peak metamorphism was attained during D_3 deformation. E-W compression at some interval between 1850 Ma and 1800 Ma caused ductile D_4 deformation, coeval with medium/high grade metamorphism in the adjacent LSGM zone.

The location of the Loimaa area at the intersection of different structural trends in southwestern Finland restricts structural and metamorphic correlation with adjacent areas. The D_2 structures in the Loimaa area may be correlated with D_1 structures in the Pori area and the D_3 structures in the Loimaa area may be correlated with D_3 in the Tampere–Vammala area whereas correlation to the south (the LSGM zone) and west is problematic. Evidently the overall structural correlation in southwestern Finland needs further work.

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