THE EVOLUTION OF THE RAAHE—LADOGA ZONE IN FINLAND: ISOTOPIC CONSTRAINTS

by

MATTI VAASJOKI and MATTI SAKKO

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The NW-SE trending Raahe—Ladoga zone in the Fennoscandian (Baltic) Shield forms a major ore-bearing district along the Archaean-Proterozoic boundary. New U-Pb mineral analyses on 17 samples and a review of pre-existing data indicate that:

(1) volumetrically minor but spatially widely distributed granitoid rocks in the central part of the belt register U-Pb ages between about 1930 and 1910 Ma ago;

(2) between 1910 and 1900 Ma ago the zone underwent intermediate-acidic volcanism associated with base metal mineralization;

(3) nickel-bearing early orogenic gabbros were emplaced between 1900 and 1880 Ma ago;

(4) intrusion of syntectonic granitoids, hypersthene granites and plagioclase porphyrites/ophitic gabbros all along the belt was accompanied by the peak of regional metamorphism in the central and northwestern parts between 1890 and 1875 Ma ago;

(5) posttectonic rocks were emplaced in the central and northwestern parts between 1880 and 1850 Ma ago;

(6) regional metamorphism peaked and migmatization occurred in the southeastern part of the belt between 1830 and 1810 Ma ago;

(7) posttectonic rocks in the southeastern part were emplaced about 1800 Ma ago.

These results together with previous isotope data from Finland suggest that the new Svecofennian crust was rapidly generated between 1930 and 1850 Ma ago by the collision of a Proterozoic oceanic plate with an Archaean continent. Subsequent underplating in a W-E direction south of 62°N resulted in migmatization at about 1820 Ma and the emplacement of the ca. 1800 Ma post-tectonic granites in southern Finland.

Results on metapelites indicate that whereas zircon tends increasingly to lose its radiogenic lead as the temperature of regional metamorphism rises, total resetting is not achieved, not even at the garnet-cordierite-biotite grade. Only intense migmatization and/or immediate contact effects of hypersthene granites, approaching temperatures of 800°C, are sufficient to totally reset the U-Pb record of zircons.

Key words: absolute age, U/Pb, zircon, monazite, igneous rocks, metamorphic rocks, tectonics, Parikkala—Raahe zone, Proterozoic, Vihanti, Pielavesi, Joroinen, Rantasalmi, Kiuruvesi, Finland

Matti Vaasjoki and Matti Sakko, Geological Survey of Finland, 02150 Espoo, Finland

INTRODUCTION

The Raahe—Ladoga zone (see the index map on p. 6) has been recognized as a major tectonic, ore-bearing structure for quite some time (e.g., Kahma, 1973 and 1978). The main geological feature of the zone is the transition from predominantly Archaean rocks in the northeast to a Proterozoic domain in the southwest over a zone a few tens of kilometres wide (Simonen *et al.*, 1978) and characterized by a succession of NW-SE trending faults associated with pairs of similarly trending gravimetric minima and maxima (Elo, pers. comm.). Isotopically, the zone becomes apparent through the transition of the Nd composition of Svecokarelian granites within the ca. 1900 Ma age group, as the ε(Nd) values of the Karelian granitoids indicate a substantial input of Archean crustal material, whereas the Svecofennian granitoids southwest of the zone seem to be derived mainly from a Proterozoic mantle source (Huhma, 1986). The lead isotopic composition of the most prominent base metal ore deposits and prospects along the zone appears to be almost constant over a 300 km stretch containing leads relatively low in ^{207}Pb (Vaasjoki, 1981).

Several economic and subeconomic base metal deposits lie within the Raahe—Ladoga zone (Kahma, 1973; Helovuori, 1979; Huhtala, 1979; Mikkola, 1980) and are hosted by metavolcanic and metasedimentary rocks emplaced during the lower Proterozoic Svecokarelian orogeny (Simonen, 1971 and 1980). The base metal zone extends from the Gulf of Bothnia to Rautalampi, and at its northeastern end is surrounded by Svecofennian infra- and supracrustal rocks. In its central part it continues adjacent to the boundary of the Archaean craton and the Proterozoic orogenic rocks. Further southeast, it skims the border of the equally Proterozoic Svecofennian and Karelian rock groups, the latter being penetrated by Archaean gneiss domes forming part of the Presvecokarelian basement (Kouvo, 1958; Kouvo and Tilton, 1966; Wetherill *et al.*, 1962; Gaál, 1980) which led Koistinen (1981) to postulate a northeast dipping suture zone in this area. Superimposed on the base metal district and likewise parallel to the Archaean-Proterozoic boundary, a number of nickel deposits and prospects associated with differentiated mafic intrusions form the Kotalahti Ni-Cu belt (Tontti, 1981).

This report deals with three different parts of the Raahe—Ladoga zone and aims to establish their evolutionary differences with the aid of isotopic determinations. The Vihanti—Pyhäsalmi base metal district is of key economic interest and the Pielavesi—Rautalampi area has been previously studied chiefly for its block movements (Marttila, 1976; Korsman *et al.* 1984). The third geologically different domain is the progressively metamorphosed Joroinen—Sulkava area (Korsman, 1977). In the following pages each of these areas will be treated separately and at the end a synthesis of the tectonic evolution along the Raahe—Ladoga zone will be attempted.

FROM VIHANTI TO PYHÄSALMI: THE DOMAIN OF MAJOR BASE METAL DEPOSITS

Geological setting and previous isotopic studies

Throughout the Vihanti—Pyhäsalmi area the terrain is flat and outcrops are scarce as most of the area is covered by Pleistocene tills and bogs. Nevertheless, from previous work (e.g., Wilkman, 1931; Salli, 1964; Gaál *et al.*, 1974; Rauhamäki *et al.*, 1978) and ongoing mapping it seems that at least two cycles of volcanic activity have occurred in the district. The first one, associated with the base metal ores, was extensively deformed during the Svecokarelian orogeny (Rouhunkoski, 1969; Rauhamäki *et al.*, 1978; Helovuori, 1979; Huhtala, 1979). The lower volcanic sequence is penetrated by early tectonic gabbros and syntectonic granitoids, and there are occasional occurrences of intraformational conglomerates. These in turn are overlain by the younger volcanic sequence, which was terminated by the emplacement of hypabyssal plagioclase porphyrites. At Vihanti, the ore-bearing sequence is cut by fault-controlled diabase dykes. There are also a few apparently post-tectonic granites with unknown contacts in the district.

Most of the rocks within the lower volcanic unit of the Vihanti—Pyhäsalmi area are dacitic in composition and, at Vihanti, are intimately associated with the base metal mineralization consisting of separate pyritic and sphalerite-galena orebodies (Rouhunkoski, 1968; Rauhamäki *et al.*, 1978). At Pyhäsalmi, the immediate country rocks of the ore are mica schists of presumed sedimentary origin, but voluminous volcanic rocks occur in that area as well. Over the years, many theories have been advanced concerning the genesis of the mineralization, but current opinion favours a volcanic-exhalative origin for the Vihanti deposit in particular (Rauhamäki *et al.*, 1978) and for other ores of the Vihanti—Pyhäsalmi ore zone in general (Helovuori, 1979; Huhtala, 1979). The total reserves of the Vihanti deposit, including mining since 1954, have been estimated at 15.6 Mt of sphalerite-galena ore containing 7% Zn, 0.8% Pb and 0.5% Cu and 15.9 Mt of pyrite ore mined for its sulphur content (Isokangas, 1979). Corresponding figures for Pyhäsalmi (including mining since 1962) are 31 Mt at 2.3% Zn, 0.7% Cu and 36% S (ibid.).

At Vihanti and Pyhäsalmi, isotopic studies have been carried out on galenas from the various deposits, on whole rock samples and mineral separates from the Pyhäsalmi area, on uranium-bearing apatite gneisses at Vihanti and some isolated granitoids southeast of Pyhäsalmi.

Kouvo and Kulp (1961) demonstrated that galena samples from this region differed in isotopic composition from those of either the Outokumpu or Svecofennian type of mineralization, and Vaasjoki (1981) suggested that a regional lead isotopic signature prevailed in galena samples within the main sulphide ore belt. Irrespective of the models used (Stacey and Kramers, 1975; Cumming and Richards, 1975) the galena model ages for the various samples range from 1880 to 1970 Ma and are identical within the inherent uncertainties of the lead evolution models. Apart from one sample collected from the fringe of the ore proper, galenas from the Vihanti mine proved to be isotopically homogeneous. In the Pyhäsalmi area, a granite gneiss registers a U-Pb zircon age of 1930 ± 15 Ma, a synorogenic granodiorite is 1880 ± 15 Ma old and a plagioclase porphyrite has a minimum $^{207}Pb/^{206}Pb$ age of 1875 Ma (Helovuori, 1979). Pb-Pb whole rock analyses from a sample suite from the mine area suggest an age of 1910 ± 30 Ma for the volcanic rocks associated with the sulphide mineralization (ibid.). A study of the uranium-bearing apatite gneisses from the hanging wall of the Vihanti base metal orebodies (Vaasjoki *et al.*, 1980) showed that the culmination of the regional metamorphism associated with the Svecokarelian orogeny occurred in this area about 1880 ± 5 Ma ago. Granodiorite cobbles from an intraformational conglomerate at Settijärvi have been dated at 1888 ± 7 Ma (Marttila, 1987).

Sample material

An attempt to sample the volcanic rocks of the Lampinsaari sequence at Vihanti for zircon U-Pb analyses failed as no zircon extracts were obtained despite the large sample size (about 100 kg for each of the four samples). Thus all the samples for mineral U-Pb analyses from the Vihanti area are from rocks outside the mine area or from dykes cross-cutting the Lampinsaari sequence.

Sample A780 represents an early orogenic basic intrusion, known locally as the Alpua gabbro, which is well exposed in its central part. Experience shows that most zircon usually occurs within the more acidic differentiates of basic intrusions (Kouvo, 1976) and therefore, the sampling was conducted from a dioritic portion that clearly forms a part of the gabbro intrusive.

The syntectonic granodiorites are believed to be the oldest felsic rocks in the Vihanti area, and it has been suggested that some of them at least could be remobilized portions of the nearby Archaean craton. In the outcrop where sample A781 (locally known as the Hirsikangas granodiorite) was collected, the rock is an even grained, slightly orientated gneissose granite, with magnetite-bearing pegmatite veins and sporadic uralitized portions that may be partly assimilated xenoliths of older basic rocks.

About 200 m southwest of the sampling site for A781, there is a plagioclase porphyrite consisting of (1—2)x(2—5) cm large, zoned (bytownite-andesine) phenocrysts in fine-grained groundmass similar to the early orogenic gabbros. The contact relationship to the syntectonic granodiorite has not been established because of glacial overburden. Sample A898 was taken to see whether or not the plagioclase porphyrite was related to the locally widespread gabbros.

Another syntectonic plutonic rock in the area is sometimes called a ''granodiorite with potassium feldspar phenocrysts'' (Salli, 1964) and sometimes a ''microcline granite'' (Rauhamäki *et al.*, 1978). The sample for the zircon age determination (A782) was collected at Korpi from the main type, which at the sampling location consisted of abundant, slightly orientated potassium feldspar phenocrysts (1x2 cm) in a groundmass of medium-grained quartz, biotite, amphibole and oligoclase-andesine.

At Käpylä, 12 km south of the Vihanti ore field, there is a large intrusion of homogeneous, unorientated biotite granite. The rock contains some rare, fine-grained autoliths, occasional potassium feldspar phenocrysts (0.5—1.0 cm) that very rarely have a 0.1—1.0 mm thick oligoclase rim. The rock is apparently post-tectonic and a sample (A899) was taken to determine the duration of the main tectonic activity in the Vihanti area.

A titanite separate from the fault-controlled diabase dyke (A776) transgressing the mineralized sequence was analysed for its lead isotopic composition and uranium and lead concentrations, and two zircon fractions were analysed from a pegmatitic granite dyke (A938) that transgresses the Lampinsaari sequence but is cut by the diabase dyke.

Results and discussion

The results are presented in Figures 1, and 2 and Tables 1 and 2. From these it is evident that:

1) At 1901 ± 12 Ma, the Alpua gabbro is the oldest intrusive rock in the Vihanti area;
2) The felsic syntectonic intrusive rocks (A781, A782) register overlapping zircon ages (1874 ± 13, 1860 ± 13) within experimental error, and thus may be coeval;
3) The geologically youngest rock, the apparently posttectonic Käpylä granite (A899) registers the oldest age (1887 ± 4 Ma) of all the felsic intrusive rocks;
4) The titanite separate from the diabase dyke (A776) transgressing the Lampinsaari sequence is only slightly discordant and yields a $^{207}Pb/^{206}Pb$ age of 1861 ± 5 Ma which must be a minimum and is only slightly lower than

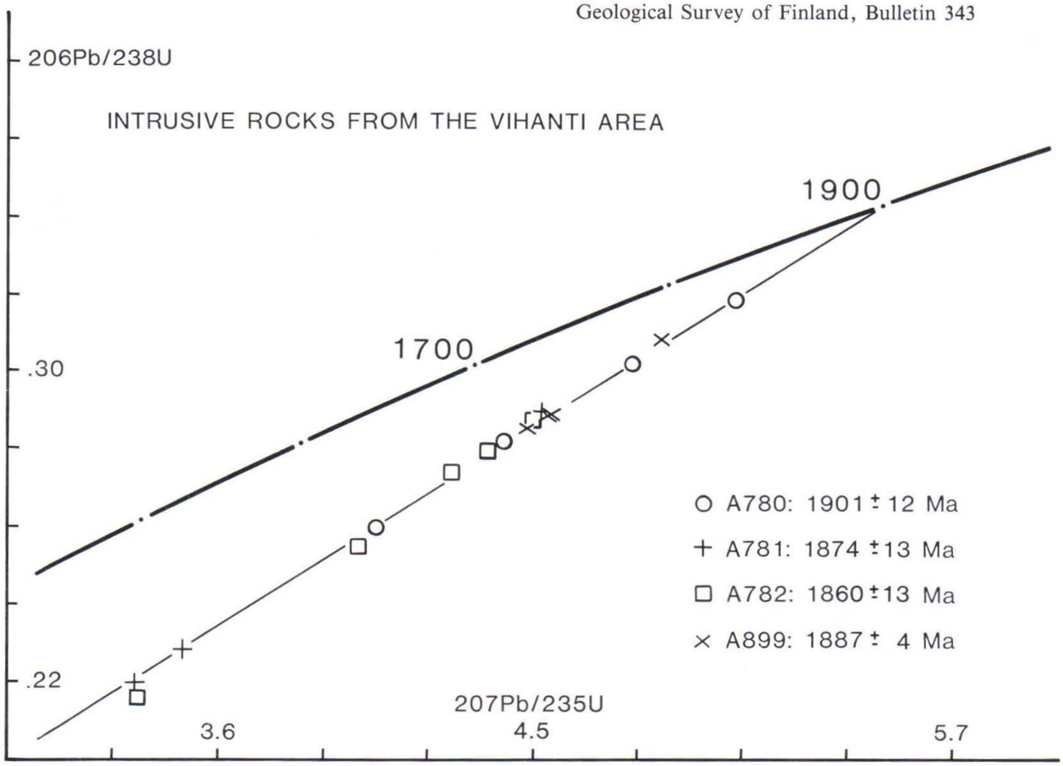

Figure 1. A concordia plot depicting the results for intrusive rocks of the Vihanti area.

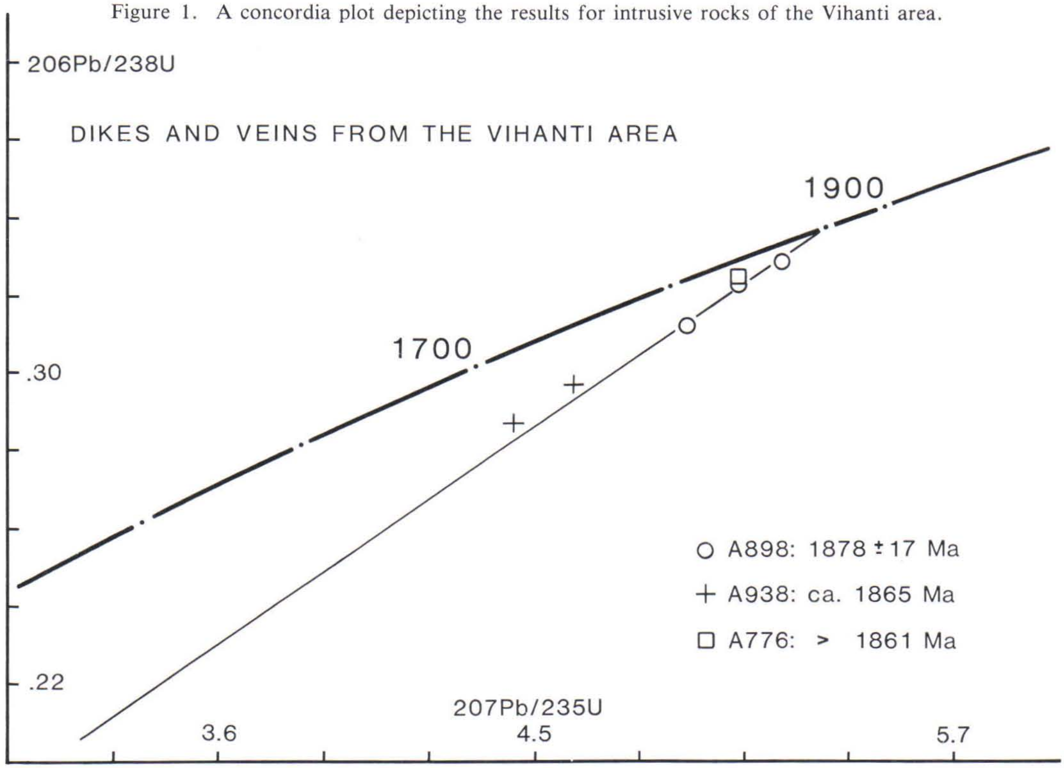

Figure 2. Mineral U-Pb results for posttectonic dykes and veins in the Vihanti area.

Table 1. U-Pb analyses of minerals from the Vihanti area.

Sample Fraction		Concentrations		^{206}Pb	Lead ratios, 206 = 100		
		^{238}U	Pb(tot)	^{204}Pb	^{204}Pb	^{207}Pb	^{208}Pb
A780 — Alpua gabbro							
A780A	4.0—4.2/—100	876.8	279.21	14550	.00687	11.595	8.997
B	+4.2/—100	522.9	173.47	18730	.00534	11.644	8.948
C	3.8—4.0/—100	1470.0	430.48	11000	.00909	11.500	8.991
D	3.6—3.8/—100	1764.3	477.85	13020	.00768	11.421	9.401
A781 — Hirsikangas granodiorite							
A781A	+4.2/—100	617.2	184.40	49980	.00200	11.386	8.063
B	4.0—4.2/100—150	1047.8	249.19	1860	.05376	11.844	10.973
C	4.0—4.2/magnetic	998.7	229.19	1460	.06849	12.041	11.110
A782 — Korpi porphyritic granite							
A782A	3.6—3.8/—200	1425.7	323.17	691	.14472	13.192	14.530
B	3.8—4.0/100—200	1213.7	319.25	772	.12953	13.141	12.127
C	4.0—4.2/100—200	913.4	257.15	997	.10030	12.655	10.623
D	+4.2/100—200	547.1	161.15	1380	.07246	12.478	9.556
E	+4.2/—200	606.6	175.40	1060	.09434	12.623	11.274
A898 — Hirsikangas plagioclase porphyrite							
A898A	+4.3	434.9	160.44	2521	.03966	12.027	16.709
B	4.2—4.3	683.2	245.44	2739	.03651	11.944	16.003
C	4.0—4.2	933.8	331.80	2761	.03621	11.961	18.678
A899 — Käpylä posttectonic granite							
A899A	+4.6	299.9	99.50	1731	.05775	12.248	10.988
B	4.2—4.6	496.6	154.17	1696	.05896	12.217	10.813
C	4.0—4.2	698.6	216.67	1367	.07312	12.391	11.666
A938 — Pegmatite vein							
A938A	4.3—4.6	315.4	105.94	627	.15949	13.458	13.795
B	4.2—4.3	663.3	217.10	615	.16235	13.465	14.789
A776 — Diabase dike							
A776A	titanite	94.2	30.17	629	.15898	13.529	7.854

All data on zircons unless otherwise indicated. Concentrations in µg/g, corrected for blank.

that of the intrusive granites. This result is identical, within experimental error, with the result from the pegmatite dyke (A938) for which two zircon fractions suggest a diffusion model age (Wasserburg, 1963) at about 1865 Ma;

5) The plagioclase porphyrite (A898) has a zircon age of 1878 ± 17 Ma;
6) There is no indication that any older material was incorporated into any of the igneous rocks in the Vihanti area.

The lead isotopic composition of the zinc-lead ores at Vihanti plots well below any average global lead evolution curve on the $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram involving the uranogenic lead isotopes (Fig.3), a feature that has led several authors (Stacey *et al.*, 1978; Vaasjoki, 1981) to argue that a considerable amount of mantle derived lead was incorporated into the Vihanti ores. However, on the $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram (Fig.3) depicting lead evolution in the complete U-Th-Pb system, the data plot slightly above the average evolution curve, indicating, assuming the plumbotectonics interpretation of Doe and Zartman (1979), that the lead derived from the lower crust. Thus the lead

Table 2. U/Pb ratios and apparent radiometric ages for zircons and titanite from the Vihanti area.

Sample	Atomic ratios			Apparent ages (Ma)		
	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	T(6/8)	T(7/5)	T(7/6)
A780A	0.3025	4.7845	0.1150	1704	1781	1880
B	0.3183	5.0787	0.1157	1782	1831	1891
C	0.2817	4.4177	0.1138	1600	1720	1862
D	0.2595	4.0511	0.1133	1487	1645	1853
A781A	0.2890	4.5240	0.1135	1637	1736	1857
B	0.2281	3.4974	0.1112	1325	1527	1819
C	0.2193	3.3613	0.1112	1278	1496	1819
A782A	0.2159	3.3700	0.1132	1260	1497	1851
B	0.2548	4.0024	0.1140	1464	1634	1865
C	0.2740	4.2701	0.1148	1562	1687	1878
D	0.2869	4.4999	0.1151	1626	1731	1882
E	0.2793	4.3723	0.1151	1589	1702	1882
A898A	0.3285	5.2045	0.1149	1831	1853	1878
B	0.3221	5.0846	0.1145	1799	1833	1872
C	0.3119	4.9334	0.1147	1750	1807	1875
A899A	0.3078	4.8668	0.1147	1730	1796	1874
B	0.2885	4.5417	0.1142	1633	1738	1867
C	0.2851	4.4813	0.1140	1616	1727	1864
A938A	0.2967	4.6099	0.1127	1675	1751	1843
B	0.2868	4.4417	0.1123	1625	1720	1837
A776A	0.3245	5.0815	0.1138	1812	1834	1861

Atomic ratios corrected for common lead.
6/4:15.15; 7/4:15.15; 8/4:34.90

isotopic data imply that the lead of the orebodies evolved at a relatively deep level before their emplacement. A corollary is that if we assume a syngenetic interpretation for the zinc-lead orebodies, the magmas of the volcanic host rocks must have been generated at similarly deep levels. If that is the case, the predominantly dacitic composition of the Lampinsaari volcanics (Rouhunkoski, 1968) favours a lower crustal derivation.

At Pyhäsalmi the immediate host rocks of the ores are mica schists presumed to represent metamorphosed pelitic sediments (Helovuori, 1979). However, a whole rock Pb-Pb isochron from the local volcanic suite passes through the galena lead isotopic composition and is thus consistent with a cogenetic origin of the sulphides and the volcanic rocks. Also, the sulphur isotopic composition within the massive ore is relatively uniform ($\delta^{34}S$ ranges from +5 to +10 per mil CDT, averaging 7.5 per mil; Helovuori, 1979) thus suggesting a significant input of seawater to a magmatic source. Therefore a submarine volcanogenic origin for the Pyhäsalmi deposit seems highly probable. Relatively old sulphur isotopic analyses from a limited number of samples from Vihanti (Rouhunkoski, 1968) exhibit values similar to those from Pyhäsalmi, but unfortunately the only up to date sulphur isotope data from the Vihanti area are from wall rocks (Rehtijärvi *et al.*, 1979), not the ores.

The radiometric age data from the Vihanti and Pyhäsalmi areas coincide in many ways: the ages of the gabbros are similar, the plagioclase porphyrites from both districts are coeval within experimental error and, apart from some slightly younger titanites found at Pyhäsalmi, there is only one noteable difference: at 1930 ± 15 Ma, the zircon age from the tonalitic Kettuperä gneiss

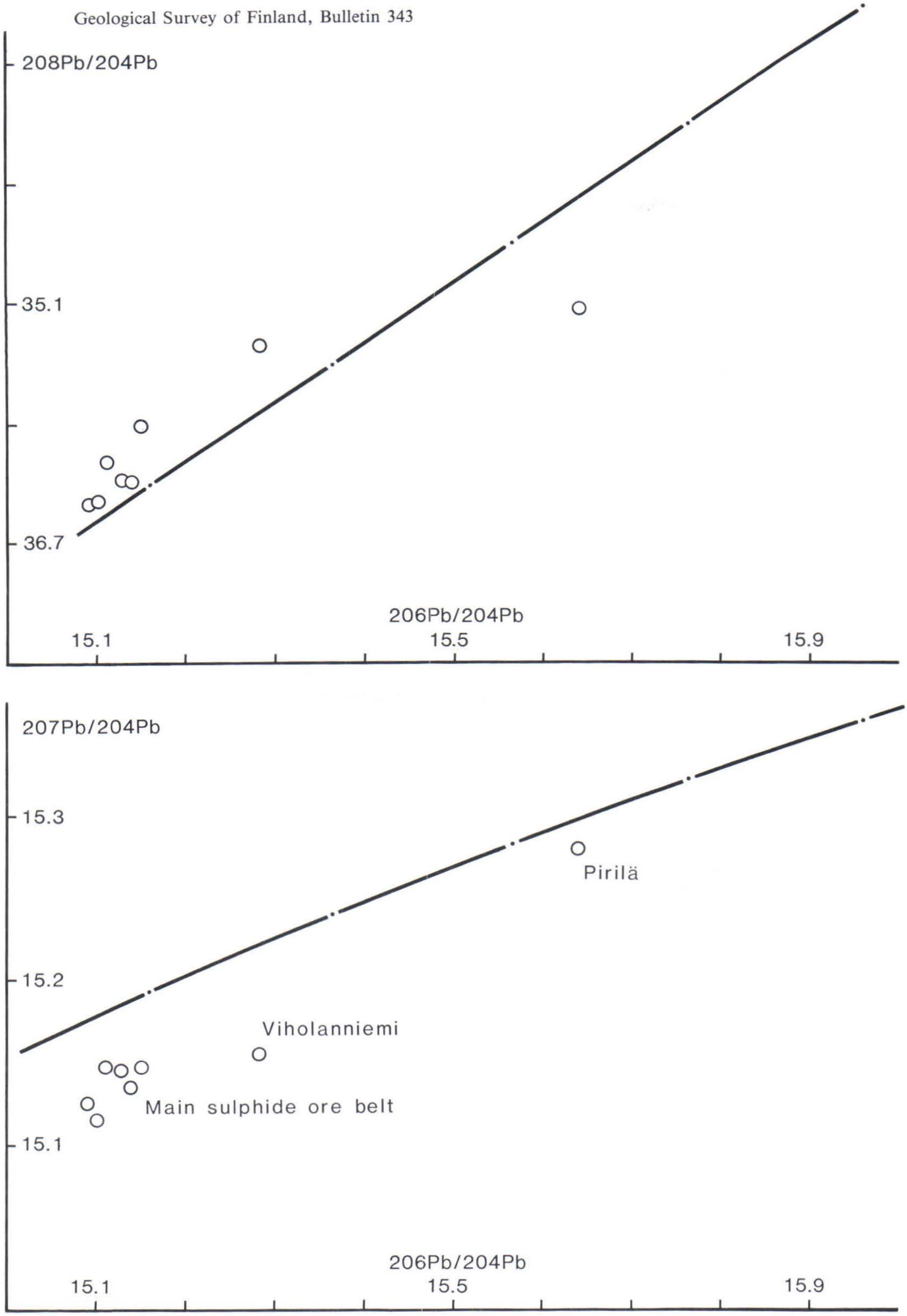

Figure 3. Average lead isotopic compositions of galenas from base metal ores and prospects within the Raahe—Ladoga zone.

at Pyhäsalmi (Helovuori, 1979) is about 50 Ma older than any of the intrusive granitoids at Vihanti.

The similarities in the radiometric ages, the lead isotopic compositions and the sulphur isotope values observed at Vihanti and Pyhäsalmi suggest that these base metal deposits were formed by similar mechanisms. Combined with other geological information, the isotopic evidence implies that the sulphides were formed by volcanogenic seafloor exhalations about 1900 Ma ago and were deformed during the intrusion of gabbros and granitoids between 1900 and 1880 Ma ago. The regional metamorphism associated with the intrusive activity peaked at 1880 Ma, and no significant remobilization of the sulphides has occurred since that time.

The cobbles from the Settijärvi conglomerate are 1888 ± 7 Ma old and demonstrate that syntectonic granitoids were incorporated into this rock and thus suggest that they were extensively weathered (Marttila, 1987). As there is some outcrop evidence suggesting that the conglomerates are intersected by the plagioclase porphyrites, the results imply relatively rapid large scale vertical movements. The ages of the Käpylä granite and the fault controlled pegmatite and diabase dykes at Vihanti suggest that all tectonic activity in the northwestern part of the Raahe—Ladoga zone had ceased by 1860 Ma at the latest but possibly already by 1875 Ma.

PIELAVESI AND RAUTALAMPI: BLOCK MOVEMENTS AND CONTACT METAMORPHISM

Geological setting and previous isotopic work

There are two basic differences between the Pielavesi—Rautalampi and Vihanti—Pyhäsalmi districts: first, the Pielavesi—Rautalampi district lies much closer to established Archaean outcrop than the latter, and, secondly, it is characterized by well defined block structures evident in abrupt changes in the metamorphic mineral paragenesis (Korsman et al., 1984; Hölttä, 1988). K-Ar analyses (Haudenschild, 1988) suggest that these movements lasted for a considerable period after the culmination of the Svecokarelian orogeny. These features make it impossible to establish a clear cut stratigraphy for the area. As a general rule, the supracrustal sequence consists of acid and basic volcanic rocks underlying a metasedimentary pile (Marttila, 1981). The intrusive rocks comprise gneissose tonalites, gabbros, hypersthene granites and porphyritic granites.

Base metal mineralization occurs frequently throughout the Pielavesi—Rautalampi area. Though none of the prospects are currently economically viable, some of them are a fair size. Thus Säviä at Pielavesi has proven reserves of 4 Mt at 1.1% Cu and 2.5% Zn (Salli, 1983), Kalliokylä at Kiuruvesi contains 1.5 Mt at 0.4% Cu and 2.5% Zn, and Hallaperä (Kiuruvesi) has about 3.5 Mt at 0.5% Cu and 1.1% Zn (Marttila, 1981). Other prospects extensively investigated are Kangasjärvi (Kumpuselkä) at Keitele and Pukkiharju at Rautalampi. The galena lead isotopic compositions are identical to those of Vihanti and Pyhäsalmi and internally are homogeneous at least within the Kangasjärvi mineralization (Table 3).

The oldest rocks encountered in the Pielavesi—Rautalampi area are tonalitic gneisses, one of which has been dated at 1922 ± 12 Ma (Korsman et al., 1984). These rocks are regarded as the basement of the overlying volcanosedimentary supracrustal sequence (e.g. Papunen, 1986). The

Table 3. Lead isotopic analyses of galenas from the Raahe—Ladoga zone.

Occurrence	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
Lampinsaari, Vihanti	15.150	15.147	34.896
Pyhäsalmi, Pyhäjärvi	15.111	15.147	34.835
Säviä, Pielavesi	15.102	15.115	34.771
Jylhä, Pielavesi	15.128	15.145	34.805
Kangasjärvi, Keitele	15.140	15.135	34.804
Pukkiharju, Rautalampi	15.091	15.125	34.765
Viholanniemi, Joroinen	15.284	15.155	35.030
Pirilä, Rantasalmi	15.640	15.280	35.090

Data normalized to the accepted values of CIT and NBS SRM981 standards.

supracrustal formation is intruded by both diorites and hypersthene bearing granites which have been dated at 1880—1890 Ma (Salli, 1983; Korsman et al., 1984). According to Marttila (1976 and 1981), these are cut by ophitic gabbros and diabases registering a zircon U-Pb age of 1886 ± 5 Ma. A titanite from a skarn at Rytky has an almost concordant U-Pb age of 1870 Ma (Marttila, 1976).

The quartz diorite from Molkanjärvi, dated at 1882 ± 4 Ma has been analysed for its Nd isotopic composition and has an ε(Nd) value of 1.4—0.9 signifying a fairly substantial input of early Proterozoic mantle material (Huhma, 1986). A sample from the Onkivesi granite/granodiorite (ibid.) right on the Archaean-Proterozoic contact, has a significantly lower ε(Nd) at 3.6—0.9 but is still not as negative as the theoretical value of 10 for 2700 Ma old continental crust at 1900 Ma.

Sample material

Marttila (1976) reported a total fraction zircon U-Pb age for the Lammasaho granodiorite (A600). This determination has now been improved by analysing four fractions separated from the original material.

The Laajamäki quartz diorite (A83) intersects the supracrustal formation in the central part of the Pielavesi map sheet. About 70% of the rock consist of oligoclase-andesine. The amount of quartz is minor, and hypersthene, diopside and biotite are the mafic constituents. The texture of the rock is hypidiomorphic and the weathered surfaces are often a brownish colour. Accessory minerals are apatite, epidote and zircon (Salli, 1983).

The southern margin of the roundish Vaaraslahti granite (Salli, 1983) is in contact with metapelitic rocks, which exhibit a definite contact metamorphic zoning (Hölttä, 1988). The effects of contact metamorphism were studied on two samples, one taken from the contact (A1026) and one from the garnet-cordierite zone (A1025).

In order to compare the timing and effects of metamorphism within the various blocks of the Pielavesi—Rautalampi area a garnet-cordierite-orthopyroxene rock was sampled at Sahinperä (A1028) in the Pielavesi block (cf. Korsman *et al.,* 1984; Hölttä, 1988).

East and north of the township of Rautalampi there is a body of granitoid rocks stretching ESE—WNW over a distance of 5 km and averaging 0.5 km in width. It contains 1—20 cm thick amphibolitic sheets and, judging from these, has been subjected to the same deformational phases as the enclosing metasedimentary-metavolcanic sequence. A sample (A1075) of the granitoid portion was collected at Pyöreänsuonvuori to clarify the age relationship of this rock to the Rastinpää intrusion (Korsman *et al.,* 1984) and the local metavolcanic rocks.

Results and discussion

The results for the Pielavesi—Rautalampi area are summarized in Tables 4 and 5 and Figures 4 and 5. They demonstrate that:

1) At 1923 ± 4 and 1914 ± 4 Ma, respectively, the quartz diorite from Laajamäki (A83) and the sill-like granitoid body represented by the sample from Pyöreänsuonvuori (A1075) are the oldest in the present material;
2) Of the two metapelite samples from the contact aureole of the Vaaraslahti intrusion, the one farther from the contact (A1025) contains a zircon with ages in excess of that of

Table 4. U-Pb analyses of zircons from the Pielavesi—Rautalampi area.

Sample	Fraction	Concentrations		$^{206}Pb/^{204}Pb$	Lead ratios, 206 = 100		
		^{238}U	Pb(tot)		^{204}Pb	^{207}Pb	^{208}Pb
A83 — Laajamäki granodiorite							
A83A	4.3—4.5/HF (*)	452.3	144.65	27100	.00369	11.774	8.867
B	4.3—4.5 (*)	467.0	149.28	8762	.01141	11.886	9.172
C	4.2—4.3/HF (*)	650.0	182.84	12842	.00779	11.806	9.505
D	4.3—4.55/abr	457.7	152.22	8734	.01145	11.891	9.421
A0600 — Lammasaho quartzdiorite							
A600A	+4.2/150—200 (+)	1392.8	424.65	2948	.03392	11.736	9.826
B	+4.3	495.6	164.20	4502	.02221	11.580	11.221
C	4.2—4.3/HF	614.0	210.05	5202	.01922	11.593	11.293
D	4.0—4.2	1093.0	413.41	342	.29266	15.228	20.129
E	4.0—4.2/HF	878.1	303.10	4536	.02205	11.626	11.150
A1025 — Yijäkönmäki metapelite, 80 m from contact							
A1025A	+4.53	399.1	168.85	1132	.08827	13.755	31.184
B	4.3—4.53	510.3	198.82	1468	.06808	13.512	11.044
C	4.2—4.3	732.7	300.41	703	.14225	14.607	13.428
D	4.0—4.2	1337.7	527.74	1001	.09988	13.901	12.981
E	monazite	2460.6	3876.73	—	—	11.548	421.67
A1026 — Yijäkönmäki metapelite, contact to hypersthene granite							
A1026A	4.2—4.3	1004.1	360.75	9413	.01062	11.686	12.017
B	4.3—4.6	451.5	209.67	16318	.00613	11.646	47.066
C	4.0—4.2	448.4	565.22	4419	.02263	11.755	13.651
D	monazite	1793.9	3211.1	105670	.00095	11.503	507.07
A1028 — Sahinperä garnet-cordierite-orthopyroxene rock							
A1028A	+4.58	319.2	160.9	373600	.00027	11.504	66.49
B	4.3—4.58	319.1	105.12	52740	.00190	11.551	2.161
C	4.2—4.3	702.0	225.32	26700	.00375	11.530	.775
D	4.0—4.2	1019.1	318.45	18170	.00550	11.509	.661
E	monazite	2607.1	4396.90	30760	.00325	11.500	461.02
A1075 — Pyöreänsuonvuori granitoid							
A1075A	+4.5/abraded	151.5	53.48	—	—	11.708	7.360
B	+4.5	163.7	55.48	—	—	11.675	7.079
C	4.3—4.5/+100	304.8	104.61	50610	.00198	11.694	7.809
D	4.2—4.3/+100	567.5	198.76	28280	.00354	11.684	10.143
E	4.0—4.2/100—200	1000.4	304.20	15560	.00643	11.593	6.713

All data corrected for blank. Concentrations in µg/g.
(*) From Haudenschild (1985)
(+) From Marttila (1976)

Table 5. U/Pb ratios and apparent radiometric ages for zircons and monazites from the Pielavesi—Rautalampi area.

Sample	Atomic ratios			Apparent ages (Ma)		
	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	T(6/8)	T(7/5)	T(7/6)
A0083A	.3060	4.9470	.1172	1722	1810	1914
B	.3048	4.9300	.1173	1715	1808	1915
C	.2903	4.6830	.1170	1644	1764	1911
D	.3159	5.1123	.1175	1769	1838	1916
A0600A	.2879	4.4762	.1128	1631	1726	1844
B	.3103	4.8253	.1128	1742	1789	1844
C	.3203	5.0043	.1133	1791	1820	1853
D	.3071	4.7675	.1126	1726	1779	1842
E	.3233	5.0497	.1133	1805	1827	1852
A1025A	.3318	5.7579	.1126	1847	1940	2041
B	.3571	6.2084	.1261	1968	2005	2044
C	.3612	6.3373	.1273	1987	2023	2060
D	.3531	6.1234	.1258	1949	1993	2039
E	.3389	5.3953	.1155	1881	1884	1887
A1026A	.3346	5.3261	.1154	1860	1873	1886
B	.3368	5.3700	.1156	1871	1880	1890
C	.3224	5.0906	.1145	1801	1834	1872
D	.3318	5.2563	.1149	1847	1861	1878
A1028A	.32560	5.1687	.1150	1818	1847	1880
B	.33449	5.3151	.1153	1860	1871	1884
C	.32993	5.2219	.1148	1838	1856	1876
D	.32150	5.0687	.1143	1797	1830	1869
E	.33767	5.3338	.1146	1875	1874	1873
A1075A	.34229	5.5254	.1171	1897	1904	1912
B	.32936	5.3038	.1168	1835	1869	1907
C	.33146	5.3318	.1167	1845	1873	1906
D	.33164	5.3209	.1164	1846	1872	1901
E	.29644	4.7029	.1151	1673	1767	1881

Atomic ratios corrected for common lead.
6/4:15.15; 7/4:15.15; 8/4:34.90

the intrusion, while the zircons in the immediate contact (A1026) have been reset to give the age of the intrusion within experimental error.

3) The monazite from A1025 has an age of 1887 ± 4 Ma and the one from A1026 an age of 1879 ± 3 Ma;
4) The zircons from the regionally metamorphosed garnet-cordierite-orthopyroxene rock at Sahinperä yield a U-Pb age of 1889 ± 13 Ma whereas its monazite has an age of 1873 ± 3 Ma;
5) At 1853 ± 12 Ma, the granodiorite from Lammasaho (A600) is the youngest of the present sample material.

The result for the Laajamäki quartz diorite (A83) is basically a three point isochron, as two of the data points crowd together and thus the abraded fraction, D, is determinative for the age. However, the results for both A83 and A1076 are coeval with the well established age of 1922 ± 12 Ma for the Rastinpää tonalitic gneiss (Korsman *et al.,* 1984). Considering that Laajamäki is located on the northeastern side of the Iisvesi fault but that Rastinpää and Pyöreänsuonvuori lie on the southwestern side, it seems certain that

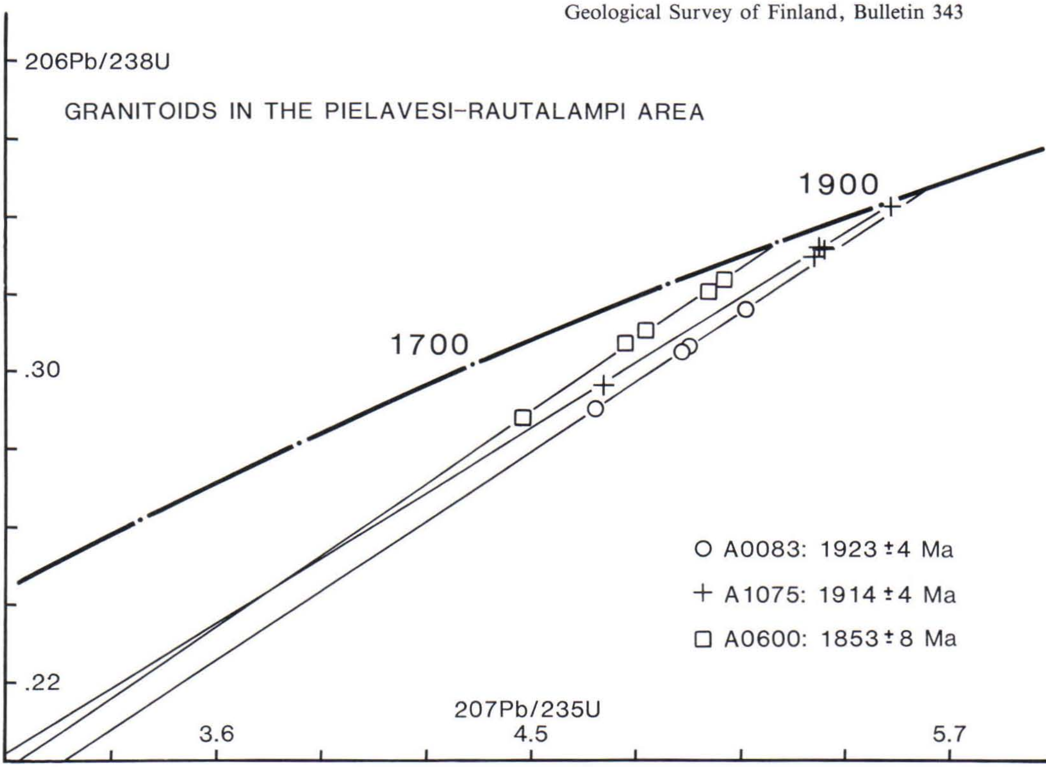

Figure 4. Zircon U-Pb results for granitoid rocks of the Pielavesi—Rautalampi district.

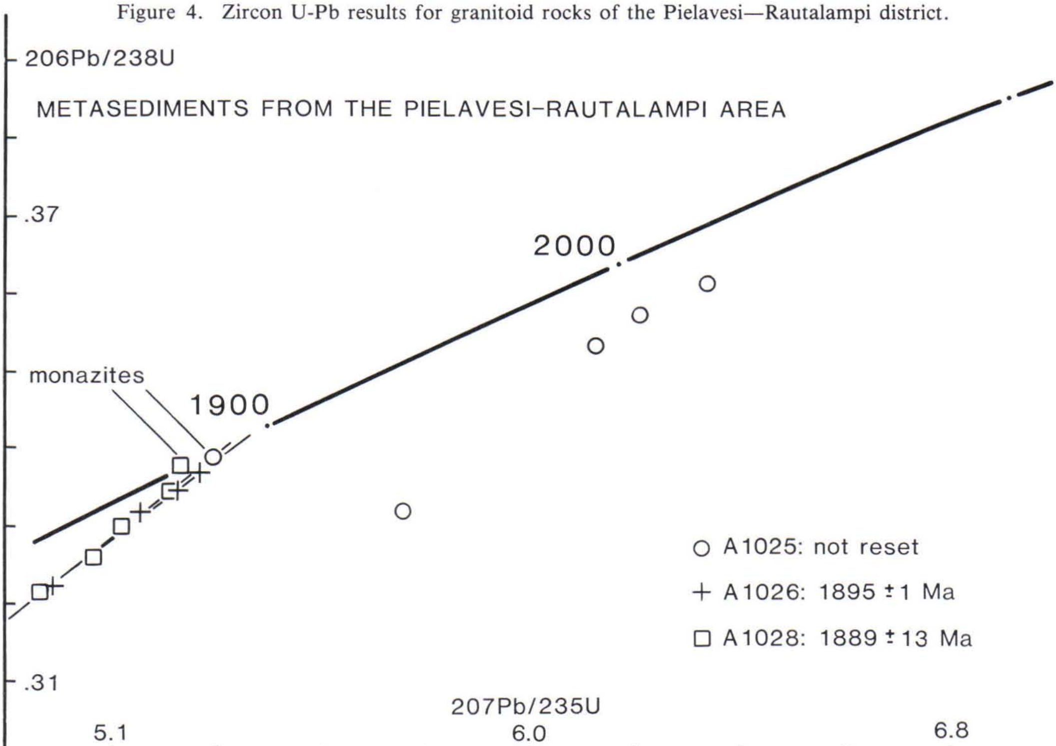

Figure 5. Mineral U-Pb determinations on metavolcano(?)metasedimentary rocks in the Pielavesi—Rautalampi area.

the ca. 1920 Ma granitoid activity was relatively common throughout the Pielavesi—Rautalampi area.

The metapelite samples (A1025, A1026) from the contact aureole of the Vaaraslahti intrusion register different zircon ages with 1892 ± 5 Ma for the contact sample (A1026) and older ages for the one farther away (A1025). As is usual for zircons from a dry environment, the analyses from the contact sample are relatively concordant. The monazite age from A1025 is, at 1887 ± 4 Ma, marginally lower than the ages for the zircons. Thus, within experimental error, the zircons from the contact sample and the monazite from the metapelites farther away are coeval with the zircons of the intruding hypersthene granite dated at 1884 ± 5 Ma (Salli, 1983) and reflect the temperature peak of these rocks. The somewhat lower monazite ages for the intrusion (1874 Ma) and the contact sample (1878 Ma) can be interpreted as the cooling age of the granite. As the probable blocking temperature of monazite is 500—600°C (Odin, 1976), the monazite results indicate that no regional thermal event exceeding that temperature range has occurred within the Vaaraslahti block since probably 1885 Ma ago and certainly not since 1875 Ma ago.

The results from the two metapelite samples from the contact aureole of the Vaaraslahti intrusion thus demonstrate that a K-feldspar-garnet-cordierite grade of metamorphism is insufficient to totally reset detrital zircons. Only at the immediate contact of the hypersthene granite, where the temperature has approached 800°C, has total resetting occurred.

The zircons from the garnet-cordierite-orthopyroxene rock at Sahinperä (A1028) register a U-Pb age of 1889 ± 13 Ma. As Sahinperä lies within the Pielavesi block, this result may not be directly applicable to the regional metamorphism within the Vaaraslahti block, but even so it demonstrates that the time difference between the intrusion of the hypersthene granites and the preceeding regional metamorphism within the Pielavesi—Rautalampi area is very short. The monazite age of 1873 ± 5 Ma may be interpreted as reflecting the regional cooling through the 500—600°C range.

The ophitic gabbro at Tuli-Toiviainen and its kindreds exhibit chilled margins towards their country rocks. They generally occur as dykes and laccoliths, and, according to Marttila (1976 and 1981) they crosscut the hypersthene-bearing granitoids. The age of the gabbro at Tuli-Toiviainen, 1886 ± 5 Ma, is consistent with this notion but also requires rapid regional cooling, a constraint supported by the monazite data.

The age of the Lammasaho granodiorite (A600) is surprisingly low as this rock has been considered part of the gneiss complex forming the basement of the volcano-sedimentary sequence (Marttila, 1976 and 1981). To overcome the stratigraphic difficulty, Marttila (ibid.) attributed the young zircon age of the rock to remobilization of the basement during the late orogenic phase of the Svecokarelian folding. This problem could possibly be solved by Sm-Nd analyses. However, for the present study the question of the stratigraphic position of the Lammasaho granodiorite is immaterial. As the rock exhibits very little or no deformation, the zircon age of 1853 ± 12 Ma sets a minimum for the end of the tectonic activity within the Lampaanjärvi block.

FROM JOROINEN TO SULKAVA: EFFECTS OF PROGRESSIVE METAMORPHISM

Geological setting and previous isotopic studies

The southeastern end of the Raahe—Ladoga zone forms part of the socalled Savo schist belt, which is transitional between the Svecofennian and Karelian supracrustal domains in Finland.

The progressive metamorphism in the area was discovered as a result of the 1:100,000 geological mapping program and has since been studied in detail by Korsman (1977) and Korsman *et al.* (1984).

The metamorphic grade in the area increases from north to south and has been established on the basis of mineralogical observations on metapelites. Particularly in the northern mica schist zone the metapelites have preserved features of their turbiditic origin such as graded bedding, which gradually disappear as the metamorphic grade increases. Thus bedding is still recognizable in the K-feldspar-sillimanite zone, but graded bedding has been obliterated by the formation of sillimanite. Migmatization sets in within the garnet-cordierite-biotite zone, where K-feldspar is often poikiloblastic. In the garnet-cordierite-sillimanite zone the proportion of migmatizing granite increases, the gneissose sections are often separated from each other by granitic segregations and the K-feldspar is no longer poikiloblastic. Furthermore, the cordierite is often pinitized and K-feldspar sericitized while the lower grade zones are almost devoid of signs of retrograde alteration. Korsman *et al.* (1984) estimate a geothermal gradient of about 50°C/km for the area.

The stratigraphic position of volcanic rocks in the Joroinen—Sulkava area is not absolutely certain. Acidic rocks mapped as quartz-feldspar schists and of probable volcanogenic origin underlie mafic-ultramafic lavas that occasionally exhibit pillow structures (Kousa, 1985). The metavolcanic sequence probably lies stratigraphically below the metapelite sequence and is in any case definitely older than the syntectonic tonalites.

The Joroinen—Sulkava area differs from other parts of the Raahe—Ladoga zone in the paucity of base metal mineralization, although Ni ores associated with gabbros occur with regular frequency (cf. Tontti, 1981). Thus the gold prospect at Pirilä and the Zn target at Viholanniemi are the only known occurrences in the area which contain any base metal accumulations worth speaking of; neither one of them is considered economic at present.

Korsman *et al.* (1984) reported a number of U-Pb zircon ages from the Rantasalmi—Sulkava district. The tonalite at Tuusmäki, which precedes the progressive metamorphism, gives a normal Svecokarelian syntectonic age of 1888 ± 15 Ma. The migmatite from Säviönsaari, in the centre of the Sulkava thermal dome, yields ages of 1810 ± 7 and 1833 ± 16 Ma for palaeosome and neosome, respectively. Monazites in the same samples are between 1820 and 1840 Ma old. A definitely post-tectonic granodiorite from Hirvensalo registers a zircon age of 1802 ± 22 Ma and contains an almost concordant titanite dated at 1764 ± 19 Ma. In the Haukivesi block, just northeast of the study area, there are hypersthene granites dated to be about 1890 Ma old (Patchett and Kouvo, 1986). The whole rock ε (Nd) value of -0.6 and zircon ε (Hf) result of $+1.6$ suggest that a large part of the material for this rock is of mantle derivation.

About 2 km west of the sampling site for A1013 (see next section), a trondhjemitic tonalite at Saunakangas has been dated at 1903 ± 10 Ma (Huhma, 1986). However, geophysical data suggest that this rock may not belong to the same block with the progressively metamorphosed rocks of the Joroinen—Sulkava area (Korsman *et al.*, 1988).

Sm-Nd analyses of the Saunakangas tonalite give an ε(Nd) value of 3.3 ± 0.5, which is suggestive of an origin from a depleted mantle source (Huhma, 1986). Similarly, about 15 km east of the Joroinen—Sulkava area, the Ni-bearing Laukunkangas gabbro has been dated at 1880 Ma and exhibits an ε(Nd) value of 0.2 ± 0.5 (ibid.). A sample from the post-tectonic Puruvesi granite, about 30 km farther east, has a zircon U-Pb age of 1797 ± 19 Ma (Nykänen, 1983) and exhibits an ε(Nd) value of -6.9 ± 0.8, which is indicative of a significant Archaean crustal component in its source (Huhma, 1986).

The few ore lead isotopic analyses from the Joroinen—Sulkava area differ markedly from the others in the Raahe—Ladoga zone (Table 3). Moreover, the leads from Viholanniemi and Pirilä are different from each other: Pirilä has a lead isotopic composition similar to the mineralizations of the central Finnish batholith area (cf. Vaasjoki, 1981), whereas the Viholanniemi lead is intermediate between this group and the leads of the Vihanti—Pyhäsalmi ore zone.

Sample material

Partly to study the effects of regional metamorphism on the detrital zircons in the metapelites and partly to investigate their provenance, two samples were collected from the metapelites. The one from Vuotsinsuo (A15) represents the mica schist zone and the one from Härkälä (A1022) the northern part of the garnet-cordierite-biotite zone. The latter sample also contained some monazite.

A quartz-feldspar schist from Viholanniemi was sampled as a representative of the acid volcanic rocks. The porphyric texture of the rock (A1013) is evident in hand specimen, and flow textures around the feldspar phenocrysts can be observed in thin section. This rock is also host to the Viholanniemi base metal mineralization.

A reddish granite intersects the metapelites in the K-feldspar-sillimanite zone exhibiting structures typical of the D_3 deformation phase close to the Pirilä gold prospect (Kilpeläinen, 1988). A sample from this granite (A1099) was collected to determine a minimum age for this deformation phase.

Results and discussion

The results are summarized in Tables 6 and 7 and Figures 6 and 7. It is apparent that:

1) The zircons from both metapelites (A15-Vuotsinsuo and A1022-Härkälä) are definitely

Table 6. U-Pb analyses of samples from the Joroinen—Sulkava area.

Sample	Fraction	Concentrations		$^{206}Pb/^{204}Pb$	Lead ratios, 206 = 100		
		^{238}U	Pb(tot)		^{204}Pb	^{207}Pb	^{208}Pb
A15 — Vuotsinsuo micaschist (metapelite)							
A0015A	+4.5/+200	145.9	58.52	2358	.04241	14.950	18.012
B	4.3—4.5/+100	289.1	112.78	3318	.03014	14.379	14.389
C	4.2—4.3/+200	525.1	189.69	3431	.02915	14.217	13.160
A1022 — Härkälä cordierite gneiss (metapelite)							
A1022A	4.3—4.55	475.7	169.17	2662	.03756	13.036	9.228
B	4.2—4.3	855.4	295.66	1551	.06448	13.278	10.092
C	4.0—4.2	1314.7	437.45	1056	.09474	13.565	11.473
D	monazite	7613.8	5292.83	86850	.00115	10.997	139.617
A1013 — Viholanniemi metarhyolite							
A1013A	+4.55	149.7	52.22	5616	.01780	11.876	12.613
B	4.3—4.55/+200	236.9	80.47	8024	.01246	11.799	14.715
C	4.2—4.3/—150	495.9	139.26	3924	.02548	11.876	18.793
A1099 — Pirilä granite							

Concentrations in µg/g; corrected for blank.

Table 7. U/Pb ratios and apparent radiometric ages for samples from the Joroinen—Sulkava area.

Sample	Atomic ratios			Apparent ages (Ma)		
	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	T(6/8)	T(7/5)	T(7/6)
A0015A	.34557	6.8561	.1439	1913	2092	2274
B	.34792	6.7055	.1398	1924	2073	2224
C	.32573	6.2109	.1383	1817	2005	2206
A1022A	.33368	5.7642	.1253	1856	1941	2033
B	.31995	5.4730	.1241	1789	1896	2015
C	.30235	5.1203	.1228	1702	1839	1997
D	.31877	4.8260	.1098	1783	1789	1796
A1013A	.32249	5.1743	.1164	1801	1848	1901
B	.30919	4.9587	.1163	1736	1812	1900
C	.24695	3.9270	.1153	1422	1619	1885

Atomic ratios corrected for common lead.
A1013: 6/4:15.15; 7/4:15.15; 8/4:34.9
A15 and A1022: 6/4:15.68; 7/4: 15.46; 8/4:35.27

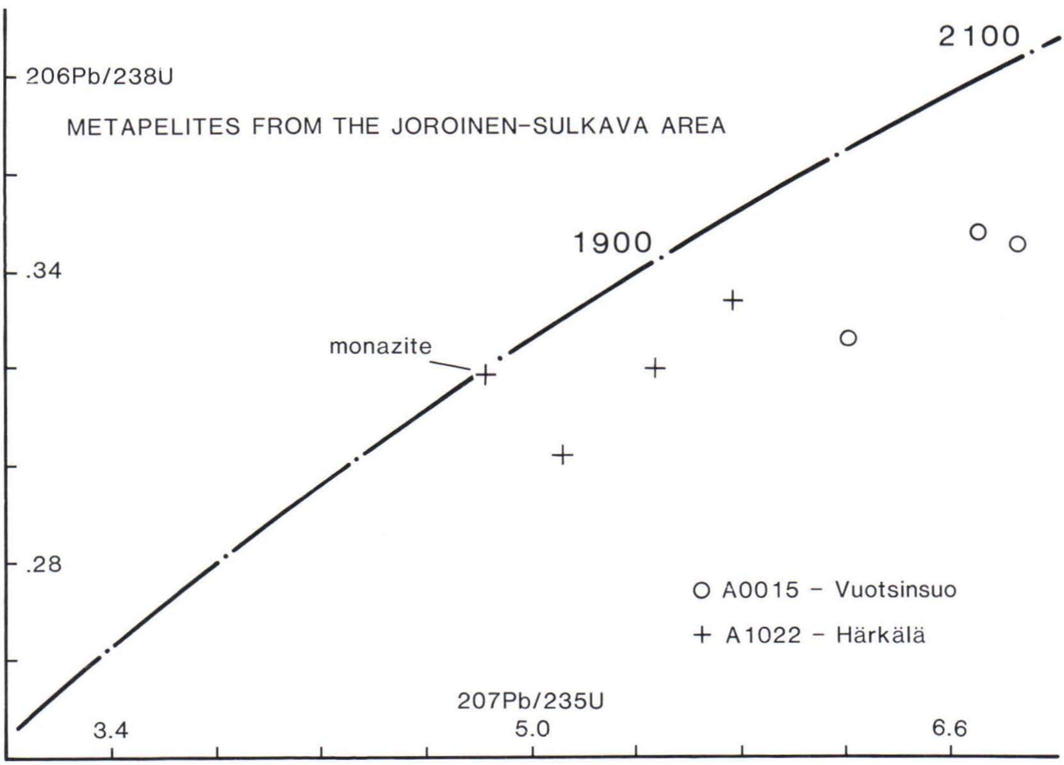

Figure 6. Zircon and monazite U-Pb data from the metapelites of the progressively metamorphosed Joroinen—Sulkava district.

older than the intruding granitoids, with the less metamorphosed sample (A15) containing the fractions with the oldest apparent ages;

2) The monazite of the sample from the garnet-cordierite-biotite zone (A1022) has an almost concordant age of 1796 ± 5 Ma;

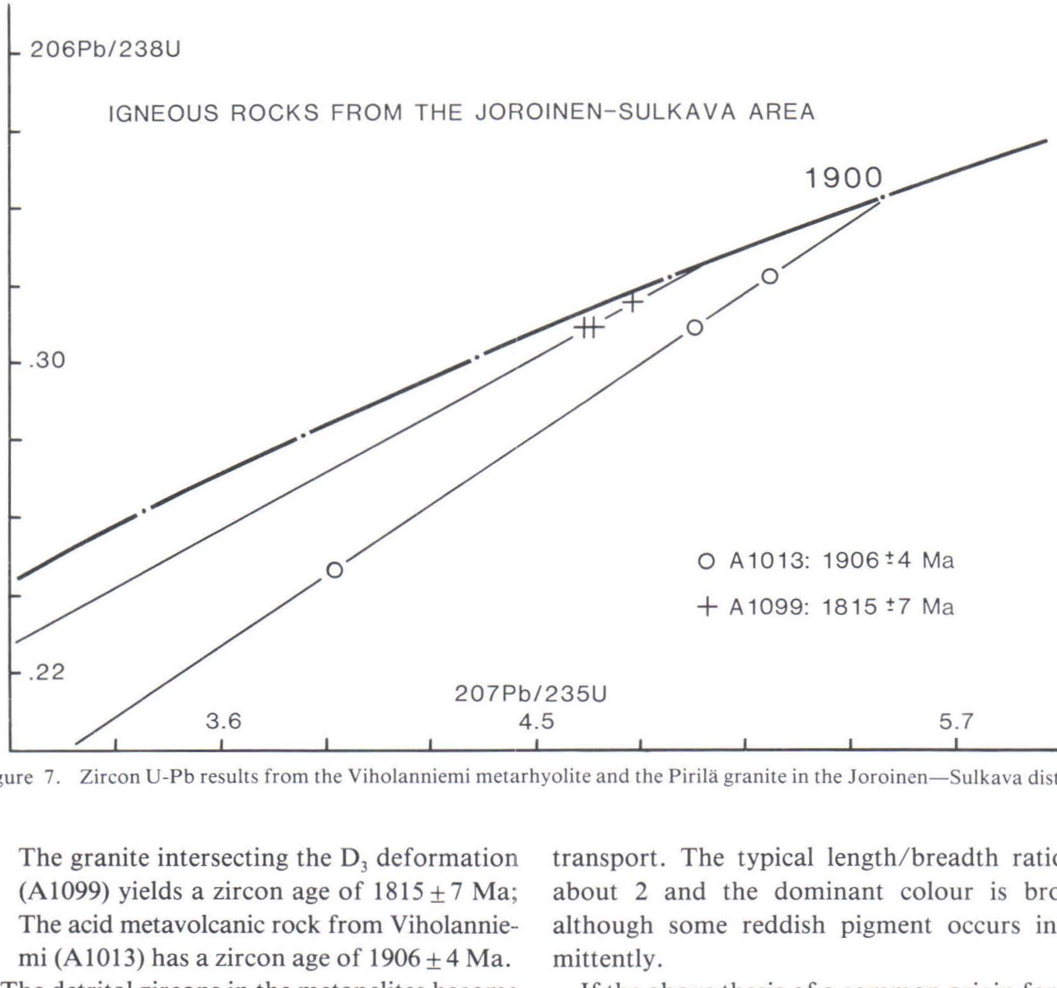

Figure 7. Zircon U-Pb results from the Viholanniemi metarhyolite and the Pirilä granite in the Joroinen—Sulkava district.

3) The granite intersecting the D_3 deformation (A1099) yields a zircon age of 1815 ± 7 Ma;
4) The acid metavolcanic rock from Viholanniemi (A1013) has a zircon age of 1906 ± 4 Ma.

The detrital zircons in the metapelites become progressively younger as the grade of regional metamorphism increases. Thus the $^{207}Pb/^{206}Pb$ ages of the zircons range from 2200 to 2300 Ma in the mica schist zone (A15) but are no more than about 2000 Ma (A1022) in the garnet-cordierite-biotite zone. In the middle of the Sulkava thermal dome (Säviönsaari, Korsman *et al.*, 1984), the zircons are totally reset in granulite facies metamorphism.

We believe that the zircons have roughly the same origin. This is largely because of their morphological similarities, as in all the fractions analysed the majority of the crystals consist of semi-euhedral tetragonal prisms terminating in pyramid surfaces rounded by abrasion during transport. The typical length/breadth ratio is about 2 and the dominant colour is brown although some reddish pigment occurs intermittently.

If the above thesis of a common origin for the zircons in all the metapelites of the Joroinen—Sulkava area is accepted, it follows that the $^{207}Pb/^{206}Pb$ age of the least metamorphosed zircons, namely those in sample A15, is the minimum age for their original material. As even the zircons of A15 must have lost some lead, it is unlikely that they derive from the Jatulian (2100—2200 Ma) volcanic rocks although albititic varieties of these rocks contain appreciable amounts of zircon. The next youngest candidates could be the feldspar porphyries (2350 Ma) that cross-cut the Sumi—Sariolian rocks in Soviet Karelia (Kratz *et al.*, 1976). However, although Pekkarinen (1979) recorded the possible existence of these rocks in Finland, they are not widespread, and the zircons from neither A15 nor A1022 have a morphology indicating volcanic origin. Thus the most likely source for the zircons of the metapelites in the Joroinen—Sulkava area is the granitoids of the Archaean basement (>2500 Ma) complex.

We emphasize that the above reasoning applies only to zircons, which during weathering and transport are among the most resistant phases. As demonstrated by Huhma (1987), the ε(Nd) values for both the Kalevian turbidites and the metasediments of the Tampere region imply that they consist principally of material younger than the Archaean. Although no Sm-Nd data exist for the Joroinen—Sulkava area, it is hardly conceivable that their ε(Nd) values would differ significantly from those of the other Svecokarelian metasediments. Considering the distribution of REE among the various mineral species commonly present in pelitic sediments, we conclude that the bulk of the metasediments in the Joroinen—Sulkava area probably contain material derived from rocks formed from mantle material between 2100 and 1910 Ma ago.

As at Vaaraslahti, it is apparent that even an intense garnet-cordierite-sillimanite-biotite grade of metamorphism is insufficient to totally reset the zircon U-Pb systems. Only at Säviönsaari in the centre of the Sulkava thermal dome, where the temperature has exceeded 750°C, were the zircons reset.

The preceding results agree with findings from the Tampere schist belt, where detrital zircons in a micaschist inclusion in the centre of a 1880 Ma old granodiorite still exhibit an age of about 2300 Ma (Wetherill *et al.*, 1962). Similarly, a granulite at Myösäjärvi in Lapland (Meriläinen, 1976) contains two zircon generations of slightly different ages, the detrital one being older whereas the younger one has the same age as the metamorphic monazite. Thus the finding that zircons retain most of their radiogenic lead during prolonged heating at temperatures exceeding 600°C (Mattinson, 1982) may be amended by the statement that in many cases lower granulite facies conditions are insufficient to totally reset detrital zircons.

This conclusion does not necessarily contradict other findings (e.g., Gebauer and Grünenfelder, 1976) which indicate that some zircons have lost large amounts of radiogenic lead already in greenschist facies conditions. In these cases the resetting is not total. Furhtermore, excessive amounts of OH-groups (e.g., Grünenfelder *et al.*, 1968) and common lead (Vaasjoki, 1977) seem to influence the degree of discordancy of zircons. It is therefore probable that zircons with initially disturbed lattices are more easily reset during metamorphism than those whose structure approaches the ideal configuration. We therefore conclude that the partial resistivity of zircons in high-grade metamorphism is rather the rule than the exception.

For the acid metavolcanic rock at Viholanniemi (A1013), neither its age nor its association with base metal mineralization presents any surprise. However, the galena lead isotopic composition of the mineralization is somewhat different from that prevailing in the Raahe—Ladoga zone from Vihanti to Rautalampi. Even more significantly, at Pirilä, only 15 km east-southeast of Viholanniemi, the galena lead is typical of the Central Finnish batholith area. Thus, over a relatively short distance, there is a marked increase in the crustal component in base metal mineralization.

CRUSTAL EVOLUTION ALONG THE ARCHAEAN-PROTEROZOIC BOUNDARY

Several authors have proposed plate tectonic models for the Fennoscandian (Baltic) Shield. Piirainen (1975) suggested a suture dipping NE along the Archaean-Proterozoic boundary, while

Hietanen (1975) proposed a suture in the same environment and an island arc embracing the SW Finnish and central Swedish Svecofennian supracrustal rocks. Gaál and Gorbatschev (1987) suggested a basically Andean type collision model for the Svecokarelian (in their terminology the Svecofennian) orogeny.

On the other hand, for lack of evidence, many geologists contest the applicability to the Precambrian of plate tectonic principles as observed today (e.g., Windley, 1977; Miyashiro *et al.*, 1982). Thus Kröner (1977) and McWilliams and Kröner (1981) have suggested a mechanism for Proterozoic intraplate orogenies whereby, after the initial break-up of the continental plate, all processes of subsidence and deformation would occur within the continental mass, and the horizontal distances during the processes would be relatively short. Recently Edelman and Jaanus—Järkkälä (1983) and Brannigan (1987) have interpreted tectonics in southwestern Finland using such a model. The one point most students of the Precambrian seem to agree on is that during the early development of the Earth, crustal nuclei were formed that served as "backbones" of the continents (e.g., Moorbath, 1985).

Sm-Nd results obtained by Huhma (1986) indicate that the roughly 1.9 Ga old granitoids of the Karelian province, lying northeast of the Raahe—Ladoga zone, contain a significant crustal component presumably derived from the Archaean, whereas equivalent rocks of similar age in the Svecofennian province exhibit much higher ε(Nd) values and therefore must contain a large Proterozoic mantle component. Similarly, a study of sedimentary provenances indicates short crustal residence times for the Proterozoic material in the Tampere Schist Belt, whereas rocks from the Karelian province contain a higher Archaean component (Huhma, 1987). Thus these Sm-Nd results strongly suggest that juvenile Proterozoic material was rapidly added to the margin of the Archaean plate during the Svecokarelian orogeny, which culminated about 1.9 Ga ago.

A similar picture emerges from lead isotopic studies of galenas. Leads from the Svecofennian domain and the ophiolite-related Outokumpu deposit are deficient in ^{207}Pb and form a definite linear trend. The Karelian and Archaean leads found northeast of the Raahe—Ladoga zone, in contrast, contain an excess of this isotope, and therefore appear to be derived from an older crustal source (Vaasjoki, 1981). Furthermore, palaeomagnetic data from the Raahe—Ladoga zone and the Archaean Varpaisjärvi area are fundamentally different from each other (Neuvonen *et al.*, 1981) and hence argue for different geological and thermal histories for these two areas.

Should an intraplate orogenic model be correct for the Svecokarelian orogeny, one would expect the isotopic data to be rather symmetrically distributed. A probable distribution would be a relatively narrow belt of mantle-related lead and neodymium signatures enclosed on either side with data indicating fairly long crustal residence times for their material. However, the data from Finland are definitely asymmetric in distribution. In southern Sweden, there is no indication of any crust older than Svecokarelian, as the younger Gothian and Sveconorwegian provinces have been added to the margin of the Fennoscandian Shield (Lindh, 1987). The only report of older material within the Svecofennian province concerns some ancient cores in zircons of the Vehmaa rapakivi batholith (Vaasjoki, 1977) which at best can represent only a very minor volume of the Svecofennian province. For the evolution of the Fennoscandian Shield in lower Proterozoic times isotopic results overwhelmingly favour a model according to which juvenile material was added to an older crustal core at an active continental margin. The mechanism may have been similar to the plate tectonic processes commonly observed in Phanerozoic environments.

However, though plates and tectonics must have existed throughout the Precambrian, exact analogies cannot and should not be drawn from present observations. One fact that should be borne in mind is that, according to geological evidence, the conductive geotherm for continental crust in the Precambrian was similar to that observed today and, consequently, because of higher heat generation, recycling of oceanic plates must have been faster (e.g., Bickle, 1978).

Park *et al.* (1984) and Park (1985) have suggested that over a period of about 50—100 Ma successive island arcs were formed in the Svecokarelian domain with progressively younger volcanism towards the south. This model may be close to the truth, but its timing is not supported by radiometric data from volcanic rocks. The relatively few data available so far (Helovuori, 1979; Aho, 1979; Patchett and Kouvo, 1986; Kähkönen *et al.*, in press; this study) indicate two periods of volcanic activity, 1910—1900 and 1890—1880 Ma. Both volcanic pulses have occurred at least in the Pyhäsalmi and Tampere areas, and no volcanism younger than 1880 Ma associated with the Svecokarelian orogeny has been detected even in the south Finnish archipelago. Thus there is no indication of measurable age differences between various volcanic areas in the Svecofennian domain of Finland.

The radiometric data from the Raahe—Ladoga zone presented in this paper suggest that there were differences in the tectonic evolution in its various parts. As is demonstrated by the present results, at Vihànti the earliest intrusive rocks were emplaced 1900 Ma ago and all granitoid activity, including that of clearly posttectonic granites terminated about 1860 Ma ago, while the regional metamorphism culminated about 1880 Ma ago. Pegmatites and diabase dykes emplaced in fault zones cutting all phases of ductile deformation register titanite and zircon U-Pb ages of about 1860 Ma. Thus all orogenic movements in the Vihanti area seem to have occurred within a time interval no longer than 40 Ma.

From Pyhäsalmi to Pielavesi and Rautalampi, tonalitic gneisses preceding all deformation and of probable mantle derivation have been dated at 1920 Ma. Early orogenic gabbros as well as syntectonic granitoids cutting the first deformation phase yield ages of about 1890—1880 Ma. As demonstrated by the Lammasaho granodiorite, tectonic activity in this area ceased about 1850 Ma ago.

However, in the Rantasalmi—Sulkava area, migmatitic rocks were still being generated 1830 Ma ago and the first posttectonic granitoids were emplaced about 1800 Ma ago. Thus, in this area, the tectonic processes lasted up to 50 Ma longer than elsewhere in the Raahe—Ladoga zone.

As is evident from Fig. 1. on p. 6, in the Vihanti area, which displays short-lived tectonic activity, the Raahe—Ladoga zone is surrounded by Svecofennian rocks, whereas the prolonged tectonism in the Savo Schist Belt occurred in an area where the Archaean craton beneath the Karelian formation was probably in direct contact with the accreting crust. The situation here is similar to that encountered in the Wopmay orogene in Canada (Hoffman and Bowring, 1984) where, after a short culmination of the main orogenic activity, a second compressional phase resulted in further deformation and reactivation of the newly formed continental crust. It thus seems feasible to suggest that whatever the accretionary mechanism was at the margin of the Archaean continent, it led to a thickening of the newly forming material at the immediate zone of collision. This in turn resulted in higher temperatures in the piles of juvenile material giving rise to prolonged periods of heating.

Another factor worth consideration is that the Savo schist belt forms the hinge between the Raahe—Ladoga zone and the Svecofennian supracrustal formations running east-west through southern Finland and central Sweden (Ward, 1987). Thus the differences in isotope geology of the Joroinen—Sulkava area and the rest of the Raahe—Ladoga zone may be the result of a process that affected only southern Finland and central Sweden, leaving other parts of the continental crust generated during the Svecokarelian orogeny untouched.

Volcanic rocks, early orogenic gabbros and syntectonic granitoids throughout Finland southwest of the Raahe—Ladoga zone register a relatively narrow range of ages from 1910 to 1870 Ma (e.g., Patchett *et al.,* 1981; Patchett and Kouvo, 1986; Hopgood *et al.,* 1983; Häkli *et al.,* 1979; Welin *et al.,* 1983). However, south of the line from Puruvesi to Pori (or about 62°N) ages of posttectonic granitoids range from 1820 to 1760 Ma (Vaasjoki, 1977; Simonen, 1982; Welin *et al.,* 1983; Korsman *et al.,* 1984, Nykänen, 1988); no posttectonic activity younger than 1850 Ma has been recorded north of the line. Similarly, only in the southern Svecofennides do migmatizing granites occur in the 1810—1840 Ma age range (e.g., Korsman *et al.,* 1984; Hopgood *et al.,* 1984; Huhma, 1986). Thus there is a distinct likelihood that a major tectonic boundary exist in southern Finland along the 62°N line.

Examining the concept of a separate southern Finnish block still further, we note that the next significant addition to the crust of the Fennoscandian Shield occurred during the Gothian episode about 1750—1650 Ma ago (Claesson, 1987; Lindh, 1987) in central and southern Sweden. The contact of this domain with the pre-existing continental crust trends almost N-S, and consequently the thrust from an oceanic plate should have come (in present day directions) from almost due west.

Our suggestion therefore is that the Svecofennian crust, as it is exposed today, was generated by a collision of Proterozoic oceanic and Archaean continental plates between 1930 and 1850 Ma ago. This plate movement stagnated at about 1850 Ma but was, in the present-day southern Finland, immediately followed by eastward underplating of the newly formed continental crust, resulting in its migmatization around 1830—1810 Ma ago, the emplacement of the 1800—1780 Ma posttectonic granitoids and the formation of new continental crust in the Gothian area of Sweden from 1750 Ma onwards. Thus the present isotopic data from the Raahe—Ladoga zone could be interpreted as being indicative of several different plate tectonic processes closely related in time and space.

CONCLUSIONS

From studies of metapelites it is evident that, although detrital zircons tend to lose increasing amounts of radiogenic lead as the metamorphic temperature rises, even garnet-cordierite-biotite and garnet-cordierite-sillimanite grades of metamorphism do not result in the total resetting of zircon U-Pb systems. Our evidence from the contact of the Vaaraslahti intrusion and the Sulkava thermal dome suggest that temperatures in excess of 750°C are required to totally reset zircons.

The main geochronological groups along the Raahe—Ladoga zone may be summarized as follows:

1930—1910 Ma: volumetrically minor but spatially widely distributed granitoid rocks in the central part of the belt from Pyhäsalmi to Rautalampi;

1910—1900 Ma: intermediate-acidic lavas associated with base metal mineralization all along the belt;

1900—1880 Ma: early orgenic gabbros with associated nickel ores;

1890—1875 Ma: intrusion of syntectonic granitoids, hypersthene granites and plagioclase porphyrites/ophitic gabbros all along the belt accompanied by the peak of regional metamorphism in the central and northwestern parts;

1880—1850 Ma: emplacement of posttectonic rocks in the northwestern and central parts of the belt;

1830—1810 Ma: peak of regional metamorphism and migmatization in the southeastern part of the belt;

1805—1795 Ma: emplacement of posttectonic rocks in the southeastern part of the belt.

These results, together with other isotope data from Finland, suggest that the new Svecofennian crust was rapidly generated between 1930 and 1850 Ma ago by the collision of a Proterozoic oceanic plate with an Archaean continent. Subsequent underplating in a W-E direction south of 62°N resulted in migmatization at about 1820 Ma and the emplacement of the ca. 1800 Ma posttectonic granites in southern Finland.

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Appendix 1. Description of analytical methods.

The minerals were separated from crushed rock samples (10—100 kg) using a shaking table, heavy liquids (methylene diodide, Clerici's solution) and magnetic procedures. The sizes of the analysed fractions varied from 1 to 10 mg, being typically about 5 mg. Some fractions were treated with the air abrasion technique (Krogh, 1982) in order to produce more concordant data. The fractions were finally purified by careful hand-picking and washed in an ultrasonic bath in 3 N HCl prior to dissolution.

Lead and uranium were dissolved and purified as described by Krogh (1973). However, for titanites and monazites uranium was separated from iron by hexone extraction. The lead from some of the galenas was purified using the anodic electrodeposition technique described by Gulson and Mizon (1979).

Linear regressions on the concordia diagrams were calculated according to the method of York (1969) assuming an accuracy of 1 ± % for the U/Pb ratios and a 90% error correlation.

Appendix 2. Locations of analysed samples.

Sample	Location	Map sheet	Northing	Easting
A 780	Alpua, Vihanti	243408	7148.62	561.58
A 781	Hirsikangas, Vihanti	243405	7141.72	554.20
A 782	Korpi, Vihanti	243402	7140.80	548.99
A 776	Lampinsaari, Vihanti	243405	7145.80	555.37
A 898	Hirsikangas, Vihanti	243405	7141.60	554.05
A 899	Käpylä, Vihanti	243404	7136.00	553.20
A 938	Lampinsaari, Vihanti	243405	7146.15	555.46
A 600	Lammasaho, Kiuruvesi	332311	7058.80	492.60
A 83	Laajamäki, Pielavesi	331408	7028.70	481.00
A1025	Yijäkönmäki, Pielavesi	331409	7031.55	488.13
A1026	Yijäkönmäki, Pielavesi	331409	7031.41	488.07
A1028	Sahinperä, Kiuruvesi	332307	7042.95	482.03
A1075	Pyöreänsuonvuori, Rautalampi	322312	6945.40	495.68
A 15	Vuotsinsuo, Joroinen	323303	6888.29	544.39
A1022	Härkälä, Rantasalmi	323308	6872.80	562.80
A1013	Viholanniemi, Joroinen	323401	6890.16	540.50
A1099	Pirilä, Rantasalmi	323306	6881.20	555.80