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Early Proterozoic Karelian and Svecofennian formations and the Evolution of the Raahe-Ladoga Ore Zone, based on the Pielavesi area, central Finland

by Elias Ekdahl

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Studies in the Pielavesi district, in the boundary zone between the Svecofennian and Karelian domains, support the notion that their development was, at least to some extent, coeval. Consequently, it is not possible to draw any clear time-stratigraphic boundary between the two domains. The whole boundary zone, including the Pielavesi district, can be correlated with the succession in northern Ostrobothnia, where the lowermost graywacke deposits represent Jatulian sedimentation. The sequence in the Pielavesi area can be considered to contain the type sequence of the Savo schists, which have traditionally regarded as Svecofennian. Depositional basement is exposed in a number of structural culminations and consists of polydeformed ortho- and paragneisses whose age is likely to be greater than the 1925 Ma date indicated by U-Pb zircon studies.

The Pielavesi sequence commences with low-K arc tholeiites followed by extensive deep-marine sediments with graphite-rich and volcanogenic intercalations. These are overlain by the shelf-type Savijärvi Suite, which contains various dolomites and impure carbonate rocks, calc-silicate lithologies, felsic volcanics, cherts, minor iron formations and graphitic schists, and a distinctive uranium-phosphorus (U-P) horizon. This association is widespread throughout the whole transitional zone between Vihanti and Virtasalmi area, and is correlated with the Karelian Marine Jatulian association of the Kuopio district, which thus represents an epicontinental facies association within the Savo schists.

A plate tectonic model is presented in which continental break-up recorded by Marine Jatulian deposition led to the formation of the Kainuu-Outokumpu Back Arc between 2.08-1.97 Ga, as a result of NNE-directed subduction processes. The Outokumpu assemblage contains ophiolitic complexes and rift-related sediments and the extensive ca. 1.97 Ga Cyprus-type mineralization in the Outokumpu district is considered to be a consequence of extremely slow sea-floor spreading rates, compared to the environment in which the Jormua ophiolite was formed. The Pyhäsalmi Island Arc (PIA) at the Archean craton margin ($\geq$1921 Ma) records the initiation of compressive deformation and Kuroko-type Zn-Cu-Pb deposits formed during this time within the upper part of a bimodal island arc volcanic sequence (Säviä Suite). In the Pielavesi district the Kalliokylä conglomerate horizon records a hiatus between the Savijärvi and Säviä Suites.

The overlying Koivujoki Suite in the Pielavesi area consists predominantly of metagraywackes that are correlated with similar sequences in the Virtasalmi and Haukivesi regions and they are probably the stratigraphic equivalents of both the Rahkamäki Formation in the Kuopio area and the Upper Kalevian Savo Province mica schists of the Outokumpu district. The Kotajärvi Metalava (1882 Ma) is younger than at least part of the Koivujoki Suite metagraywackes and was coeval with the volcanism in the Pihtipudas district, immediately to the west of Pielavesi.

Still further west, in central Ostrobothnia, even younger conglomerates, graywackes and polymodal volcanics occur and these may be correlated with the Vieremä-Haajainen deposits, which outcrop in a narrow zone between Pielavesi and the Archean basement to the northeast. The Savo schists can be traced eastwards as far as the Höytiäinen province, while to the west they terminate against the gravity trough, where they are juxtaposed against the Pihtipudas sequence. They can also be followed as far south as Sulkava, but elsewhere, boundaries are more difficult to define.

The development and metallogenesis of the Raahe-Ladoga Zone (RLZ) represents a Wilson cycle recording prolonged extension culminating in continental break-up and sea-floor spreading, and subsequent closure and collision during the Svecofennian Orogeny. The metallogenic evolution of the zone has been subdivided into five distinct provinces and stages. The extensional preorogenic stage does not appear to have been significant from the viewpoint of ore formation but the so-called pre- to early orogenic stage was associated with the Marine Jatulian sediment-hosted U-P horizon and iron formations, as well as magmatic-hosted Fe-Ti-V mineralization. The most economically important ores are the Cyprus- and Kuroko-type deposits formed during the early orogenic stage. The synorogenic Kotalahti Nickel Belt (KNB 1.89-1.88 Ga) is closely related to the geosuture along the active continental margin, which is also characterized by I-type calc-alkaline intrusions that host relatively minor Cu±Mo±Au±W porphyry-type mineralization and epigenetic (1.88-1.85 Ga) Au-As occurrences. The metallogeny of the synorogenic stage in the Raahe-Ladoga Zone thus corresponds to Cordillera-type orogenesis.

Continued subduction to the NNE resulted in the generation of extensive NW-trending dextral shear and fault systems, along with conjugate NE-sinistral zones and subsidiary N-trending dextral faults. This caused NW-directed tectonic transport in the suture zone, whereas the Kainuu-Outokumpu Back Arc was thrust towards the NNE as a major nappe structure.

It is proposed that the Svecofennian Orogeny represents three successive subduction-related events that record similar metallogenic histories, namely the Pyhäsalmi-, Skellefte-Tampere- and Orijärvi-Bergslagen Arcs. Each of these domains contains early orogenic Kuroko-type stratabound Zn-Cu-Pb deposits, at ≥1.92 Ga, 1.89 Ga and 1.89-1.86 Ga respectively, followed by syn- to late orogenic Ni-Cu deposits, porphyry-type Cu±Mo±Au±W mineralization and epigenetic Au-As occurrences. The Skellefte-Tampere Arc is also characterized by post-orogenic Li-Be-Sn-Nb-Ta pegmatites.

Key words (GeoRef Thesaurus, AGI): metamorphic rocks, stratigraphy, plate tectonics, subduction, island arcs, metallogeny, Ladoga Bothnian Bay zone, Proterozoic, Paleoproterozoic, Karelian, Svecofennian, Marine Jatulian, Pielavesi, Finland.

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INTRODUCTION

The Geological Survey of Finland (GSF) and Outokumpu Co. (now Outokumpu Finnmines Co.) have been actively exploring for base metal mineralization in the Vihanti-Pyhäsalmi-Pielavesi region, which lies within the Main Sulphide Ore Belt in central Finland (Kahma, 1973), for several decades. The Vihanti Zn-Cu-Pb deposit was discovered in 1948, followed by the Zn-Cu ore at Pyhäsalmi in 1958. Serious exploration in the Pielavesi district commenced in the early 1950's, on the basis of a mineralized sample submitted by an amateur prospector; this finally resulted in the discovery of the Säviä Cu-Zn deposit in 1966. Meanwhile, the Kangasjärvi Zn-Cu ore had been found in 1964. In addition, several dozen smaller volcanic-hosted Zn-Cu prospects, as well as a number of Ni-Cu occurrences associated with mafic intrusions have been investigated.

The effectiveness of exploration in the Pyhäsalmi-Pielavesi region has been continually impeded by an inadequate understanding of stratigraphic relationships. This results from a combination of intense deformation, high metamorphic grade and the paucity of distinctive or laterally persistent marker horizons. The area belongs to the so-called Savo Schist Belt (Fig. 1), which straddles the boundary zone between the Svecofennian and Karelian orogenic domains. The complexity of this zone has made correlation between the two domains difficult, so that an improved understanding of lithostratigraphy is of general geological as well as economic importance.

In 1978 the exploration division of the GSF began an intensified exploration effort in the Pielavesi area, with an additional objective being to clarify the relationships between various lithological units. The region also fell within the area investigated during 1983-1986 in connexion with the Integrated Multimethod Mineral Resource Prediction Project (Gaál, 1988).

This study represents the results of field work carried out by the writer in the Pielavesi area, during the 1970's and 1980's, in the course of routine exploration and in while coordinating exploration activities for the Central Finland Regional Division of the Geological Survey.

Location and physiography of the study area

The Pielavesi area is located some 100 km to the northwest of Kuopio, and is situated within the province of North Savo, near its boundary with the provinces of Central Finland and Ostrobothnia. The study area lies predominantly within the Pielavesi 1 : 100 000 map sheet (3314), but does extend northwards into the Kiuruvesi sheet (3323) and southwards into the Vesanto sheet (3313); this makes a total area of 1100 km , of which some 220 km is occupied by lakes.

The principal physiographical feature of the region is the basin occupied by the lake Pielavesi, which has a mean level of 102 m above sea level. Topography is subdued, but highly variable, as is typical of North Savo in general, with numerous hills exceeding 180 m in elevation interspersed with many small lakes. Much

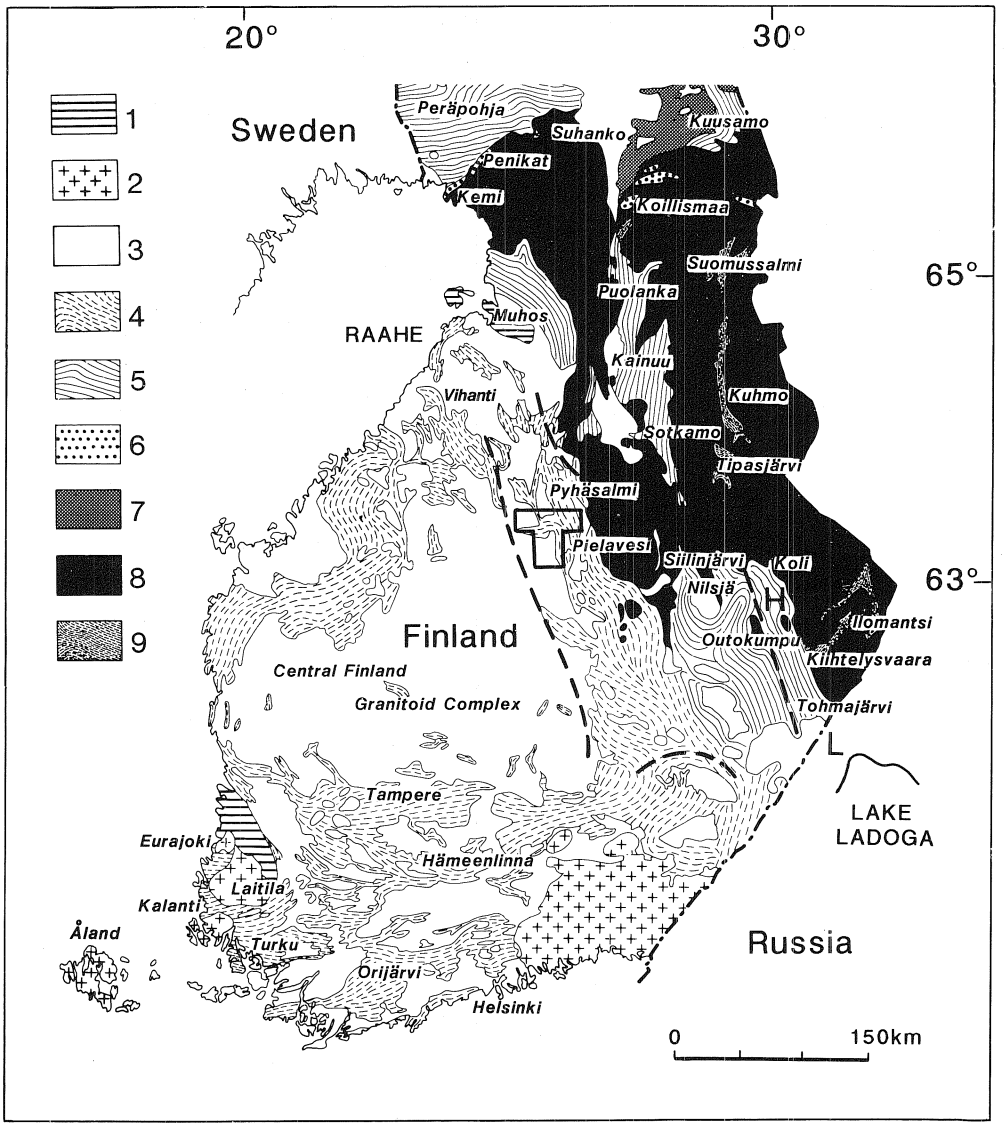


Fig. 1. Location of Pielavesi study area (solid line) and Savo Schist Belt (broken line) on simplified geological map of Finland (after Simonen, 1980; Papunen and Vorma, 1985). H-L = Höytiäinen-Ladoga basin. 1 = Jotnian sediments; 2 = Rapakivi granites; 3 = Svecokarelian granitoids; 4 = Svecofennian supracrustal rocks; 5 = Karelian supracrustal rocks; 6 = Presvecokarelian layered intrusions; 7 = Lapponian supracrustal rocks; 8 = Archean granitoids; 9 = Archean greenstone belts.

of the bedrock is covered by till, with outcrops being generally more abundant in the central and northern parts of the study area. The best exposures however, are along the rocky shorelines of Pielavesi and on the numerous small islands within the lake.

The influence of glaciation upon the landscape is evident from the prominent northwest

orientation of waterways, eskers and, in the southwestern part of the region, drumlin fields (Ekdahl, 1981). Glacial erosion has also to some extent accentuated pre-existing topographical features and fracture systems. Narrow valleys and ridges often reflect tightly folded zones, or more generally antiforms and synforms.

Previous investigations

The central part of the study area is known as the Säviä schist belt, which continues northwards as far as Kiuruvesi, while the western part of the area is contiguous with the Ruotanen schist belt, which can be traced from Pyhäsalmi southeastwards via Koivujärvi to Kangasjärvi (Fig. 2). The Säviä and Ruotanen belts converge both northwards and southwards, forming overall an oval shaped complex.

The stratigraphy of the Ruotanen belt has been inferred from the vicinity of the Pyhäsalmi mine, as well as from further south at Koivujärvi, which has been considered as a key area in this respect. Talvitie (1959) correlated the "quartzites", calc-silicate lithologies and sulphidic schists at Koivujärvi with the epicontinental sequence surrounding the Kuopio basement domes (Väyrynen, 1954; Preston, 1954). The interpretations of Nikander (1976) and Helovuori (1976) differ from each other, principally with respect to the relationship between the above-mentioned lithologies and the volcanic horizons containing the stratabound ore deposits. Nikander attempted to resolve this problem by reversing the relative sequence at Ruotanen (Table 1). This interpretation of the Koivujärvi sequence (Nikander, 1976) is supported by the structure of the Ruotanen schist belt at Kangasjärvi, where a stratabound sulphide ore body is located stratigraphically above a calc-silicate sequence that can be traced around a major synformal structure.

Uncertainty also surrounds the nature of the stratigraphic sequence of the carbonate, calc-silicate and graphitic schists in the Pielavesi and Kiuruvesi district (Table 2). The interpretation for the eastern part of the Pielavesi region and Lampaanjärvi, resulting from exploration work by Outokumpu Co. (Haga, 1980; Isohanni, 1982) is consistent (Table 2) with that for the Säviä schist belt (Makkonen, 1981; Ekdahl, 1981). Marttila (1976, 1981) considered the zone of ortho- and paragneisses extending southwards from Kiuruvesi to Pielavesi to represent a basement gneiss complex, which underwent metamorphism and anatexis during Svecofennian orogenesis, resulting in their present migmatitic appearance. According to Marttila (1981) the Vieremä-Haajainen sequence can be correlated with the Kalevian elsewhere in eastern Finland.

The Pielavesi region has also been subdivided into a number of distinct metamorphic blocks (Hölttä, 1988). Granulite facies metamorphism took place at higher pressures in the Pielavesi region and was attained some 30-40 Ma before that in the migmatite zones of southern Finland (Korsman et al., 1984; Korsman et al., 1988).

The Integrated Multimethod Exploration Project, covering the adjoining map sheets Pyhäjärvi (3321), Kiuruvesi (3323), Pielavesi (3314) and Pihtipudas (3312) was aimed at developing methods for mineral resource prediction by integrating existing geoscience data. Ore showings were demonstrated to be preferentially situated within cordierite-anthophyllite-, diopside- and graphite bearing-gneisses. The metasediments were originally deposited upon a granitic gneiss basement, with graphite-bearing gneisses lowermost, overlain by calc-silicate gneisses and amphibolites. The graphite-bearing gneisses show a distinct positive correlation between graphite and Cu, Zn, Ni, Co, U and V abundances (Tiainen, 1988).

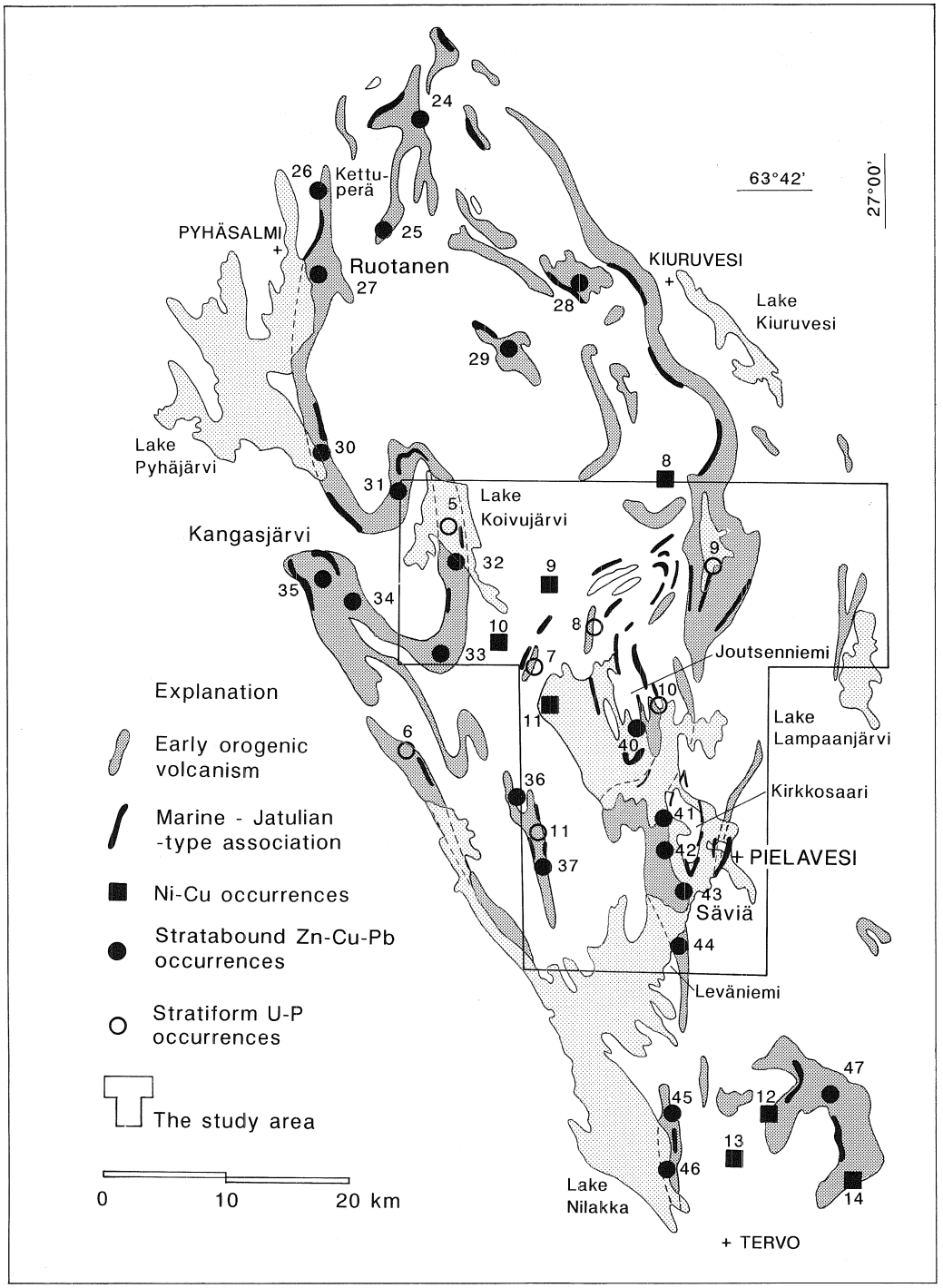


Fig. 2. Principal geological and metallogenic features of the study area. The Säviä schist belt forms the eastern part and the Ruotanen schist belt the western part of an oval-shaped complex.

Table 1. Stratigraphy of the Ruotanen schist belt.

Koivujärvi (Nikander 1976)	Ruotanen (Helovuori 1976) East
	E. Stratabound ore
Mafic agglomerate, amphibolite and associated cummingtonite-anthophyllite- and garnet gneisses	D. Mineralized zone - mica-cordierite gneisses - cordierite-anthophyllite rocks - leptite gneisses
Lapilli tuffs	C. Felsic volcanics (=leptites)
Amphibolite	B. Mafic volcanics
Carbonate and calc-silicate lithologies Leptites	A. Normal metasediments - amphibolites - dolomites - calc-silicate rocks - quartzites - amphibolites - graphitic gneisses - micaceous gneisses
Felsic tuffite	- amphibolites
Mica gneisses	
Unknown basement	West

Table 2. Stratigraphy of the Pielavesi and Kiuruvesi districts.

Pielavesi (Isohanni 1982)	Kiuruvesi (Marttila (1981)
	Vieremä-Haajainen sequence - mica schists - graywackes - conglomerate
Upper mica gneisses - sillimanite gneisses - hornblende gneisses - arkosites	Sedimentary sequence - hornblende-mica gneisses - graphitic schists - calc-silicate and carbonate rocks
Volcanites - mafic, intermediate tuffs and tuffites - calc-silicate rocks	Volcanic sequence - mafic, intermediate and felsic lavas, tuffs and tuffites - calc-silicate and graphite-bearing lithologies
Lower mica gneisses - pale, graphite- and sulphide-bearing gneisses	Basal weathering sequence - quartz- and feldspar-rich gneisses
Mobilized basement complex	Basement gneiss

Methods and objectives of the study

The purpose of this study is to describe the stratigraphic and structural development and geotectonic setting of the Pielavesi region. The study area represents a section through the Säviä schist belt and the western part of the Ruotanen schist belt, which is a crucial part of the Savo Schist Belt when comparing Karelian stratigraphic sequences with those of the Svecofennides.

The accompanying 1 : 100 000 Stratigraphic map of the Pielavesi area (appended map 1) is based on Geological Survey data dating back to the original surveys of the 1950's, as well as Outokumpu exploration reports. All thin sections (170) examined were prepared at the Geological Survey. Age determinations were made at the GSF isotope laboratory. Except for a few taken by the author the outcrop photographs have been taken by J. Väätäinen.

81 samples were collected by the author from metavolcanic formations in the study area for geochemical analysis, along with some from other lithologies as well. Samples were 1-2 kg in size and weathered rinds were sawn off prior to crushing. XRF analyses were performed at the GSF in Espoo using a Philips PW14-80 spectrometer and data were analyzed using the HST package on a Microvax 3600 computer. Diagrams were prepared with the aid of Scientific Graphics Coplot Software. Tectonomagmatic interpretations of the volcanic formations is used to support inferences concerning tectonic development.

A specific objective of the study was to examine mineralization within the context of regional geological evolution, including a synthesis of metallogenesis in the whole Raahe-Ladoga Zone and its development from a plate-tectonic standpoint. A metallogenic map at a scale of 1 : 800 000 compiled by the author is also presented, based principally on data acquired by the GSF and Outokumpu Co. (appended map 2). The tectonic interpretations are based on published information as well as field observations and low-altitude geophysical data, manipulated using DISIMP software. The main emphasis of the study is approached from an exploration viewpoint.

GEOLOGICAL SETTING

Savo Schist Belt

The Savo Schist Belt has not been precisely defined, in either a geological or geographical sense. It broadly refers to a region 50-150 km wide and 300-350 km long, straddling the boundary between the Karelides and Svecofennides.

The sediments surrounding the Kuopio and Kotalahti basement domes are correlated with the Karelian. Aumo (1983 b) subdivided the Kuopio sequence into the following units: basalmost, Sariolan conglomerates (Lippumäki FM), overlain by Lower Jatulian quartzites (Vuorimäki FM), with the Upper Jatulian being represented by Marine Jatulian sediments (Petonen FM) and volcanics (Vaivanen FM) and uppermost, Kalevian flysch sediments (Rahkamäki FM). Väyrynen (1954) likewise regarded the Savo schists to the west of Kuopio as Karelian.

The rocks to the west and southwest of Kuopio were originally turbiditic sediments with volcanic, carbonaceous and graphitic interlayers. The geology is characterized by complicated tectonics with multiple folding and faulting. The turbidites are intruded by tonalites and recrystallized into veined gneisses under upper

amphibolite to granulite facies conditions.

On the basis of lithology and stratigraphy Hautala (1968) correlated the schists surrounding the Venetpalo gneiss dome in the Kärsämäki area with the Karelian formations, even though the deformation style is more reminiscent of the Svecofennian domain. He further considered the schists at Kärsämäki represent typical Savo schists. It is difficult to define the northwestern border of the Savo schists towards the Vihanti area. In the west, the belt is intruded by the central Finland batholith complex and in southern Savo, by the Pihlajavesi and Puruvesi granites. The boundary between the Savo schists and the Tampere Schist Belt is almost impossible to define, except in a broad geographical sense. To the east, the Savo schists can be traced across the Suvasvesi Shear Zone as far as the Höytiäinen province (Fig. 1). Thus, the boundaries of the Savo Schist Belt remain open to the northwest, southwest and southeast.

In recent years, the Savo schists have been increasingly regarded as Svecofennian (Papunen and Vorma, 1985; Laajoki and Luukas, 1988; Gaál and Gorbatschev, 1987; Nurmi et al., 1991; Luukas, 1991 a; Laajoki, 1991). Luukkonen and Lukkarinen (1986) essentially followed the boundaries defined by Simonen (1980), regarding the bulk of the Savo schists as belonging to the Svecofennian Bothnian Group, whose depositional age was believed to broadly coincide with that of the Kalevian. Nevertheless, since the rocks of the Kuopio area have distinct Karelian affinities, while at the same time belonging to the Savo schists, the region is best considered as representing a transitional zone between the Svecofennides and Karelides.

Svecokarelian suture zone

Hietanen (1975) introduced a completely different approach to the use of the traditional terms Karelian, Svecofennian, Svionian and Bothnian, such that in her plate tectonic interpretation, the Svecofennian domain represented an oceanic plate being subducted beneath the Archean craton. Plate tectonic interpretations have subsequently been readily widely accepted (Piirainen, 1975; Mäkelä, 1980; Koistinen, 1981; Gaál, 1982, 1986, 1990; Park et al., 1984; Gaál and Gorbatschev, 1987; Ward, 1987; Luukas, 1991 a; Park, 1991). It is notable, that Simonen (1953) previosly concluded, that the volcanics of the Tampere district had accumulated in volcanic arc systems.

According to Hietanen (1975), the Svecokarelian fault zone separates the continental Karelian block from the oceanic Svecofennian domain, while Koistinen (1981) termed this zone a geosuture. The Savo schists occupy a central position within this suture, which according to the plate tectonic model, represents the complex deformation of marine sediments and the products of arc volcanism, along with high-grade metamorphism and widespread stratabound synvolcanic mineralization (cf. Gaál, 1986). As subduction progressed, volcanism changed from tholeiitic to calc-alkaline in character (Mäkelä, 1980). The composition of sediments and volcanics is otherwise dependent upon geotectonic setting and proximity to arcs or continental crust; there is no evidence for Archean basement beneath the Svecofennian domain (Huhma, 1986; Vaasjoki and Sakko, 1988), while the marginal zone of the craton is characterized by basement domes.

The intense negative gravity anomaly associated with the Raahe-Ladoga Zone could correspond to the position of the trench, while it is reasonable to associate the actual arc itself with the adjacent positive gravity anomaly in a manner analogous to that presented for the Aleutian arc (Le Pichon et al., 1973). The deep seismic SVEKA profile (Luosto et al., 1984)

revealed the existence of faults extending into the mantle along the border of the Archean craton, which have controlled the movement of potassium-rich material. The abundance of granitoids in this zone may also account for the negative gravity anomalies. The average crustal thickness in the Fennoscandian Shield is around 45 km, whereas within the Raahe-Ladoga Zone, including the Pielavesi area, it is 55-60 km. The regional positive gravity anomaly in the Pielavesi area also increases in intensity eastwards, consistent with the existence of a mafic body at shallow depth (Elo, 1981).

The plate tectonic model presented by Park et al. (1984) involved progressive southwards migration of arcs including the Skellefte (2000-1950 Ma) and Tampere (1950-1900 Ma) volcanic arcs. Continental basic igneous activity (2.3-2.05 Ma) related to a phase of tensional activity preceded break-up stage, which was represented by the genesis of the Outokumpu assemblage associated with a mantle diapir. Obduction of the Outokumpu back arc basin took place during the later orogenic phase. The model of Gaál (1986), involving a marginal basin and magmatic arc system on the active continental margin, during subduction towards the east was subsequently modified (Gaál, 1990), involving a reversal of subduction after collision. Thinning of continental crust led to marginal basin formation (Outokumpu rift zone). Westwards subduction of oceanic crust was associated with the formation of oceanic island arcs, intruded by tonalites at 1930-1900 Ma, along with the obduction of Outokumpu association back onto the foreland. Subsequent reversal of subduction resulted in the development of a continental margin magmatic arc between 1900-1870 Ma, with associated massive sulphide deposits.

Ward (1987) contended that evidence for an Andean-type magmatic arc on the craton was lacking, and that the structural vergence along the craton margin was consistent with subduction towards the southwest, followed by or converging with another zone of opposite polarity. Collision zones in the western Pacific, such as the Japanese archipelago and the Banda arc and Papua were regarded as possible present day analogs. Park et al. (1984) also drew comparisons with the Cordillera of western North America, emphasizing in particular the possibility of major strike-slip motion.

Vihanti ore zone

The Pielavesi district belongs in a metallogenic sense both to the southern part of the Vihanti Zinc Ore Zone as well as to the Kotalahti Nickel Belt (Gaál, 1972), which form parts of the Main Sulphide Ore Belt (Kahma, 1973). Thus, the Pielavesi area is of central significance in deciphering the stratigraphic position of the Zn-Cu-Pb ore deposits of central Finland. Therefore it is necessary to consider stratigraphic relationships in the wider context of the Vihanti-Pyhäsalmi zone. The Raahe-Ladoga Zone will be examined in its entirety later in the study, since also includes certain types of deposits which are not found in the Pielavesi region.

In the Lampinsaari area at Vihanti, migmatitic mica gneisses comprise the lowest exposed unit. The earliest stage of volcanism was felsic and took place within a marine setting, indicated by their association with dolomites and calc-silicate lithologies; together these constitute the Lampinsaari Suite (Rauhamäki et al., 1980). The felsic volcanics consist of homogeneous quartz porphyries, banded tuffs and tuffites (Table 3). Cordierite gneisses, quartzites and other quartz rocks in the area have been interpreted, on the basis of chemical and isotopic data, as being derived from hydrothermally altered felsic volcanic protoliths (Rouhunkoski, 1968). A banded structure is commonly

Table 3. Stratigraphy of the central Ostrobothnia and the Lampinsaari complex at Vihanti.

Central Ostrobothnia (Gaál et al. 1974)	Lampinsaari (Rauhamäki et al. 1978, 1980)	
The upper geosynclinal unit - Metaturbidites - Quartzite, meta-arkose - Conglomerate		
Discordance		
Lower geosynclinal unit - Mafic and felsic volcanics - Metagraywackes - Mafic volcanics - Epicontinental metasediments	Felsic and mafic dykes Mica gneiss Felsic volcanic-sedimentary formation - Quartz porphyry - Graphite tuff - Calc-silicates, dolomites - Dolomites, calc-silicates - Felsic volcanics - Felsic volcanics calc-silicates, dolomites	Lead-silver-gold Zink ore Disseminated copper ore Pyrite ore Uraninite-apatite mineralization
Great discordance		
Basal complex	Mica gneiss, with amphibolite intercalation	

seen in the main ore, which stratigraphically overlies a disseminated type of chalcopyrite-pyrrhotite ore also containing lenses richer in pyrite.

The volcanic sequence at Lampinsaari commenced under marine conditions, which were critical for the formation of a distinctive uranium and phosphorus (U-P) horizon which, according to Rauhamäki et al. (1980), can be traced over a wide area regionally, occurring consistently stratigraphically beneath sulphide mineralization of the Vihanti-type.

The Zn-Cu deposits of the Pyhäsalmi-Pielavesi region all appear to lie at approximately the same stratigraphic level (Huhtala et al., 1978). The lowest supracrustal unit is a biotite-plagioclase gneiss which contains sporadic tuffaceous intercalations. The proportion of volcanogenic material gradually increases, at first in association with exhalative deposits and calcium carbonates. These are followed by mafic, intermediate and felsic volcanics, which pass transitionally upwards into sediments with graywacke affinities. The Ruotanen ore body is zoned and retains relict primary features (Helovuori, 1979). Ore formation was related to calc-alkaline rhyolitic and rhyodacitic volcanism, with the main Zn-Cu lode, in which pyrite is the dominant sulphide, occurring within a sericitic zone hosted by garnet-cordierite-anthophyllite rocks.

The Säviä sequence, including volcanics, calc-silicate and graphite-bearing sediments forms a discontinuous horizon as far as Tervo where it turns eastwards, due to the effects of folding, and then southwards again (see appended map of the Raahe-Ladoga Ore Zone). Within the continuation of this zone, at Karttula, to the west of Kuopio Fagerström (1990) has identified subalkaline hornblende-bearing gneisses which are interpreted as having originally formed in a volcanic arc setting.

Further west, the Rautalampi area forms another distinctive unit, in which the Pukkiharju volcanic exhalative Cu-Zn deposit nevertheless resembles mineralization in the Pyhäsalmi-Pielavesi district, in terms of isotopic compositions and alteration patterns (Lahtinen, 1988), as well as in being associated with bimodal volcanism. Volcanogenic sediments, quartzofeldspathic schists, carbonates and calc-silicate rocks and graphitic schists are also

Table 4. Stratigraphy of the Virtasalmi-Juva district (modified from Hyvärinen, 1969; Pietikäinen, 1986 and Makkonen, 1988).

- Garnet-cordierite-sillimanite-mica gneisses
- Mafic and felsic volcanics with garnet-bearing skarns
- Carbonate rock, quartz-feldspar gneisses, calc-silicate rocks, cherts, iron formations, U-P horizon, graphitic schists
- Veined, locally graphite-bearing mica gneisses
Unknown basement

present, with the latter, according to Lahtinen, forming a coherent marker horizon. The stratigraphy of the Pukkiharju area is not known in full, but the carbonate- and graphite-bearing units are situated between a volcanic unit and a sequence of mica gneisses containing aluminous gneiss intercalations.

The next major occurrence of volcanic lithologies to the south is in the Virtasalmi-Juva region and again the stratigraphic sequence (Table 4) is broadly consistent with that established in the Vihanti-Pyhäsalmi zone.

Garnetiferous skarns and mineralization resembling that at Hällinmäki are present in the lower part of the diopside amphibolite unit, which is interpreted as representing coeval carbonate precipitation and volcanism (Hyvärinen, 1969). The Hällinmäki ore was interpreted as a pneumatolitic contact skarn associated with a quartz diorite intrusion, while Suvanto (1983) surmised that the fluids responsible for skarn mineralization were related to the volcanism.

The protoliths to the skarn sequence at Hällinmäki commenced with the deposition of clay or volcanic material, accompanied by sporadic limonite, calcite, siderite and possibly also silica. Hyvärinen (1969) considered that the principal minerals in the garnetiferous horizons, such as andradite and hedenbergite could have been derived from such precursors. According to Lawrie (1992) the amphibolites altered to massive calc-silicate lithologies which is reflected in a gradual increase of calcic garnet (andradite) at the expence of plagioclase, together with a rapid change from diopside to hedenbergite pyroxene. Scapolite often occurs in intensely altered units instead of plagioclase. Lawrie interprets magnetite-rich bands as admixtures of tuffaceous material and chemical precipitates formed at the site of exhalative mineralization. The amphibolites of the Hällinmäki area are classified as tholeiitic or calc-alkaline Al-rich basalts corresponding to island arc basalts (Suvanto, 1983) or within plate basalts or to E-type MORBs (Lawrie, 1992).

According to Koistinen (1978), stratigraphic position is paramount with regard to mineralization in the Virtasalmi district, although some structural control is also evident from the en echelón disposition of the ore bodies. There is no evidence in the open pit at Hällinmäki for any mineralizing post-depositional intrusions, rather the distinct banding of the ore indicates an early origin (Koistinen, 1978; Suvanto, 1983). Immediately to the north of Hällinmäki, skarns occur in association with volcanic breccias and agglomerates. The carbonate unit is a potential marker horizon throughout the area.

A total of at least ten Cu-Zn prospects have been identified throughout the zone of volcanics passing through Hällinmäki, over a distance of 50-60 km. In the vicinity of Hällinmäki they tend to be more cupriferous and associated with mafic volcanics while in the southern part Zn-Cu-Pb occurrences are hosted by felsic volcanics with hydrothermal alteration zones (Makkonen, 1988, 1989, 1991). In the Virtasalmi district felsic volcanics are subordinate to mafic lithologies and hydrothermal activity has been less significant than in the Pyhäsalmi-Pielavesi region.

Ore deposits in the Hällinmäki belt are stratiform and occur at a comparable stratigraphic level to those at Säviä, Lampinsaari and Pyhäsalmi. Ore formation at Hällinmäki was evidently closely connected with a carbonate-

graphite schist-iron formation association in which sulphides probably precipitated as a result of hydrothermal fluid activity rather, than during alteration of amphibolites to massive calc-silicate lithologies (cf. Lawrie, 1992). The Vihanti ore zone as a whole is characterized by arc dominated volcanism and may therefore be referred to as the Vihanti-Pyhäsalmi-Virtasalmi ore zone or in short, the Pyhäsalmi Island Arc (PIA). The term skarn has been applied liberally throughout the whole PIA zone to virtually all regionally metamorphosed calc-silicate horizons irrespective of their genesis. However, the main stratigraphic marker horizon throughout the PIA zone is in fact this association consisting variously of carbonates,calc-silicates, the U-P-horizon, chert, iron formations, graphite schists and quartz-feldspar gneisses. It occupies a distinct stratigraphic position immediately preceding volcanic arc magmatism and stratabound ore formation, although at Lampinsaari it was also partly coeval with mineralization.

DEPOSITIONAL BASEMENT IN THE PIELAVESI AREA

Archean basement has not been found to the southwest of the Raahe-Ladoga Zone, while that presently exposed along the craton margin was extensively disrupted during the Svecofennian orogeny; which term is used here in preference to Svecokarelian Orogeny (Simonen, 1980), as advocated by Gaál and Gorbatschev (1987).

Wilkman (1938) described much of the Pielavesi region as gneissose granites, while Helovuori (1976, 1979) interpreted the Kettuperä gneiss (1932 Ma) to the east of the Ruotanen schist belt as local depositional basement and exented this interpretation to include the basement to other Svecofennian and Bothnian formations. Subsequently it has been shown that the Kettuperä gneiss contains both orthogneisses and paragneisses (Marttila 1993). Marttila (1976) considered that the gneissose granites containing supracrustal enclaves in the eastern part of the Kiuruvesi district also formed part of the regional basement complex, arguing that they were subsequently remobilized during the Svecofennian Orogeny. Characteristically this unit contains remnants of supracrustal schists and are extensively recrystallized granodiorites or quartz diorites. They are however distinct from the Kettuperä gneiss, which more closely resembles the Kirkkosaari and Leväniemi gneiss domes in the Pielavesi district.

The oldest known Svecofennian crust is found within the Raahe-Ladoga Zone and has been dated at 1.93-1.92 Ga (Helovuori, 1979; Korsman et al., 1984; Patchett and Kouvo, 1986; Huhma, 1986; Vaasjoki and Sakko, 1988). Equivalents of these can also be recognized in the central part of Ostrobothnia, at Veteli and at Evijärvi (A. Lonka and E. Marttila, pers. comm., 1991; Vaarma, 1990). These basement gneisses (Fig. 3) are typically dome-shaped, as at Veteli, Nivala, Kärsämäki, Pyhäjärvi and Pielavesi. In the Savo Schist Belt, granitoids belonging to this age group form the lowermost structural unit and are regarded as the depositional basement (Helovuori, 1979; Ekdahl, 1981; Papunen, 1986; Vaasjoki and Sakko, 1988). Welin (1987) interprets the age of these rocks to indicate that tectonomagmatic processes commenced somewhat earlier along the craton margin than elsewhere, while considering that Svecofennian sedimentation began somewhere between 2.00 and 1.93 Ga. Because of the apparent scarcity of mafic and ultramafic rocks and absence of ophiolites,

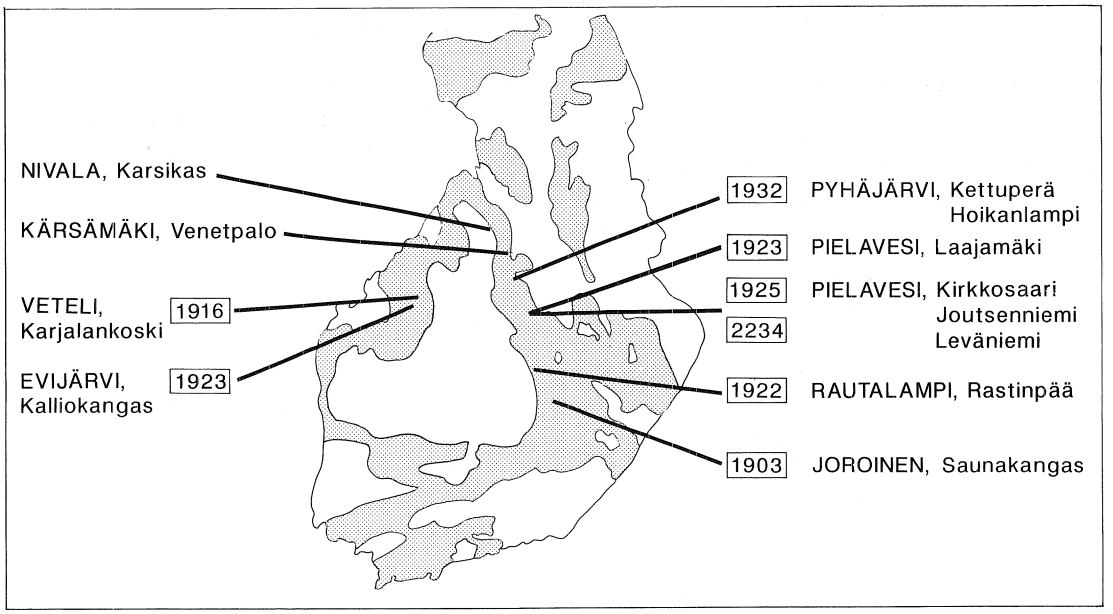


Fig. 3. Tonalitic gneisses of age 1.93-1.92 Ga considered to represent depositional basement to the rocks of the study area and related U-Pb zircon data (references in text). Svecokarelian schists shown in gray.

Welin regarded the tectonic setting as continental rather than oceanic and furthermore, correlated the Svecofennian graywacke sequences, which contains Jatulian and Archean detritus, with the Kalevian deposits of the craton margin. The isotopic composition of the Svecofennian sediments is in consistent with a predominantly Archean provenance (Huhma, 1986). Instead according to Huhma, rifting of the craton, recorded by Jatulian magmatism, led to the development of the Svecofennides and rapid crustal growth around 1.9 Ga. Svecofennian sediments represent a mixture of juvenile, collision-related Proterozoic material and older, craton-derived detritus, deposited in a volcanic arc setting. Lu-Hf and Sm-Nd studies (Patchett et al., 1981; Huhma, 1986) indicate that the Svecofennides were formed during the early Proterozoic, and that they are not underlain by Archean basement. However, studies in the Tampere area (Kähkönen, 1989) indicated significant crustal evolution preceding the volcanic activity (1.9 Ga).

Gaál and Gorbatschev (1987) proposed that subduction commenced around 2.00-1.95 Ga, by which stage some kind of substrate for sedimentation existed. Likewise, Vaasjoki and Sakko (1988) argued that Svecofennian crust formed during plate convergence involving Proterozoic oceanic crust and Archean continental crust, between 1.93-1.85 Ga.

The 1.93-1.92 Ga tonalitic gneisses within the marginal zone represented by the Savo Schist Belt thus occupy a central position in the debate concerning the nature of the regional basement. Some of them include paragneiss enclaves and they are clearly of significance in the context of Svecofennian stratigraphy.

The Kirkkosaari, Joutsenniemi and Leväniemi domes

Definition and distribution

The gneisses of above mentioned age group are exposed in the Pielavesi area as a series of domal culminations at Joutsenniemi, Kirkkosaari and Leväniemi (appended map 1). The domes are mantled by the supracrustals of the Pielavesi sequence. Surrounding graphite- and sulphide-bearing schists are multiphase folded and form symmetrical structures reflected by geophysical anomalies (Fig. 4).

The Joutsenniemi dome in the northern part of the study area is about 15 km in length and 1-2 km wide. In the central part a gentle depression is present, with the dome lithologies bering partly obscured by schists of the cover sequence. Synformal depressions are also developed at the margins of the dome, resulting in narrow zones of cover rocks alternating with dome gneisses. Horizontal to gently plunging folds and lineations are well exposed in bluff sections at Salosaari, along the western margin of the dome (Fig. 5). The central part of the Joutsenniemi dome was intruded at a later stage by an ultramafic intrusion and deformed along associated fracture systems. The intrusion is also clearly discernible as a positive gravity anomaly and, on the basis of gravitational modelling, it appears that the density contrast between the peridotite and the dome lithologies persists to a depth of at least 800 m (Elo, 1981).

The Kirkkosaari dome, in the central part of the study area, is about 6 km long and 2 km wide. At Leväniemi, however, the supracrustal sequence is exposed only in the eastern limb of a narrow syncline, although the dome is likely to be the southerly continuation of the Kirkkosaari structure. The Joutsenniemi and Kirkkosaari domes with manteling supracrustal cover sequences are key sections in establishing the regional stratigraphy of the Pielavesi district.

Structurally the gneiss domes represent the deposit exposed sections in the study area, and no evidence has been found to suggest that have intruded discordantly into the overlying supracrustal rocks. The domes were intruded by mafic dykes which are interpreted as being related to the lowest volcanic phase.

Lithologies

Ortho- and paragneiss

Both the Joutsenniemi and Leväniemi domes consist of relatively homogeneous orthogneisses (Fig. 6 a) with modal compositions corresponding principally to tonalite-quartz diorite (App. 1). They tend to be gneissose and intensely oriented, with a dominant N-trending axial plane schistosity.

The orthogneisses are medium-grained and homogeneous and are very pale on weathered surfaces. Plagioclase (An_{20-30}) comprises up to 50-70% of the rock and quartz from 20-40%, with K-feldspar being less than 5%, although somewhat more abundant at Leväniemi than at Joutsenniemi. Plagioclase has been to some extent altered to sericite, carbonate and saussurite. Biotite and hornblende generally comprise less than 10% of the rocks and accessory minerals include opaques, apatite, zircon and sphene.

In contrast, the Kirkkosaari dome consists principally of compositionally heterogeneous paragneisses, although at Sikoniemi, on the western side of the dome, orthogneisses resembling those at Joutsenniemi are present. Ortho- and paragneisses appear to have transitional contacts, making it difficult to ascertain whether or not the orthogneisses represent deformed granitoids or reworked paragneisses,as was argued for similar lithologies at Kiuruvesi by Marttila (1976). The Kirkkosaari paragneisses resemble nebulitic migmatites in which the neosomes may be the result of anatexis (Fig. 6 b). These were subsequently intruded by a younger phase of reddish granitic veins. Potassium feldspar is noticeably more abundant at

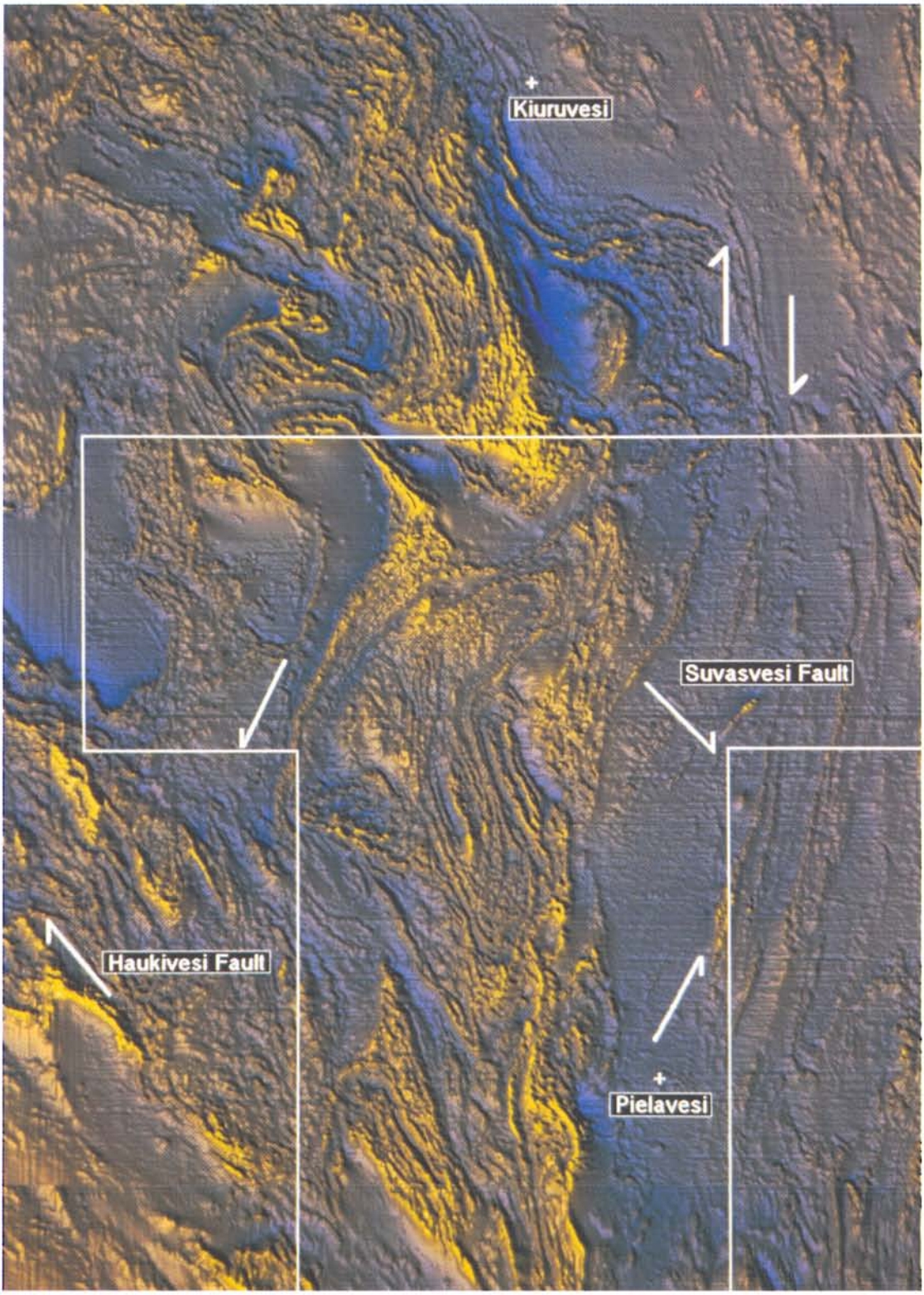

Kiuruvesi
Suvasvesi Fault
Haukivesi Fault
Pielavesi

Fig. 5. Joutsenniemi axial culmination at Salonsaari. Hammer is parallel to almost horizontal fold hinge.

Kirkkosaari than at Joutsenniemi. Deformation has clearly been intense and took place within the plastic regime. The protoliths to the paragneisses were presumably hornblende- and mica gneisses and amphibolites. The paragneisses consist of about 50-60% plagioclase (An_{30-40}), 15-20% quartz and 15-30% biotite and hornblende, with accessory K-feldspar, apatite, sphene, zircon and opaques. Garnet is also typically present and modally , the gneisses correspond to quartz-rich granodiorites. The Kirkkosaari paragneisses are similar to those at Haapamäki (Hoikanlampi) in the Pyhäjärvi area and at Venetpalo in the Kärsämäki district (Marttila, 1993). The Pielavesi gneiss domes are nevertheless somewhat deficient in K_2O compared with the gneisses at Kettuperä and Hoikanlampi (App. 1). The Joutsenniemi, Kirkkosaari and Leväniemi orthogneisses are richer in plagioclase and quartz than the Archean Pieni Neulamäki gneiss at Kuopio, and conversely, contain less K-feldspar (Aumo, 1983 b). The Pielavesi gneisses also have higher SiO_2 and Fe_2O_3 abundances than the Pieni Neulamäki gneisses, while Al_2O_3, MgO, Na_2O and K_2O are correspondingly lower. The Kirkkosaari and Joutsenniemi gneisses closely resemble those of the Rastinpää gneiss at Rautalampi, both petrographically and geochemically.

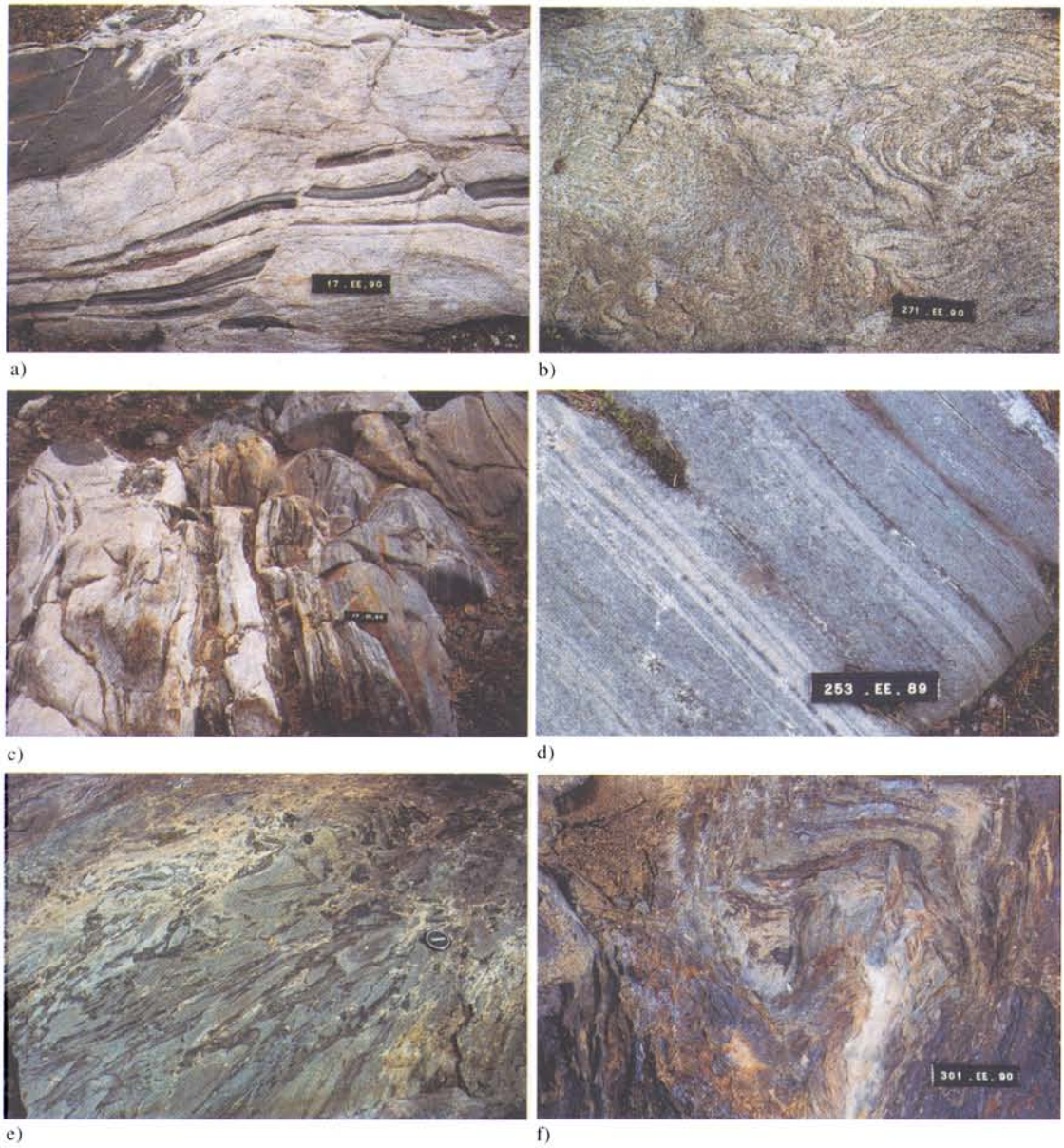
Mafic inclusions

The Pielavesi gneiss domes also contain mafic enclaves whose compositions match those of the lowermost metavolcanic unit (Salo Metavolcanite) of the overlying cover sequence. Evidently they represent the intrusive equivalents of the volcanics which have been disrupted and brecciated during subsequent deformation, with sporadically preserved trains of boudinaged fragments suggesting an originally dyke or sheet-like morphology (Fig. 6 a). The dykes are narrow and no chilled margin effects have been found. Fragmented dykes are concordant with the foliation in the host gneisses, and they have been folded together with them. The mafic enclaves are fine-grained and massive, consisting of 50-70% hornblende and 10-40% plagioclase (An_{35-60}); occasionally relict augite has been found within hornblende. Scapolite has locally formed in places at the expense of plagioclase and accessory minerals include sphene, biotite, epidote (which also occurs as distinct bands), opaques and quartz.

Contacts

The contact between the gneiss domes and the overlying Salo Metavolcanite is exposed at Salonsaari, at the western margin of the Joutsenniemi dome and also along the western margin of the Kirkkosaari dome. In both areas cover and basement appear to be concordant (Fig. 6 c), although in places the contact is

Fig. 4. The Pielavesi study area (outlined) is located between the dextral Suvasvesi and Haukivesi shear zones. Domal structures are visible on the aeromagnetic map, which also shows the general ductile style of deformation. Sinistral NE-trending antithetic conjugate structures record somewhat more brittle deformation. High-resolution pixel map is based on low-altitude aeromagnetic data with oblique illumination from the southwest.



a) b) c) d) e) f)

Fig. 6. The Pielavesi supracrustal sequence is underlain by (a) strongly foliated orthogneiss with basic inclusions as at Joutsenniemi, or (b) nebulitic paragneiss as at Kirkkosaari. Contact (c) is sharp, concordant and strongly deformed. The lowermost lithodeme of the Salonsaari Suite is the (d) Salo Metavolcanite, which consists principally of lava with a metamorphic banding structure. Metalava at Vuonamonsaari (e) contains strongly deformed pillow-like features. The Salo Metavolcanite is followed by (f) graphitic schist as seen at Kalmalahti near Kirkkosaari. Outcrop identification bar is 20 cm in length.

gradational over a distance of several meters rather than abrupt, with alternating layers of gneiss and metavolcanite. This is caused either by isoclinal folding, or by the presence of subvolcanic dykes. In proximity to the contact the basement gneisses are to some extent remobi-

lized, with apophyses intruding fractures in the metavolcanites. Adjacent to the contact the metavolcanites contain more hornblende. The contact zone is typically 2-3 m thick and is intensely deformed and brecciated.

PIELAVESI SUPRACRUSTAL SEQUENCE

Stratigraphic subdivisions and nomenclature

Väyrynen (1933, 1954) subdivided the Karelian formations of eastern Finland into the Sariola, Jatuli and Kaleva groups, with the lowermost Sariola-type conglomerates being overlain by Kainuu-type Jatulian quartzites and so-called Marine Jatulian carbonaceous phyllites and dolomites. The Jatulian sequence is in turn overlain disconformably by Kalevian conglomerates, phyllites and mica schists.

Subsequent investigations (Piirainen, 1969, 1975; Nykänen, 1971; Silvennoinen, 1972; Laajoki, 1973; Pekkarinen, 1979) have demonstrated that a number of episodes of mafic magmatism took place during Jatulian sedimentation; some of the units previously interpreted as metadiabases have been reinterpreted as predominantly volcanic, including lavas, tuffs and tuffites. The Jatulian was also preceded by a major phase of mafic magmatism (Alapieti, 1982; Silvennoinen, 1991).

Meriläinen (1980) subdivided both the Jatuli and Sariola groups into upper, middle and lower units (Fig. 7). Luukkonen and Lukkarinen (1986) followed Meriläinen (1980) in including in central Finland the former Marine Jatulian sequence, which consists of dolomite, black schist, iron formations, mafic volcanics and hypabyssal intrusions, within the Upper Jatulian. This includes in many places also a distinctive uranium- and phosphorus-bearing horizon. The Kaleva has been subdivided into lower and upper subgroups such that the former represent preorogenic sedimentation and volcanism and the latter, syn- to late orogenic sedimentation, volcanism and intrusion.

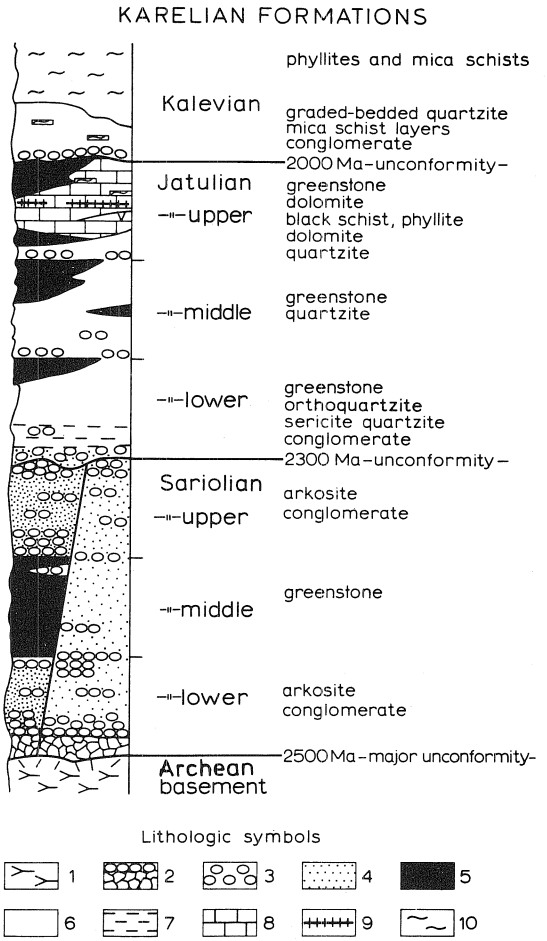


Fig. 7. Simplified schematic stratigraphic section of Karelian formations after Meriläinen (1980). 1 = Archean basement; 2 = Basal conglomerate; 3 = Conglomerate; 4 = Arkosite; 5 = Greenstone; 6 = Orthoquartzite; 7 = Sericite quartzite; 8 = Dolomite; 9 = Black schist; 10 = Phyllite and mica schist.

Generally, the Kaleva is characterized by phyllites, mica schists and mica gneisses, with graphitic deposits, arkoses and quartzites present towards the base of the sequence. Luukkonen and Lukkarinen (1986) did not include any magmatic units within the Kaleva in Karelia, except for those at Outokumpu and Jormua and they placed the boundary between the Svecofennian and Karelian supergroups in central Finland broadly along the westernmost known outcrops of Archean rocks. Thus they regard the pelitic and psammitic metasediments, gneisses, mafic, intermediate and felsic volcanics, dolomites, calc-silicate lithologies, quartztites, arkosites and black schists of the Savo Schist Belt to the west of Kuopio as belonging principally to the lower part of the Bothnian (Kalevian) Group. Otherwise, traditional classification of the Svecofennian according to Simonen (1980) has been as follows:

Upper Svecofennian	argillaceous sediments
Middle Svecofennian	mafic volcanics (lavas and pyroclastics) with intercalated sediments, including conglomerates
Lower Svecofennian	immature sediments (arkoses and graywackes) with thin carbonate intercalations and some mature arenites

The Jatulian and Kalevian boundary has been recently re-evaluated in the Kainuu Schist Belt, where iron formations are now regarded as Lower Kalevian rather than Marine Jatulian (Gehör and Havola, 1986; Kontinen, 1986, 1987; Laajoki, 1988 a,b) and evidently occur as separate horizons intercalated with turbiditic graywackes and phyllites rather than form a distinct association with dolomites. The Lower Kalevian in this area is separated from the Jatulian by a disconformity overlain by a breccia-like conglomerate. In North Karelia too, it has been argued that at least part of the Upper Jatulian and Lower Kalevian sequences are coeval and represent lateral facies variations, such that a distinct transgressive disconformity at the boundary between the two groups need not always be present (Ward, 1987, 1988). Laajoki (1988 a,b) has interpreted the Karelian formations in terms of various tectofacies, where the term tectofacies refers to the whole range of sediments and extrusive or intrusive units formed during a particular tectonic phase.

The precise Karelian stratigraphic sequence has not yet been finalized, nor is it clear that a single blanket stratigraphy is applicable to the region, or whether instead there were several independent depositional basins. Nevertheless it is worthwhile to retain the usage of the terms Jatulian, Marine Jatulian and Kalevian in a chronostratigraphic sense. There is so far no better alternative than Marine Jatulian to describing the transition between the Jatulian and Kalevian sequences.

Lithostratigraphic interpretations of the Pielavesi area are made more difficult due to the intensity of deformation and metamorphism. A single unit may vary in appearance according to the degree of deformation and changing mineral compositions over relatively short distances and migmatization has moreover obscured or obliterated primary lithological boundaries.

The lithodemic approach to lithological classification, as recommended by the North American Commission on Stratigraphic Nomenclature (1983) and as applied to the metamorphosed Salahmi-Pyhäntä area (Laajoki and Luukas, 1988), has also been adopted for the present study. A number of lithodemic units have been defined and ranked into suites, with emphasis on recognizing distinctive volcanic and sedimentary associations that relate to specific processes or stages of tectonic development. The relationship between the four major lithodemic units, or suites, in the Pielavesi area

has been determined just as for conventional lithostratigraphic units, with the oldest being the Salonsaari Suite, overlain successively by the Savijärvi, Säviä and Koivujoki Suites. The stratigraphic sequence is applied primarily to the Säviä schist belt, although correlation with the adjoining Ruotanen schist belt appears feasible.

Individual lithodemes have not always been named because their recognition and precise definition is often difficult. However, certain distinctive units, such as the Salo Metavolcanite, Kalliokylä Conglomerate and Kotajärvi Metalava have been separately named.

Salonsaari Suite

Definition, subdivision and distribution

The Salonsaari Suite is the lowest supracrustal unit recognized in the Pielavesi area and can be traced northwards towards Kiuruvesi. It has been named after the island Salonsaari, at the western margin of the Joutsenniemi dome.

The Salonsaari Suite is both volcanogenic and sedimentary in nature, with a mafic volcanic unit at the base, known as the Salo Metavolcanite. This is overlain by a graphitic schist which passes transitionally into migmatitic mica gneisses. Thin graphite-bearing and tuffaceous intercalations suggest relatively deep water conditions with sporadic volcanic activity.

The most complete section of the Salonsaari Suite is exposed on the western margin of the Joutsenniemi dome at Salonsaari itself, as well as on adjacent islands and the eastern margin of the Kirkkosaari dome. In the northern part of the Säviä schist belt the mica gneisses are intensely deformed, with a banded, tightly folded and migmatitic appearance, and in some places is effectively a granodioritic gneiss with only local primary lithological features remaining. In practice it has proven impossible to identify and represent at map scale the precise boundaries between mica gneisses and granodiorites and quartz diorites of intrusive origin, since in many places the mica gneisses effectively become schollen migmatites. The upper part of the Salonsaari Suite is characterized by a greater abundance of graphite- and carbonate-bearing horizons. The lowermost migmatized mica gneisses in the Ruotanen schist belt are also analogous with those of the Salonsaari Suite.

Lithologies

Salo Metavolcanite

The Salo Metavolcanite weathers to a dark green color and consists of dense, massive lava in its lower part, while the upper part is more clearly stratified, suggesting a pyroclastic or epiclastic origin (Fig. 6 d). Alternatively, this banding could represent metamorphic differentiation since it is locally discernible in lava-like units as well. Some units 10-30 cm thick containing relict uralite phenocrysts are also found and at Vuonamonsaari, highly deformed pillow lavas are possibly present, forming a unit 3-4 m thick (Fig. 6 e).

The fine to medium grained lava units consist principally of plagioclase (An_{40-70}), hornblende and pyroxene. Hornblende usually forms aggregates and mostly represents the breakdown and recrystallization of pyroxene. Accessory minerals include opaques, biotite, apatite, garnet, sphene and quartz, in addition to sericite, carbonate, saussurite and epidote, which occur as alteration products of plagioclase. The banding in the upper tuffaceous units results from variations in abundances of plagioclase and hornblende, or plagioclase and clinopyroxene.

Graphite schists

A persistent, sharply bounded graphite schist horizon directly overlies the Salo Metavolcanite and is readily discernible from geophysical data. The graphite schists are intensely deformed and tightly folded (Fig. 6 f) and are traversed by shear zones containing pegmatite veins. Mineralogically the schists consist of quartz, plagioclase, graphite and K-feldspar, with additional pyrite and pyrrhotite; anomalous Cu or Zn concentrations have not been recorded.

Migmatitic mica gneisses

The variably migmatized mica gneisses of the Salonsaari Suite are the dominant supracrustal lithology throughout the whole Pielavesi and Kiuruvesi region (Fig. 8 a-f). They have variously been referred to as quartz-feldspar gneisses, biotite-plagioclase gneisses, migmatitic mica and hornblende gneisses, sillimanite-mica gneisses, metasediments, and lower mica gneisses (Marttila, 1976, 1981; Nikander, 1976; Huhtala et al., 1978; Huhtala, 1979; Haga, 1980; Makkonen, 1981; Isohanni, 1982). On weathered surfaces the rocks are light gray and they are strongly foliated and banded. They typically consist of light-colored units from 5-10 cm in thickness alternating with thin biotite-rich lithons. Granitic and pegmatitic veins are generally concordant with respect to lithological layering and have clearly been deformed along with their host rocks. There exist also ptygmatic veins forming usually flowing structures in tectonized zones. Hornblende is not abundant but plagioclase (An_{30-40}) and K-feldspar are common, as is quartz. To the north of Jylhä and at Piekkälä, garnet and cordierite porphyroblasts are common, while to the northeast of Sulkavanjärvi, sillimanite is present. Garnet is ubiquitous, as in almost all supracrustal units in the Pielavesi district. The protoliths to the mica gneisses are regarded as sediments derived from weathering of a mica-poor granitic gneiss basement (Marttila, 1981; Isohanni, 1982). Where primary lithological layering is evident, psammitic and pelitic units alternate with thin tuffaceous layers (1-10 cm) of mafic to intermediate composition (Fig. 8 b).

No precise boundary can be drawn between the mica gneisses and the highly deformed quartz diorites that intrude them, while in the eastern parts of the study area, it is difficult to establish whether they actually represent a basement complex (Marttila, 1976; Isohanni, 1982) rather than migmatized supracrustal gneisses. However, the relationship between intrusive quartz diorites, orthogneisses, paragneisses and the lower mica gneisses of the Salonsaari Suite can be determined more convincingly in sections exposed adjacent to the basement domes at Salonsaari itself and at Kirkkosaari.

Most of the quartz diorites and granodiorites of the Pielavesi area appear to have formed by anatexis of the stratigraphically lowermost mica gneisses; typically they form planar bodies concordant with lithological layering in the supracrustal gneisses and do not necessarily represent a distinct magmatic event. The transition from mica gneiss through to intrusive quartz diorite is well exposed throughout the whole of the Salonsaari Suite. The transition into quartz granodiorites via schollen and nebulitic migmatites is well displayed in extensive exposures along lake shores (Fig. 8 a-f).

Schollen migmatites

Mica gneisses are also transitional into schollen migmatites, particularly in the vicinity of the Kirkkosaari and Joutsenniemi domes; mafic to intermediate and calc-silicate enclaves, often roundish and highly contorted, are distributed throughout a quartz dioritic or granodioritic groundmass (Fig. 8 c-f). These granitoids are usually hypersthene-bearing and are distinctly foliated. The schollen migmatites formed in response to intense deformation along D_3 stage NW-trending dextral and NE-

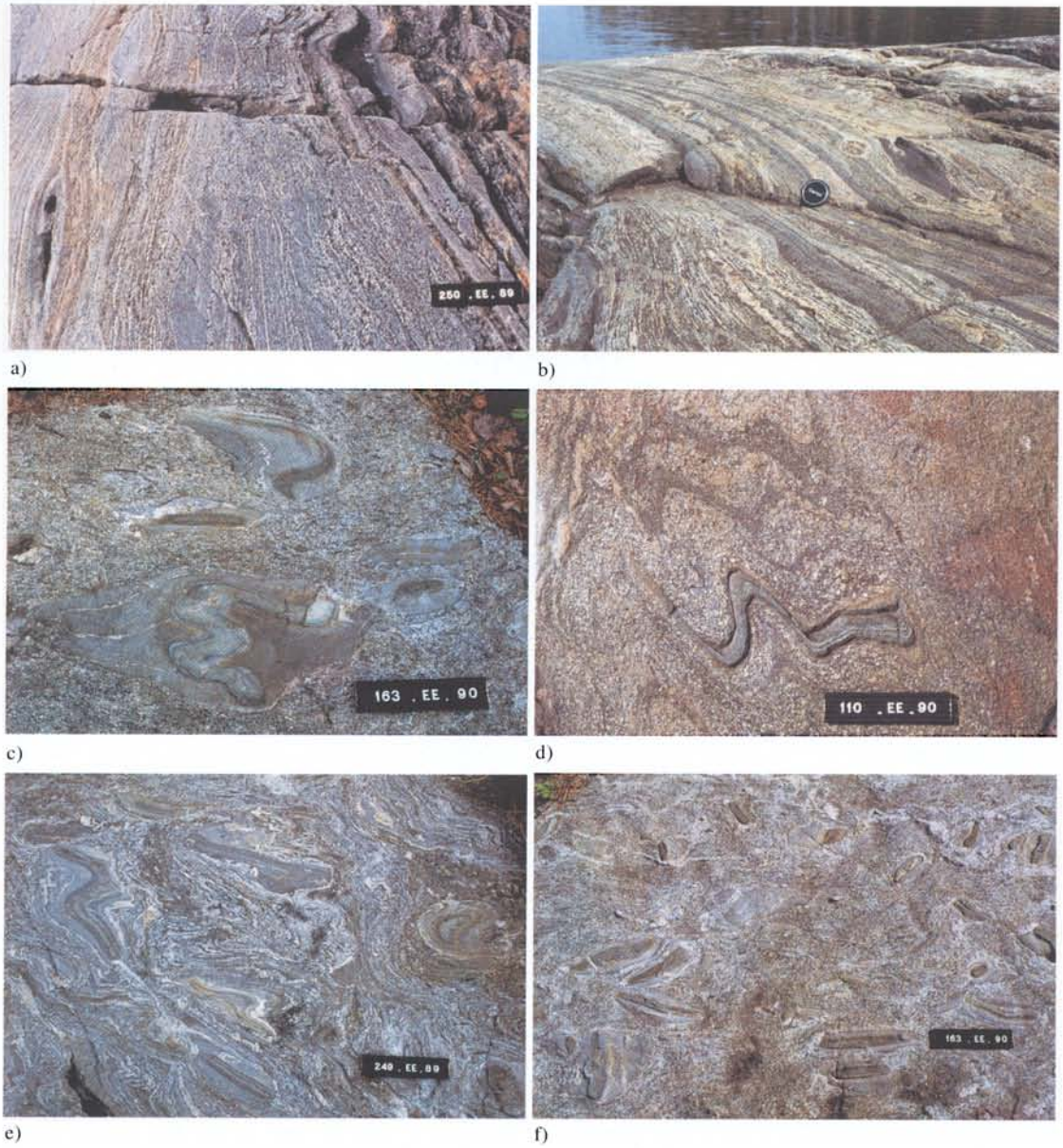


Fig. 8. Migmatitic mica gneisses of the Salonsaari Suite contain (a) narrow graphitic and calc-silicate layers and (b) volcanic intercalations. Within shear zones the mica gneisses are transformed into schollen migmatites (c-f). Photographs a and b from Salonsaari; c, d and e from Ruokokangas; f from Lehtisaari.

trending sinistral shear systems, so that the groundmass too is gneissose. Some enclaves retain evidence of an earlier, F_2 stage of folding (Fig. 8 c), although some mobilization of granitic magma took place during D_2 deformation as well (Fig. 8 d). The schollen migmatites of the Pielavesi area closely resemble, in both the general appearance and structural style, the

trondhjemite-charnockite schollen migmatites of the Haukivesi district (Gaál and Rauhamäki, 1971).

Contacts and thicknesses

The lowest unit of the Salonsaari Suite, namely the Salo Metavolcanite appears to have been essentially a sequence of lava flows that is apparently concordant with the underlying gneiss domes, although the contact is clearly intensely deformed (Fig. 6 c). Dyke intrusion presumably took place during the same phase of magmatism, now represented by boudinaged fragments or isolated mafic enclaves within the gneisses. The Salo Metavolcanite, with a total thickness varying between 10-50 m is overlain by the graphitic schist unit some 5-20 m thick. The contact between these two units is exposed at Kalmalahti, on the eastern shore of Kirkkosaari, and is very sharp, whereas the contact between the graphite schists and the overlying mica gneisses is more gradual. The presence of tuffaceous intercalations throughout the mica gneisses indicates intermittent but persistent volcanic activity during sedimentation. Migmatitic mica gneisses are transitional into granodiorites and quartz diorites, while the schollen migmatites are characteristic of synorogenic deformation. Interpretation of the cross section indicates that the total thickness of the Salonsaari Suite may attain 1000 m.

Savijärvi Suite

Definition, subdivision and distribution

The type area for the Savijärvi Suite is situated on the northwestern margin of the Joutsenniemi dome. Wilkman (1938) first drew attention to the pyrrhotite-bearing quartzites and overlying calc-silicate and dolomite units in this area.

The Savijärvi Suite is, like all units in the Pielavesi area, inclined towards the northwest, dipping at 45-60 degrees to the southeast, and is upward facing. Because of the rapid variations in lithologies, it has not been possible to establish a precise stratigraphical sequence within the Savijärvi Suite. The Suite overlies graphite-bearing migmatitic mica gneisses, commencing typically with quartz-feldspar gneisses and quartzite-like metacherts. Calc-silicate and graphite-bearing units alternate with each other, while the abundance of carbonate-rich horizons appears to increase upwards, reaching their maximum thickness in the Savijärvi axial depression. Carbonate layers are also present at Koivujärvi, Sulkavanjärvi and Uiveronlahti, associated with calc-silicate lithologies and felsic volcanics. The Savijärvi Suite also contains a uranium-phosphorus horizon. At Kalliokylä, some 6 km southeast of Savijärvi, a conglomerate unit is exposed, belonging to the upper part of the Suite.

The Savijärvi Suite is found throughout the whole study area, although it is somewhat thinner in the south and represented only by narrow graphitic and calc-silicate horizons. The Savijärvi Suite forms a valuable marker horizon since the presence of iron sulphides and graphite enable it to be readily traced on geophysical maps. The Savijärvi Suite, including the U-P horizon, is also present in the Koivujärvi area, where the Ruotanen schist belt is exposed in a synclinorium overturned towards the northwest. Somewhat further north, in the Maaseläinlahti-Ollinniemi area, a felsic volcanic - dolomite - calc-silicate rock - black schists association, reminiscent of that at the Vihanti deposit, is present, containing in addition a silicate facies iron formation (Mäki, 1987).

From Koivujärvi the association can be traced southwards and then northwestwards via

Lahnasjärvi towards Kangasjärvi, form where it continues as a discontinuous unit southeastwards to Teerimäki and Haasiakangas. The leptites and quartzites at Koivujärvi correspond to the felsic volcanics and quartz-feldspar gneisses of the Savijärvi Suite and pass transitionally via banded tremolitic leptites to calc-silicate rocks. The carbonate and calc-silicate horizons at Koivujärvi occur at relatively low stratigraphical levels within the volcanogenic sequence representing, according to Nikander (1976), the initial stages of volcanism. At Kangasjärvi, the calc-silicate rocks occur stratigraphically beneath the stratabound ore.

Lithologies

Felsic metavolcanites

The Savijärvi Suite contains felsic volcanic lithologies that have variously been described as quartz-feldspar gneisses, quartzites, leptites or felsic tuff (Wilkman, 1938; Talvitie, 1959; Nikander, 1976; Ekdahl, 1976; Haga, 1980; Äikäs, 1988 f). Narrow mafic units may also be present. Boudinaged calc-silicate layers are common. The felsic volcanics weather to a pale or yellowish color whereas the calc-silicate lithologies tend to have a greenish blue. The metavolcanics are fine grained and generally banded, as a result of both their initial planar lamination as well as the presence of calc-silicate bands (Fig. 9 a). Principal minerals are quartz, plagioclase (An_{25-50}), K-feldspar and commonly also biotite and sericite; accessory minerals include apatite, epidote, sphene, zircon and opaques. The felsic volcanics are gradational into calc-silicate rocks, reflected by a progressive increase in the proportion of diopside-tremolite-carbonate horizons. Minor amounts of iron sulphides and graphite are also typically present, particularly where rocks are sericitized and contain cordierite. At Koivujärvi, where the volcanics are in a relatively good state of preservation, primary porphyritic and pyroclastic features have been recognized.

Carbonaceous and calc-silicate rocks and cherts

Carbonate-bearing rocks are pale to greenish when fresh but weather to dark brown and have a very irregular, though distinctly schistose appearance (Fig. 9 b). The purest layers contain about 60-70 % carbonate, with the remainder being silicate minerals such as serpentine, diopside, chondrodite, tremolite, phlogopite and apatite. The chondrodite has been almost completely altered to serpentine and mica. Disseminated iron sulphides and graphite are invariably present in the carbonate rocks. The carbonate at Koivujärvi has been determined as calcite, comprising 40-80 % of the rock (Nikander, 1976). Higher up in the Ruotanen sequence, carbonates have been analyzed and shown to be calcitic dolomites, with a CaO/MgO ratio varying between 1.5 and 2.3 (Mäki, 1987). Similar calcitic dolomites are also present further south at Rautalampi (Lahtinen, 1988). At Savijärvi calcite is dominant, with dolomite occurring as relicts within grains of calcite or as isolated large crystals (Äikäs, 1988 f). Chemically, the Savijärvi carbonates are calcitic dolomites with Cao/MgO ratios varying between 2 and 3 (Kojonen, 1987). At Sulkavanjärvi, carbonate is most abundant in certain narrow layers where it can comprise up to 80% of the rock.

Calc-silicate rocks are banded and typically greenish and are usually relatively resistant to weathering, forming irregular surfaces of high relief. Principal minerals present are diopside, plagioclase (An_{40-70}), tremolite, carbonate and opaques, while accessory phases include quartz, apatite, biotite and zircon. Calc-silicate minerals are also present as large porphyroblasts in otherwise fine-grained lithologies or as narrow and irregular bands within carbonates and felsic volcanics (Figs. 9 c and d). Laminar cherts, sometimes described as quartzites or quartzite schists, are also associated with these lithologies (Fig. 9 e). The chert layers are usually thin, with a maximum thickness of 20 cm and are usually extensively fractured, with quartz having been strongly recrystallized. In

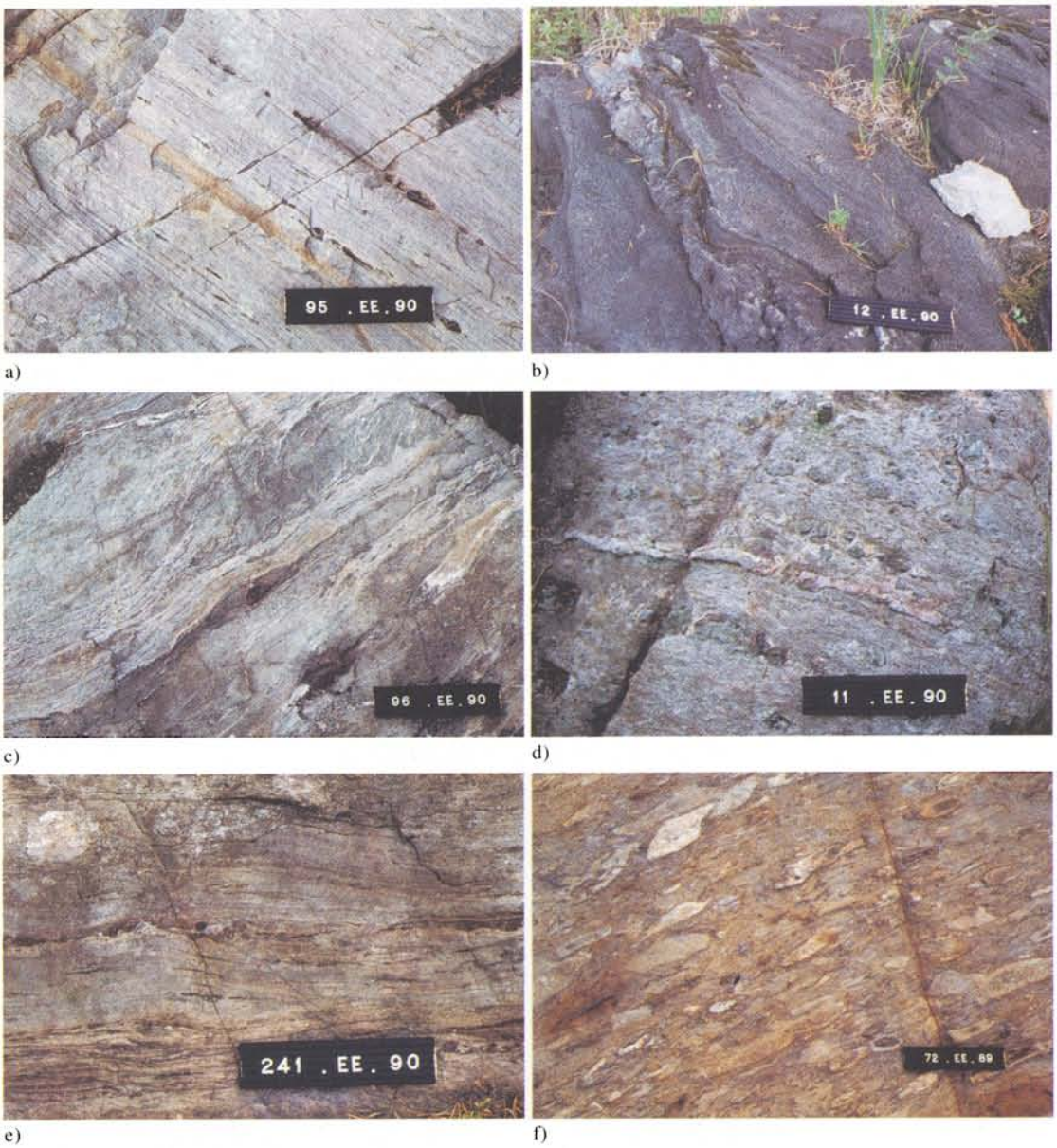


Fig. 9. The Savijärvi Suite contains (a) graded felsic metatuffs or tuffites with thin calc-silicate horizons; (b) carbonate rocks (view towards north, with vergence of thrusting towards northwest); (c) calc-silicate rocks with narrow cherty layers; (d) felsic calc-silicate rocks (porphyroblasts 1-3 cm in size); (e) chert bands in felsic metatuffs and (f) the Kalliokylä Conglomerate. Photographs a and c from Heinämäki; b and d from Uivero; f from Kalliokylä.

the southern part of the study area at Jokijärvi, near Tervo, within the Säviä schist belt, sulphide-bearing chert layers up to several meters in thickness are associated with calc-silicate rocks.

The majority of carbonate rocks in southern

and central Finland consist of calcite with sporadic dolomitic interbeds and presumably represent open marine, deep water deposition (Pekkala, 1987). This contrasts with those in eastern and northern Finland which were evidently formed in shallow, restricted basins with evaporites and locally under brackish conditions, favouring the precipitation of dolomite, or early diagenetic dolomitization of primary calcite. The widespread presence of graphitic schists in eastern Finland is also consistent with deposition in isolated depositional basins. In the boundary zone between the Karelian and Svecofennian formations, carbonates are present as thicker accumulations containing both dolomitic and calcitic layers which, according to Pekkala (1987), suggest variable conditions of formation. Lunden (1987) also described some substantial carbonate deposits from southern Finland, of which the most pure are the Otamo dolomites at Siikainen and Huutokoski at Joroinen. The differences between Svecofennian and Karelian carbonates are thus probably not as significant as was formerly maintained (Simonen et al., 1978).

Uranium- and phosphorus-bearing (U-P) horizon

The Savijärvi Suite includes a stratiform uranium-phosporus horizon which has been recognized in both the Säviä and Ruotanen schist belts. Accordingly, Äikäs (1988 f) has employed the term "Pielavesi phosphatic basin" and considered deposition to have taken place in a manner analogous to that at the Vihanti deposit at Lampinsaari. The U-P horizon within the Savijärvi Suite occurs as narrow layers intercalated with carbonates and calc-silicate rocks as well as quartz-feldspar gneisses. Apatite tends to be aligned within and parallel to lithological banding whereas calc-silicates and calcite are more randomly oriented. Uraninite and sulphides occur as inclusions within apatite, while the calc-silicate rocks also contain some molybdenum and tungsten. Apatite- and graphite-bearing layers represent primary lithological banding in what was originally dolomitic carbonate (Äikäs, 1988 f).

Black schists

The graphitic schists within the Savijärvi Suite do not form any particularly coherent or distinctive marker horizon, although graphite contents tend to generally increase through the upper part of the Salonsaari Suite and into the Savijärvi Suite. Graphite is common as narrow discontinuous concentrations within carbonates, calc-silicate rocks and felsic volcanics and is commonly accompanied by pyrite and pyrrhotite. Graphite-bearing horizons are not particularly thick nor are carbon and sulphur contents particularly high but they are sufficiently abundant and extensive as to make the Savijärvi Suite distinct from its surroundings. Evidently they record deposition under reducing conditions. Graphite and suphide disseminations have been concentrated during deformation into fractures and faults.

Kalliokylä Conglomerate

At Kalliokylä a polymictic conglomerate (Fig. 9 f) some 3-5 m thick is intercalated with calc-silicate rocks and graphitic schists and can be traced for over a kilometer along strike, although in places it is rather indistinct. The matrix of the conglomerate resembles fine-grained banded tuff, and consists of plagioclase, hypersthene, hornblende, biotite and quartz; accessory minerals are apatite, opaques and carbonate. The matrix also contains irregular lenses and veins of quartz. Concordant clasts are highly elongate, from 5-25 cm in length and only 1-5 cm in sections perpendicular to the lineation (Fig. 9 f). The clasts are predominantly quartz-feldspar gneisses, sometimes with calc-silicate banding, true calc-silicate lithologies and graphitic schists. In addition, somewhat more mafic hornblende gneiss and coarse-grained granodiorite fragments are present,

commonly containing cordierite and hypersthene. The Kalliokylä Conglomerate is very local and possibly represents rapid syndeformational erosion and deposition.

Thicknesses and nature of contacts

The rapid variations in lithology within the Savijärvi Suite indicate very changeable depositional environments and repeated brief volcanic interludes. The abundance of volcanic material does however, increase upwards, with a gradual transition via massive mafic volcanic units into the overlying Säviä Suite.

The most complete section through the Savijärvi Suite is exposed in the Savijärvi axial depression, where it is some 100-150 m thick. At Savijärvi, the carbonate-rich units form discontinuous lenticular horizons varying in thickness from several centimeters up to 20 m whereas at Koivujärvi they are typically 0.2-0.5 m thick. At Sulkavanjärvi, the Savijärvi Suite is up to 200-300 m in thickness but carbonate rocks here are subordinate to calc-silicate rocks and felsic volcanics with calc-silicate banding. At Uivero the carbonate rock comprises 5-10 m of a total thickness for the Suite of about 100 m, with felsic volcanic units several tens of meters thick being predominant. Several profiles have been drilled through the Savijärvi Suite (Fig. 10), where it dips moderately towards the southeast and overlies migmatitic gneisses and quartz diorites of the Salonsaari Suite. The contact between the two suites is gradual, with a progressive upwards increase in the abundance of carbonate, calc-silicate and graphitic lithologies. The boundaries between the carbonate and calc-silicate units are also transitional, in contrast to those with felsic volcanics, even though the latter do contain some calc-silicate bands. In the Savijärvi section, the overlying Säviä Suite is represented by mafic metavolcanics, which are in turn overlain by biotite-hornblende gneisses of the Koivujoki Suite, with the drilling profiles extending as far as the Kotajärvi Metalava (Fig. 10). The presence of the Kalliokylä Conglomerate suggests rapid deposition during a tectonically active phase prior to or initiating the volcanism recorded by the Säviä Suite.

Säviä Suite

Definition, subdivision and distribution

The distribution of the main sequence of volcanic rocks throughout the Pyhäsalmi-Pielavesi area defines an ovoid region about 100 km from north to south and some 40 km from east to west (Fig. 2). These volcanics and associated lithologies comprise the Säviä Suite, named after the Säviä Cu-Zn deposit.

The Säviä Suite consists of mafic, intermediate and felsic metavolcanics, as well as their hydrothermal alteration products. At Säviä, Pangansalo and Koivujärvi the Suite commences with mafic volcanics, lavas and tuffs. Some of the banded diopside-bearing mafic metavolcanics may represent deformed pillow lavas, as have been described from the Haukivesi district (Gaál and Rauhamäki, 1971) and from Virtasalmi (Pietikäinen, 1986). Primary features, including agglomeratic and fragmentary lava structures, are more abundantly preserved at Koivujärvi. At Säviä itself, the mafic volcanics are followed by a major intermediate sequence with numerous minor felsic intercalations. Intermediate metatuffs are also present in the Hirvijärvi and Uivero areas, in the Pangansalo zone and at Lampaanjärvi. They are less abundant at Koivujärvi, where felsic pyroclastics immediately overlie the mafic volcanic sequence.

The mafic sequence is best exposed at Paloniemi to the north of Säviä, from where they

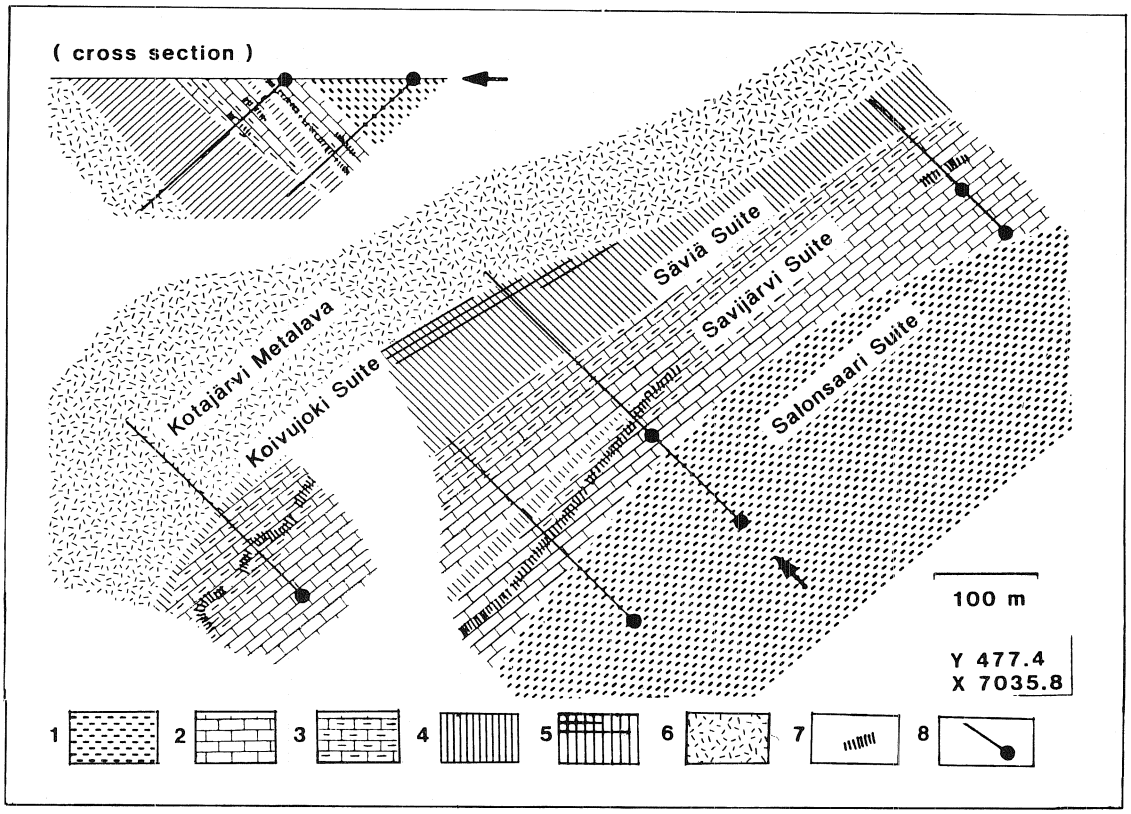


Fig. 10. Vertical cross-section through the Savijärvi Suite. 1 = Migmatitic mica gneiss (Salonsaari Suite); 2 = Carbonate rock; 3 = Quartz-feldspar gneiss and calc-silicate intercalations in carbonate rocks; 4 = Amphibolite (Säviä Suite); 5 = Mica-hornblende gneiss (Koivujoki Suite); 6 = Kotajärvi Metalava (Koivujoki Suite); 7 = U-P horizon; 8 = Drill hole. Modified after Äikäs (1988d). Drill profile is towards NW, so that arrow indicates top of overturned sequence.

can be traced northwards as thinner, evidently more distal horizons. The stratigraphic position of the Säviä Suite volcanics can also be ascertained further north in the Joutsenniemi profile as well as both east and west of Uivero and in the Savijärvi drill profile (Fig. 10). The sequence occupies a relatively open synform at Pangansalo, while at Koivujärvi the Ruotanen schist belt verges towards the northwest, in places almost flat-lying (see sections on the appenden map 1).

Hydrothermal alteration processes in the Säviä Suite are indicated by the presence of sericitic, silicified and garnet-cordierite-anthophyllite lithologies. Alteration within the Säviä zone was most intense at Leväniemi, Säviä itself and at Paloniemi. Garnet-cordierite-anthophyllite rocks are also present at Koivujärvi but an especially thick development of altered rocks occurs at Kangasjärvi. In the Pangansalo-Haapajärvi zone however, there is no evidence of such alteration. The Säviä base metal deposit and related mineralizations are situated near the top of the Säviä Suite, within more or less hydrothermally altered felsic volcanics.

Lithologies

Mafic metavolcanics

The volcanic sequence within the Säviä Suite commences with mafic metavolcanics, com-

monly represented by uralite-porphyritic lavas (Fig. 11 f). Individual flows are typically separated from each other by tuffitic intercalations and may be up to several meters in thickness. The uralitic porphyries are dark green in outcrop and tend to be homogeneous and only weakly oriented. Relict uralite phenocrysts are 1-5 mm in diameter, in a fine-grained (<0.5 mm) groundmass. Primary augite is generally present in the cores of the phenocrysts, surrounded by hornblende. Plagioclase (An_{50-60}) is the other principal mineral present and in certain layers it is recognizable as highly altered relict phenocrysts. The abundance of plagioclase is also observed to increase upwards through individual flows. In addition, the groundmass contains opaques, sphene, biotite, epidote and sericite.

The boundaries of the mafic lava flows at Koivujärvi show evidence of fragmentation (Nikander, 1976), with fragments varying from several centimeters to several decimeters in size.

Banded mafic metalavas forming flows 2-3 m thick and containing pillow structures are exposed on the island Kokkosaari to the north of Paloniemi. The pillows range from 20 cm to 40 cm in diameter and are highly strained (Fig. 11 e). No vesicles or amygdales have been recognized and it is not possible to determine the younging direction of the pillows. The pillows consist principally of hornblende, plagioclase and diopside, while the interstices between pillows are predominantly pale green and are composed of epidote and diopside.

The more typical metalavas are fine-grained with a weak mineral alignment and consist principally of hornblende and plagioclase, although the former has been locally replaced by cummingtonite; accessory phases include opaques, sphene, apatite and sericite. Primary brown hornblende has been found at Uivero. Individual flows tend not to be of great thickness and they alternate with strongly foliated pyroclastic units. Diopside-rich bands are also common, reflecting the abundance of carbonate precipitation within the volcanic sequence.

On weathered surfaces the metatuffs are dark green with distinct banding. The major minerals present are plagioclase (An_{40-65}) and hornblende, with additional diopside and accessory quartz, opaques, apatite, sericite and zircon. Hornblende sometimes occurs as aggregates whose outlines resemble pseudomorphs after phenocrysts. The banded appearance is due to variations in the relative abundance of hornblende and plagioclase or, in some cases, diopside and plagioclase. A small amount of garnet is also typical in the mafic lithologies and at Paloniemi garnet and diopside form irregular narrow seams to the extent that the rock has been referred to in the field as garnet skarn. These macroscopically resemble the skarns within the lower parts of the diopside amphibolites in the vicinity of the Hällinmäki ore deposit (Hyvärinen, 1969). The metavolcanics are invaded by thin garnet-quartz-carbonate veins, and sporadically contain disseminated pyrrhotite or pyrite.

Based on their mineral parageneses and metamorphic grade, the volcanics have been variously described as amphibolites, hornblende gneisses, amphibolite skarns and hornblende-cummingtonite gneisses (Laitakari, A.J, 1968).

Intermediate metavolcanics

The most volumetrically significant unit within the Säviä Suite consists of fine-grained metatuffs and metatuffites that, owing to rapid variations in composition, usually display distinct banding (Figs. 11 c and d). Layers are typically from 1-25 cm thick. Principal minerals are plagioclase (An_{35-65}), hornblende, cummingtonite and quartz, while accessory phases include biotite, apatite, magnetite, zircon, sphene, carbonate, epidote and sericite. Hypersthene is also widespread but this is a result of recrystallization under high-grade metamorphic conditions. Hydrothermal alteration, indicated by the presence of sericitization and

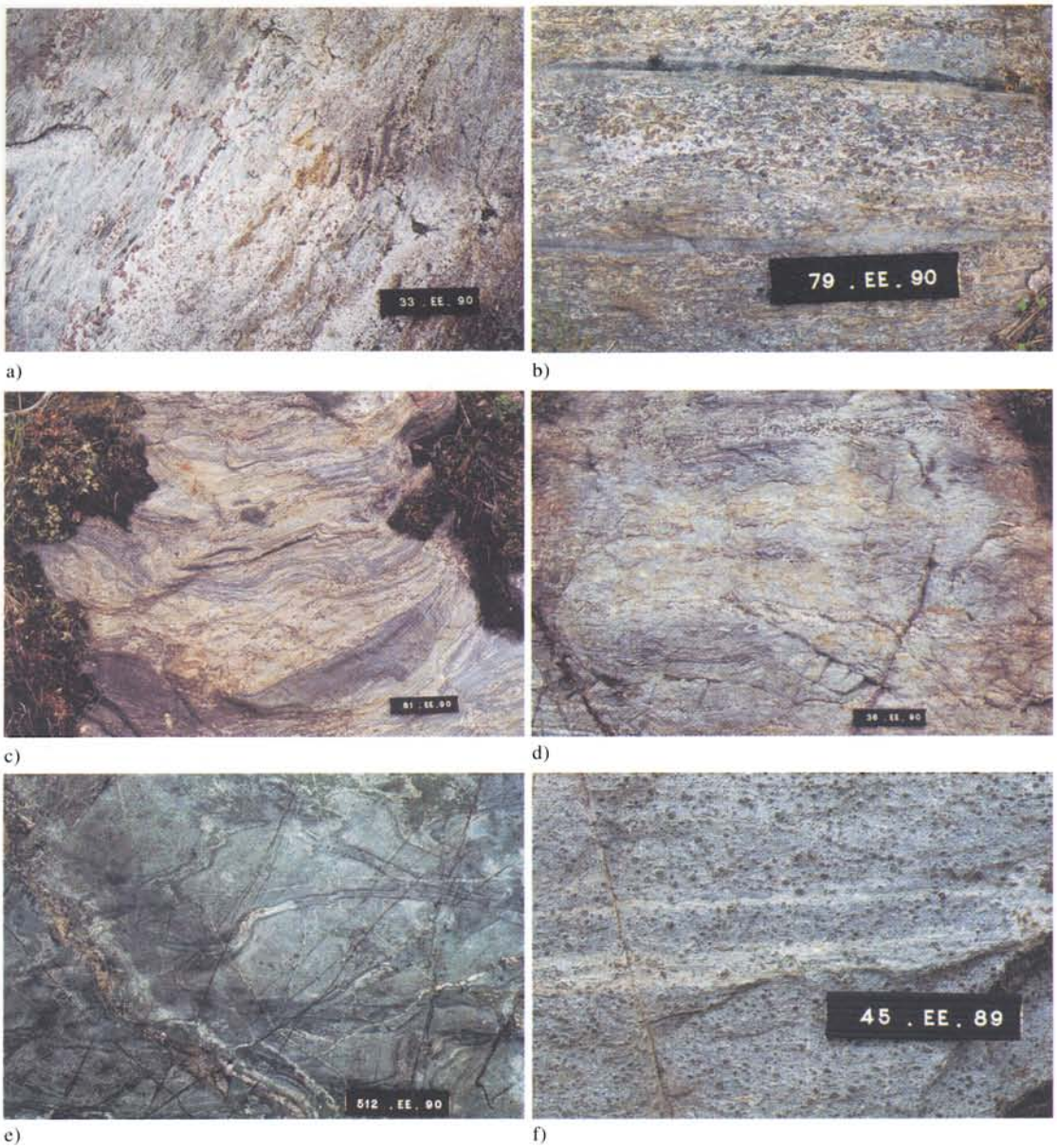


Fig. 11. The stratigraphic sequence within the Säviä Suite is, from top to bottom (a) garnet-cordierite-sillimanite gneisses (originally felsic volcanics), (b) stratiform garnet-cordierite-anthophyllite rocks, (c-d) intermediate to felsic metatuff containing altered horizons, (e-f) deformed pillow lavas and uralitic porphyry. Photographs a-c from Paloniemi; d from Uivero; e from Kokkosaari and f from Haapakoski.

bands of garnet-cordierite-anthophyllite rock tend to increase upwards through the sequence (Fig. 11 b). These intermediate lithologies have previously been termed hornblende gneisses, hornblende-cummingtonite gneisses, cummingtonite gneisses, or cummingtonite-hypersthene

gneisses (Laitakari, A.J, 1968).

The intermediate metatuffs at Paloniemi contain lapilli-like features, consisting of lenses of diopside, quartz, plagioclase and carbonate; presumably they represent highly deformed primary carbonate horizons, rather than true lapilli.

A narrow agglomeratic intercalation is present within the sequence at Pangasalo (Fig. 12 a), consisting of a banded intermediate tuffaceous matrix and clasts, now highly elongated, measuring 1-5 cm perpendicular to the lineation and 20-25 cm parallel to it. Clasts consist predominantly of uralite- and plagioclase porphyry and felsic volcanics. The unit is sharply bounded and only some 2 m in thickness.

Felsic metavolcanics

It has not been possible to identify individual felsic volcanic units in the Säviä-Paloniemi area. Instead, felsic pyroclastic layers (Fig. 12 b) alternate regularly with more mafic units. The felsic volcanics (Fig. 11 a) have also been referred to as quartz-feldspar gneisses, quartzites, cordierite-sillimanite-garnet gneisses or biotite- and hornblende-poor gneisses (Laitakari A.J, 1968; Huhtala, 1979).

In places the cordierite gneisses retain some evidence of their original pyroclastic layering and in the Ruotanen area, sericite quartzites also have Ti/Zr ratios consistent with their origin as felsic volcanics, pyroclastics and quartz porphyries (Mäki, 1986). Drilling profiles at Paloniemi also intersected porphyritic lithologies. Nikander (1976) considered that lapilli tuffs in the Koivujärvi area also contained pumaceous fragments, although they now consist of highly elongate lenticular aggregates of quartz and K-feldspar. The felsic pyroclastic deposits at Pangansalo are also better preserved than at Säviä (Fig. 12 c).

The felsic tuffs typically consist of quartz, plagioclase (An_{20-30}), K-feldspar and biotite, with accessory opaques, apatite, sphene, hornblende, epidote and sericite. K-feldspar is relatively uncommon at Paloniemi, whereas it is rather abundant at Koivujärvi and Pangansalo. The tuffs weather to a pale buff color and tend to be fine-grained and banded, with plagioclase and K-feldspar occurring as phenocrysts. The felsic volcanics are commonly sericitized and usually contain weak disseminations of pyrite and pyrrhotite.

Garnet-cordierite-anthophyllite gneiss

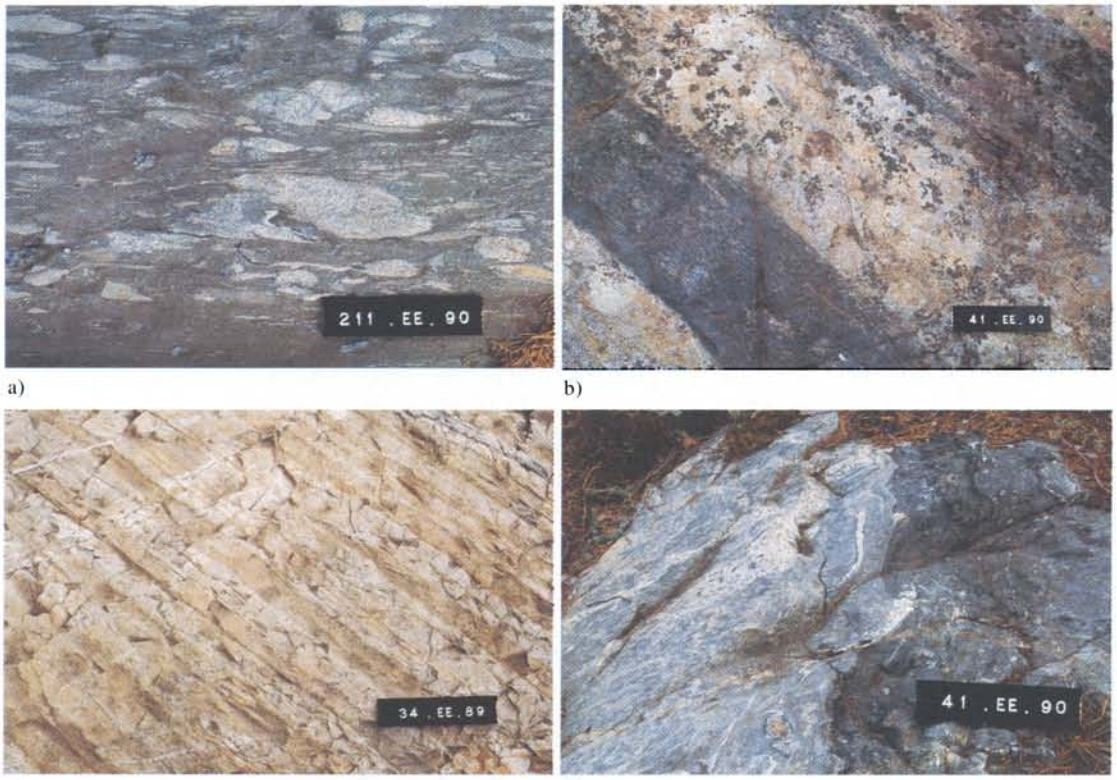
Iron- and magnesium-rich garnet-cordierite-anthophyllite and/or sillimanite-bearing lithologies are most abundant in the Paloniemi-Säviä-Leväniemi zone, which is some 12 km in length and is of particular significance with regard to its mineralization potential.

The garnet-cordierite-anthophyllite rocks differ chemically from the metavolcanics in having lower Ca and Na abundances and conversely, higher Fe and Mg abundances. This difference is attributed to intensive hydrothermal alteration of mafic volcanics (Makkonen, 1981; Mäki, 1986; Rasilainen, 1991). Evidently Ca and Na were leached from the volcanics and Mg was added by interaction with sea water. The garnet-cordierite-anthophyllite rocks occur within the upper part of the volcanic sequence and are stratabound.

On the basis of variations in plagioclase abundance, Makkonen (1981) subdivided these lithologies at Leväniemi, in the Säviä zone, into plagioclase-rich and plagioclase-poor varieties. The garnet-cordierite-anthophyllite rocks occur as narrow intercalations within intermediate metatuffs and as thicker massive units; the former type tend to belong to the plagioclase-rich group and appear to be situated at a somewhat lower stratigraphic level.

Contacts and thicknesses

The Säviä Suite begins abruptly with mafic, predominantly submarine lavas, although the profiles drilled at Savijärvi and Paloniemi in-



a) b) c) d)

Fig. 12. Säviä Suite (a) intermediate agglomerate with clasts of plagioclase- and uralite-porphyry and felsic volcanics; (b) felsic pyroclastic with mafic layer; (c) laminated felsic pyroclastic and (d) contact between mafic and intermediate lithology. Photograph a from Pangansalo, b and d from Paloniemi and c from Lampurinkangas.

dicate that some skarn and mica gneiss intercalations persist. The contact with the underlying carbonate-bearing sediments is conformable and has been intersected by drilling profiles at Säviä, Paloniemi and Savijärvi. The existence of the Kalliokylä Conglomerate near the boundary nevertheless indicates considerable relief and erosion prior to extrusion of the Säviä volcanics. At Kokkosaari the mafic metavolcanics directly overlie mica gneisses with a sharp, tectonized contact.

The mafic metavolcanics of the Säviä Suite attain a maximum thickness of 200 m in the Säviä and Paloniemi sections. The transition to intermediate lithologies is distinct, although mafic intercalations occur sporadically throughout the entire intermediate and felsic succession (Fig. 12d). Boundaries between individual lithological units are sharp. The total thickness of the Säviä Suite volcanic sequence at Paloniemi is of the order of 500 m, whereas at Savijärvi it is a mere 30- 40 m thick. At Koivujärvi it is difficult to estimate the thickness due to the prevalence of gently plunging folds, although preseumably several hundred meters of section are present. The existence of lapilli tuff and agglomeratic features at Paloniemi and Pangansalo, as well as at Koivujärvi, indicates proximity to a volcanic vent. At a larger scale, this probably formed part of a more extensive chain of volcanic centers.

The garnet-cordierite-anthophyllite rocks

generally form layers from several centimeters to several meters thick which together attain a combined thickness of 100-150 m in the vicinity of the Säviä ore deposit and nearly as much at Leväniemi, Paloniemi and Kangasjärvi within the Ruotanen zone. Where altered rocks originally consisted of compositionally distinct units, primary lithological boundaries are still discernible. Felsic metavolcanics are thus represented by sericitized garnet-cordierite-sillimanite gneisses.

Koivujoki Suite

Definition, subdivision and distribution.

The Koivujoki Suite takes its name from the mica gneisses exposed at Koivujoki, in the northwestern part of the study area. They are distinctly different from the underlying migmatitic gneisses in being less intensely deformed and containing more biotite and hornblende but fewer graphitic schist intercalations. In addition they are commonly graded, resembling graywackes and arkosites, with sporadic diopside amphibolite intercalations. The hornblende-mica gneisses at Länsi-Säviä to the south of Koivujoki are also correlated with the Koivujoki Suite. A further distinct zone of these younger mica schists occurs in the eastern part of the study area, at Lampaanjärvi, from where it can be traced northwards along the western margin of the Vieremä-Haajainen zone. Marttila (1981) considered that the rocks of the former zone stratigraphically underlie those of the latter. A smaller area of these relatively young and less strongly deformed mica gneisses occurs in the eastern part of the Säviä zone, at Penttilänlahti and Haapakoski; Pääjärvi (1991) has also described similar lithologies from Tuppilahti, on the Vesanto map sheet, immediately to the south of the present study area.

Rather than forming a series of separate intrusive bodies, the rocks described by Salli (1983) as gabbros and diorite actually belong to a coherent unit outcropping over an area of more than 200 km^2 and consisting predominantly of metalavas and associated gabbroic rocks. They are evidently correlative with the less voluminous mafic intercalations (both flows and sills) within the mica gneisses of the Koivujoki Suite. This unit has been named the Kotajärvi Metalava and can be traced as far northwards as Sulkavanjärvi where it resembles quartz diorite in appearance and is consequently more difficult to recognize. The best-preserved exposures are at Länsi Laukkala and Saarela, although the Saarela gabbro itself is a younger, unrelated intrusion.

Lithologies

Mica-hornblende gneisses

The mica gneisses of the Koivujoki Suite weather to a dark gray and are distinctly graded, showing distinct alternations between psammitic and pelitic lithologies on a scale of 5-20 cm. Granitic veins are common but the effects of deformation are less intense than in underlying rock units (Fig. 13 a). Disseminated magnetite is present within the lower parts of the Koivujoki mica gneisses. The more mafic composition of the gneisses compared to those of the Salonsaari Suite reflect derivation from a correspondingly more mafic source. An exception to this is in the vicinity of Penttilänlahti where detritus may have been supplied by erosion of the extensive felsic volcanics of the Pangansalo zone. Although they have been strongly recrystallized, the mica gneisses retain some features indicative of their origin as graywackes.

The principal major minerals present, in highly variable relative proportions, are plagio-

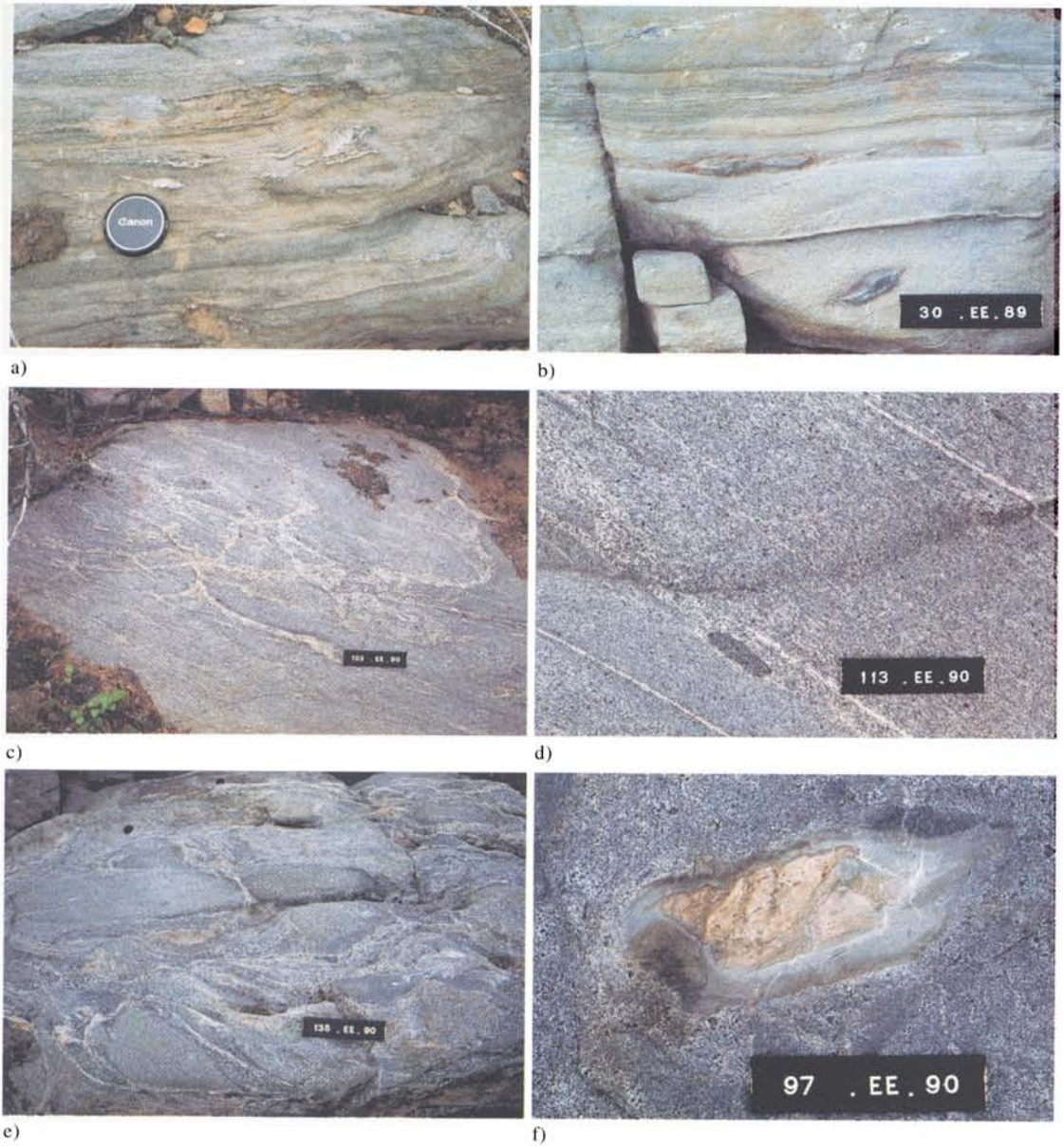


Fig. 13. The Koivujoki Suite comprises (a) graywackes that are often hornblende-bearing mica gneisses and (b) graded arkosic graywackes. The Kotajärvi Metalava is also coeval with these deposits and consists of (c) breccia-like structures, (d) foliated but massive lava-like variants, (e) strongly deformed relict pillow lava or breccia structures, and (f) xenoliths of Savijärvi Suite felsic volcanics. Photgraph a from Koivujoki, b from Haapajärvi, c, d and f from Ruokokangas, e from Hämeensaari.

clase, quartz, biotite and hornblende; accessory minerals include K-feldspar, garnet, zircon, sillimanite, cordierite and apatite. Relict sillimanite occurs within porphyroblasts replaced by muscovite. Andalusite occurs throughout the Lampaanjärvi zone, while garnet is ubiquitous.

K-feldspar is less abundant than in the mica gneisses of the Salonsaari Suite. Mafic metavolcanic intercalations are typical of the upper part of the gneiss unit and graphite-bearing layers are also present, although they are not laterally persistent.

Meta-arkosites

Feldspathic graywackes are relatively abundant at Lampaanjärvi, Koivujoki and to the north of Haapajärvi. They weather to a very pale color and are characterized by well-developed grading, with coarser quartz- and feldspar-rich basal parts and biotite- and muscovite-rich upper parts defining units from 5-30 cm in thickness (Fig. 13 b). In addition to K-feldspar, the principal minerals present are biotite, muscovite, plagioclase and quartz; the latter two minerals may also occur as isolated larger clasts. Primary depositional features are no longer recognizable, although in the Haapajärvi area, where the arkosic units are most abundant, some possible relict slumping structures have been observed.

Kotajärvi Metalava and associated metagabbros

The "Kotajärvi Diorite" of Salli (1983) appears to form an extensive, presumably extrusive unit that has been deformed together with enclosing supracrustal units and which occurs at a consistent stratigraphic level. At Länsi Laukkala, to the south of Kotajärvi, lithologies interpreted as metalavas outcrop as a gently dipping unit many square kilometers in extent.

Typically the metalavas weather to dark green or brown and contain hornblende crystals, often with relict pyroxene cores. A brecciated appearance, due to irregular networks of granitic veins (Fig. 13 c) is also characteristic and may be a primary feature. Generally the undisrupted metalavas are fine-grained and massive although more anisotropic layers containing aggregates of hornblende are occasionally present (Fig. 13 d). In the Saarela area, in proximity to the inferred source of eruption, the brecciated structures are more abundant (Fig. 13 e). The dominant minerals in the Kotajärvi Metalava are plagioclase (An_{10-25}), hornblende, clinopyroxene and, to a lesser extent, K-feldspar and quartz. Opaques, sphene and apatite are typical as accessory minerals.

Medium-grained gabbroic variants are also associated with the Kotajärvi Metalavas, occurring at the same stratigraphic level. At Tommonmäki, such gabbros are present as an almost flat-lying intrusion tens of meters thick within the metalava sequence, although the contacts between the two units are not exposed and it is therefore difficult to establish their exact relationship to one another. The gabbroic units are subophitic, consisting of plagioclase (An_{45-60}), clinopyroxene, orthopyroxene and hornblende, the latter generally being an alteration product after pyroxene. Accessory minerals include biotite, apatite, opaques, sphene and quartz.

Mafic metavolcanics

The mica gneisses of the Koivujoki Suite contain thin mafic metavolcanic intercalations that have been referred to in the field as diopside amphibolites. More intermediate compositions are also present, as are sporadic uralite porphyry horizons. The latter are closely associated with mafic and ultramafic lenses that are of significance because of their Ni-Cu potential. Layered intrusive sills indicate intermittent magmatic activity concomitant with deposition of the graywackes. At Lampaanjärvi and Länsi Säviä similar, narrow metavolcanic horizons also occur within the mica gneisses.

Contacts and thicknesses

The contact between the mica gneisses of the Koivujoki Suite and the underlying Säviä Suite volcanics has not been seen in outcrop while the present distribution of the younger gneisses, as separate domains on either side of the

Säviä schist zone gives the impression of an allochthonous relationship; subsequent tectonic movements have controlled the preferential preservation of the gneisses as currently exposed. Otherwise the Koivujoki Suite mica gneisses are bounded by concordant granodiorite and quartz diorite intrusions.

In the Savijärvi drilling profiles (Fig. 10), the Kotajärvi Metalavas are separated from the underlying Säviä mafic volcanics by a 30 m thick mica-hornblende gneiss unit that represents clastic sedimentation. Therefore, the metalavas are at least in part younger than the mica gneisses. The contact zone between the metalava and the gneisses is marked by a hornblende-rich selvage some 10 cm thick; similar features occur at the base of the metalavas in the Länsi Laukkala area, where they they are up to several meters in thickness. At Hämeenluoto the Kotajärvi Metalava appears to conformably overlie intermediate and mafic pyroclastics belonging to the Säviä Suite. The diopside-bearing amphibolite and uralite porphyry units occurring as intercalations within the mica gneisses at Lampaanjärvi, Länsi Säviä and Koivujoki, which represent synsedimentary volcanism, are also correlated with the Kotajärvi Metalava. The Kotajärvi Metalava also contains enclaves of felsic volcanics and calc-silicate rock presumably derived from the underlying Savijärvi Suite (Fig. 13 f), as well as larger mafic metavolcanic inclusions from the Säviä Suite.

The Kotajärvi Metalava varies from 50-100 m in thickness and at a regional scale essentially forms a folded antiformal sheet with gentle plunges and older units exposed in its core. The gabbroic units possibly represent subvolcanic intrusions, occurring at approximately the same stratigraphic level and having undergone the same deformational history. Geochemically however, the gabbros are distinct from the metalavas.

The graywacke-like mica gneiss unit may be up to 2000 m thick in the Koivujoki area. Similarly, the mica gneisses at Lampaanjärvi form a very thick unit that continues northwards along the southwestern margin of the Vieremä-Haajainen Formation.

INFRACRUSTAL LITHOLOGIES

Distribution of units and lithologies

Ultramafic rocks

The most prominent ultramafic unit in the study area is the coarse-grained plagioclase pyroxenite outcropping in the core of the Joutsenniemi dome, which forms a massive body about 4 km in length and 700 m in width. Relict clinopyroxene is common within the central parts of hornblende crystals and aggregates, varying from 0.5-3 cm in diameter. These comprise up to 90% of the rock, the remainder consisting of plagioclase (An_{50-80}). Accessory minerals include biotite, opaques, apatite and zircon. Thin granitic veins traverse the pyroxenite but otherwise the degree of deformation is relatively low and it may represent syntectonic intrusion (Fig. 14 a). The ultramafic unit contains some pyrrhotite and chalcopyrite but nickel is not anomalous.

Smaller ultramafic lenses of similar appearance and composition occur to the north of Haapajärvi, at Säviä and in the area between Paloniemi and Kirkkosaari; in these bodies too, disseminations of pyrrhotite and chalcopyrite are ubiquitous. Other ultramafic occurrences will be discussed in a later section in connection with Ni-Cu mineralization.

Fig. 14. The Joutsenniemi dome is intruded by (a) an ultramafic body cut by coarse-grained granitic veins. The majority of the Pielavesi quartz diorites are palingenetic (b and d) and contain enclaves of older metasediments. Pegmatitic dykes (c) represent relatively brittle deformation (view towards north). Photograph a from Joutsenniemi, b from Salonsaari, c from Hämeensaari and d from Leväniemi.

Gabbros

The most significant gabbro intrusions are those at Saarela, Tyypekki and Teerimäki. The Saarela Gabbro is a tear drop shape in plan, some 9 km long and a maximum of 3 km wide at its northern end. Two faults bisect the gabbro (trending 300° and 010°), with the area to the south of the faults being exposed on only a few islands. The Saarela gabbro also includes some olivine-bearing and anorthositic variants, while near the eastern margin of the intrusion, serpentinites that may be of significance for Ni-Cu mineralization are present; the mineralized glacial erratics that have been found in the area are probably derived from this unit.

The medium-grained Tyypekki gabbro contains abundant fine-grained gabbro and graphite shist enclaves, which are considered to represent spalling or erosion of the intrusion margin (Pääjärvi, 1991). The Tyypekki Gabbro is about 3x2 km in area. The main gabbro body at Teerimäki is an almost 2 kilometer long boomerang-shaped intrusion with a strong SE-trending lineation. Hornblende-biotite gabbro occurs at the margins white the central part of the intrusion consists of hyperstene- and hyperstene-cummingtonite gabbro. Smaller satellitic gabbroic bodies occur along strike from the main intrusion and additionally at Pikonniemi, within the Koivujoki mica gneisses and at Sulkavanjärvi.

The gabbros are typically medium-grained and either dark green or brown on weathered surfaces, depending on prevailing mineral compositions. Plagioclase (An_{45-60}) and both orthopyroxene and clinopyroxene are the dominant minerals present, with accessory phases including biotite, cummingtonite, opaques, apatite, sphene and occasionally quartz. Plagioclase is generally euhedral and gives the rock an ophitic or subophitic character and hornblende occurs as a replacement product after pyroxene. On the basis of its mineralogy, the rock can therefore be classified as a gabbro, or more specifically, hornblende gabbro or norite.

Quartz diorites and granodiorites

The regional geology of the Pielavesi and Kiuruvesi region is dominated by the abundant quartz diorite and granodiorite intrusions that are generally concordant with layering in the supracrustal rocks. In the eastern part of the area they have been interpreted as basement complexes (Marttila, 1976). The two lithologies commonly merge with one another, forming a compositional continuum that cannot be subdivided at map scale. They also usually contain indistinct relicts of older gneisses. Compositional variations presumably reflect differences between source materials and degrees of melting (App. 1). Although they are usually concordant, distinctly discordant intrusive contacts have also been observed. In the Pielavesi area it is difficult to recognize an external origin for the quartz diorites and granodiorites; rather they appear to be the products of in situ partial melting.

The quartz diorites and granodiorites are generally medium-grained have a blebby or banded gneissic appearance (Fig. 14 b), where not resembling more homogeneous plutonic lithologies (Fig. 14 d). Mineral compositions and the abundances of mafic minerals vary appreciably. Felsic minerals include quartz and plagioclase (An_{20-30}) and generally K-feldspar as well, while mafic phases present are biotite and hornblende. Epidote, apatite, sphene and zircon occur as accessories, along with garnet, suggesting an excess of Al_2O_3, consistent with derivation from a metasedimentary source.

Pyroxene-bearing quartz diorites and granodiorites, indicative of higher temperatures and pressures during intrusion or subsequent metamorphism, are present in the vicinity of Joutsenniemi, Vaaraslahti and Sulkavanjärvi. Enclaves of mafic and intermediate supracrustal lithologies are abundant, in a groundmass that may be described as brownish or greenish, easily weathering charnockite. These lithologies are analogous to the charnockites occurring in the Kiuruvesi and Haukivesi districts (Marttila, 1976; Gaál and Rauhamäki, 1971). Wahl (1963) already recognized their spatial relationship to the Raahe-Ladoga Zone.

Granites

Very coarse-grained porphyritic pyroxene granites have a broadly similar distribution to the pyroxene-bearing quartz diorites throughout the Vaaraslahti-Joutsenniemi-Sulkavanjärvi area, occurring typically within NW-, N- or NE-trending zones. The rocks are prone to weathering, with a friable, saprolitic texture extending to depths of a meter or more and in outcrop they tend to be a darker greenish or brownish color than other granitoids. Grain sizes vary from 0.5-2.0 cm and pyroxene is present as hypersthene. The Vaaraslahti hypersthene granite may have formed by anatexis of predominantly pelitic metasediments. In places K-feldspar phenocrysts have been overgrown by plagioclase (Hölttä, 1988). Hölttä considered that the relatively dry hypersthene granite magma had formed under granulite facies conditions and intruded the rocks at the margins of the charnockitic province during somewhat lower grade metamorphic conditions. Coarse-grained porphyritic microcline granites are also found to the west of Koivujärvi at Tuomijoki, where they are exposed on a number of small islands within the lake Pielavesi.

Exposures of even-grained granites are abundant to the north of Kirkkosaari and east of the town of Pielavesi; the latter typically contain enclaves of schists, granodiorite and more mafic plutonic lithologies. The granites have intruded preferentially into zones of weakness and in places they are deformed with mylonitic textures.

Dyke rocks

Narrow mafic dykes, typically only some tens of centimeters thick, occur throughout the entire study area. Compositionally they correspond to the gabbros within area. The narrower dykes are fine-grained and massive, while the thicker ones possess an ophitic texture. Marttila (1976) described extensive laccolith-like diabases from the Kiuruvesi area. At Teerimäki the younger diabase dykes clearly intrude the main gabbro massif. In the NE-trending Koivujoki zone, ophitic diabase dykes are associated with lenticular mafic and ultramafic intrusive bodies.

Pegmatitic dykes are usually simple in form, coarse-grained and are often quartz-rich in their central parts. They tend to follow late-stage, brittle structural trends and are clearly discordant with respect to previously folded and deformed lithologies. A good example of this is in the Koivujoki mica gneiss zone, where a swarm of pegmatite dykes strikes towards 025°-030° (Fig. 14 c). Another preferred trend of pegmatite intrusion was N-S.

Contact relationships

The contact between plagioclase pyroxenite and orthogneisses is exposed at Joutsenniemi, where the grain size of the ultramafic unit progressively decreases over a 2-3 m interval approaching the contact, ultimately forming a massive fine-grained lithology about a meter thick. In addition, ultramafic veins, generally concordant with the regional domal structure, are found within the orthogneisses. The group of ultramafic bodies aligned in a NW-trending zone in the Jylhä-Kirkkosaari area may also be interpreted as fracture-controlled (Damstén and Ekdahl, 1989). These intruded into both the Säviä Suite volcanics and the Kirkkosaari tonalitic basement and they may be correlated with the larger gabbro bodies in the study area.

There are no obvious contact metamorphic effects where ultramafic bodies have intruded the graywacke-like gneisses of the Koivujoki Suite; rather contacts appear to be more tectonic than intrusive in nature. Although the larger gabbroic bodies tend to be spatially associated with the quartz diorites and granodiorites they clearly truncate them. The Teerimäki Gabbro for example, is clearly younger than the Säviä Suite volcanism, hydrothermal alteration and mineralization (see later discussion). A similar relationship is evident in the Tervo area where enclaves of both felsic and mafic volcanics, as well as garnet-cordierite rocks occur within the Saarisenjärvi gabbro (Ekdahl, 1974).

The Tyypekki gabbro and the ultramafic lenses in the Koivujoki gneisses contain abundant graphitic gneiss enclaves, as do the surrounding quartz diorites. The mutual relationship between the gabbro and the quartz diorites has not been established due to poor exposure but at Tervo a chilled gabbro contact with quartz diorites has been observed (Ekdahl, 1974).

The diabase dykes in the Kiuruvesi area cut all other rock types (including the pyroxene granodiorites), with the exception of the porphyritic and pyroxene-bearing granites (Marttila, 1981). In the Säviä schist belt, mafic dykes occur that post-date volcanism and hydrothermal altered garnet-cordierite-anthophyllite rocks, as well as most phases of deformation (Makkonen, 1981). It is demonstrable in the Koivujoki area that pegmatites are youger than

the diabase dykes.

Although generally concordant, the granodiorites and quartz diorites locally intrude discordantly across lithological layering in their country rocks. Since, however, they represent complex and prolonged or multiple episodes of emplacement, it is difficult to assess their precise age relationships. The porphyritic granites contain a wide variety of country-rock inclusions; in places these granites have intruded within deformation zones after the main phase of folding, as indicated by the presence of folded enclaves in various stages of assimilation.

GEOCHEMICAL CLASSIFICATION AND TECTONOMAGMATIC SETTING OF THE PIELAVESI SEQUENCE

Alteration processes

Chemical alteration of volcanic rocks can take place during submarine volcanism as well as during subsequent regional metamorphism. Elements such as Ti, P, Zr, Zn, Cr, Y, Nb and the rare earth elements (REE) are considered to be relatively immobile during most hydrothermal alteration processes and during metamorphism, while K, Cs, Pb, Ba, Al, Fe, Ca and Si are less so (Condie, 1982).

Altered rocks have alkali metal abundances and element ratios that deviate from those in unaltered rocks and can therefore be identified by comparison against the compositional spectrum for igneous rocks (Hughes, 1972). According to these parameters, approximately 40% of the samples analyzed from Pielavesi have been subjected to synvolcanic hydrothermal alteration processes (Fig. 15 a). These correspond in particular to the garnet-cordierite-anthophyllite rocks, as well as to mineralized units. Subsequent granulite- and amphibolite facies metamorphism have further modified primary features and compositions.

The ternary system of Davis et al. (1978) enables alteration effects to be recognized on the basis of major element data (Fig. 15 b). If the three most deviant samples are excluded, then the mafic metavolcanites of the Pielavesi area can be classified according to their tectonic setting using the criteria presented by Davis et al. (1978).

The Savijärvi Suite volcanics in particular show a great range in chemical composition, which can be attributed to the abundance of tuffaceous material. It is nevertheless evident that there has been considerable element mobility during alteration, with some samples enriched in Na, falling within the spilitic field on the diagram, while conversely, others are enriched in potassium. On the MgO/10 - CaO/Al_2O_3 - SiO_2 diagram, which represents the extent of metamorphic alteration (Davis et al., 1978), it is apparent that many samples have relatively high CaO/Al_2O_3 ratios, reflected in the development of calc-silicate assemblages during metamorphism.

The metavolcanics from the Säviä Suite are relatively unaltered, with only slight variations in Na and K contents, the samples having been selected from the best preserved parts of the schist belt, where there was no evidence of mineralization.

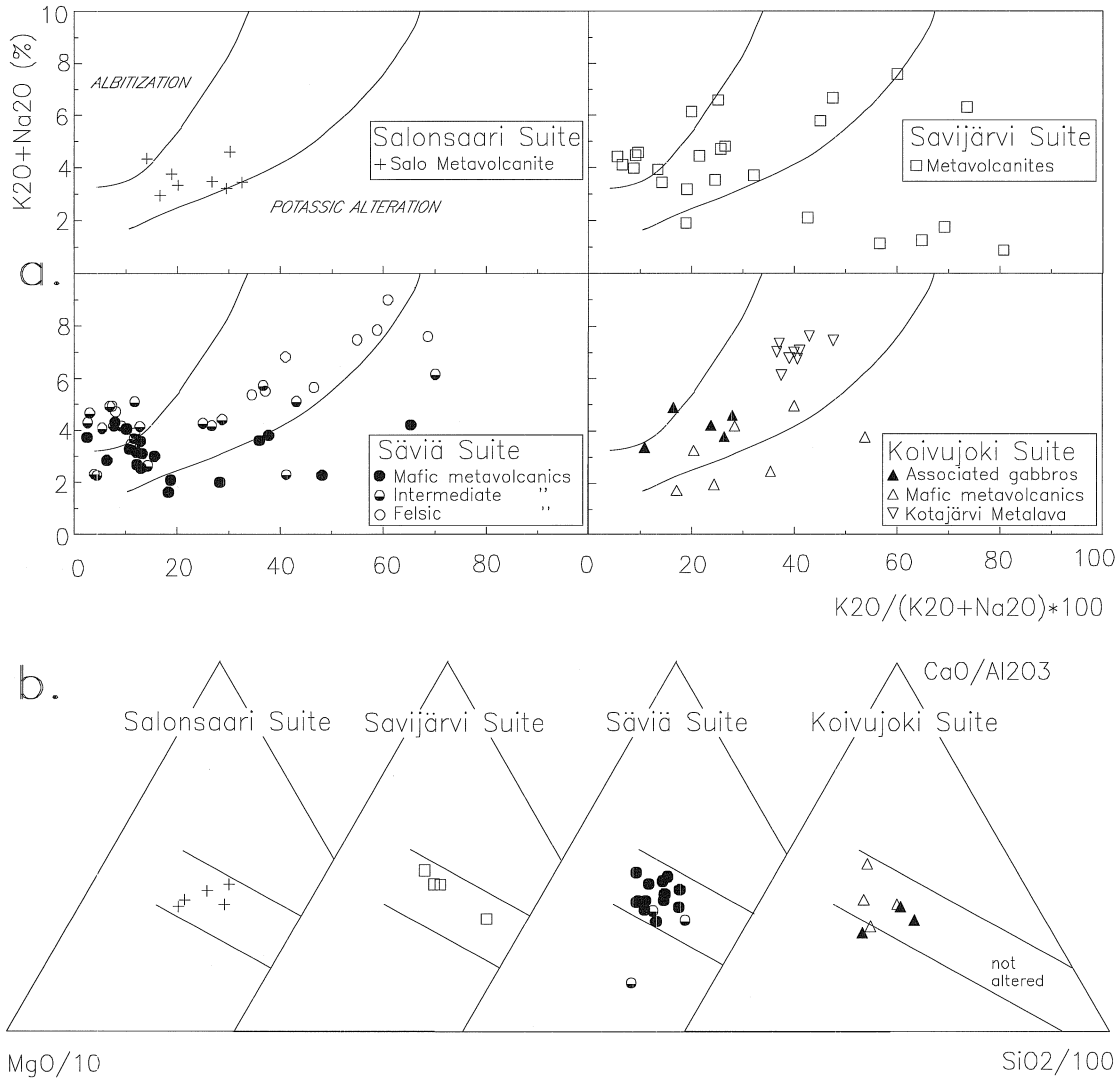


Fig. 15. Degree of alteration of Pielavesi metavolcanics shown on the (a) Hughes (1972) igneous spectrum diagram and (b) basaltic alteration triangular plot of Davis et al. (1978).

Geochemistry of volcanogenic rocks

A general appraisal of the geochemistry of the Pielavesi metavolcanics has been carried out using the AFM triangular diagram (Irvine and Baragar, 1971) and the Jensen cation diagram (Jensen, 1976), (Fig. 16 a,b) and Na_2O+K_2O/SiO_2 diagrams of Le Maitre (1984), (TAS, Fig. 17 a) and Kuno (1966), (Fig. 17 b). The chemical characteristics of the various suites have been portrayed by means of the Fe_2O_3(total)/MgO diagram (Fig. 18) of Pharoah and Pearce (1984) as well as the K_2O/SiO_2 diagram (Fig. 19) of Peccerillo and Taylor (1975). The volcanism was clearly bimodal, with the mafic rocks invariably following the

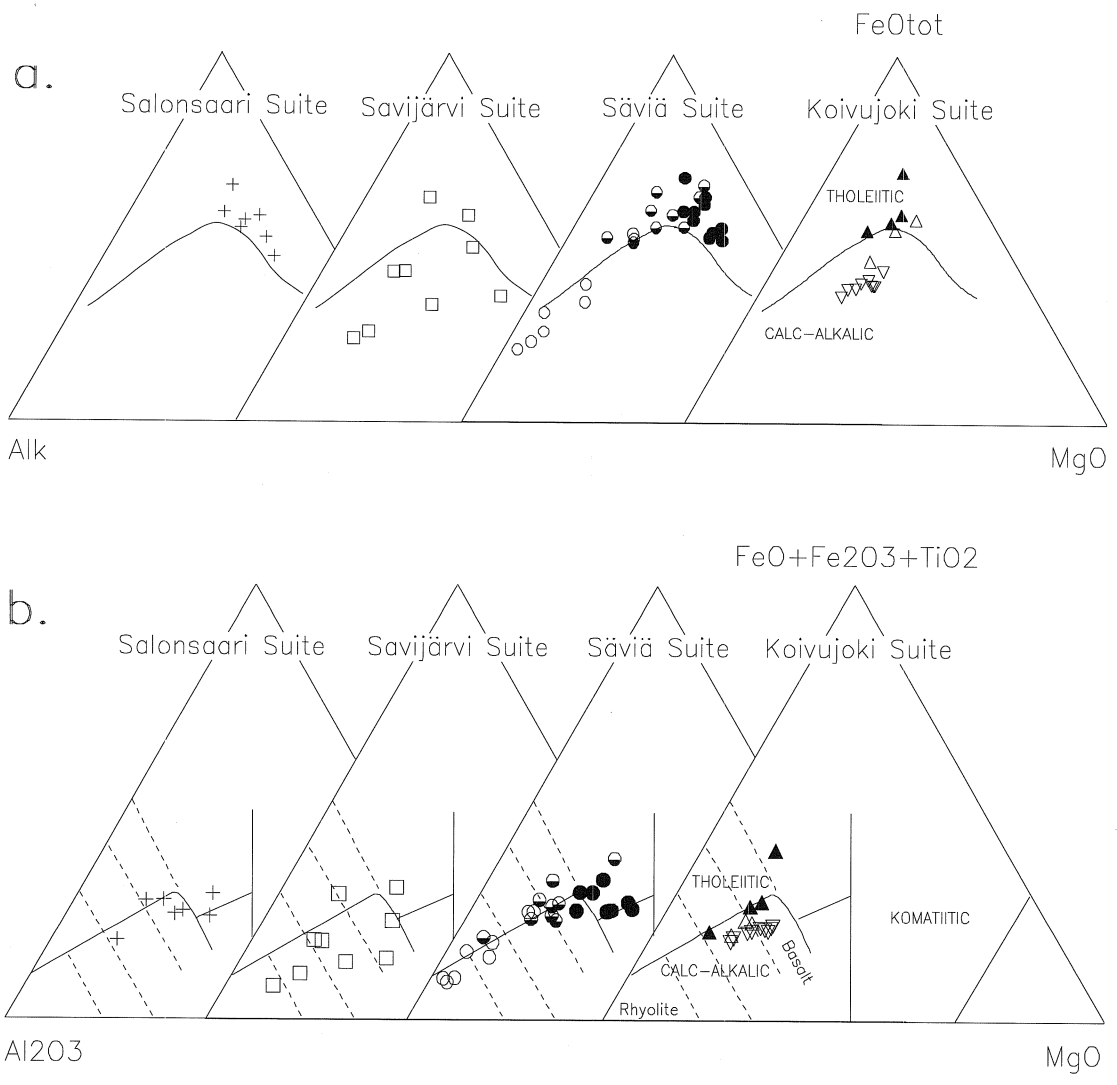


Fig. 16. Geochemical characteristics of Pielavesi metavolcanics on (a) AFM diagram after Irvine and Baragar (1971) and (b) the Jensen cation plot (Jensen, 1976). Symbols as in Figure 15.

tholeiitic trend. The felsic volcanics and some of the intermediate lithologies on the other hand are calc-alkaline in composition.

The plate tectonic setting of volcanic rocks is to some extent reflected in their compositions, such as the abundance of SiO_2 and variations in the ratios of Ti, Zr, Hf, P, K, Sr, Y, Cr, Ni, Th, Nb and REE. Mid-ocean ridge basalts (MORB) are for instance, typically subalkaline and low in K, Cs, Rb, Ba, Th, U and light REE whereas ocean island basalts (OIB) are characteristically subalkaline to alkaline and richer in Ti, Zr, Nb, Ta and the light REE. Volcanism in magmatic arcs and at convergent continental margins is characterized by the association basalt-andesite-dacite-rhyolite (Pearce and Cann, 1973; Wyllie, 1981; Pearce, 1982; Holm, 1982, 1985; Gill, 1981; Leeman,

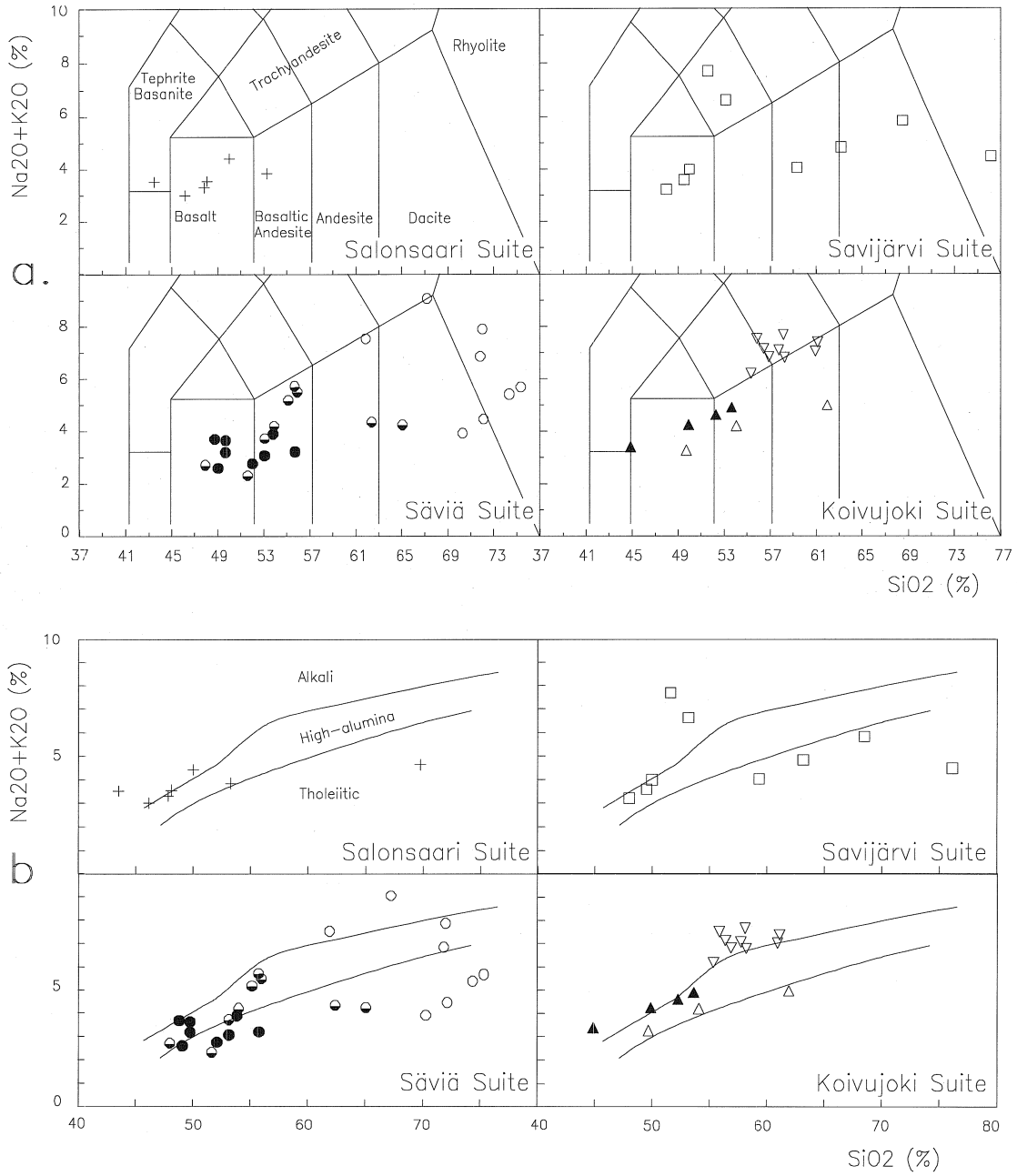


Fig. 17. Geochemical characteristics of Pielavesi metavolcanics on (a) the TAS diagram (Le Maitre, 1984) and the (b) $(Na_2O+K_2O)/SiO_2$ diagram of Kuno (1966). Symbols as in Figure 15.

1983).

The analysis of the tectonomagmatic affinities of the Pielavesi volcanics is based on the samples that were unaltered according to the

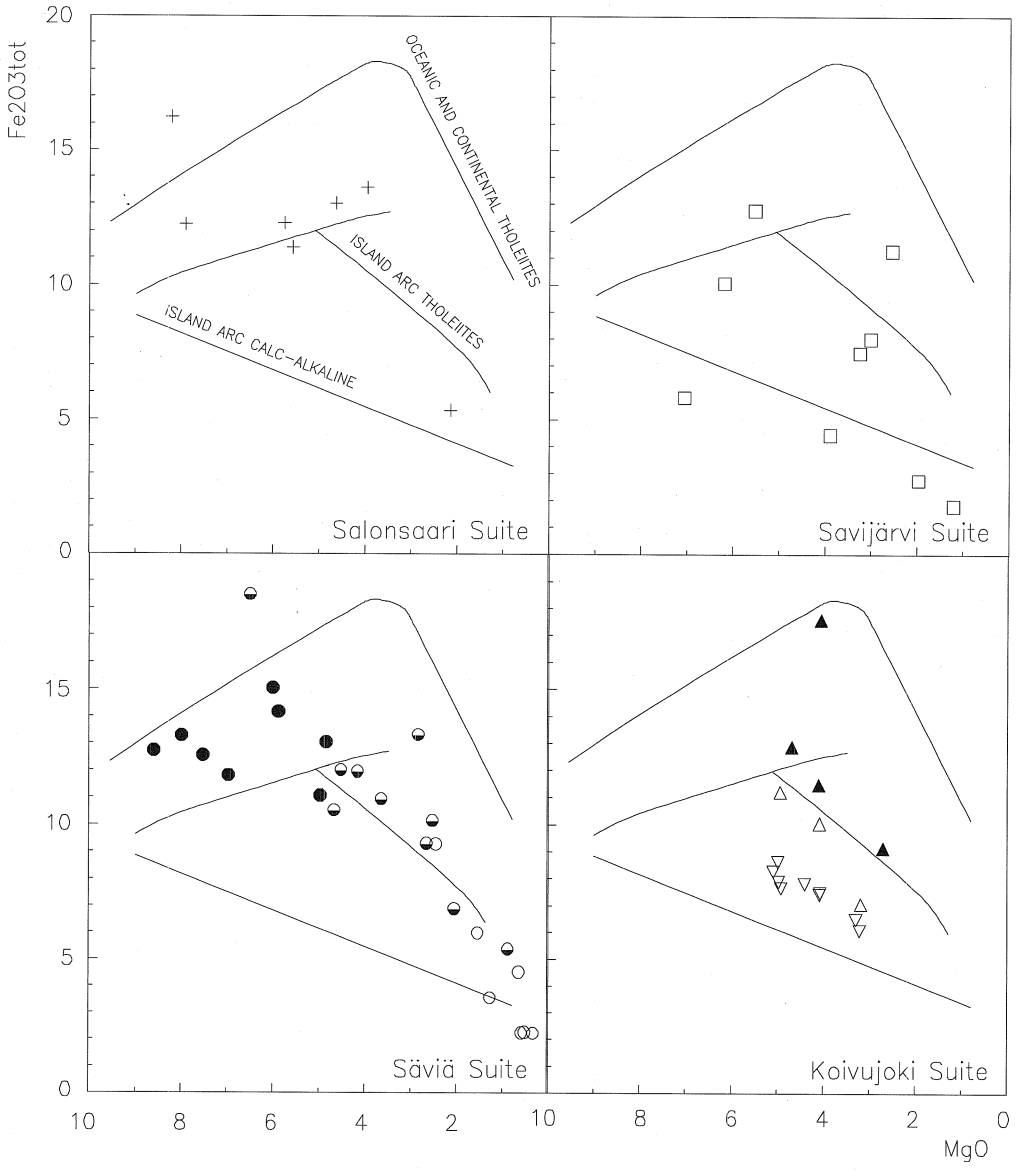


Fig. 18. Classification of Pielavesi metavolcanics according to Pharaoh and Pearce (1984). Symbols as in Figure 15.

criteria of Davis et al. (1978). The high field strength elements Ti, Zr, Y and Nb are not usually transported effectively in aqueous fluids, except perhaps where the activity of certain complexing agents such as fluorine is high, and therefore tend to remain unaffected in rocks which have been subjected to metasomatic alteration processes. According to Pearce and

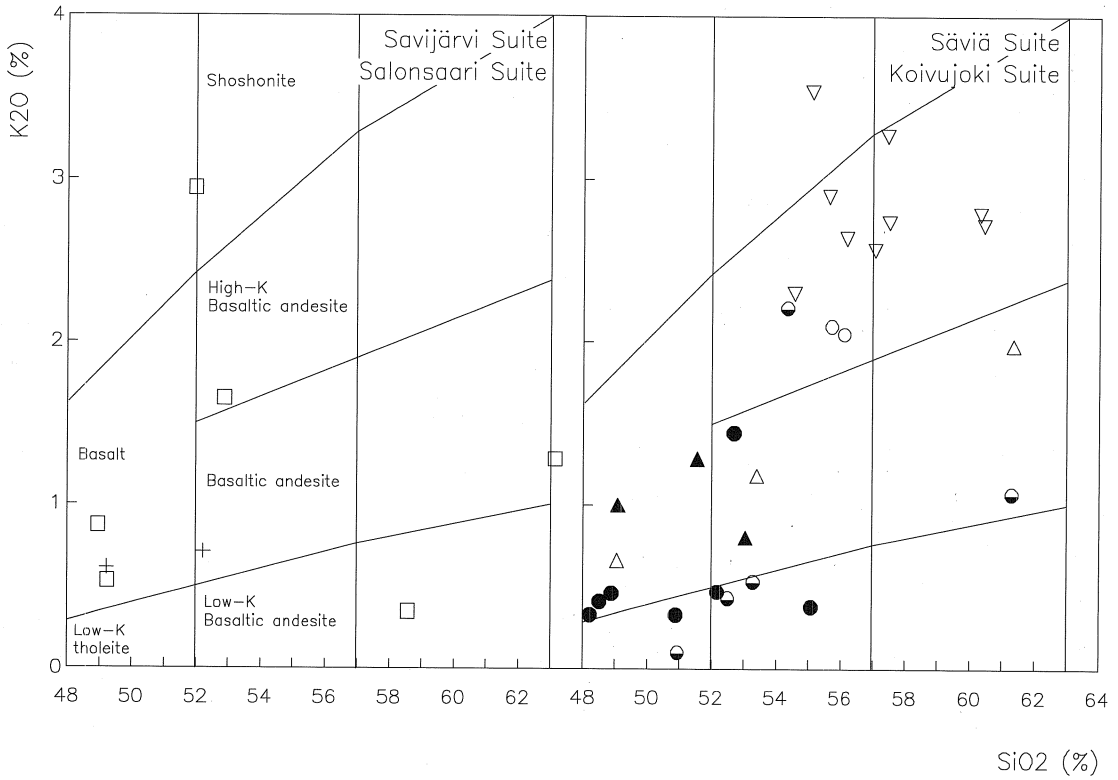


Fig. 19. K_2O/SiO_2 diagram (Peccerillo and Taylor, 1975) for Pielavesi metavolcanics. Symbols as in Figure 15.

Norry (1979), these elements are valuable in studying metamorphosed basalts; most basalts erupted in within plate settings can, for example, be recognized by their high Ti/Y and Zr/Y ratios (Pearce and Gale, 1977). Likewise, most basalts erupted in island arc settings can be distinguished from MORB by their lower absolute abundances of Ti, Zr and Nb (Pearce, 1975).

Salonsaari Suite

The volcanics of the Salonsaari Suite correspond compositionally to tholeiitic basalts (Figs. 16 and 17), with alkali contents within range of the median composition defined by Kuno (1966). On the basis of their potassium contents, they fall close to the low-K tholeiitic (LKT) field in Figure 19. According to the Pharaoh and Pearce (1984) Fe_2O_3 (total)/MgO covariation diagram, the Salo Metavolcanite represents island arc tholeiitic (IAT) magmatism, an interpretation supported by the major component diagrams as well. On the basis of the immobile elements Zr, Ti and Y, however, the Salo Metavolcanites lie within the low-K tholeiite and calc-alkali basalt fields (Fig. 21). The basalts thus appear more indicative of a plate margin rather than intraplate environment. The proportion of Sr in the Zr-Ti-Sr diagrams is relatively variable and generally high, with some samples falling outside the appropriate field. This suggests some degree of alteration and therefore diminishes the reliability of these samples as discriminators of tectonic setting.

Savijärvi Suite

With the exception of two samples, the Savijärvi Suite metavolcanics are calc-alkaline in nature (Figs. 16 and 17). On the TAS diagram (Fig. 17 a), the rocks lie along the subalkaline trend, whereas on the MnOx10-TiO_2-P_2O_5x10 diagram (Fig. 20 a), they plot principally within the island arc tholeiite field, towards the P-rich end. High phosphorus contents are possibly a result of volcanism in association with shallow water sedimentation characterized by carbonates and other chemical deposits. One of the Savijärvi samples actually falls within the island arc alkali basalt field but on the Fe_2O_3(total) versus MgO fractionation diagram (Fig. 18 a), the rocks broadly speaking show island arc affinities. The marked variations in composition are attributed to their mixed volcano-sedimentary origins and depositional environment. The low Zr contents and Zr/Y ratio of volcanic arc basalts results from a high degree of partial melting of a depleted source (Pearce and Norry, 1979).

Trace element abundances of the Savijärvi Suite indicate that the metavolcanics do not represent ocean floor basalts or within plate basalts (Figs. 21, a,b,c and 22 a,b,c). Rather, on the basis of Zr, Ti, Y and Sr characteristics, they show a distinct volcanic arc trend and can mostly be classified as plate margin lavas. One sample, because of its high TiO_2 content and another, having a high Zr/Y ratio plot in the within plate field. When the rocks are compared on the Zr/Y-Zr diagram, a number of samples appear to be within plate lavas, as a result of their relatively high Zr abundances.

Säviä Suite

On the basis of major element compositions, the Säviä metavolcanics fall into distinct groups (Fig. 16 a); mafic and intermediate lithologies represent basalts and andesites having a tholeiitic fractionation trend, whereas the felsic volcanics are principally dacites and rhyolites of calc-alkaline affinity. On the TAS diagram however, all samples lie within the subalkaline field (Fig. 17 a). There are relatively few samples of andesitic composition, in comparison to basaltic or more felsic lithologies. Manganese, Ti and P abundances of the Säviä basalts are characteristic of island arc tholeiites, although two samples fall within the calc-alkaline field in Figure 21 a and a further sample, because of its high Ti content, plots within the MORB field. On the basis of their SiO_2 and K_2O abundances (Fig. 19), the rocks may also be defined as low-K tholeiites.

When classified according to Zr, Ti, Y and Sr abundances, the Säviä basalts appear to be island arc tholeiites with a distinct island arc compositional trend (Fig. 20 a,b,c and 21 a,b,c). The Zr-Ti/100 versus Sr/2 diagram reveals a low-K tholeiitic trend, although the wide scatter represents Sr mobility during hydrothermal alteration. With only two exceptions the basalts are consistent with a plate margin setting.

Koivujoki Suite

The Koivujoki Suite consists of two distinct groups, the more mafic of which includes gabbros and belongs to the tholeiitic trend, in contrast to the more felsic group, comprising the Kotajärvi Metalava and other mafic metavolcanics, which exhibit calc-alkaline character. Discrimination using trace elements has only been attempted for mafic samples that have not been demonstrably hydrothermally altered. On the TAS diagram (Fig. 17 a) it is evident that the more gabbroic lithologies lie approximately within the basaltic field, whereas the Kotajärvi Metalava falls in the trachyandesite field, clearly indicating alkali enrichment. On the K_2O/SiO_2 diagram (Fig. 19), the rocks fall within the high-K field and on the basis of Mn, Ti and P abundances, they could have formed in an island arc milieu. The same inference may be drawn from the Fe_2O_3(total)/MgO diagram (Fig. 18) and the AFM diagram (Fig. 17 a), with the gabbroic lithologies having island arc tho-

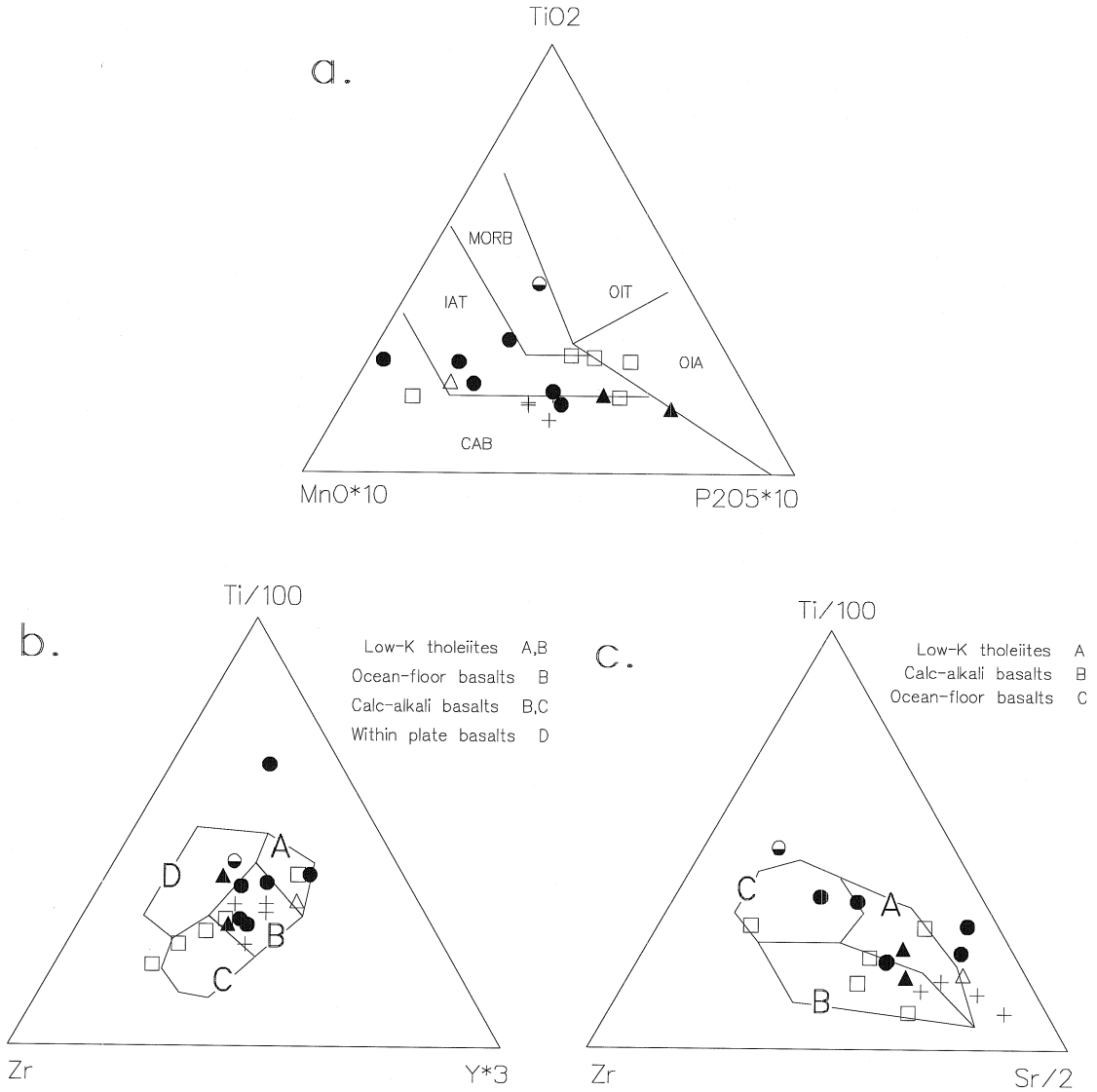


Fig. 20. Tectonic setting of Pielavesi metavolcanics according to discrimination diagrams of (a) Mullen (1982) and (b and c) Pearce and Cann (1973). Symbols as in Figure 15.

leiitic trends and the Kotajärvi Metalava showing a calc-alkaline trend.

The metavolcanites associated with the Koivujoki Suite mica gneisses are subalkaline and tend to follow the calc-alkaline trend, although a large number of samples were excluded because of their high degree of Ca enrichment; no other evidence for alteration was, however, found. An island arc origin is suggested by the Mn, Ti and P abundances in Figure 21, as well as by the Fe_2O_3(total)/MgO diagram (Fig. 18), which more specifically indicates an arc tholeiite trend. The potassium abundances of the unaltered samples are typical for basalts.

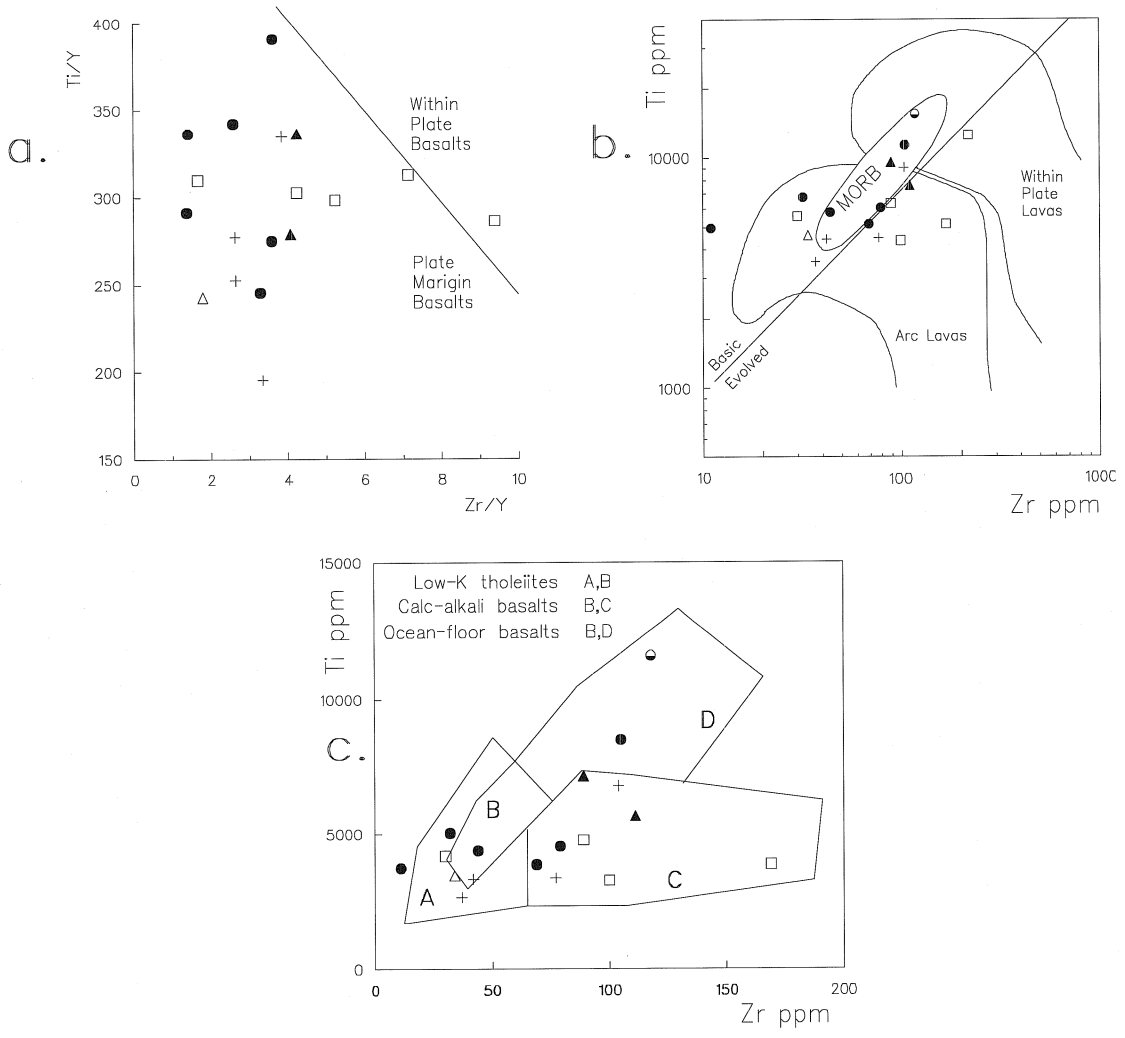


Fig. 21.Tectonomagmatic setting of Pielavesi metavolcanics according to diagrams of (a) Pearce and Gale (1977), (b) Pearce (1982) and (c) Pearce and Cann (1973). Symbols as in Figure 15.

According to their Zr, Ti, Y and Sr abundances (Figs. 20 b,c and 21 a,b,c) both of the above groups have the characteristics of island arc and plate margin basalts. The Zr/Y-Zr diagram is nevertheless more ambiguous and suggests a within plate setting. Pearce and Cann (1973) proposed that volcanic arc basalts form a continuum which is gradational from low-potassium tholeiites through calc-alkaline compositions to shoshonites, with the tholeiites representing earlier stages of evolution near to the trench. Tholeiitic and alkali basalts on the other hand are also widespread in ocean floor and continental settings.

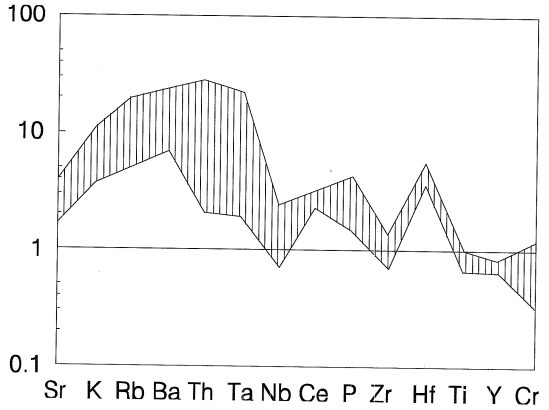


Fig. 22. MORB-normalized trace element data for Pielavesi metavolcanics (Pearce, 1983).

Discussion

The mafic volcanics of the Säviä Suite were erupted within a carbonate-dominated sedimentary setting. Submarine conditions are indicated by the presence of pillow lavas in many places. Banded mafic lithologies evidently represent deformed lavas and pyroclastic deposits. At Säviä the mafic volcanics follow a highly variable felsic and intermediate succession in which the proportion of felsic material is less than that in the Lampinsaari and Ruotanen zones.

The TiO_2, Ni and Cr abundances (App. 1 and 2)of the Pielavesi basalts are rather typical of island arc associations as also the basalts at Pyhäsalmi, Kangasjärvi, Karttula, Rautalampi and Virtasalmi (Mäki, 1986; Rasilainen, 1991; Fagerström, 1990; Lahtinen, 1988; Suvanto, 1983). The basalts at Pielavesi are not particularly primitive, although it is difficult to draw definitive conclusions due to the effects of metasomatism and metamorphism. Presumably these processes would affect the MgO contents of the rocks, which is one of the more sensitive indicators of the degree of evolution of mafic magmas.

On the basis of MORB-normalized trace element data, the Pielavesi basalts most closely show affinities with active plate margin volcanism (Fig. 22). The Pielavesi basalts have enrichments in Sr, K, Rb, Ba, Th, Ce and P, which are also characteristic of subduction related processes (Pearce, 1982). Elevated Ta and Nb values may reflect derivation from a mantle magma that was enriched in incompatible elements; this however is a feature more typical of within plate basalts.

The Salo metavolcanites have higher Al_2O_3, Na_2O, K_2O and P_2O_5 abundances than those of the 2.2 Ga, 2.1 Ga and 1.97 Ga mafic rocks in North Karelia (Vuollo et al., 1992). Conversely, Fe_2O_3 and TiO_2 contents are lower than in the 2.1 Ga Fe-tholeiites which are considered to have been erupted and intruded in a continental setting. The lowermost metavolcanic units at Pielavesi differ geochemically from all other lower Proterozoic rocks in eastern Finland.

The Jatulian volcanism that followed the deposition of the Marine Jatulian in the Kuopio and Siilinjärvi district represents more primitive mantle-derived melts than the Säviä volcanics; according to Lukkarinen (1990) they were erupted in a cratonic rift or rifted continental margin environment. The TiO_2, and Fe_2O_3(total) contents of the Säviä Suite basalts

are lower than those of mafic metavolcanites in the Vaivainen Formation at Kuopio as well as in those of the Parkkila Member at Siilinjärvi (Kousa, 1986). The TiO_2, Al_2O_3 and Na_2O contents of the Säviä mafic volcanics are likewise lower than those of the Marine Jatulian Ottola Formation (Pekkarinen and Lukkarinen, 1991). The Fe_2O_3 contents of the Säviä volcanics are somewhat higher than those of the Rahasmäki Formation 1.97 Ga at Nilsiä (Paavola, 1984). The mafic lavas at Säviä (particularly samples 24, 26, 27 and 30 in App. 1) have very similar TiO_2, Fe_2O_3(total), K_2O and P_2O_5 abundances to those of the 1.97 Ga tholeiites in North Karelia (Vuollo et al., 1992).

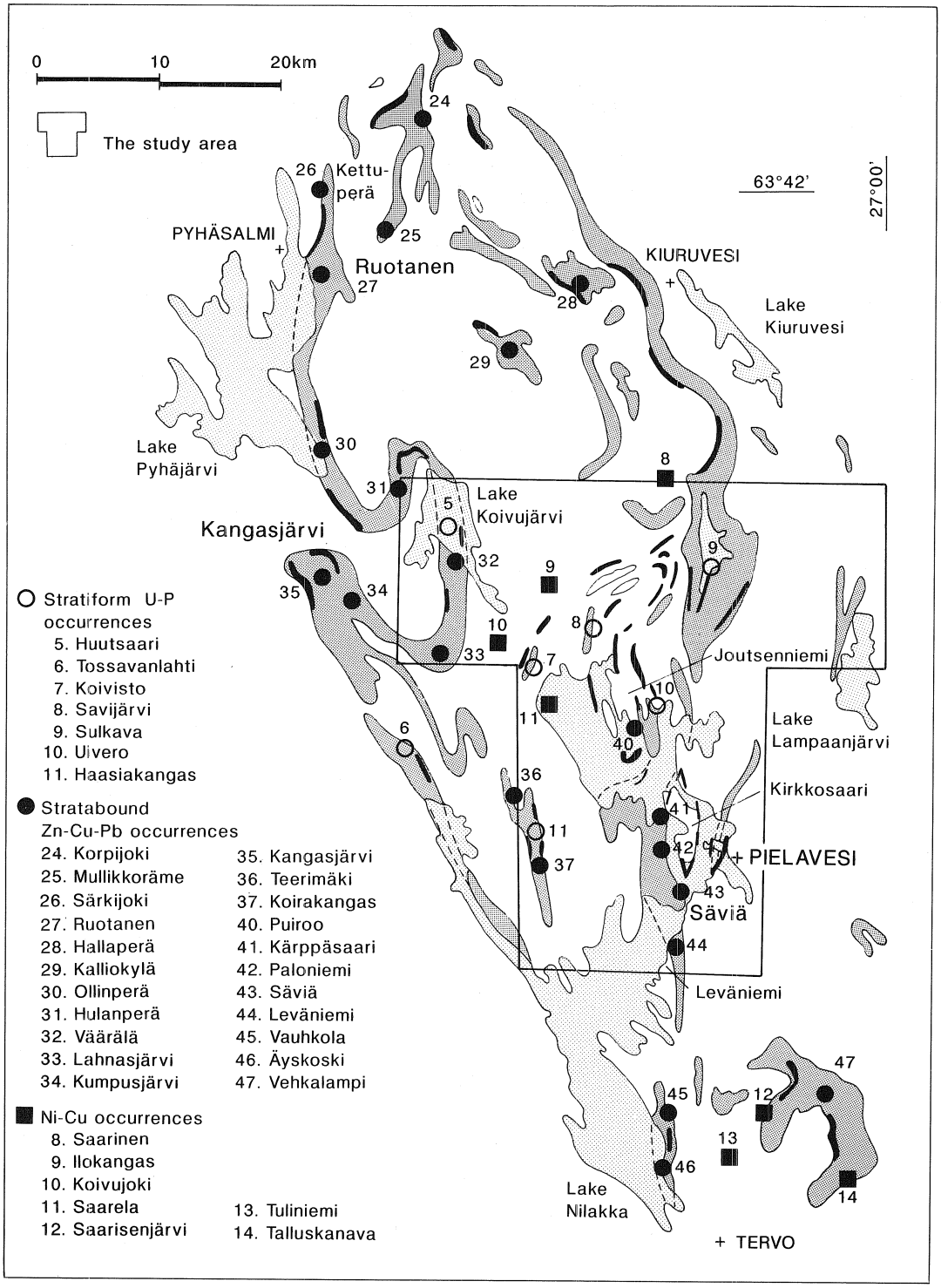
MINERALIZATION IN THE PIELAVESI DISTRICT

Three separate and significant mineralizing epochs can be recognized in the Pielavesi district, each of which corresponds to a distinct phase in the regional geological evolution (Fig. 23). First to form were the stratiform U-P prospects within the Savijärvi Suite. Evidently uranium and phosphorus deposits formed in separate basins along the entire continental margin and do not necessarily represent a contiguous lithological horizon. In many places iron formations are spatially associated with these deposits but in the Pielavesi district they are not widely developed. The second, and economically most significant, stage of mineralization immediately succeeded the previous one and was related to island arc volcanism and associated hydrothermal alteration processes; it too, was essentially stratabound in nature. The third phase of mineralization is represented by synorogenic mafic intrusions and sills which are prospective for Ni and Cu. The various ore occurrences within the study area will be treated below within the context of overall geological developments, with emphasis being placed upon their geological charateristics. Comparisons will also be made with similar deposits outside the Pielavesi area. The numbers given to individual prospects in the text refer to their location in Figure 23.

Stratiform U-P deposits.

The precipitation of phosphate from sea water takes place under conditions that are also favourable for the formation of other chemical sediments, including cherts, carbonates, and may also be accompanied by black shales and clastic deposits (Sheldon, 1984). Marine phosphorus is ultimately of terrestrial origin and its abundance increases as a function of depth. The most favourable conditions for the formation of phosphate deposits involve a continuous supply of organically derived phosphorus and the upwelling of phosphorus-enriched waters into oxygenated shelf environments, where solubility decreases with increasing temperature, pH and salinity (Gulbrandsen, 1969). Phosphate is typically deposited within extensive basins up to hundreds of kilometers in extent, on continental shelves and the upper parts of continental slopes, generally at depths of less than 300 m (Sheldon, 1981). Due to sensitivity to changes in the above-mentioned parameters, consider-ably lateral and vertical variations in thickness of deposits occurs.

Phosphorites are traditionally defined as


0 10 20km
The study area
63°42'
27°00'
PYHÄSALMI
Kettu-
perä
Ruotanen
KIURUVESI
Lake
Kiuruvesi
Lake
Pyhäjärvi
Kangasjärvi
Lake
Koivujärvi
Joutsenniemi
Lake
Lampaanjärvi
Kirkkosaari
PIELAVESI
Säviä
Leväniemi
Lake
Nilakka
TERVO
Stratiform U-P occurrences
5. Huutsaari
6. Tossavanlahti
7. Koivisto
8. Savijärvi
9. Sulkava
10. Uivero
11. Haasiakangas
Stratabound Zn-Cu-Pb occurrences
24. Korpijoki
25. Mullikkoräme
26. Särkijoki
27. Ruotanen
28. Hallaperä
29. Kalliokylä
30. Ollinperä
31. Hulanperä
32. Väärälä
33. Lahnasjärvi
34. Kumpusjärvi
35. Kangasjärvi
36. Teerimäki
37. Koirakangas
40. Puiroo
41. Kärppäsaari
42. Paloniemi
43. Säviä
44. Leväniemi
45. Vauhkola
46. Äyskoski
47. Vehkalampi
Ni-Cu occurrences
8. Saarinen
9. Ilokangas
10. Koivujoki
11. Saarela
12. Saarisenjärvi
13. Tuliniemi
14. Talluskanava

consisting of at least 50% apatite, which corresponds to around 20% P_2O_5 (Pettijohn, 1975). In practice however, the amount of phosphate is often lower, and Äikäs (1988 b) has applied the term phosphorite to deposits in Finland containing in excess of 5% P_2O_5, while describing rocks with anomalous phospate (>0.3% P_2O_5) as phosphatic.

Proterozoic uranium-bearing phosphate deposits in Finland are characterized by U contents of 0.03-0.04% and P_2O_5 contents between 3-5%. Uranium shows strong correlation with fluorapatite (Rehtijärvi et al., 1979; Rehtijärvi, 1983). Apatite occurs as disseminations, streaks and bands and as thin metaphosphorite intercalations and lenses. Uranium is a characteristic element in marine phosphorites and is probably incorporated with phosphorus during precipitation (Aikäs, 1988 b). The uranium-bearing minerals include apatite, uraninite, sphene and uranium-thucolite. Uranium abundances of Proterozoic phosphorites are appreciably less than those of Cambrian and younger deposits (Rehtijärvi, 1983). Uranium-phosphorus deposits are typically layered and stratabound, with individual phosphoritic horizons generally being only 0.5-10 cm in thickness, even though they may belong to phosphatic sequences up to hundreds of meters thick. Graphite and sulphides are usually present in Finnish phosphatic deposits but the association is characterized by dolomite/limestone, calc-silicate rocks, cherts, black schists, felsic volcanics and quartzites (Äikäs, 1988 b).

Type occurrence

Savijärvi

The most significant U-P occurrence in the study area, at Savijärvi (locality 8 in Fig. 23), is situated on the northwestern limb of the Joutsenniemi dome, within an axial depression. Mineralized lithologies include phosphatic calc-silicate gneisses, carbonates, quartz-feldspar gneisses and metaphosphorites. The mean P_2O_5 content of the rocks analyzed from Savijärvi is 1.39%, while U contents are around 90 ppm, giving a P_2O_5/U ratio of about 208 (Äikäs, 1988 f). The highest P_2O_5 value measured is 12%, so that true phosphorites are also present, even though they only form sporadic layers 0.5-1 cm thick, or lenses up to 2 cm thick in carbonates. The metaphosphorites are dark gray and fine-grained, consisting of apatite, carbonate, calc-silicate minerals and graphite. Other phosphatic lithologies contain bands and disseminations of apatite. Apatite tends to follow lithological banding whereas calc-silicate minerals and calcite may truncate it (Äikäs 1988 f). Apatite is also commonly contains fractured, which are filled with uraninite and sulfides. Uranium occurs within uraninite, which is present within the apatite-rich horizons. Microscopically the Savijärvi uraninite is indistinguishable form that at the Vihanti Lampinsaari deposit, in being subhedral, homogeneous, unaltered and somewhat oval shaped in cross section (Äikäs, 1988 f). Many grains are however euhedral and are associated with pyrrhotite, pyrite, galena, molybdenite and graphite.

Other occurrences

The Säviä zone contains further U-P occurrences at Koivisto, Uivero and Sulkavanjärvi (localities 7, 10 and 9 respectively in Fig. 23). Some 6 km to the southwest of Savijärvi, at Koivisto, an analogous sequence of uranium-bearing and phosphatic rocks is exposed in a anticlinal structure. Scheelite is also present and the mineralized sequence is overlain by mafic and intermediate volcanic lithologies belonging to the Säviä Suite, and the Kotajärvi

Fig. 23. Location of mineralization within the study area. Other symbols as in Figure 2.

Metalava. The occurrence is underlaying by the migmatitic mica gneisses.

The Savijärvi Suite is exposed along the eastern margin the Joutsenniemi dome at Puiroo, below a distal Zn-occurrence. The same horizon is also present in the Palonen area and in the anticlinal zone between Uiveronlahti and Kalliokylä. The Uivero association overlies mica gneisses and consists of some 30-50 m of dolomitic carbonates, calc-silicate rocks and felsic tuffs, and can be traced along srike for several kilometers; the unit is classified as phosphatic and has a P_2O_5/U ratio of about 100 (Äikäs, 1988 e). The distribution of both U and P, as well as other features of the occurrence, resemble those at Savijärvi. The sequence is overlain by mafic and intermediate volcanic rocks of the Säviä Suite, garnet-cordierite-anthophyllite rocks and the Kotajärvi Metalava, with associated metagabbros containing metavolcanic fragments derived from the Säviä Suite.

In the Ruotanen schist belt, U-P occurrences have been found at Huutsaari, Tossavanlahti and Haasiakangas (localities 5,6 and 11 respectively in Fig. 23). The Huutsaari U-P prospect is associated with calc-silicate rocks, evidently derived from a calcic felsic pyroclastic protolith (Äikäs 1988 d). Uranium-bearing bands usually alternate with layers 1-3 cm thick consisting of tremolite and biotite in addition to quartz and plagioclase. Radioactive horizons contain on average 145 ppm U but very little Th is present. At Huutsaari the P_2O_5/U ratio is about 230 (Äikäs, 1988 d). In the northerly continuation of the Koivujärvi area, at Maaselänlahti, calc-silicate rocks have been found to contain elevated P_2O_5 abundances, while the phosphate content of the whole zone is anomalously high (Mäki, 1987). To the south of Kangasjärvi, at Tossavanlahti, a number of locally derived glacial erratics rich in U and P have been found, which presumably belong to the same stratigraphic horizon as the other occurrences of the Pielavesi district (Äikäs, 1988 a).

A number of targets associated with calc-silicate lithologies were also identified by the Multimethod Mineral Resource Prediction Project (Gaál, 1988), such as Sopenkylä, Kukkomäki, Kirkkosaari and Haasiakangas, with the latter occurrence being classified as belonging to the phosphatic calc-silicate group, owing to its anomalously high phosphorus content. Because of the repetition of the Savijärvi Suite in numerous fold structures, it is to be anticipated that a large number of U-P occurrences remain to be discovered.

Stratabound Zn-Cu-Pb deposits

The majority of Precambrian massive sulfide ores formed during the felsic-dominated stage of volcanic cycles (Sangster and Scott, 1976; Scott, 1988). The deposits are generally restricted to a relatively narrow stratigraphic interval compared with the total duration of volcanic activity. The lowermost part of such sequences is typically dominated by pillow lavas, followed by flows of predominantly andesitic composition and finally, by dacitic to rhyolitic volcanics. The volcanic sequences are usually enclosed by locally derived sediments and the overall setting may represent primitive arcs in the Archean, or more mature arcs in the Phanerozoic, such as for example, the Kuroko deposit in Japan.

The massive sulphide deposits of the Säviä and Ruotanen zones also occur at a distinct level in the upper part of the island arc association. At Pielavesi the sulphide ores are located towards the felsic top of the main volcanic sequence. Mafic volcanic lithologies also underlie the stratabound mineralization in the Ruotanen belt (Table 1) as shown by recent studies at Mullikkoräme some kilometers to the north of the Ruotanen deposit (Luukas, 1992). The

ore forming stage was followed by other mafic flows belonging to the next volcanic cycle (Huhtala, 1979; Luukas 1992). The rhyolites at the Vihanti Lampinsaari deposit are considered to represent the lower part of the preserved stratigraphical sequence. The stratabound deposits of the Pyhäsalmi Island Arc (PIA) zone correspond most closely to the island-arc hosted Kuroko-type in the fourfold classification of Sawkins (1976) which distinguishes Kuroko-, Cyprus-, Besshi- and Sullivan-types based on deposit characteristics, including both composition and tectonic setting. Nevertheless, similarities with the late Archean Abitibi Belt in Canada are also apparent, where the volcanic sequence becomes progressively more felsic upwards (Scott, 1988).

Kuroko-type deposits were previously considered to have formed in relatively shallow (<500 m) water depths but Ohmoto (1983) showed that depths up to or possibly exceeding 2000 m were more reasonable. According to Franklin et al. (1981), Kuroko-type deposits formed at depths of at least 600 m. At Pielavesi, and indeed throughout the whole of the island arc zone, depths are more likely to have been nearer 500 m, given that volcanism took place directly after deposition of the relatively shallow water Savijärvi Suite sediments. The graphitic sediments would have proveded a suitably reduced environment for the precipitiation of sulfides.

The Zn-Cu-Pb deposits of the PIA-zone have been interpreted as the products of submarine hydrothermal systems (Huhtala, 1979; Makkonen, 1981; Rehtijärvi, 1984; Mäki, 1986; Lahtinen, 1988). The main evidence for this being the preservation of sedimentary structures within the stratiform massive part of the ore. Notwithstanding the intense deformation and metamorphism at Lampinsaari, Ruotanen and Säviä, primary layered features are still recognizable. Circulating fluids are capable of transporting metals as chloride complexes together with reduced sulfur (H_2S, HS) to produce massive sulfide ore bodies as long as they are neutral to weakly acid and are of high enough temperature (Scott, 1988). Measured temperatures for the emission of hydrothermal fluids are in the range 275-350° C and chemical and isotopic analyses show that the fluids are marine in origin, although they have been slightly modified by interaction with basalt (Scott, 1988). In the Ruotanen area, the increased MgO/CaO ratio is a clear indicator of the degree of alteration of mafic volcanics (Mäki, 1986), as is the increase in the K_2O/Na_2O ratio of felsic volcanics.

The proximal parts of Phanerozoic ore deposits are characterized by footwall alteration pipes and stringer zones, whereas alteration effects in the hanging wall are usually only observed if another deposit has formed in the overlying strata. A simultaneous upwards and lateral decrease in the Cu/Cu+Zn+Pb ratio is explained by the remobilization of earlier deposited sulfides by continual hydrothermal flow (Franklin et al., 1981). That massive ore deposits and associated garnet-cordierite-anthophyllite assemblages in Finland formed through the action of hydrothermal fluids is also widely accepted (Huhtala, 1979; Helovuori, 1979; Mäki, 1986; Lahtinen, 1988; Mäkelä, 1989). Nevertheless the garnet-cordierite-anthophyllite rocks appear to be the counterparts to alteration assemblages beneath Phanerozoic ore deposits, which are characterized by sericitization, chloritization, various clay minerals, biotite formation, silicification and carbonation (Franklin et al., 1981). The ore deposits and supracrustal rocks of the PIA zone have been strongly deformed so that feeder pipes and stringer zones and primary zoning features have been largely obliterated and are consequently difficult to identify. The massive ores of the zone generally consist of pyrite and pyrrhotite, sphalerite, chalcopyrite and galena, with smaller amouts of gold, silver and barium.

The Paloniemi-Säviä-Leväniemi area represents proximal volcanic activity, as indicated by the locally thick agglomerate-lapilli tuff sequences, pillow lavas, intense alteration and

massive pyritic and pyrrhotite-bearing stratabound ore deposits. More distal sulfur-zinc occurrences are found at Puiroo, some 13 km north of Säviä and at Vauhkola, 15 km to the south. Distal ore processes are characterized by their great lateral extent, the relative lack of hydrothermal alteration in the vertical direction, overall thinness, distinctly banded structure, a marked increase in the Zn/Cu ratio and the intercalation of calc-silicate rocks and graphitic sediments with volcanic lithologies.

In addition to Ruotanen, Kangasjärvi also represents a proximal setting. More distal deposits are present at Kumpusjärvi, Lahnasjärvi and some 10-15 km further south at Teerimäki and Koirakangas.

Proximal-type occurrences

Säviä

The ore deposits in the Paloniemi-Säviä-Leväniemi area (localities 42, 43 and 44 respectively in Fig. 23) occur within hydrothermally altered felsic pyroclastic deposits in the upper parts of the Säviä Suite. The host rocks to the Säviä Cu-Zn ore body are so-called "ore quartzite", cordierite gneiss and quartz-feldspar gneiss (Laitakari A.J., 1968). The "ore quartzite" presumably represents exhalative cherts, which are common in Phanerozoic deposits.

Hydrothermal alteration resulted in the transformation of primary volcanic minerals into an assemblage consisting of chlorite and montmorillonite, which were further recrystallized during metamorphism to give rise to the present garnet-cordierite-anthophyllite association. Ore deposition took place during the same hydrothermal processes, during which metals were liberated from silicates and precipitated as sulfides in proximity to volcanic vents (Makkonen, 1981). The garnet-cordierite-anthophyllite rocks usually contain weak sulfide disseminations and in the Ruotanen area they clearly show higher abundances of Zn, Cu, S, Ba and Ag than unaltered volcanic lithologies (Mäki, 1986).

The Säviä ore body was originally a stratiform massive sheet 3-5 m thick, with underlying disseminated and veined stinger ores. This sheet, along with the enclosing Säviä Suite, was folded and tectonically thickened during isoclinal F_1 phase affected also by tight F_2 folding, so that it is now some 20-30 m thick. Underlaying garnet-cordierite-anthophyllite rocks are located on either side of the tightly folded ore body (Figs. 24 a and b). Separate ore lenses were formed during D_3 deformation, characterized by an intense lineation plunging 40-50° to the southeast, so that there exists a strong tectonic, in addition to primary stratigraphic control. The Säviä deposit has been deformed, refolded and overthrusted towards the northwest, in a manner analogous to the northeastwards translation of the Outokumpu ore bodies (Koistinen, 1981).

The Säviä deposit retains evidence of the compositional zoning characteristic of stratabound mineralization (Table 5). The regular alternation of different ore types, namely compact ore - brecciated ore - disseminated ore - brecciated ore - compact ore - disseminated ore reflects repeated folding of a single sequence.

The pyrite/pyrrhotite ratio of the Säviä copper deposit increases from the stringer ore near the base of the deposit upwards through the massive and disseminated or banded sulfur- and zinc-rich upper parts (Horizon A in Table 5). The banded structure is in fact only discernible in this upper part where coarse pyrite and sphalerite alternate with each other. Individual

Fig. 24. Cross-section through the (a) northern and (b) southern end of the Säviä ore body. The lower part of the profile is schematic, depicting an isoclinal synform, with an originally thin ore horizon having been tectonically thickened during F_1 and F_2 folding.

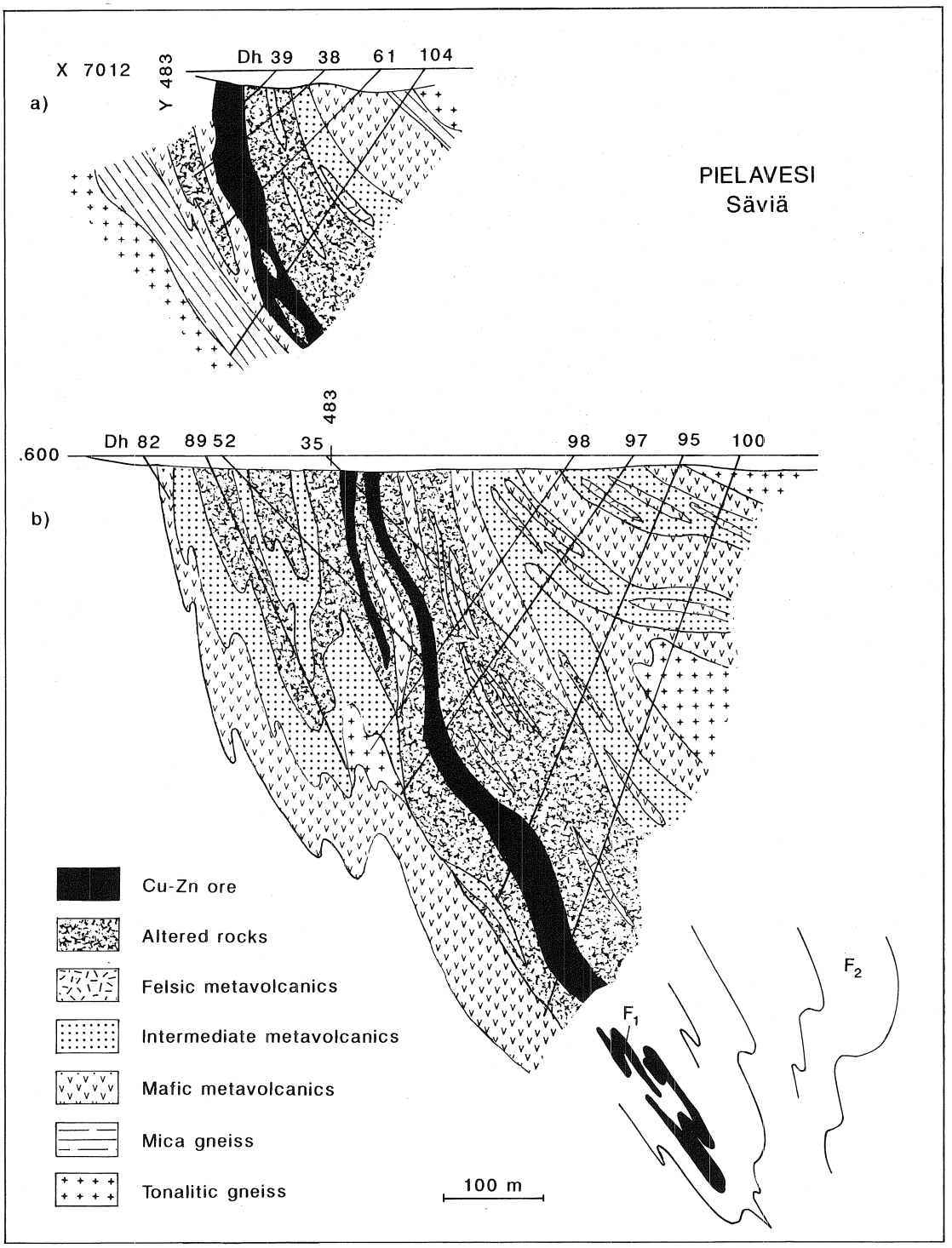

X 7012
Y 483
a)
Dh 39 38 61 104
PIELAVESI
Säviä
483
.600
Dh 82 89 52 35 98 97 95 100
b)
Cu-Zn ore
Altered rocks
Felsic metavolcanics
Intermediate metavolcanics
Mafic metavolcanics
Mica gneiss
Tonalitic gneiss
100 m
F1
F2

Table 5. Morphological and mineralogical zoning through restored primary vertical cross-section of the Säviä ore deposit, based on drillcore data.

Top	A Pyrite-sphalerite (chalcopyrite, galena) - compact banded layers
	B Pyrite-chalcopyrite-pyrrhotite (sphalerite, barite) - compact and brecciated ores (massive ore)
	C Pyrrhotite-chalcopyrite - veined and disseminated ore (stringer ore)

pyritic layers may be up to several meters in thickness and may contain some interstitial pyrrhotite, chalcopyrite and sphalerite. Galena also occurs as sporadic disseminations and veins throughout the pyritic ore. Fine-grained barite and carbonate is also present. This kind of ore is most characteristic of the southern part of the deposit, particularly in the westerly limb of the synformal structure.

The primary massive ore at Säviä consists of compact and brecciated ore (Horizon B in Table 5). Primary features have not been found, due to tectonic transposition and recrystallization. Massive horizons occur to either side of the main ore and are from 2-8 m thick. The ore consists of pyrite, chalcopyrite and pyrrhotite, with occasional sphalerite. Angular fragments of host rock suggest remobilization of ore while country rocks were deforming in a brittle manner. Chalcopyrite increases in abundance downwards, being at a maximum in the massive ore and it is also distinctly enriched where the ore has been brecciated. Moderately elevated Au and Ag contents have been recorded from the main ore body and small amounts of barite, carbonate and anhydrite occur in association with copper ore, the latter mineral sometimes forming crystals up to 2 cm across (Laitakari A.J. 1968).

The veined and disseminated ore, which has been interpreted as stringer ore (Horizon C in Table 5) constitutes the footwall to the massive ore body although, as a result of isoclinal folding, it is now encountered in both the footwall and the hanging wall, as well as within the main ore body itself. The stringer ore consists essentially of a pyrrhotite-chalcopyrite-pyrite paragenesis. The basal parts of the main ore and the footwall are particularly rich in copper disseminations, while the disseminated and veined ore types are also notable for their relatively high Au and Ag contents. Stringer-type disseminated ore is also present adjacent to the Ruotanen deposit (Mäki, 1986) but Pb and Ag are only found in alteration zones outside the main ore body. The Lampinsaari ore body is also zoned, with anomalous Pb, Au, Ag and Ba concentrations above the massive zinc ore.

The Säviä deposit is estimated to contain around 4 Mt of ore, with 1.1% Cu and 28% S. A separate sulfur-zinc ore body contains an additional 1 Mt, with 2.0% Zn and 33.0% S (Laitakari A.J. 1968). The Säviä deposit grades on average, 0.2 gt^{-1} Au and 5-10 gt^{-1} Ag.

Kangasjärvi

The Kangasjärvi deposit (locality 35 in Fig. 23) occurs within the southerly extension of the Ruotanen schist belt and is hosted by a succession of mafic, intermediate and felsic volcanogenic rocks and their altered derivatives. Rasilainen (1991) interpreted the sequence as having formed in a convergent plate margin setting. Country rocks include altered cordierite-sillimanite-garnet-anthophyllite gneisses, cummingtonite-bearing gneisses and calc-silicate rocks. The volcanic sequence is characterized by magnesium enrichment and by the absence of carbonate minerals, indicating a proximal Kuroko-type environment of deposition (Rehtijärvi, 1984).

The Kangasjärvi deposit has also been overthrust towards the northwest and has a highly elongate, flattened profile, plunging towards the southeast at about 45°. The ore is clearly stratabound and pyrite and sphalerite are present as layered bands and massive aggregates, with chalcopyrite and pyrrhotite being rare. Gangue minerals include quartz, barite, muscovite, chlorite and diopside. Galena, gold

and silver has been found at the margins of the ore body, while the abundance of zinc and the Zn/Cu ratio both fall in the lower part of the deposit (Rehtijärvi, 1984). Hydrothermal alteration increases in intensity towards the footwall of the deposit and is characterized by increases in Fe, Mn, Mg, P, Ba, Au, S, Zn and Co concentrations and concomitant decreases in the abundances of Ca, K and Sr (Rasilainen, 1991). The Kangasjärvi deposit is analogous with the Pyhäsalmi ore body in terms of both its metal abundances and its structural position. The deposit was mined in 1986 by Outokumpu, who extracted some 86 kt of ore averaging 5.4% Zn, 0.06% Cu and 38% S, and concentrated it at their Pyhäsalmi plant.

Distal occurrences

Drilling at Puiroo (locality 40 in Fig. 23), within the Säviä zone, intersected a narrow pyrite-sphalerite mineralization which was traced along strike for 1.5 km. The thickness of the horizon nowhere exceeded 2 m, with Zn grading at 1-2%. The occurrence corresponds stratigraphically with the Säviä volcanic sequence, although it represents more sedimentary dominated facies variants. Hydrothermal alteration perpendicular to layering was not observed.

At Vauhkola (locality 45 in Fig. 23), some 15 km south of Säviä, the Säviä zone is still recognizable, though discontinuous, consisting of variably migmatized mica gneisses, amphibolites, calc-silicate rocks, graphite-bearing schists and garnet-cordierite gneisses. Pyrite and pyrrhotite are both present as either disseminations or bands in calc-silicate rocks and quartz-feldspar gneisses. A strongly deformed mineralized zone several hundred meters in length has been identified, with a maximum thickness of 5 m and Zn contents remaining below 2%, with slightly anomalous amounts of Pb and Ag as well. A still smaller mineralization has been located at Äyskoski (locality 46 in Fig. 23), about 2 km south of Vauhkola.

The compact pyrite-pyrrhotite Kumpusjärvi mineralization is only about a meter thick and occurs within the Ruotanen schist belt, some 3 km eastsoutheast of Kangasjärvi (locality 34 in Fig. 23). It is hosted by a calcareous quartz-rock which is part of the association comprising calc-silicate rocks, cordierite gneisses and mafic to intermediate volcanic lithologies.

The Lahnasjärvi and Hirvikangas occurrences (locality 33 in Fig. 23) are situated about 7 km southeast of Kangasjärvi, where the regional stratigraphic sequence may be correlated with that at Koivujärvi. Stratabound zinc-rich horizons are present within the upper parts of felsic volcanic units and host lithologies include calc-silicate rocks, cherts, cordierite- and sillimanite-bearing gneisses and quartz-feldspar gneisses (Nikander, 1981). Mineralizations are rather thin (0.5-1.0 m) and can be traced over distances of hundreds of meters, with Zn contents ranging up to 2-3%. The banded appearance of the mineralization is due to alternations of pyrite and sphalerite, while some chalcopyrite, galena and molybdenite is also present. Gahnite occurs at Lahnasjärvi, as indeed at Kangasjärvi. At the Hirvikangas occurrence traces of Pb, Ag, Au and Co have been found (Nikander,1981). The Väärälänniemi occurrence (locality 32 in Fig. 23) consists of pyrite-pyrrhotite-chalcopyrite mineralization within felsic volcanic rocks.

The discontinuous schist belt can be followed for about 20 km to the southeast of Kangasjärvi, where a similar association of mica gneisses, hornblende gneisses and garnet-cordierite-sillimanite gneisses occurs along the margin of the Teerimäki gabbro intrusion (locality 36 in Fig. 24). Here the calc-silicate rocks also contain apatite and sphalerite-pyrite-barite mineralization was encountered during drilling of cordierite-sillimanite gneiss enclaves within the gabbro intrusion. Traces of Cu, Ag and Au have also been found. The Teerimäki occurrence has been compared with the Ransko complex in Czechoslovakia (Nikander, 1983), where quartz-cordierite-sil-

limanite gneisses intruded and surrounded by gabbro also contain sphalerite-barite ore deposits. The Ransko ore is interpreted as an epigenetic hornfels deposit, where the intrusion of younger quartz diorite magmas into a mafic complex has caused intense progressive alteration and the formation of quartz hornfels (Nemek and Holub, 1980). In the case of Teerimäki however, it appears more probable that mineralization is of a distal exhalative kind, subsequently intruded by the Teerimäki gabbro, so that hydrothermal alteration, ore formation and regional metamorphism all predate gabbro emplacement. Mafic intrusions throughout the district commonly contain inclusions of country rock. The Ransko complex have also been interpreted as resulting from interaction between xenoliths and magma (Watkinson and Mainwaring, 1978).

Distal type occurrence is containing pyrite, pyrrhotite and small amounts of sphalerite, chalcopyrite and galena also present at Koirakangas, to the south of Teerimäki (locality 37 in Fig. 23).

Ni-Cu deposits

The distribution of Svecofennian Ni-deposits has been interpreted as defining a ring surrounding the Central Finland Granitoid Batholith Complex (Häkli, 1971; Piirainen, 1975; Papunen and Vorma, 1985; Papunen, 1986) or alternatively, being related to a NW-SE dextral wrench fault and shear system (Gáal, 1972, 1982, 1985, 1986; Talvitie, 1975,). The Ni-deposits of the Kotalahti belt were considered to have been derived from tholeiitic magmas generated by partial melting of subducted oceanic crust. Their present structural geometry is characterized by dextral transcurrent shear zones with associated drag folds and they were accompanied by medium to high grade metamorphism (Gaál, 1982, 1986).

In the Kotalahti lithological series (Mäkinen, 1987), the first intercumulus phase after olivine and orthopyroxene is clinopyroxene, later followed by plagioclase. Thus, the differentiation series includes peridotites, norites and gabbros and orthopyroxene-bearing gabbros dominate over ultramafic intrusions. According to Mäkinen (1987), the Kotalahti-type magmas were produced by partial melting but were somewhat richer in incompatible elements and crystallized at higher pressures than the Vammala-type magmas. This also implies a lower temperature and lower degree of melting for the parent magma to the Kotalahti type.

Studies in southern Savo suggest that either lowering of temperature or contamination by and assimilation of country rock were most significant in causing separation of a sulphide melt phase (Makkonen, 1992). Experimental petrological investigations have also shown how the addition of silica can dramatically decrease the solubility of sulphur in a silicate melt (Poulson and Ohmoto, 1990). According to Naldrett (1989), sulfide separation may take place by batch equilibration, if the solubility limit of sulphur is simultaneously exceeded throughout the whole magma, in which case sulfides equilibrate with the silicate melt and sink to the floor of the magma chamber. A further theory invokes fractional segregation, where the relative proportion of sulphur in a sulfide-saturated silicate melt gradually increases, for instance during the crystallization of olivine.

On the other hand, if a magma intrudes sulfide-rich sedimentary country rocks, as was evidently the case in the Kotalahti belt, then the assimilation of such sediments could promote the separation of sulphide phases from the melt (Gaál, 1990; Makkonen, 1992).

The Ni deposits in southern Savo are considered to have formed via batch equilibration at a relatively early stage (Makkonen, 1992). They tend to occur in the basal parts of layered

mafic bodies intruded into predominantly turbiditic sedimentary sequences. The intrusions represent either single stage differentiation series or two-stage pulses of magma. The Ni content of the Juva occurrence is directly proportional to the MgO content of the parent magma, which ranges between 5.5-11.2% Conversely, the Cu/(Cu+Ni) ratio decreases as MgO increases.

As a result of deformation during and after emplacement, the mafic intrusions are typically steeply dipping sheets or pipes. The accumulation and mobilization of sulfides took place during D_2 and particularly D_3 deformation (Makkonen, 1992).

The Laukunkangas Ni deposit is considered to be related to olivine tholeiitic magmas generated by partial melting of the upper mantle (Grundström, 1980). As with the Juva occurrence, the Laukunkangas Ni ore is associated with the basal ultramafic cumulate phase of the intrusion and moreover, the intrusion records two separate magma pulses. Synkinematic tonalite-trondhjemite magmatism accompanied the mineralized mafic intrusions, with abundant bodies of hornblende diorite and hornblende gabbro and the widespread development of schollen migmatites throughout the Kotalahti Nickel Belt (Gaál and Rauhamäki, 1971; Gaál, 1986). Mineralized bodies occur as pipes, plugs and lenses or else are of irregular form, resulting from their synkinematic character.

Two distinct groups of Ni mineralization have been identified in the Pielavesi district, the first of which is represented by layered ultramafic bodies intruding mica gneisses of the Koivujoki Suite. The second group consists of pipe-like ultramafic intrusions and gabbro plutons whose positions have been largely tectonically controlled, either within fault and shear systems, or in other dilational sites generated during folding; the Saarinen serpentinite intrusion in the northern part of the study area is typical of these. The Ni-Cu occurrences at Tuliniemi and Saarisenjärvi in the southern side of the study area also belong to this latter group, whereas the Talluskanava serpentinite is a primary layered intrusion. All mafic and ultramafic intrusions in the Pielavesi area contain a well-developed lineation plunging 40°-60° to the southeast.

Type occurrences

Ilokangas and Koivujoki

A number of ultramafic lenses have been described from the NE-trending Koivujoki zone; all are concordant with, and occur at approximately the same stratigraphic level within the mica gneisses of the Koivujoki Suite, thus being reminiscent of a primary layered intrusion or swarm of intrusions.

The mineralized Ilokangas and Koivujoki lenses (localities 9 and 10 respectively in Fig. 23) are at most 20-30 m thick and can be traced for less than 200 m along strike. Serpentinized dunites are strongly altered to tremolite-, chlorite- and talc-rich rock at the margins of the intrusions, although contacts are sharp and tectonized. Graphite, which was presumably derived form adjacent sediments, is present within the Koivujoki ultramafic rock.

Mineralization at both Ilokangas and Koivujoki is restricted to the margins of the intrusion. The Pielavesi supracrustal sequence is overthrust towards the northwest. Thus the Ni-occurrences are in an overturned position but were originally upward younging, along with the rest of the stratigraphic sequence, which is consistent with the structural interpretation.

The primary disseminated mineralization at Ilokangas (Fig. 25) was intersected by drilling in the hanging wall of the intrusion (=stratigraphic footwall).Pentlandite occurs as discrete grains and as wispy streaks in pyrrhotite, and zoned chromite is also present in small amounts within the disseminated ore.

The surficial exposures of the Ilokangas occurrence show abundant chalcopyrite and pyrrhotite in footwall (= stratigraphic hanging wall) breccia ore. This is clearly epigenetic in

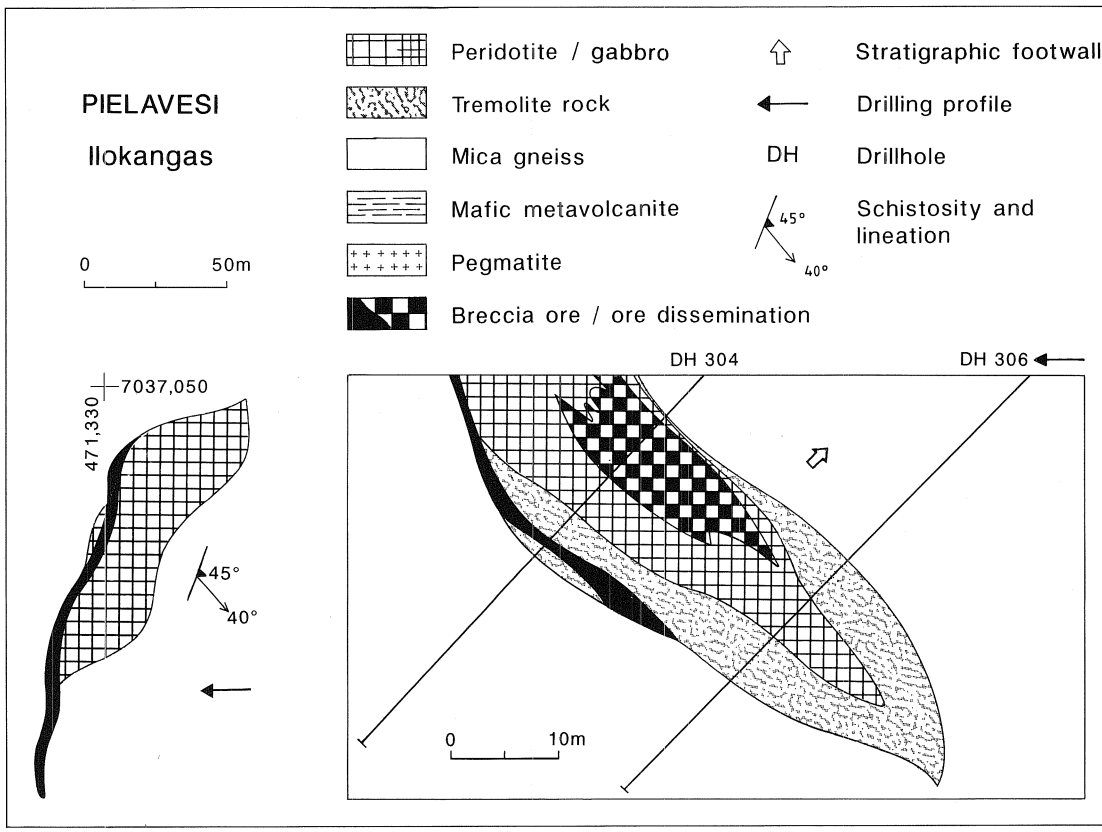


Fig. 25. Cross-section through the Ilokangas ultramafic intrusion in which the primary disseminated ore is dispersed through the stratigraphic footwall. The younger breccia ore is in another contact zone.

origin and was formed at a late stage along the contact (Ekdahl, 1988). This ore consists almost entirely of compact pyrrhotite having Ni contents around 5%. The boudinaged southern part of the ultramafic body is mineralized throughout and tends to contain higher Cu abundances; it is also traversed by Ni-Co-As veins, up to 2-3 cm thick, having the mineral paragenesis nikkolite, gersdorfite, maucherite, pentlandite, pyrrhotite, chalcopyrite, violarite and bravoite. Compact sulfide veins locally carry up to 20.6% Ni, 4.0% Cu, 2.15% Co, 18.6% S, 185 ppm As and 1.9 ppm Au, 0.14 ppm Pt and 5.0 ppm Pd. The Ni-Co-As veins represent the last ore-bearing fluids and also contain anomalously high precious metal concentrations.

The primary disseminated ore at Ilokangas has Ni contents of 0.51% and 0.29% Cu, while the secondary, brecciated ore contains 0.47% Ni and 0.54% Cu. The Cu/(Cu+Ni) ratio of the breccia ore (0.53) also indicates the mobility of Cu-bearing phases with respect to the primary disseminated ore (0.22). The deposit is however uneconomic at present, containing an estimated 20 kt of ore.

The pyrrhotite-rich mineralization associated with the ultramafic layered intrusion at Koivujoki (Fig. 26) is located very near the surface, on the hanging wall (= stratigraphic footwall) of the deposit. The stratigraphic hanging wall contains a 2-5 m thick layer of norite, indicating the differentiated nature of the intrusion. The deposit is estimated to con-

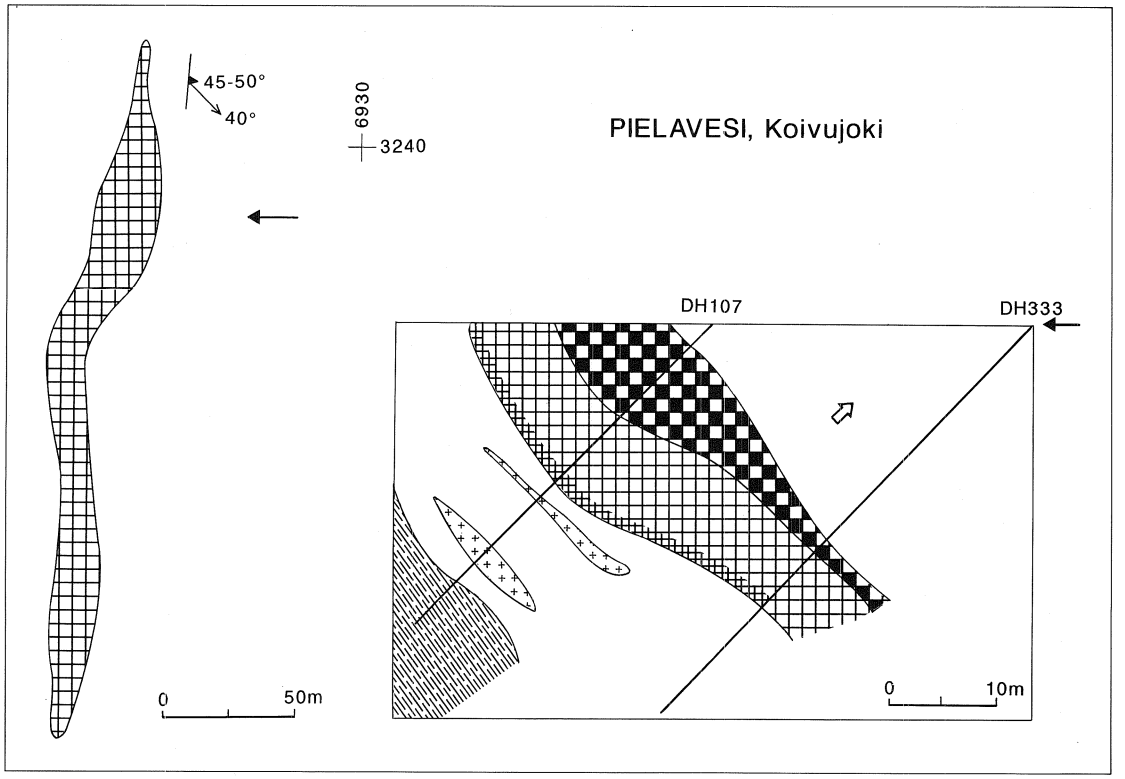


Fig. 26. Section through the Koivujoki ultramafic layered intrusion. Primary disseminated ore is located on the stratigraphic footwall. Other symbols as in Figure 25.

tain some 25 kt of ore with 0.94% Ni and 0.3% Cu and the Cu/(Cu+Ni) ratio of 0.25 corresponds to that of the Ilokangas disseminated ore.

Saarinen

The Saarinen ultramafic body (locality 8 in Fig. 23) is situated towards the northern margin of the study area and is a lenticular intrusion some 400 m in length and about 150 m thick. Clinopyroxene, orthopyroxene and secondary amphibole are present in addition to olivine and plagioclase is also sporadically observed. Olivine has been replaced by serpentine and magnetite. Country rocks to the intrusion are typically migmatitic graphite-bearing mica gneisses, which also occur as fragments within the intrusion. Hybridization effects and pegmatite-like zones occur at the contact.

Along the steeply dipping eastern margin a 20 m thick dissemination of chalcopyrite, pyrrhotite and pentlandite occurs which, on the basis of drillcore data, appears to be discordant to the trend of the intrusion (Fig. 27). The abundance of olivine nevertheless increases towards this eastern margin. The distribution of Cr is also consistent with gravitational crystal fractionation processes, with preferential removal of Cr from the melt during crystallization of orthopyroxene; chromite is accordingly absent. The Saarinen ocurrence is rather small, containing about 75 kt of ore with 0.16% Ni and 0.26% Cu. The Cu/(Cu+Ni) ratio of the deposit is 0.62, which is considerably higher than in the previously discussed layered intru-

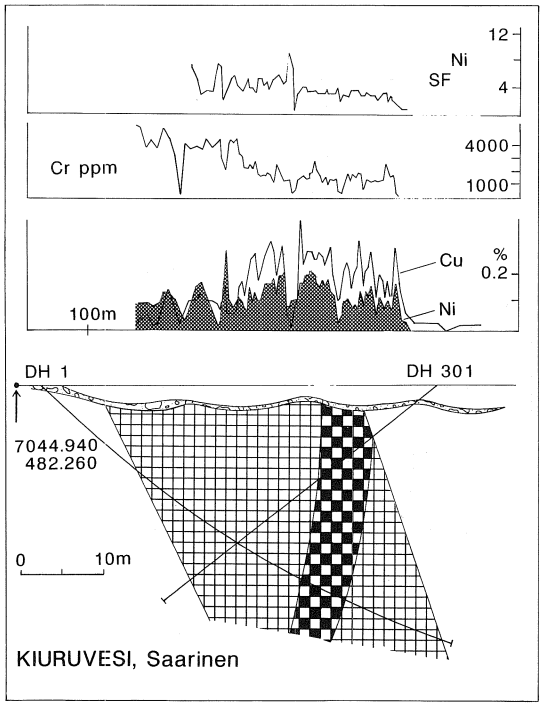


Fig. 27. Section through the Saarinen ultramafic intrusion. Nickel abundances of sulphide phases in the intrusion, along with Cu-, Ni- and Cr-contents are shown in upper part of diagram. Other symbols as in Figure 25.

sions. The Ni content of the sulfide phases varies between 2-4% and increases slightly form the margins towards the center of the intrusion.

Other occurrences

The Saarela pluton (locality 11 in Fig. 23) consists principally of norite and olivine gabbro, with hornblende gabbro occurring at the margins. Ultramafic lithologies are evidently present as well, since Ni- and Cu-bearing serpentinite erratics have been found on certain islands in Lake Pielavesi; one of these boulders also contains anomalous PGE concentrations, namely 0.31 ppm Pt and 1.90 ppm Pd. Because of the amount and depth of water however, follow-up investigations have not been practicable.

Three Ni occurrences are known in the Tervo area, in the southern side of the study area, namely Tuliniemi, Saarisenjärvi and Talluskanava (localities 13, 12 and 14 respectively in Fig. 23). The noritic lens at Tuliniemi is situated within a N-trending zone of migmatitic graphite-bearing gneisses and is apparently concordant and elongate parallel to the dominant regional fabric lineation. The intrusion contains three parallel sulphide-rich horizons with Ni contents of 5-6%. The occurrence contains a total of around 90 kt of ore averaging 0.34% Ni and 0.15% Cu, with a Cu/(Cu+Ni) ratio of 0.31.

The Saarisenjärvi body is a tectonically controlled multiple intrusion consisting of peridotites and pyroxenites, olivine gabbros and hornblende-cummingtonite gabbros (Ekdahl, 1974). Weak Ni and Cu sulphide disseminations are present within the ultramafic parts of the intrusion. Partly assimilated country rock fragments are common within the intrusion.

The ultramafic Talluskanava intrusion (Fig. 28) is concordant with respect to the enclosing migmatitic garnet-bearing mica gneisses and is about 500-600 m in length and 100-150 m thick. The intrusion, which consists of peridotite (30% olivine and 70% orthopyroxene) lies along a trend towards 055°, which also contains a number of other small ultramafic pods. The contacts with the country rocks are sharp and the intrusion contains the regional tectonic lineation, which southeastern plunges at 50° (Nurmi, 1974). The intrusion is overturned towards the northwest such that a 4-7 m thick sulphide lens near the upper contact now effectively forms the footwall. The contact with the mica gneisses is serpentinized and also contains talc. Sulfide minerals include pyrrhotite, chalcopyrite, pentlandite and cubanite. Overall the Talluskanava occurrence is estimated to contain 150 kt of ore with 0.32% Ni and 0.20% Cu. A richer section can however be delineated, containing 55 kt of ore averaging 0.53% Ni and 0.24% Cu. The respective Cu/(Cu+Ni) ratios of 0.38 and 0.31 are similar to those for Koivujoki

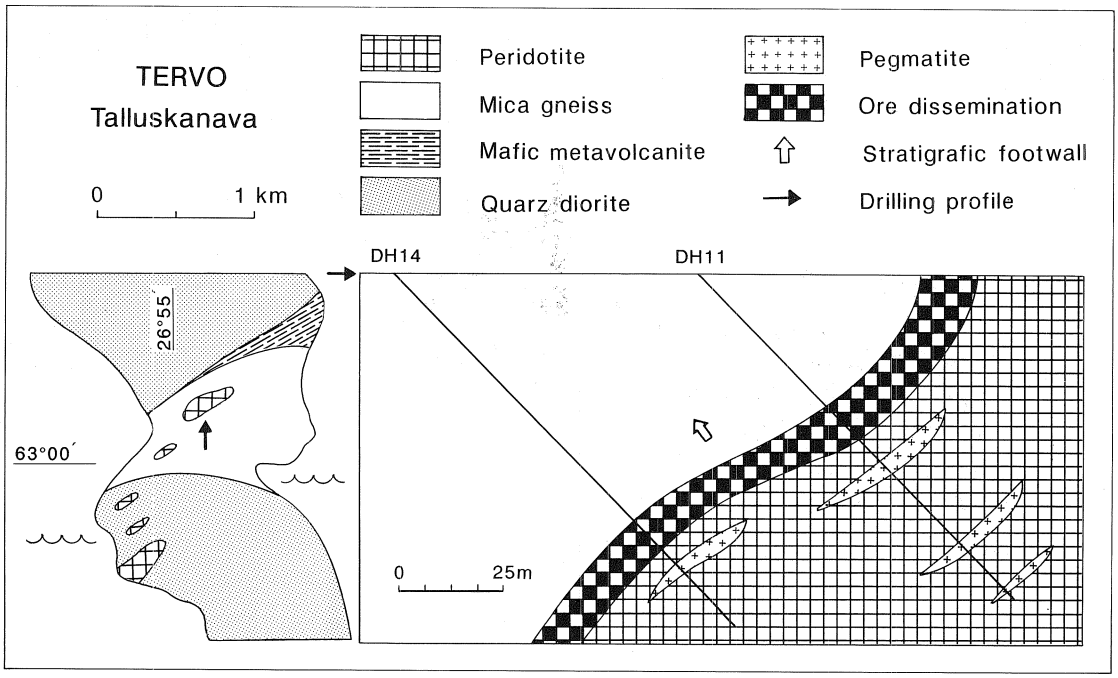


Fig. 28. Section through the Talluskanava ultramafic layered intrusion, which is located structurally in a zone overturned towards the northwest.

and Ilokangas.

The highest forsterite content determined for the twenty olivine grains analyzed from the Talluskanava intrusion was 82.1 (Nurmi, 1974), on the basis of which the initial MgO/(MgO+FeO) ratio of the magma was determined as 0.602 (mol% using $K_D^{melt/olivine} = 0.33$, after Roeder and Emslie (1970). Spinifex-textured rocks having this ratio contain 9.8% MgO, while on the basis of experimentally derived calibrations the forsterite compositions correspond to a value of 10.2% MgO. However, the MgO content of the intrusion is approximately the same as that of basalt magma. A similar result was obtained for intrusions in the Juva district (Makkonen, 1992), so that the Talluskanava and Koivujoki intrusions appear to be in general analogous to the layered intrusions at Juva.

TECTONICS AND METAMORPHISM

The structure of the Pielavesi district is dominated by elongate, approximately N-S trending, open to tight F_2 antiforms and synforms. Dextral NW-trending faults and ductile shear zones are also prominent, as well as NE and N-S trending fault systems, all of which overprint the earlier structures. At Pyhäjärvi, these structures were all assigned to a complex, progressive deformational event (designated D_3) by Ward (1984), expressed as shear zones and associated folds and crenulations, with some possibly antithetic or conjugate NE-trending

sinistral shear zones. Koistinen (1984 b) also recognized an independent D_4 stage of deformation in the vicinity of the Pyhäsalmi mine, during which N-S trending faults formed. Similar ductile faults and shear zones occur in the Pielavesi district and can be traced southwards towards Tervo.

First generation folds are not easily recognized in the study area and their regional distribution and significance is not known. Ward (1984) regarded D_1 and D_2 structures as both having formed during early N-NE vergent recumbent folding and thrusting. Studies of the Ruotanen ore body and its immediate surroundings in the Pyhäsalmi district (Koistinen, 1984 b; Luukas, 1991 b) have however shown that the ore body is actually located within an isoclinal F_1 fold hinge which has been refolded into an upright F_2 synform plunging southwards at 40°-60°. The competence contrast between the sulphide ore and country rock has doubtless influenced the style of deformation both at Pyhäsalmi and at Säviä, where the ore appears to have been tectonically thickened by F_1 and possibly F_2 folding, after which it was further reoriented during D_3.

For all practical purposes S_1 in the Pielavesi district is parallel to primary lithological layering S_0, while S_1 and S_2 are generally subparallel and difficult to distinguish from one another. Therefore, the regional stratigraphic map (appended map 1) shows lithological form lines representing a composite S_0-S_1 fabric. D_1 and D_2 may have been progressive in this area too, with D_2 involving gradual reorientation of structures into steeper attititudes. S_2 axial planes generally dip eastwards or southeastwards at 45°-90° throughout much of the area, with folds reclined towards the west and northwest (Fig. 29). In the northern part of the study area F_2 folds tend to be more open. Regionally D_2 is consistent with maximum compression from the SW (Ward, 1984), and this appears to be the case throughout Pohjanmaa as well (Luukas, 1991 b). The axial culminations in the Pielavesi area, corresponding to the Joutsenniemi and Kirkkosaari domes may be attributed to N-S D_3 compression.

Fig.29. Isoclinally folded Salonsaari Suite migmatitic mica gneiss with northwesterly vergence (yellow arrow indicates north). Cross-section indicates general style of folding throughout study area. Karhusuo.

The structural features recognized in the Pielavesi district represent the culmination of prolonged continuous orogenic processes. These can be interpreted in terms of northeastwards plate convergence, which led to the development of dextral transcurrent faulting and shear systems along the Raahe-Ladoga Zone as well as within the Archean basement foreland. The Pielavesi district lies almost entirely between two such shear systems, namely the Suvasvesi and Haukivesi shear zones. The dextral nature of this deformation is evident in the plastic, clockwise rotation and deflection of the Joutsenniemi dome and dextral drag folding (Fig. 30). Dilation within and between NW-trending swarms of faults or shear zones permitted intrusion of dolerite and pegmatite dikes along N-S and NE-SW directions, particularly in the Koivujoki area. Layered ultramafic bodies were emplaced prior to or during D_2, so that in some places they have been overturned. The Saarisenjärvi intrusion at Tervo formed during protracted and episodic shear along a zone trending roughly 025°-030° (Ekdahl, 1974). The Saarinen intrusion within the northern part of the study area is also controlled by shear

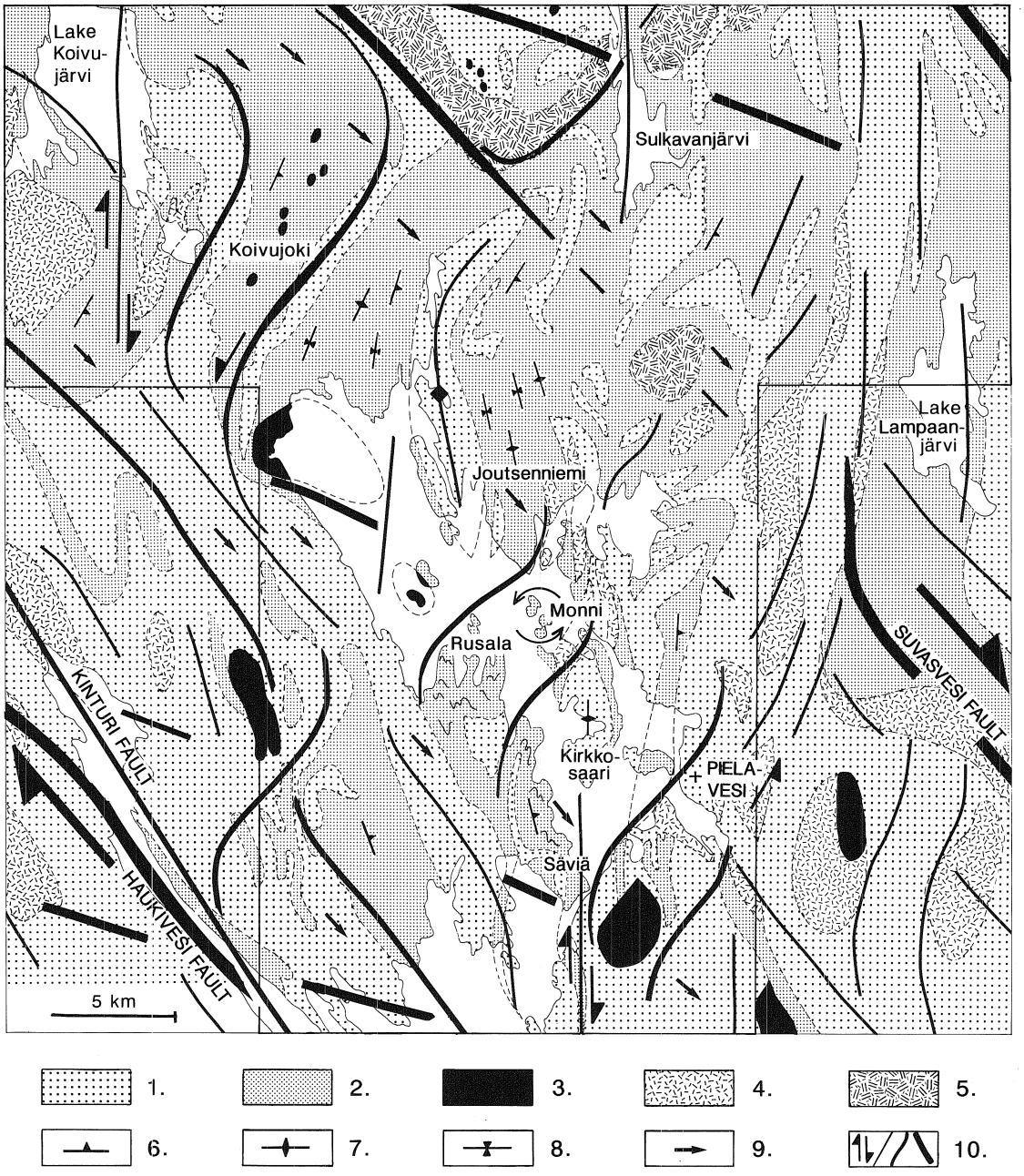


Fig. 30. The dextral Suvasvesi and Haukivesi shear zones generate clockwise rotation, drag folds and dilational fracture arrays with local development of antithetic NE sinistral structures. In the Säviä area, an additional N-trending dextral shearing is evident. 1 = Quartz and granodiorites; 2 = Volcano-sedimentary unit; 3 = Mafic and ultramafic intrusions; 4 = Granitoids, mainly porphyritic; 5 = Pyroxene-bearing granitoids; 6 = Schistosity (S_2); 7 = Axial culmination (F_2 antiform); 8 = Axial depression (F_2 synform); 9 = Lineation (L_3) ; 10 = Fault zones, major and minor tectonic features.

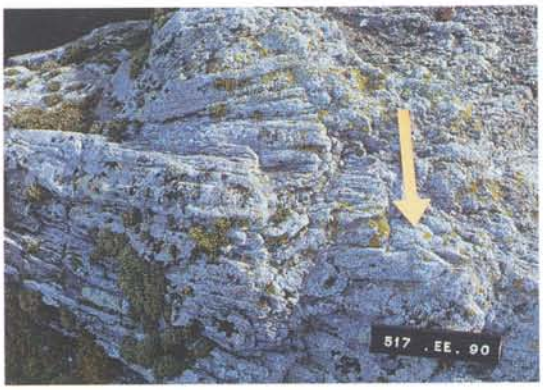


Fig. 31. NW-trending D_3 shearing and tectonic transport have resulted in a penetrative mineral lineation in Salonsaari Suite migmatites (yellow arrow indicates north). Mustilainen.

Fig. 32. Koivujoki Suite metagraywacke with $S_1 = S_2$ (horizontal direction in photograph) and NW-trending D_3 shearing and associated stretching lineation (L_3) and crenulation cleavage (yellow arrow indicates north). Koivujoki.

zones trending both 345° and 030°.

At Pielavesi D_3 deformation is expressed by the overturning of D_2 folds towards the northwest, which is also kinematically consistent with the reorientation of N-vergent early structures into dextral Phase 3 shear zones in the Pyhäsalmi district (Ward, 1984). Clockwise rotation between dextral shear zones has reoriented originally N-S S_2 structures into a NE-SW direction, producing sigmoid trends; this is seen especially well in the Koivujoki area. This deformation phase has resulted in a prominent NW-SE crenulation cleavage and SE-plunging mineral elongation lineation throughout the region (Figs. 31 and 32). Overprinting relationships are well documented in the Kortekylä area, in the northern part of the study area, where the NW-trending S_3 foliation is almost perpendicular to S_2. The pipe-like mafic intrusions in the area are oriented parallel to the regional SE-plunging mineral lineation, whereas stratiform Ni occurrences and stratabound ores (Ruotanen, Kangasjärvi, Säviä) have been remobilized by shearing and tectonic transport during D_3 so that they too form elongate lenses within NW-SE trending folds and shear zones.

Two distinct modes have been recognized within the overall NW-trending fault systems, namely 290°-310° and 320°-330°, with the former being interpreted as the earlier (Talvitie, 1975) and corresponding to the sinistral Kerimäki fault trend of Gaál (1985). The latter direction is the more important, coinciding with the trend of the Suvasvesi Shear Zone. Changes in fault trends with time may be due to overall changes in direction of convergence stresses. Ward (1984) considered that ductile early Phase 3 deformation controlled the intrusion of the Saarela gabbro and that brittle deformation was superimposed. Thus, deformation in these zones records a transition from ductile through to predominantly brittle behaviour. The Haukivesi Shear Zone passing through the southwestern part of the study area, was clearly overprinted by the brittle Kinturi Fault. It is possible, though difficult to demonstrate, that some of the fault zones further east are inherited from original rifting.

Northeasterly trending conjugate structures are recognized in the central part of the study area near the Kirkkosaari and Joutsenniemi domes, which have an inferred relative horizontal component of displacement of about 2 km. This was associated with anticlockwise rotation of the Monni Block and disharmonic folding in the Rusala area (Fig. 30). Northeast-trending structures including mylonites are

common in the Säviä-Pielavesi zone and truncate the NW-SE trends. Meridional dextral N-S structures are most prominent in the Pielavesi-Koivujärvi area and Sulkavanjärvi-Säviä zone, as well as at the eastern margin of the study area. These ductile features may correspond generally to D_3 and are characteristic of the whole PIA zone.

A triangular domain in the Sulkavanjärvi area, bounded by faults and porphyritic pyroxene granite contains disrupted and brecciated schists and metamorphosed and recrystallized pyroxene-bearing granodiorites and quartz diorites. Intrusion of mangerites and porphyritic granites postdated D_2 and presumably took place during D_3. Magma was intruded along the limbs of F_2 folds concurrently with NW-SE shearing which generated the SE-plunging lineation and upright schistosity. Feldspar is commonly oriented towards 120°, as for example at Tuomijoki and on the islands in Lake Pielavesi. Phase 3 deformation and pyroxene-bearing intrusions generally followed but locally coincided with the regional metamorphic thermal peak, which attained granulite facies grade in the Pielavesi area (Korsman et al., 1984).

The central part of the study area (the Pielavesi Block of Korsman et al., 1984) has retained granulite facies assemblages whereas the Lampaanjärvi and Korppinen Blocks to the east and west respectively, contain medium- to upper amphibolite facies assemblages (Hölttä, 1988). Granulite metamorphism in the Pielavesi Block culminated at temperatures of 800-880° C under pressures of 5.5±1.0 kb, followed by almost isobaric cooling, renewed heating and a further stage of isobaric cooling (Hölttä, 1988). The reheating event is recorded in the metamorphic aureole surrounding the Vaaraslahti hypersthene granite, which is clearly superimposed on the earlier event (Hölttä, 1988). The Vaaraslahti intrusion is also important in a number of other respects since it has been precisely dated at 1884 Ma (Vaasjoki and Sakko, 1988) and thus provides a minimum age constraint on the main phase of deformation and metamorphism (Korsman, 1989). The porphyritic pyroxene-bearing magmas generally appear to be syntectonic with respect to NE- and NW-trending Phase 3 structures. This is consistent with the abundance of country-rock enclaves containing D_2 structures in these intrusions.

During the intrusion of the hypersthene-bearing granites, the Rastinpää, Kettuperä and Kirkkosaari tonalitic gneisses also underwent retrograde metamorphism, attaining their present migmatitic appearance by 1.9 Ga (Korsman et al., 1984). U-Pb titanite ages of 1.86 Ga indicate that by this time, the presently exposed erosion surface had cooled below 400-500°C. The Lammasaho granodiorite at Kiuruvesi gives a zircon U-Pb age of 1853±12 Ma (Vaasjoki and Sakko, 1988), which is considered to record the last stage of tectonomagmatic activity within the Lampaanjärvi Block.

RADIOMETRIC STUDIES

Kirkkosaari dome

Isotopic studies have provided ages for the basement to the Pielavesi supracrustal sequence at Kirkkosaari. The characteristic banding defined by mica- and hornblende-rich layers is taken to indicate the paragneissic origin of the majority of these rocks, although due to intense deformation it is difficult to ascertain the amount of primary igneous material present. Paragneisses and orthogneisses appear to grade imperceptibly into one another and a

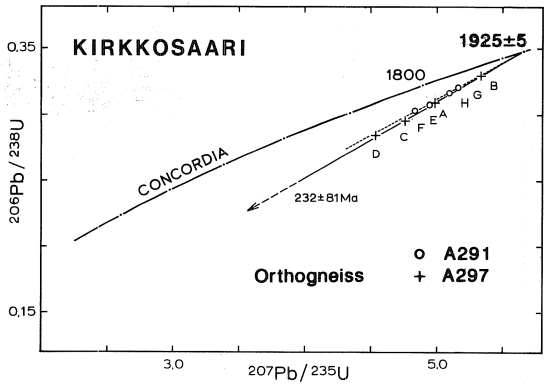


Fig. 33. Concordia plot for zircon fractions from Kirkkosaari orthogneiss (isotopic data in App. 2 A).

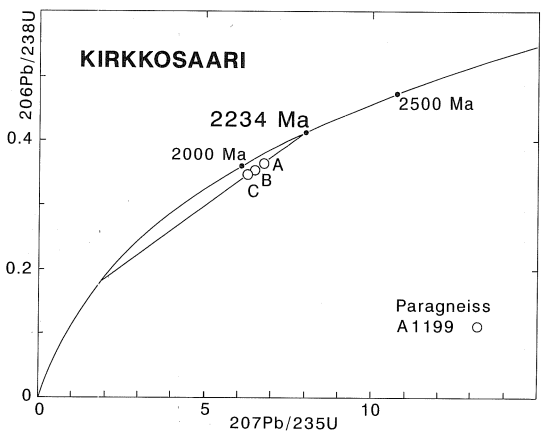


Fig. 34. Concordia diagram for Kirkkosaari paragneiss zircon fractions (isotopic data in App. 2 B).

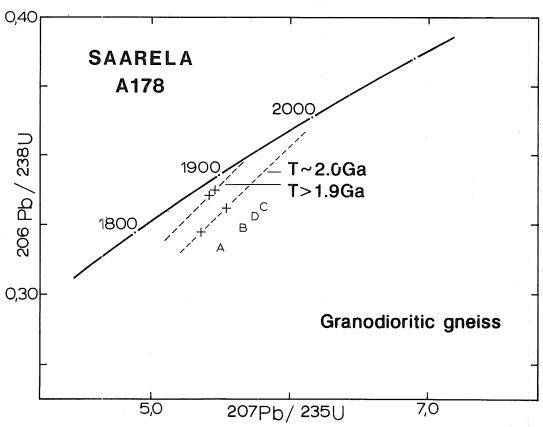


Fig. 35. Concordia diagram for various zircon populations from the Saarela granodioritic gneiss (isotopic data in App. 2 C).

sample of the latter has yielded a U-Pb zircon age of 1925±5 Ma (Fig. 33, App. 2 A). Zircons are abundant and form a distinct population of small, clear euhedral character. They are notable for their relatively small abundances of nonradiogenic lead, which is usually associated with rocks formed in K- and Pb-deficient environments such as within mafic intrusions. Water-poor charnockitic intrusions, such as the Voinsalmi hypersthene granodiorite in southeast Finland may also contain relatively pure, Pb-free zircons (Patchett and Kouvo, 1986).

A paragneiss sample from the same extensive outcrop containing the orthogneiss has also been dated and gave a result of 2234±69 Ma (Fig. 34, App. 2 B). Zircons from this sample are mostly stubby prisms, and often roundish. Three separate samples were analyzed and the results are technically precise. Interpretation is, however, complicated because similar results have been obtained from Svecofennian metasediments (Huhma, 1986; Kähkönen, 1989), where they evidently represent detrital mixing of Archean and early Proterozoic zircon population (Huhma, 1990).

The Kirkkosaari gneisses correlate lithologically and in their domal setting with the Kettuperä gneiss at Pyhäjärvi for which an age of 1932 Ma has been obtained (Helovuori, 1979; interpreted by Marttila (1993) as a metamorphic age) and the Hoikanlampi gneisses between Pyhäjärvi and Koivujärvi, which has been dated at 1873 Ma (Marttila, 1993). These apparent age discrepancies between otherwise similar lithological units indicate that caution is required in interpreting the significance of isotopic data from these highly deformed gneisses. In the case of Kirkkosaari in particular, which has been subjected to granulite facies metamorphism, the possibilty of resetting of original intrusive ages should be considered.

In the 1960's, during exploration by the Survey in the area southwest of the Saarela Gabbro, a sample of granodioritic gneiss, then considered to represent basement, was analyzed and yielded a discordia array intercepting

the concordia at 2053 Ma. Ages calculated on the basis of the diffusion model were also near 2.0 Ga (1982 Ma and 1990 Ma) and $^{207}Pb/^{206}Pb$ ratios of the discordant data gave ages between 1964 Ma and 1977 Ma, it appears that the sample may indeed approach 2.0 Ga in age (Fig. 35, App. 2 C). A sample taken recently from the same outcrop gave a somewhat younger age, nearer 1.9 Ga. The granodioritic gneiss occupies the same structural position as those interpreted by Marttila (1976) and Isohanni (1982) as granodioritic and quartz dioritic basement gneisses. On the basis of the present study, paleosomes in these gneisses are interpreted to correspond to the volcano-sedimentary unit within the Salosaari Suite, which is overlain by volcanics of the Säviä Suite.

U-P horizon

The Vihanti Lampinsaari phosphorus horizon is situated beneath the island arc volcanic series, just as in the Ruotanen and Säviä schist belts and at Virtasalmi, where it formed prior to volcanism and Cu-Zn ore deposition. Äikäs (1988d) correlated the "Pielavesi phosphatic basin" with the "Vihanti phosphatic basin" on the basis of similarities in the occurrence of uraninite, and also with the phosphatic occurrences at Siilinjärvi and Nilsiä. The Siilinjärvi phosphorites have been regarded as Marine Jatulian and thus predating Jatulian volcanism (Äikäs, 1987).

The Lampinsaari deposit at Vihanti and the Marine Jatulian Temo occurrence at Nilsiä both give dates of 1876±2 Ma for the U-P horizon (Vaasjoki et al., 1980), which is rather similar to the Pb-Pb age of 1897±7 Ma obtained for the Nuottijärvi occurrence at Paltamo. These are interpreted as metamorphic ages, with sedimentation having taken place between 1.90-2.08 Ga. The U-P horizon within the Savijärvi Suite in the Pielavesi district is assumed to correlate with these other units.

Synorogenic magmatism

Syntectonic intrusions into the Pielavesi supracrustal sequence include porphyritic pyroxene granites, granodiorites, quartz diorites and mafic and ultramafic bodies. The Vaaraslahti pyroxene granite has a U-Pb zircon age of 1884±5 Ma (Salli, 1983), which is representative of similar intrusions throughout the Raahe-Ladoga Zone, including the Haukilampi intrusion at Suonenjoki (1888 Ma, Korsman et al., 1984), and the Voinsalmi intrusion at Rantasalmi (1887 Ma, Patchett and Kouvo, 1986). The zircon age of the Lampaanjärvi granodiorite (1883 Ma), corresponds to those of 1883±3 Ma and 1882 Ma for the Palokangas intrusion at Pielavesi and Molkanjärvi quartz diorite respectively (Pääjärvi, 1991). The Pielavesi granodiorites are palingenetic, so that migmatization preceded their emplacement and crystallization.

The Kotajärvi Metalava within the Koivujoki Suite was referred to by Salli (1983) as the Kotajärvi Diorite, for which a U-Pb zircon age of 1882±2 Ma has been obtained. Calc-silicate enclaves derived from the Savijärvi Suite as well as metavolcanic enclaves from the Säviä Suite have been found in the metalava. Hornblende gneisses of the Koivujoki Suite are also seen to underlie the Kotajärvi Metalava in the Savijärvi profile so (Fig. 10) that they must in part be older. Amphibolites and ultramafic lay-

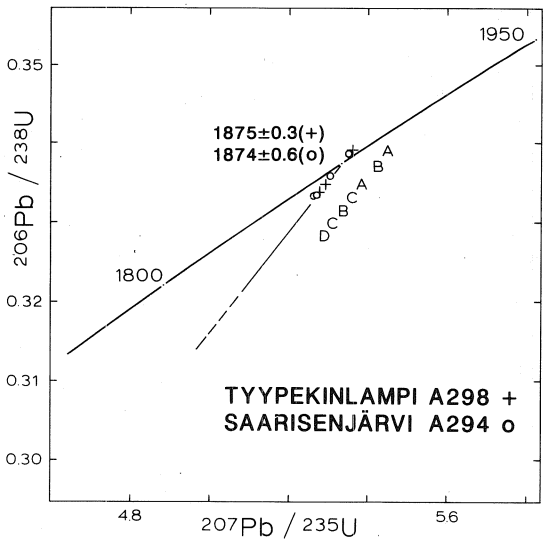


Fig. 36. Concordia plot for zircon fractions from the Tyypekki gabbro and Saarisenjärvi gabbro pegmatoid (isotopic data in App. 2 D).

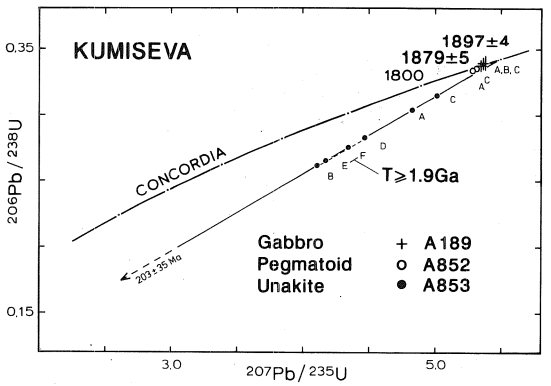


Fig. 37. Concordia diagram for zircon fractions from the Kumiseva pyroxene gabbro, gabbro pegmatoid and unakite (isotopic data in App. 2 E).

ered intrusions within the Koivujoki Suite are correlated with the Kotajärvi Metalava, which represents the last stages of volcanism in the Säviä and Ruotanen schist belts, corresponding in age with that at Pihtipudas, to the west of the study area (1883±20 Ma, Aho, 1979). A U-Pb zircon age of 1882±2 Ma form a pegmatitic gabbro at Lammaskylä, near Viitasaari (Nironen and Front, 1992) also represents the same magmatic event.

The Tyypekki pyroxene gabbro intruded supracrustal rocks and contains an abundance of schist enclaves. Zircons from the gabbro are characteristically brownish and euhedral. The U-Pb age of 1875±0.3 Ma (Fig. 36, App. 2 D) is somewhat younger than the 1885 Ma date obtained from a gabbroic pegmatoid occurring at Tulitoiviainen immediately to the north of the study area (Marttila, 1981). According to Marttila, this is the youngest unit in the region except for the pyroxene-bearing porphyritic granites, pegmatites and aplites. Other mafic intrusions representing the same magmatism as the Tyypekki Gabbro include the complex body at Saarisenjärvi at Tervo, from which a gabbroic pegmatoid representing the latest phases of intrusion (Ekdahl, 1974) has yielded an age of 1874±0.6 Ma (Fig. 36, App. 2 D). The Teerimäki Gabbro is also correlated with the Tyypekki intrusion and contains country-rock enclaves derived from Zn-Cu mineralized Säviä Suite volcanics that were clearly metamorphosed prior to disruption by the gabbro. The Tyypekki, Teerimäki and possibly also Saarela gabbros belong to a slightly younger age group than the Ni-bearing intrusions in the Kotalahti Nickel Belt.

The Kumiseva Gabbro to the northwest of the study area has a U-Pb zircon age of 1879±5 Ma (Fig. 37, App. 2 E) and is cut by the brittle Kinturi Fault (Fig. 38), which indicates the prolonged nature of transcurrent deformation. Unakites and possibly orbicular gabbros are also present in the zone (Sipilä, 1989), which may indicate crystallization at relatively high crustal levels, as has been inferred for intrusions at Kylmäkoski (Papunen, 1986). The Lahnanen orbicular quartz diorite at Viitasaari

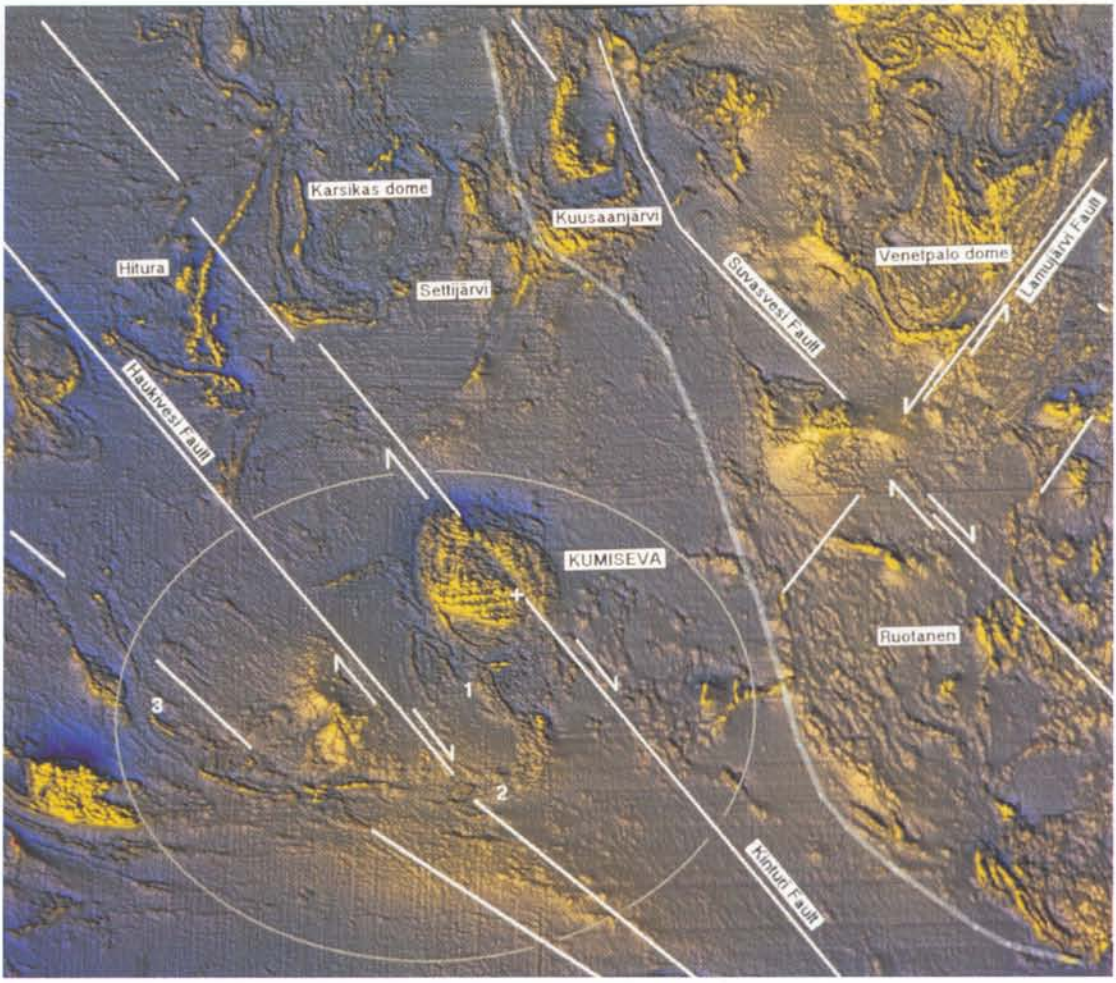


Fig. 38. Kumiseva gabbro and surrounding satellite intrusions (Kumiseva cluster) are revealed in both aeromagnetic and gravimetric datasets (1 = Nuoranen; 2 = Nuottijärvi; 3 = Ajakainen). The western boundary of the Savo Schist Belt is indicated by the thick white line. Dextral displacement parallel to Kinturi Fault transects the layered Kumiseva pluton. Locations of samples analyzed for isotopic dating are marked with a cross. Color pixel map is obliquely illuminated from the southwest.

has an age of 1876 Ma (Nironen and Front, 1992) and was formed during the same tectonic processes. Zircons separated from the unakites within the zone of more intense deformation are nevertheless somewhat older (1897±4 Ma), which is at variance with intepretations based on field relationships (Sipilä, 1989). Five fractions have been analyzed from the unakites (App. 2 E) and two of these suggest an even greater age, around 1.9 Ga. Movement along the Kinturi Fault, or other brittle reactivation of the Haukivesi Shear Zone also took place following the cessation of orogenic intrusive activity. Paavola (1988) reached similar conclusions with respect to NW-trending brittle deformation zones within early Proterozoic granitoids and gabbros intruding the Archean Iisalmi Block to the east of the study area.

Synorogenic magmatism in the Pielavesi district thus culminated at 1.89-1.88 Ga with the intrusion of mafic bodies, pyroxene granites and quartz diorites, which are mainly palingen-

ic in origin. Syntectonic pyroxene-bearing schollen migmatites also formed at this stage. Some of the pyroxene gabbros are nevertheless slightly younger around 1.88-1.87 Ga in age, by which time brittle deformation mechanisms predominated.

KARELIAN AND SVECOFENNIAN DOMAINS

The basement problem

The nature of the basement to the Svecofennian supracrustal sequences remains unknown, although an oceanic substrate is widely invoked (Hietanen, 1975; Mäkelä, 1980; Claesson, 1985; Huhma, 1986; Rickard, 1986; Gaál, 1986, 1990; Mäkelä, 1989). No direct evidence for the existence of such oceanic crust has yet been found; although Park (1988 b) suggested that the mafic complexes of the Kotalahti belt might have been remnants of a vanished Svecofennian ocean, this suggestion was later abandoned (Park, 1991). The lack of evidence for Svecofennian ophiolites admits the possibility that the region is underlain by, or at least contains large areas underlain by, crust of continental affinity. The lowest exposed graywackes in southern Finland and central Sweden contain detrital zircon populations that give convential ages ranging between 2.3-2.1 Ga (Huhma, 1986; Rickard, 1986; Welin, 1987; Kähkönen, 1989) which could be interpreted to indicate an early Proterozoic basement to the Svecofennian. Huhma (1987) argued that the Tampere graywackes in southern Finland represent a mixed provenance in which a small component was in excess of 2.3 Ga. Subsequent SHRIMP zircon studies (Huhma, 1990) showed that indeed some zircons are Archean, with the oldest being 3.44 Ga, while the majority are between 2.1-1.95 Ga. This clearly implies an early Proterozoic continental source that predates most of the presently exposed Svecofennian crust. The negative ε_{Nd} values and the extremely old T_{DM} age of the Ryggskog slate in western central Sweden are consistent with a large component of Archean material and a relatively proximal source area (Welin, 1987).

Petrographical and geochemical evidence from the Bergslagen district in central Sweden suggests that felsic volcanism was a product of anatexis of a felsic to intermediate crustal precursor having an age between 2.4-2.1 Ga (Baker et al., 1988). New Proterozoic crust was possibly generated in a series of island arcs during a phase of multiple subduction.

The 2.1-1.8 Ga Wopmay Orogen developed at the western margin of the Archean Slave Province in northwestern Canada (Hoffman, 1980) is in some respects similar to the Svecofennian. The La Bine Group of the Wopmay Orogen unconformably overlie the Hottah Terrane which has been dated as 1.92 Ga (Hildebrand, 1981). A study of detrital zircons from Hottah Terrane metasediments nevertheless indicates that the age of source rocks is between 2.4-2.0 Ga. Crust of this age was not previously recognized in Canada either, but studies based on the Wopmay Orogen and subsurface samples are changing this picture (Hoffman, 1989). Sm-Nd isotopic data from the Wopmay Orogen support the existence of crust of this throughout extensive parts of western Canada. Bowring and Podosek (1989) considered it likely that crust of this age is more widespread than previously thought in other cratons too and suggest that crustal models invoking a major phase of crustal growth at 1.9 Ga be reevaluated.

However, zircon isotopic data from Svecofennian metasediments do not support the existence of a major 2.4-2.1 Ga crustal prove-

nance in the Fennoscandian Shield and a major transfer of material from mantle to crust during the period 2.0-1.9 Ga is indicated (Huhma et al., 1991). Accordingly, the Sm-Nd data are interpreted by Huhma et al. as reflecting mixing of Archean and juvenile early Proterozoic material, rather than crustal growth at 2.4-2.1 Ga.

At present there are not sufficient isotopic data from the Kirkkosaari paragneisses to justify their being compared or contrasted with graywackes of the Tampere district. It is clear, however, that establishing the genesis of the 1.93-1.92 Ga units is fundamental to resolving questions about Svecofennian stratigraphy. The Kirkkosaari orthogneiss and paragneiss complex is regarded as basement to the Pielavesi supracrustal sequence, as are the analogous gneisses at Kettuperä in the Pyhäsalmi district (Helovuori, 1979; Marttila, 1993). The present study area continues southwards into the area covered by the Vesanto 1 : 100 000 map (Sheet 3313) where Pääjärvi (1991) considers veined and deformed tonalites and granodiorites to represent Archean basement, which can be correlated with those mapped in the southwestern part of the adjoining Lapinlahti 1 : 100 000 sheet (3332). These rocks are lithologically and petrographically similar to the orthogneisses and paragneisses in the Kirkkosaari dome. Therefore, there exists evidence for Archean basement a short distance to the east of Kirkkosaari, while some 10-20 km further east, towards Siilinjärvi and Kuopio, the presence of Archean rocks is undisputed. Since no discordance has been demonstrated between the sequences at Pielavesi and Kuopio, it is reasonable to conclude, though difficult to prove, that the basement at Pielavesi is, if not reworked Archean basement, 2.4-2.0 Ga crust, rather than juvenile material formed during the Svecofennian Orogeny. However Sm-Nd results indicate, that 1.93 Ga gneisses have only a small Archean crustal componet (H. Huhma, pers. comm., 1993). Microproble studies may shed more light on this problem in the future.

The transitional Savo schist domain

Because of their position along the margin of the Archean craton, the Savo schists contain elements of both Svecofennian and Karelian aspect, the latter being represented by the Kuopio district (Fig. 1). The island arc character of the sequence is most evident in the Pielavesi district, where the lowermost units include the Salo Metavolcanite and a thick, migmatized metagraywacke sequence containing volcanogenic intercalations (Salonsaari Suite). These are followed by shelf-type clastic and chemical sediments with minor volcanics (Savijärvi Suite), which is immediately overlain by the distinctly bimodal volcanics of the Säviä Suite, which are interpreted as the products of island arc magmatism. The Kotajärvi Metalava (1882 Ma) within the graywacke-dominated Koivujoki Suite represents the final phase of volcanic activity in the Pielavesi district.

Broad similarities with the proposed Pielavesi stratigraphic sequence are recognizable as far north as the Vihanti district and as far south as Virtasalmi (Tables 1-4). Diagnostic features include distinctive depositional environments and metamorphic grade and the absence of persistent marker horizons. The geological development of the Kiuruvesi district (Marttila, 1976, 1981) correlates well with that described for Pielavesi, except that Marttila considered the carbonate-, calc-silicate- and black schist lithologies to be present both above and below the volcanic sequence (Table 2). Subsequently, Marttila (1993) regarded the bimodal volcanism as having taken place almost immediately after the formation of the Kettuperä orthogneisses and paragneisses, while still taking the Kettuperä U-Pb zircon age (1932 Ma) as metamorphic. After this, volcanism gave way via

carbonate, calc-silicate- and graphite-bearing schists to predominantly graywacke sedimentation. Sections through the supracrustal sequence surrounding the Pielavesi domes nevertheless indicate that the shelf-type sedimentary association is restricted to a single structural level, though it may be repeated elsewhere by folding.

Island arc volcanism and stratabound ore formation throughout the entire PIA zone was preceded by the distinctive association comprising dolomite/carbonate, calc-silicate rocks, graphite schists, cherts, U-P and BIF horizons and minor volcanism. Although not everywhere complete, this horizon can be recognized over a distance of 350 km along the craton margin (Tables 1-4) and indicates that depositional conditions were similar over a wide area. In some places the sequence is thin or even locally absent and arc volcanism may have overlapped with shelf sedimentation, as was apparently the case in the Vihanti district (Table 3). From Pielavesi the arc volcanics can be traced through Itä-Karttula (Fagerström, 1990) southwards through the Virtasalmi-Juva area, where Zn-Cu ore deposits occur in the same stratigraphic milieu as those of the Vihanti-Pyhäsalmi district (Table 4). The most significant phase of volcanism with respect to mineralization was followed by predominantly graywacke sedimentation, with sporadic volcanic and graphite schist intercalations.

The stratigraphy of the Haukivesi area corresponds in a broad sense with that recognized in the Virtasalmi-Juva region. The lowest units exposed represent deep marine eugeosynclinal facies, including veined gneisses with graphite schists and calc-silicate intercalations, sulphide-bearing graphitic gneisses and diopside amphibolites (Gaál and Rauhamäki, 1971). These are overlain by conglomerates and metaturbidites. The diopside amphibolites and graphite-, cordierite- and garnet-bearing gneisses occurring to the west of Kerimäki may be correlated either with the Haukivesi eugeosynclinal sediments or the Marine Jatulian deposits of the Tohmajärvi district (Nykänen, 1975). The Kalevian metaturbidites of the Kitee-Kerimäki area may correspond to the so-called miogeosynclinal facies in the Haukivesi district and Nykänen (1975) considered that they differ from the graywackes and mica schists of the Höytiäinen zone to the east only in respect of their higher metamorphic grade. Ward (1987, 1988) recognized primary differences in depositional lithofacies between the Höytiäinen province and the sediments to the west, as well as in their subsequent deformational style and history, and further concluded that the earliest sediments in the Höytiäinen province were coeval with the Tohmajärvi mafic volcanic complex, dated by Huhma (1986) as 2.1 Ga in age. The Savo province metaturbidites (the Savo schists according to Koistinen, 1993), including those of the Outokumpu district, where they overlie the 1.97 Ga Outokumpu association, can be correlated with the miogeosynclinal metaturbidites of the Haukivesi district (Gaál and Rauhamäki, 1971; Koistinen, 1987).

The rocks of the Kuopio district represent a continental margin setting which deepened westwards. The Lippumäki Formation at Kuopio is a basal polymictic conglomerate typical of the Sariolan deposits in eastern Finland (Aumo, 1983 a,b). Jatulian facies are represented by the quartzites of the Vuorimäki Formation, which may be time equivalents of the deep water pelitic sediments of the Salonsaari Suite at Pielavesi. The problem of whether or not the lowest supracrustal units in the Salo Metavolcanite at Pielavesi can be correlated with the lowest unit at Kuopio remains unresolved. These correlations nevertheless indicate that the lowermost parts of the Pielavesi sequence are at least Jatulian in age, if not older.

Lithologically the Savijärvi Suite correlates very well with the Marine Jatulian Petonen Formation at Kuopio (Table 6). A zone of carbonate-bearing rocks, graphite schist and amphibolites has been traced from Karttula to Pielavesi in connection with mineral exploration

Table 6. Stratigraphy of the Pielavesi area compared with the Karelian sequence of the Kuopio district (after Aumo 1983 a,b) and the Svecofennian sequence of central Ostrobothnia (modified from Isohanni et al., 1980).

SVECOFENNIAN	MARGINAL ZONE	KARELIAN
Nivala - Haapajärvi (Isohanni et.al. 1980)	**Pielavesi**	**Kuopio** (Aumo 1983 a, b)
-Felsic-intermediate volcanism -Settijärvi Conglomerate	Vieremä-Haajainen Formation	
-Metagraywackes with volcanic interbeds — Pihtipudas volcanism 1883 Ma	Koivujoki Suite -Micagneiss with volcanic interbeds, -Kotajärvi Metalava — Granites, Gabbros, Quartzdiorites	Rahkamäki Fm -Micagneiss with volcanic and graphite bearing interbeds — Kaleva
-Bimodal volcanism	Säviä Suite -Altered rocks -Felsic-mafic metavolcanics	Vaivanen Fm -Mafic metalava with cherty interbeds — Jatulian volcanism
-Calc silicates, black schist	Savijärvi Suite -Dolomite, calc-silicates, chert, U-P, black schist, metavolcanites	Petonen Fm -Dolomite, calc-silicates, chert, U-P, black schist, BiF, amphibole schist — Marine Jatuli
-Migmatitic mica gneiss with graphite	Salonsaari Suite -Schollen migmatites -Migmatitic mica gneiss -Graphite schist -Salo Metavolcanite	Upper / Lower — Jatuli Vuorimäki Fm -Quarzite
		Lippumäki Fm -Polymictic conglomerate — Sariola
-Remobilized basement?	Basement domes -Ortho-and para-gneisses	Basement -Ortho-and para-gneisses — Archean

Oy Lohja Ab (Ollila, 1981, 1982). From Pulkonkoski near Maaninka, these amphibolites can be followed to Siilinjärvi, where they have been correlated with the Jatulian volcanics at Kuopio (Lukkarinen, 1990). Therefore, the bimodal volcanism of the Säviä Suite may correspond more or less to this volcanic episode.

The Lampaanjärvi mica gneisses, which form part of the Koivujoki Suite, can be traced from the Pielavesi district northwards to an area NE of Kiuruvesi, where Marttila (1981) regards them as underlying the Vieremä-Haajainen sequence, which he considered to be Kalevian, or possibly Upper Kalevian (Marttila, pers. comm., 1991). East of Lampaanjärvi however, the mica gneisses are juxtaposed directly against Archean basement.

The transitional Savo schists (Fig. 1) thus record deposition in a craton margin setting, with progressively deeper water conditions to the south and west. A major rifting episode led to the rapid onset of marine sedimentation with volcanic intervals. The western parts of the schist belt were characterized by island arc volcanism under shallower marine to emergent conditions. However, intense folding and shearing, migmatization and the intrusion of vast amounts of synorogenic magma have made it almost impossible to recognize the primary configuration of internal features, even though the region as a whole appears to form a coherent entity.

The Nivala-Haapajärvi area to the SW of the marginal zone is representative of the Bothnian association of the Svecofennian domain (Luuk-

konen and Lukkarinen, 1986). The main stratigraphic features of the Nivala area are comparable with those of Pielavesi (Table 6). The Settijärvi Conglomerate in the upper part of the sequence contains quartz diorite clasts dated at 1888±7 Ma (Vaasjoki and Sakko, 1988), which may be marginally older than the syntectonic quartz diorites of the Pielavesi district. On the basis of its stratigraphic position, the Settijärvi Conglomerate may correlate with the basal conglomerates in the Vieremä-Haajainen zone. The Settijärvi Conglomerate is overlain by bimodal volcanic rocks of the Kuusaanjärvi Formation which Kousa (1990), on the basis of lithological similarities, correlated with the volcanism at Pihtipudas, dated at 1883±20 Ma (Aho, 1979). This would therefore correspond to the Kotajärvi Metalava in the Pielavesi area (1882 Ma) and the mafic volcanism following the Ruotanen sequence at Pyhäjärvi (Huhtala, 1979; Luukas, 1992).

The Settijärvi Conglomerate is also underlain by bimodal volcanics that have not been affected by hydrothermal alteration and may also be devoid of mineralization. These in turn appear to be underlain in places by calc-silicate rocks and graphite-bearing migmatites. The problem of basement to these units centers around interpretations of granodioritic and quartz dioritic units such as the Karsikas dome at Nivala (Fig. 3). The Ostrobothnian Schist Belt consists in general of supracrustal relicts separated and intruded by a variety of granitoids, which show increasing deformation southeastwards towards the Savo schists (see Fig. 38).

The Marine Jatulian association - a key horizon

The Marine Jatulian sequence defined by Väyrynen (1933, 1954) constitute one of the most distinctive and easily recognizable stratigraphic horizons within the Karelidic Schist Belt. The Jatulian sequences can be followed from Finnish North Karelia for a great distance northwards but westwards they appear to terminate in the Kuopio area after which, according to Väyrynen, marine geosynclinal facies prevail.

Pekkarinen (1979) considered that the term 'Marine Jatulian' was not everywhere appropriate because it was not certain that marine conditions prevailed throughout the deposition of the sequence. Subsequently 'Marine Jatulian' appears to have been supplanted by 'Upper Jatulian' (Meriläinen, 1980; Simonen, 1980; Luukkonen and Lukkarinen, 1986) and refers to an association consisting of dolomites, mica schists, quartzites, sericite schists, black schists, iron formations (silicate, sulphide and oxide facies), a U-P horizon, mafic volcanics and related hypabyssal intrusions. The concept of the 'Marine Jatulian' can nevertheless be retained when referring to the transitional zone further west, with particular mention being made of the Marine Jatulian volcanism, which also becomes more prominent westwards. According to Aumo (1983b), the mafic Vaivanen Formation at Kuopio was erupted after Marine Jatulian deposition, represented in that area by the Petonen Formation. In this study, the stratigraphy of the Kuopio region will be adopted and the term 'Marine Jatulian-type' will be employed to refer to broadly similar Svecofennian lithological associations. The presence of the U-P horizon and iron formations within the 'Marine Jatulian' association lend support to this interpretation and its time-stratigraphic significance.

The U-P horizon

At present 23 separate U-P occurrences are known from the Karelian domain and 19 occur in the Svecofennian domain, if the boundary between the two domains is drawn as in Figure 1 (Table 7). All these occurrences are mutually

Table 7. U-P occurrences in the Karelian and Svecofennian domains and their characteristic features. U-P = uranium-phosphorus occurrence; P = High P_2O_5 abundance; Typical lithological association (chert, BIF = iron formation, B.S. = black schist, carb = carbonate rocks); Mn, Mo or Wenrichment. References given in text.

			Typical association					High	
Karelian domain	U-P	P	Chert	BIF	B.S.	Carb	Mn	Mo	W
TERVOLA Mustamaa	x	x	x	x	x	x			
KIIMINKI Vepsä	?	x		x	x	x	x	x	
POSIO Suonna			?	x		x			
PUOLANKA Pääkkö	?		x		x				
PUOLANKA Seppola	?		x		x	x		x	
PALTAMO Nuottijärvi	x	x			x	x			
KAJAANI Koutaniemi	x	x	x		x	x			
SOTKAMO Losonvaara	x	x		x	x	x			
SOTKAMO Tuomivaara	?	x		x	x	x	x		
NILSIÄ Tesmo	x	x		x	x	x			
NILSIÄ Temo	x	x			x	x			
SIILINJÄRVI Toso	x	x	x		x	x			
SIILINJÄRVI Kuivasteenmäki	x	x	x		x	x		x	
SIILINJÄRVI Vironniemi	x	x	x		x	x			
SIILINJÄRVI Rahvo	x	x			x	x		x	
KUOPIO Neulamäki	x	x	x	x	x	x			
VEHMERSALMI Pitkäniemi	x	x			x	x			
JUANKOSKI Viitaniemi	x	x			x	x			
POLVIJÄRVI Rukkajärvi	x	x			x	x			
POLVIJÄRVI Mantilanniemi	x	x			x	x			
LIPERI Karhunsaari	x	x	x	x	x	x			x
LIPERI Loukonsaari	x	x			x	x			x
KIIHTELYSVAARA Viistola	?	?	x	x	x	x			
Svecofennian domain	U-P	P	Chert	BIF	B.S.	Carb	Mn	Mo	W
KÄLVIÄ Ruotsalo	x	x	x		x	x			
VETELI Räyrinki	x	x			x	x			
KRUUNUPYY Småbönders	x	x			x	x			
LAPUA Simpsiö	?	x	x	x	x	x	x		
YLISTARO Vittinki	?	x	x	x	x	x	x		
RAAHE Relletti	x	x		x	x	x			
VIHANTI Lampinsaari	x	x			x	x			
VIHANTI Vilminko	x	x			x	x			
VIHANTI Nevasaari	x	x			x	x			
PYHÄJÄRVI Ollinniemi	?	x		x	x	x			
KIURUVESI Huutsaari	x	x			x	x			
PIELAVESI Savijärvi	x	x	x		x	x		x	
PIELAVESI Koivisto	x	x			x	x		x	x
PIELAVESI Uivero	x	x			x	x			
PIELAVESI Haasiakangas	x	x			x	x			
KEITELE Tossavanlahti	x	x			x	x		x	
JUVA Näärinki	x	x	x	x	x	x			x
JUVA Tutunen	x	x	x	x	x	x			
KERIMÄKI Louhi	x	x			x	x			

similar geochemically, petrographically and in their stratigraphic setting, which corresponds to the Savijärvi Suite of the study. Geochemically the U-P occurrences are analogous to each other, all being characterized by a high P_2O_5/U ratio, between 100-300 (Fig. 39) and low Th abundances, generally < 10 ppm (Rehtijärvi, 1983). The mean U contents of the occurrences

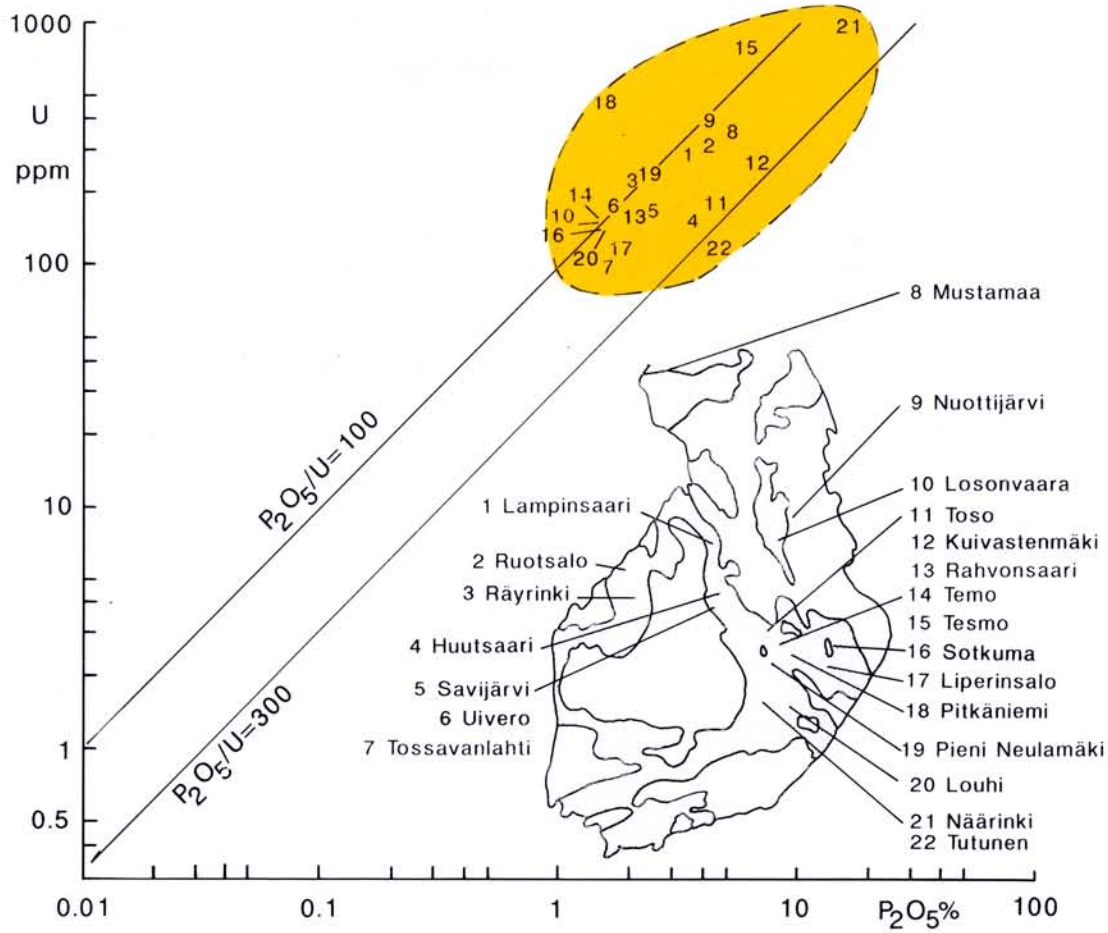


Fig. 39. P_2O_5/U distribution in Svecokarelian U-P deposits using the criteria of Rehtijärvi (1983). Locality references given in text.

are around 0.03-0.04% and P_2O_5 abundances around 3-5%. The horizon also contains sporadic enrichments in Mn, W and Mo (Table 7). Graphite is also abundant and black schists and sulphides are typically present as well. In the Svecofennian domain however, iron formations and graphite schists tend to be less abundant.

The U-P occurrences are stratiform and without exception occur within the 'Marine Jatulian' association, which strongly suggest contemporaneous deposition within an extensive contiguous basin. The phosphatic iron formations in Kainuu are genetically related to the U-P horizon, even though they are devoid of U enrichment. Äikäs (1975) reported iron formations from the Losonvaara occurrence at Sotkamo and correlated it with the Tuomivaara phosphatic iron formation and also with the Talvivaara black schists.

The U-P horizon in the Kainuu Schist Belt (including the occurrences at Losonvaara, Härmänmäki and Nuottijärvi) are hosted by calc-silicate and carbonate rocks and cherts that overlie Jatulian orthoquartzites but preceded graywacke and black schist sedimentation (Äikäs 1975, 1988 b). The U-P horizon has also been found to the west of Kajaani, at Koutaniemi, where it is associated with Marine Jatulian

facies and mineralization resembling that of the Vihanti-Pyhäsalmi type (Kataikko, 1987).

In the Savo-Karelia region the Jatulian association is generally thin, or may even be absent, so that the U-P horizon may have been deposited directly upon weathered Archean basement, as exposed for example around the Sotkuma basement inlier, and also at Liperinsalo, where U-P occurrences are associated with calc-silicate and carbonate rocks. The same features have also been observed in the Siilinjärvi area (Äikäs, 1990 a and b; 1987, 1988 g), where the U-P horizon preceded deposition of the protolith to the black schists, just as at Nilsiä, where the Marine Jatulian is also seen to overlie the Kinahmi quartzites (Äikäs, 1988 b and g, 1989).

At Kuopio too, the Petonen Formation surrounding the Pieni Neulamäki basement dome contains U-P enriched layers (U values = 98-495 ppm, P_2O_5 = 1.2-4.5%; checked by the author). The Marine Jatulian association continues towards Vehmersalmi, where a further U-P occurrence has been found (U = 0.047%, P_2O_5 = 1.462%, Th = 0.005%; checked by the author). North of the Suvasvesi Shear Zone the Marine Jatulian association can be traced towards Losomäki in the Kaavi district; Zinc occurrences have also been found in this area, in addition to mineralization within the U-P horizon at Viitaniemi (Äikäs, 1976) and within close proximity to the Outokumpu-type assemblage at Pieni Vehkalahti, and possibly also at Sorsasaari near Kuopio (Vanne, 1989).

In the Karelian domain the U-P horizon is generally located within the calc-silicate- carbonate-rock-chert association which lies on top of Jatulian quartzites or else directly on weathered Archean basement, and is overlain by black schists and Jatulian volcanics, where present. Within the marginal zone, the U-P horizon and the enclosing Marine Jatulian-type association occurs between the marine facies represented by the migmatitic mica gneisses predating the arc volcanics. The U-P horizon also preceded Zn-Cu mineralization in the Vihanti area, in the Ruotanen and Säviä schist belts, as well as in the Virtasalmi-Juva zone. The horizon may additionally be recognized in the Haukivesi zone, at Louhi, near Kerimäki, where a U-P occurrence within a carbonate-black schist sequence contains 0.014% U and 1.55% P_2O_5 (E. Räisänen, pers. comm., 1991). This sequence has been correlated with both the Tohmajärvi Marine Jatulian (Nykänen, 1975) and eugeosynclinal sediments of the Haukivesi zone (Gaál and Rauhamäki, 1971).

The U-P horizon has thus been found to represent a contiguous unit that can be recognized throughout at least a single depositional basin (Huhtala et al., 1978; Rehtijärvi, 1983; Äikäs, 1988 b, 1990 b). In principle there is no reason why correlation could not be extended from basin to basin, or throughout the whole of the Marine Jatulian of Karelia. The same applies to the U-P horizon in the Svecofennian domain, with correlation between Pielavesi and Haukivesi seeming feasible, in spite of locally poor exposure or areas where the association is not particularly well developed. The Savijärvi Suite at Pielavesi, along with the U-P horizon, is lithological identical with the Marine Jatulian. Äikäs (1987, 1988 d and f) correlated the "Pielavesi phosphatic basin" with the "Vihanti phosphatic basin" and mineralization at Siilinjärvi and Nilsiä on the basis of similarities in the occurrence of uraninite. The same geological association is also present at Pulkonkoski, near Maaninka, between Kuopio and Pielavesi. The horizon is not found to the west of the Kinturi Fault except within the Bothnian Schist Belt, where the Ruotsalo occurrence at Kälviä is associated with carbonates, calc-silicate rocks and black schists Äikäs (1988 c). This occurrence resembles that at Lampinsaari and similar features are also present at Räyrinki near Veteli and Småbönders near Kruunupyy (Lindmark, 1977).

This widespread shelf association is consistent with the proposal by Sokolov and Heiskanen (1985) according to which the accumulation of carbonates, stromatolites, red beds

and P-enriched deposits was a synchronous global event (Fig. 40). This phase of deposition was considered to be preorogenic and was followed by early orogenic magmatism. The Pb-Pb ages of the Pääkkö iron formation at Puolanka (2080±45 Ma, Sakko and Laajoki, 1975) and the Kalkkimaa dolomite at Kemi (2050 Ma, Silvennoinen, 1972; Pekkarinen, 1979) are often taken to correspond to that of Marine Jatulian deposition. In contrast the U-Pb ages for the U-P horizon at Lampinsaari and Temo near Nilsiä are 1876±2 Ma (Vaasjoki et al., 1980), while the Pb-Pb age for the Nuottijärvi iron formation at Paltamo is 1897±7 Ma. Vaasjoki et al. (1980) nevertheless considered these latter dates to represent metamorphic rather than depositional ages and regarded their true age as being somewhere between 1.90-2.08 Ga. The U-P horizon of the Savijärvi Suite is correlated with these occurrences, for which a minimum age of 1.97 Ga is indicated, on the basis of data from albite diabase dikes cutting the quartzites at Kinahmi, near Nilsiä (Paavola, 1984). Therefore, the Marine Jatulian association and the overlying Upper Jatulian mafic volcanics can be regarded as having formed between 2.08-

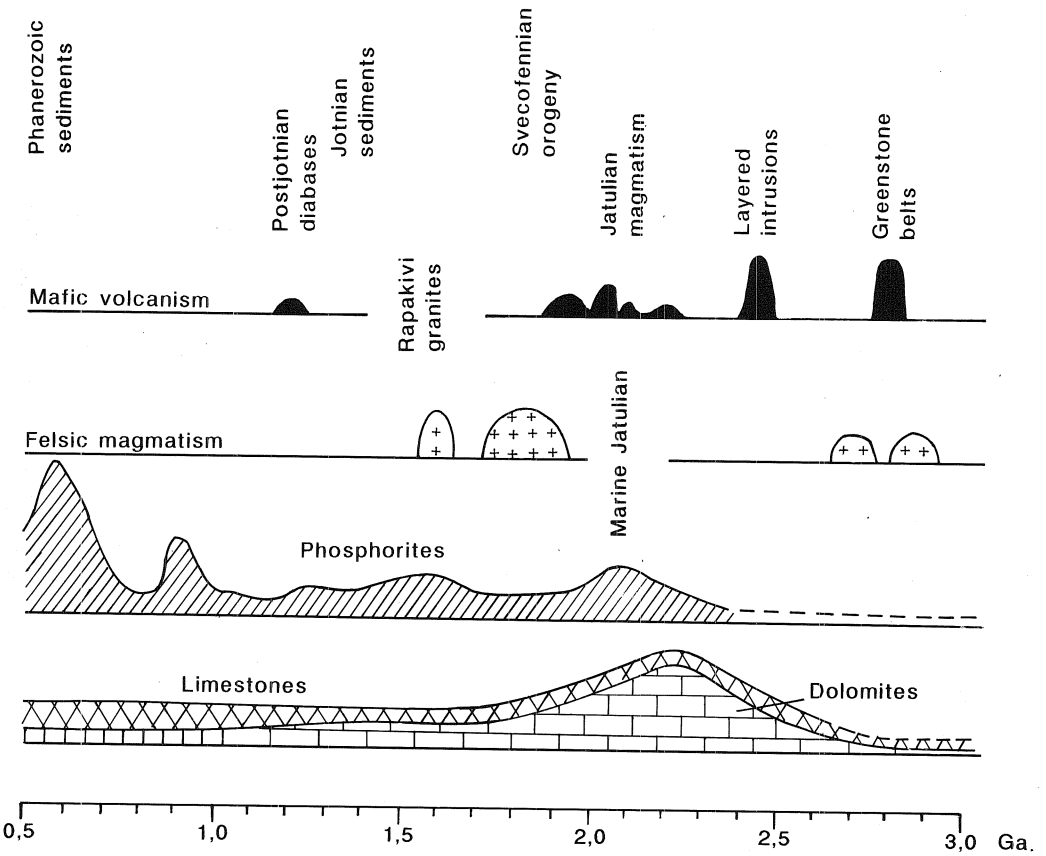


Fig. 40. Evolution of Proterozoic syngenetic dolomites, limestones and phosphorites and extent of felsic and mafic volcanism (modified after Sokolov and Heiskanen, 1985).

1.97 Ga. The U-P horizon is a widely distributed key stratigraphic unit which permits broad correlations to be carried out across the Karelian-Svecofennian transitional zone.

Iron formations

The presence of phosphorus is closely associated with the occurrence of iron formations. The Mn- and P_2O_5-rich iron formations of the Puolanka and Sotkamo districts (Pääkkö, Väyrylänkylä, Iso Vuorijärvi, Körölä, Seppola, Tuomivaara) have been interpreted as being of Superior-type and occur over a range of stratigraphic levels, consisting of mixed oxide-silicate facies or manganiferous sulphide facies (Laajoki, 1985). Precipitation of phosphorus took place under marine conditions and in some cases distinct pure phosphate laminae are present.

The iron formations of the Kainuu district, including those associated with the Talvivaara Ni-Cu-Zn deposit, have usually been assigned to the Marine Jatulian (Sakko and Laajoki, 1975; Ervamaa and Laajoki, 1977; Laajoki and Saikkonen, 1977; Havola, 1980; Ervamaa, 1980; Ervamaa and Heino, 1983; Laajoki, 1985). Recent studies at Tuomivaara have nevertheless shown that turbiditic sedimentation had commenced prior to the iron formations, which are also underlain by a breccia-like conglomerate (Gehör and Havola, 1986). The Kainuu iron formations can thus be regarded more generally as forming part of the Lower Kalevian sequence (Kontinen, 1986, 1987; Laajoki and Gehör, 1986; Laajoki and Luukas, 1988).

The deposition of the iron formations was evidently preceded by rifting of the craton margin, possibly in connexion with continental break-up between 2.2-2.0 Ga (Park et al., 1984; Kontinen, 1987). Therefore the Kainuu iron formations need not form part of the Marine Jatulian association but may instead be separated from them by a major discordance. Likewise, in Finnish North Karelia, deposition of turbiditic graywackes in the Höytiäinen-Ladoga province (H-L, Fig. 1) was inferred to have commenced as long ago as 2.10 Ga, with a possible conformable transition through Marine Jatulian into the Kalevian (Ward, 1987, 1988).

The interpretation of a discordance within the Kainuu Schist Belt is based on the recognition of breccia-like conglomerates which are overlain by turbidites, iron formations and black schists, followed again by breccias and turbidites grading into psammites (Laajoki and Luukas, 1988). It is possible that some of these features may be a result of tectonic repetition during Svecofennian thrusting. The Marine Jatulian association at Kiihtelysvaara in Finnish North Karelia also contains some phyllites and arenites and was followed, after an erosional hiatus, by deposition of Kalevian turbidites (Pekkarinen, 1979). In the transition area between Karelia and Savo, iron formations are nevertheless clearly present within the Marine Jatulian, with the hematitic horizons containing martite pseudomorphs after magnetite in the Viistola Formation at Kiihtelysvaara having been correlated with the iron formations of Kainuu by Lehto and Niiniskorpi (1977). The section at Viistola includes brecciated cherty layers with microfossils interpreted is algae in origin having been found in associated carbonates. Stromatolitic structures have also been identified in glacial erratics, although their ultimate source has not been located. The presence of iron formations within Marine or Upper Jatulian sequences of Russian Karelia has been well documented (Muradymov et al., 1986).

Silicate facies iron formations and cherts are also present within the Marine Jatulian association overlying the quartzites at both Nilsiä and Kuopio (Aumo, 1983 b; Paavola, 1984), representing the most southwesterly occurrences of the Karelian iron formations. Further north the Marine Jatulian is represented by the Suonna phosphatic iron formations at Posio (Lehto and Niiniskorpi, 1977), while in northernmost Bothnia, Superior-type iron formations and cherts occur within the Vepsä Formation

(Paulamäki, 1983).

To the west of Kuopio, a wide Marine Jatulian-type association is evident, in which iron formations were not so extensively developed. Narrow silicate facies iron formation have nevertheless been encountered within the Ruotanen schist belt at Ollinniemi (Mäki, 1987). Calc-silicate rocks and sulphide-bearing cherts at Jokijärvi near Tervo, to the south of the study area, presumably represent the same stratigraphical level. It is only much further south, between Virtasalmi and Juva, that iron formations and cherts again become conspicuous components of the carbonate - calc-silicate rock - black schist association. Both oxide and silicate facies iron formations have been found in close connexion with the U-P horizon at Näärinki near Juva (Pekkarinen, 1972; Makkonen, 1988). Between Raahe and Vihanti U-P horizons, quartzite-phyllite intercalations, black schists and iron formations have been variously reported in association with carbonate- and calc-silicate lithologies in numerous places, although stratigraphic relationships are not always clear (Nykänen, 1959; Salli, 1965; Alapieti, 1972).

Iron formations of the Karelian domain record shallow water environments in which large amounts of iron, silica and phosphorus were deposited, giving rise to a more or less coherent stratigraphical horizon that represents the time interval 2.08-1.97 Ga. Studies in Kainuu and Finnish North Karelia (Kontinen, 1986; Ward, 1987) indicate that deposition of graywackes took place before 2.0 Ga, during the time interval traditionally regarded as Jatulian, so that there is some overlap between the Marine Jatulian and Lower Kalevian. This transitional period reflects extension the craton margin, with local facies variations resulting from differential uplift and subsidence within and between different basins. Iron formations and black schists are noticeably less abundant in the Svecofennian domain but general lithological similarities and the absence of strong evidence to the contrary allows the Marine Jatulian-type association of the marginal zone to be correlated with the Marine Jatulian-Kalevian transition in the Karelian domain.

Problems of interpretation are not confined the marginal zone alone, since only after the Svecofennian basement enigma is solved, will it be possible to confidently establish the absolute stratigraphic positions of the calc-silicate rocks, carbonates, iron formations, black schists, quartz rocks and amphibolites found throughout western Finland, particularly in the Kaustinen-Veteli-Evijärvi area and further south in the Vimpeli-Lapua-Ylistaro district. According to Tuukki (1984) the quartz rocks of southern Bothnia are not primary clastic sediments but recrystallized metacherts occurring in association with phosphatic iron formations. The Simpsiö banded manganiferous cherts in particular contain very high P_2O_5 contents.

The Lapua iron formations clearly show affinities with those of southwestern Finland (Tuukki, 1984), which occur within a felsic arc volcanic setting, towards the top of the lower Svecofennian subgroup. These deposits in turn are reminiscent of the quartz-banded iron ores of the Bergslagen Province in central Sweden (Frietsch, 1980; Sipilä, 1981). However, the central Swedish iron formations consist principally of banded apatite-bearing oxide facies deposits. Sipilä (1981) attributed this difference to contrasting depositional environments. The principal difference between the Marine Jatulian iron formations of eastern Finland and those of southwestern Finland is that the latter tend to have lower graphite and phosphorus abundances.

C-isotopes

The $\delta^{13}C$ values obtained from lower Proterozoic black schists in eastern Finland generally indicate that a substantial component of graphite is organic in origin (Loukola-Ruskeeniemi, 1990 a). The isotopic compositions of graphite in 2.1-1.97 Ga black schists are very similar throughout Kainuu and Finnish North Karelia,

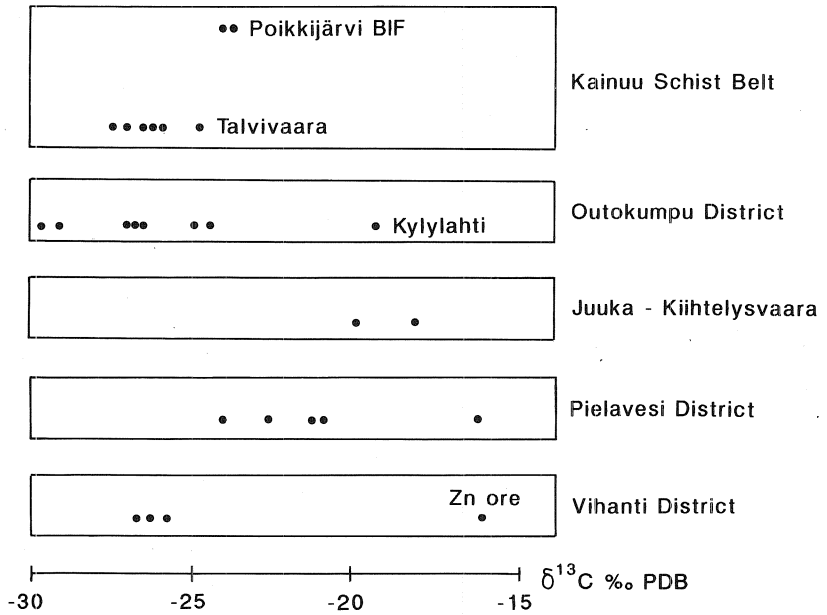


Fig. 41. Graphitic schists from the Savijärvi Suite at Pielavesi have $\delta^{13}C$ values that are very similar to those of other early Proterozoic black schists (Loukola-Ruskeeniemi, 1991). Elevated values result from metamorphism or hydrothermal processes related to mineralization.

including the Outokumpu region, and also in the Vihanti area (Loukola-Ruskeeniemi, 1989, 1990 b, 1991). Black schist samples from the Savijärvi Suite at Pielavesi (Fig. 41), including two each from Savijärvi and Paloniemi and one from Uivero, have $\delta^{13}C$ values between -24.6 and -20.9 per mil, except for one of the Savijärvi samples, which was taken from a sheared zone and has a value of -15.89 per mil. Generally metamorphism results in an increase in $\delta^{13}C$ values because of reaction with carbonate-derived CO_2 fluids. Such increases in $\delta^{13}C$ have been demonstrated for graphite schist samples from highly sheared zones in the Juuka and Kiihtelysvaara area and is attributed by Loukola-Ruskeeniemi (1991) to either fluid flow in shear zones, or general hydrothermal processes. The same explanation may be given to account for the elevated $\delta^{13}C$ values reported from Vihanti (-16 per mil) and Kylynlahti (-19 per mil). Black schists associated with the Poikkijärvi iron formation also exhibit slightly elevated $\delta^{13}C$ values, although they are rather similar to those obtained from black schists at Talvivaara and in the Outokumpu district. The Jormua, Talvivaara, Alanen, Pappilanmäki and Outokumpu black schists appear to be genetically related on the basis of their carbon isotopic compositions and are all situated at approximately the same stratigraphic level (Table 8; Loukola-Ruskeeniemi et al., 1991). The compositions of the Savijärvi Suite samples appear to overlap with those of the Kainuu-Outokumpu zone, Vihanti and Kiihtelysvaara within reasonable uncertainty limits (Fig. 41).

The $\delta^{13}C$ values of carbonates from the Savijärvi in the present study area, as well as from Tutunen at Juva and Puutossalmi near Vehmersalmi have also been determined and corroborate the existence of a distinct difference between the carbon isotopic composition of Karelian and Svecofennian carbonates (Karhu, 1989). The impure carbonates from Savijärvi have typical marine carbonate $\delta^{13}C$ values,

Table 8. The stratigraphy of the Kainuu Schist Belt compiled by Loukola-Ruskeeniemi et al. (1991) after Kontinen (1986, 1987).

Granitoids	ca. 1.85 Ga
Upper Kalevian metasediments, mica schists. etc.	< 1.96 Ga
The Jormua complex and related mafic-ultramafic rocks	ca. 1.96 Ga
Lower Kalevian metasediments: black schists, metal-rich black schists, mica schists, etc. Iron formation occurrences (2.08 Ga Sakko and Laajoki 1975)	1.96- < 2.1 Ga
unconformity	
Upper Jatulian metasediments and volcanites, black schists, dolomites	2.1-2.3 Ga
Lower Jatulian metasediments and volcanites, quartzites, arkosites, conglomerates, etc.	
unconformity	
Late Archaean granite gneiss	2.5-2.9 Ga

around 0 per mil (Table 9) and the Tutunen sample with a value of +2.86 per mil is characteristically Svecofennian (cf. Ankele and Montola samples from Virtasalmi, with values of +2.0 per mil and + 1.9 per mil respectively). In contrast, the Vehmersalmi sample ($\delta^{13}C$ = +6 per mil) compares better with Jatulian dolomites, such as at Kuopio (+6.4 per mil) and Siilinjärvi (+5.7 per mil; J. Karhu, pers. comm., 1993).

The exceptionally high $\delta^{13}C$ values obtained for Jatulian carbonates are probably primary in origin and may reflect both distinctive depositional processes and a significant organic input, enabling large amounts of organic carbon to be buried and removed from the biosphere (Karhu, 1992). A similar interpretation has been given in Zimbabwe (Schidlowski et al., 1976), where the 2.65-1.95 Ga Lomagundi association ($\delta^{13}C$ = +9.4) and its inferred correlative in the Urungwe klippe ($\delta^{13}C$ = -4.0) are have isotopic compositions analogous to those of the Jatulian

Table 9. Carbon and oxygen isotopic compositions for carbonate from carbonate rocks at Vehmersalmi, Juva and Pielavesi.

Sample	anal.no	$\delta^{113}C$ (PDB)	$\delta^{18}O$ (SMOW)	Yield # (µmol/mg)
Vehmersalmi, Puutossalmi, XRD analysis: pure carbonate (88 % dol.)				
Tot	C-291	5.93	18.06&	9.8
Cal		5.43	19.34	1.6
Dol		6.09	17.88	8.2
Juva, Tutunen, XRD analysis: calcite				
R378/216.8	C-292-1	2.86	20.01	9.0
R392/85.2	C-292-2	2.04	20.24	6.9
R377/131.1	C-292-4	1.83	20.05	8.3
Pielavesi, Savijärvi, XRD analysis: calcite impure				
R303/124.2	C-293-1	0.25	-	4.0
R303/52.5	C-293-2	-2.99	-	7.7
R303/61.1	C-293.3	-2.45	18.69	7.5

&: $\delta^{18}O$-total calculated on the basis of XRD results for calcite and dolomite
#: yield for pure calcite is 10.0 µmol/mg

and Svecofennian respectively. These sequences in Zimbabwe represent lateral facies variants and the anomalous isotopic composition of the Lomagundi carbonates would constitute strong evidence for an exceptional depositonal environment. Extensive removal of organic carbon from this particular environment is implied by the unusually high positive $\delta^{13}C$ values and is supported by the occurrence of carbonaceous argillites in close proximity to the dolomites (Schidlowski et al., 1976). This applies particularly to the Piriwiri graphitic rocks deposited prior to the Lomagundi Group and to the carbonaceous members of the Striped Slate Formation overlying the carbonates.

The black schist horizons associated with carbonate rocks and iron formations in Kainuu and North Karelia are unusually thick and have relatively high carbon contents, which account for the enrichment in the $\delta^{13}C$ values of the associated carbonates. In contrast, Svecofennian sedimentation took place in relatively deeper open ocean conditions and thick graphitic schist horizons are rare or absent. The graphite schists of the shelf association are also relatively thin and may indicate that this environment was not sufficiently rich in nutrients or was too deep to support a large biomass. The variability in isotopic compositions is not yet sufficiently well established as to detect distinct stratigraphical controls on variations.

Correlations and discussion

Assesment of the relationship between the Karelian and Svecofennian domains depends largely on an understanding of the nature and tectonic setting of the Savo Schist Belt, of which the Pielavesi area is representative. The recognition of a suture zone would support the interpretation of the Svecofennian domain as exotic with respect to the Archean block to the east. On the other hand the recognition of similarities between the stratigraphic sequence in the Pielavesi district and the Jatulian-Kalevian deposits of the Karelian domain suggest coherent development. The relationship between magmatism and sedimentation within the Karelian domain is well preserved in the Kuusamo district and Finnish North Karelia (Table 10). Internal stratigraphic details will not be considered here since they have been examined comprehensively in recent years in the Puolanka district by, amongst others, Laajoki (1988 a, b), and in the Koli area by Kohonen and Marmo (1992). Of interest too is the position of the Northern Ostrobothnian Schist Belt, which has been correlated with the Kalevian sequence of the Karelian domain.

In the Kuusamo district volcanic rocks are more prominent in the lower part of the stratigraphic sequence (Lower Lapponi and Lower and Middle Jatuli; Silvennoinen, 1972). The most extensive volcanic phase, represented by Greenstone Formation I is correlated temporally with the 2.4 Ga layered mafic intrusions. These have no counterparts in Finnish North Karelia, where however, four younger phases of mafic volcanism and dike intrusion have been recognized, namely 2.2 Ga, 2.1 Ga, 2.05 Ga and 1.97 Ga (Pekkarinen and Lukkarinen, 1991; Vuollo et al., 1992). Volcanism within the Marine Jatulian (around 2.05 Ga) appears to become more important in the west, towards Savo. However, the paucity of isotopic age data and the complexity of deformation in this region make correlations increasingly difficult (Table 10).

The Jäkäläniemi albite diabase intrusion at Kuusamo has an age of 2206 Ma (Silvennoinen, 1991), thus constraining the age of Greenstone Formation III to approximately the same time interval as the karjalitic intrusions of Finnish North Karelia. The Dolomite Formation at Kuusamo correlates with the Marine Jatulian (Table 10), and its age is approximately dated

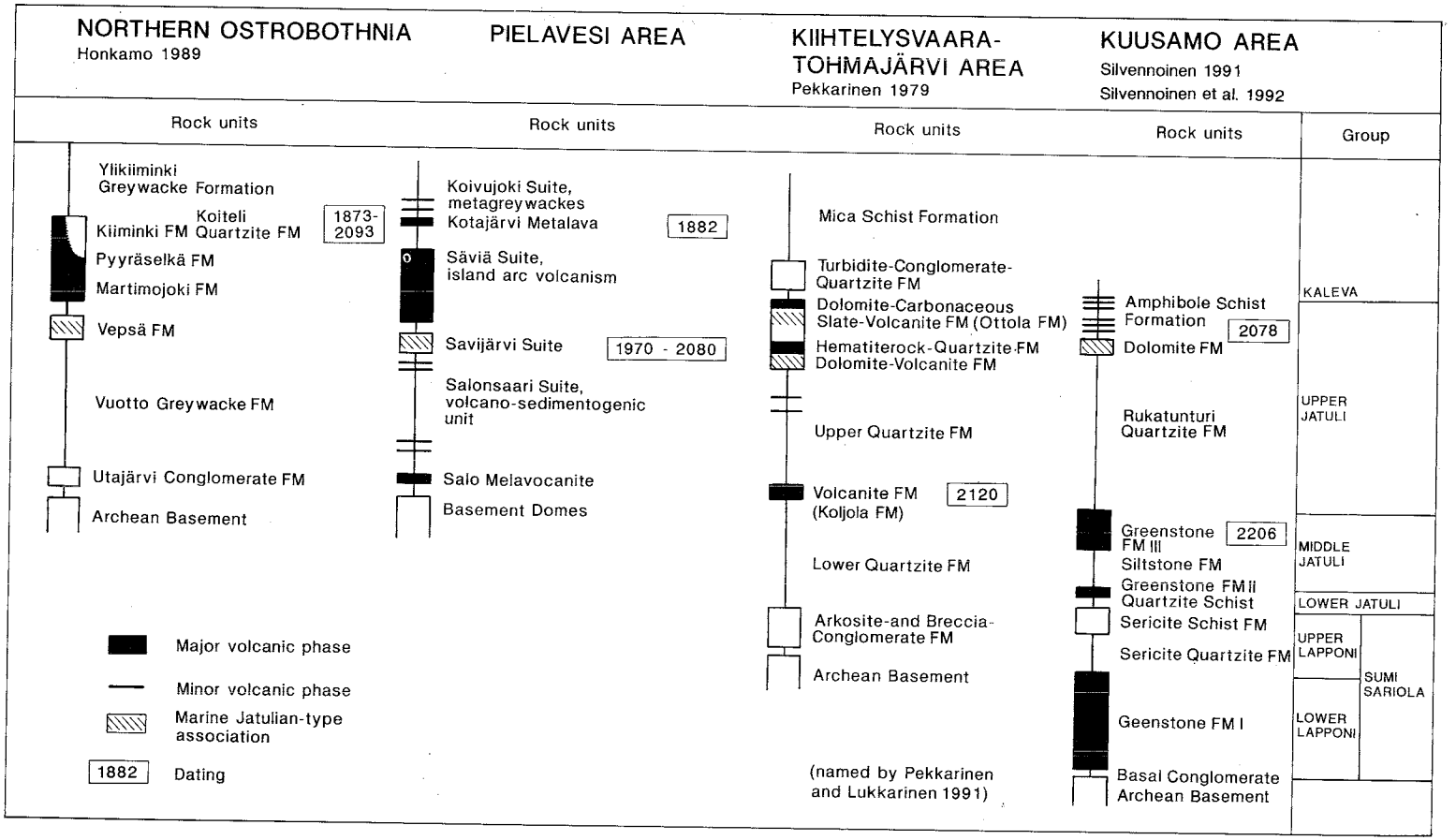


Table 10. Simplified stratigraphic sequences for northern Ostrobothnia, Pielavesi, North Karelia and Kuusamo, stressing the role of volcanism.

by the Viipus dyke (2078 Ma), which cuts the underlying Rukatunturi Quartzite Formation. In North Karelia the Jatulian quartzites are cut by tholeiitic diabase dykes that are coeval (1.97 Ga) with the Horsmanaho gabbro in the Outokumpu assemblage and the Jormua ophiolite (Koistinen, 1981; Kontinen, 1987), as well as the Rahasmäki metadiabase at Nilsiä (Paavola, 1984). The color and morphology of zircons from the Rahasmäki dyke are typical of Jatulian igneous zircons, which may call into question the rather arbitrary assignment of 2.O Ga as the upper limit of Jatulian magmatism. On the other hand, the Rahasmäki metadiabase does not correspond chemically with either the 1.97 Ga or 2.1 Ga dykes of North Karelia. Since the lower boundary of the Kalevian may be older (2.1 Ga) than originally thought, and Jatulian and Kalevian facies may temporally overlap, this may not be a problem. However, the relationship between the Outokumpu assemblage and Jatulian volcanism remains a significant problem, especially since they occur in close geographic and stratigraphic proximity to one another.

Volcanism in northern Ostrobothnia and the PIA zone appear to have occupied similar stratigraphic positions, although this was evidently more of an oceanic setting than Jatulian volcanism in the Kuopio-Siilinjärvi area (Table 6). The latter has been correlated with the volcanic Ottola Formation, which was coeval with Marine Jatulian sedimentation in the Kiihtelysvaara area (2050 Ma; Pekkarinen and Lukkarinen, 1991). On the other hand the Tohmajärvi complex has been dated at 2.1 Ga (Huhma, 1986) and is tentatively correlated with the lower volcanic unit at Kiihtelysvaara, known as the Koljola Formation (Pekkarinen, 1979; Pekkarinen and Lukkarinen, 1991), which is thus also coeval with the 2.1 Ga Fe-tholeiitic dikes in Karelia (Vuollo et al., 1992).

The Pielavesi and northern Ostrobothnian sequences show many features in common suggesting the stratigraphic correlation between the two regions may be justified (Table 10). The Salonsaari Suite is accordingly correlated with the Vuotto Graywacke Formation which overlies the Archean basement and lowermost sediments in northernmost Ostrobothnia (Honkamo, 1988, 1989). This unit is succeeded by the Vepsä Formation which includes dolomites, calc-silicate rocks, cherts, iron formations and black schists. This may be correlated with the Savijärvi Suite at Pielavesi. The overlying mafic metavolcanic rocks of the Martimojoki, Pyyräselkä and Kiiminki Formations correspond to the Säviä Suite in the Pielavesi area where however, because of its geotectonic setting, volcanism was bimodal rather than exclusively mafic. Felsic volcanic clasts from the Koiteli Conglomerate Formation in northern Ostrobothnia have been dated at 2093±35 Ma (Honkamo, 1988), which imposes a maximum age limit on deposition. The clasts are most probably derived from the immediately preceding volcanic units (Honkamo, pers. comm., 1992). The Yli-Kiiminki Formation graywackes are correlated with the Koivujoki Suite at Pielavesi. The Koiteli Formation in northern Ostrobothnia is intruded by a gabbro dated at 1873±4 Ma, which coincides with the age of intrusion of the Tyypekki Gabbro (1875 Ma) in the present study area and is only marginally younger than the Kotajärvi Metalava (1882 Ma).

The sedimentary sequence in northern Ostrobothnia were deposited directly upon Archean basement, with the characteristic Jatulian continental and epicontinental deposits being absent. This suggests a rapid transition from continental erosion to a marine depositional environment and implies that the Vuotto Graywacke Formation represents Jatulian deep water sedimentation, possibly in an oceanic setting, consistent with the age of the clasts in the overlying Koiteli Formation. In these respects correlation with the graywackes of the Höytiäinen province in North Karelia is invited, although it may represent an even earlier phase of sedimentation. Gaál (1990) considered that the well preserved turbidites, including

Table 11. Broad-scale correlation of events in the Svecofennian and Karelian domains.

SVECOFENNIAN DOMAIN (Oceanic)			KARELIAN DOMAIN (Continental-epicontinental)	
Sed. phase	Magmatism	Plate tectonic model	Magmatism	Sed. phase
UPPER SVECO-FENNIAN	Late orogenic granitoids Synorogenic magmatism	Collision orogeny	Synorogenic granitoids	1800 KALEVIAN
MIDDLE SVECO-FENNIAN	Early orogenic magmatism	Subduction with island arc - and back arc volcanism	Tholeitic volcanism and diabases	2000
LOWER SVECO-FENNIAN	New Svecofennian crust Preorogenic magmatism	Break up	Jatulian magmatism Fe-Tholeites Karjalites	Marine Jatulian JATULIAN
	?	Rifting	Layered intrusions	2300 SARIOLAN 2500
?			ARCHEAN BASEMENT	

those of the Höytiäinen province, represent relatively late foredeep basin fill postdating ophiolite emplacement. Whether or not this is so, they at least partly overlie the Marine Jatulian association and may therefore correlate more readily with the Yli-Kiiminki Graywacke Formation in northern Ostrobothnia, instead of the older Vuotto Formation. A parallel evolution for the Savo and north Ostrobothnia schist belts was also advocated by Honkamo (1985). The presence of a U-P horizon and a minor development of iron formation within the Vepsä Formation in northern Ostrobothnia further highlight the importance of this lithological association as a stratigraphic marker horizon.

The Savo Schist Belt represents a more or less coherent geological entity, which has also undergone a relatively unified metamorphic history, as implicit in the recognition of the Kiuruvesi-Haukivesi complex by Korsman et al., 1988. The Koivujoki Suite metagraywackes with their minor carbonaceous intercalations are comparable lithologically and in their overall position with the upper turbidites of the Virtasalmi and Haukivesi regions. The Haukivesi turbidites were preceded by an association reminiscent of the Marine Jatulian (Nykänen, 1975), although they are actually rather similar to the mica schist and gneisses of the Savo province in the Outokumpu district (Loukola-Ruskeeniemi et al., 1990), which are considered to be Upper Kalevian, along with the Nuasjärvi group in Kainuu (Kontinen and Sorjonen-Ward, 1991). These sediments record rapid deposition and thorough mixing of detritus within a major depositional system. The

Koivujoki Suite sediments at Pielavesi may also represent this kind of deposition and differ from the more felsic sediments in the Salonsaari Suite, which are strongly deformed and migmatized. The synorogenic Kotajärvi Metalava (1882 Ma) is younger than at least some of the uppermost graywackes in the Pielavesi region. The Koivujoki Suite resembles the Rahkamäki Formation in the Kuopio district in its general setting. Lower Kalevian Formations have not yet been recognized from the Savo province nor can they be correlated with the thick graywacke sequences underlying the bimodal volcanics in the Pielavesi area.

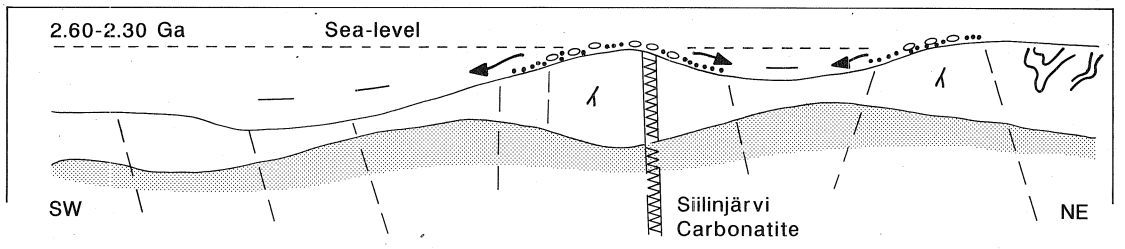
It is difficult to precisely define the southwestern limit of Archean basement at the boundary between the Karelides and Svecofennides (cf. Simonen et al., 1978) and to place a sharp boundary between the Jatulian and Kalevian sequences (Gaál and Gorbatschev, 1987). The Savo schists, straddling as they do the rifted margin of the Archean craton, contain a wide variety of rock types formed in diverse geotectonic settings. Park (1991) for instance concurs when considering that the metasedimentary and metavolcanic rocks of the Pielavesi - Kiuruvesi area belong broadly to the lower Bothnian Group deposited between 2.2-1.93 Ga. The rocks of the Kuopio area are effectively the same age as the Savo schists and the occurrence of an association in the marginal zone that is completely analogous with the Marine Jatulian, including the presence of the U-P horizon, provides further support for some kind of contiguity between the Svecofennian and Karelian domains both geographically and temporally (Table 11). Does a precise and definite boundary between the two domains really exist? The presence of Jatulian sediments and volcanic rocks in the Svecofennian clearly implies that its history began before the currently accepted date of 1.93-1.92 Ga. Most of the geological development of the Pielavesi area predates the 1883±20 Ma volcanism at Pihtipudas (Aho, 1979) immediately to the southwest of the Kinturi Fault, which also coincides with a major gravity trough. If, alternatively, the Kirkkosaari-type gneisses are regarded as truly 1.93 Ga in age, then the whole Pielavesi sequence may be correlated with the Kalevian, which again leaves unanswered the question as to where the boundary between the Karelian and Svecofennian is to be drawn. It has not so far been possible to locate this line in the field.

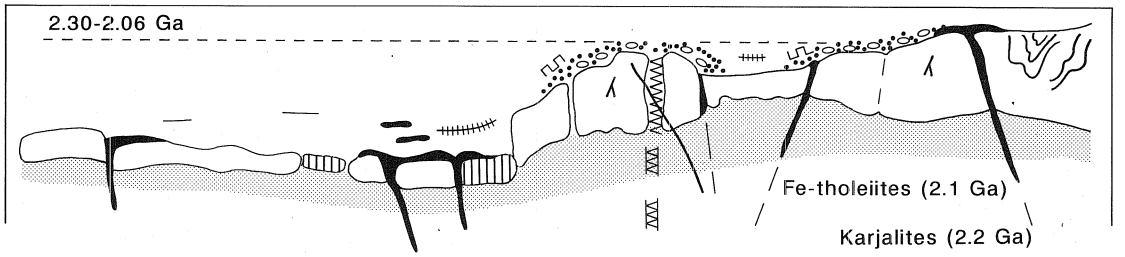
A PLATE TECTONIC MODEL

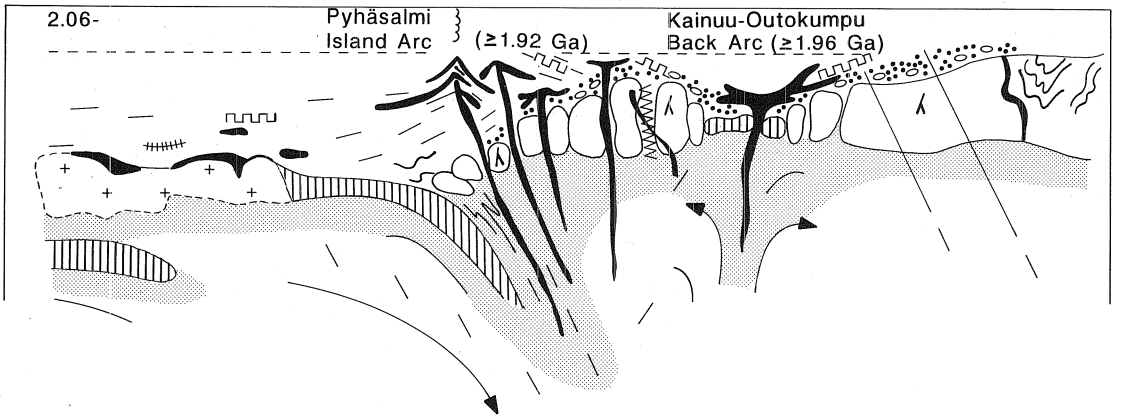
Tectonomagmatic evolution

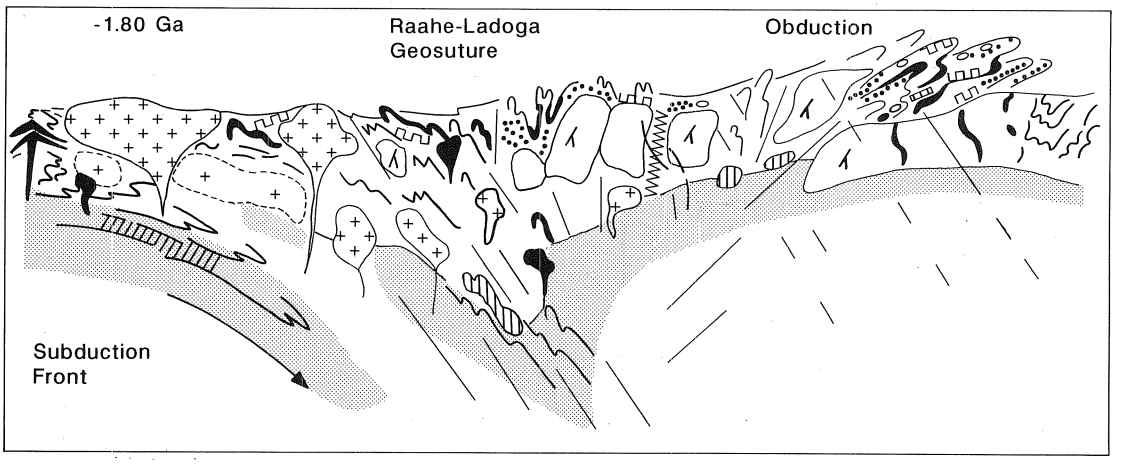
The overall evolution of the early Proterozoic continental margin is comparable with a Wilson cycle (Wilson, 1968), when considering that extension culminated in continental break-up and sea-floor spreading, followed by convergence and closure during the Svecofennian Orogeny. The earliest stages of rifting are recorded by the 2.44 Ga tholeiitic Koillismaa layered intrusions, which were interpreted as forming in an ENE-trending aulacogen. The Lower Lapponian and Sumian volcanic rocks of central Lapland, Kuusamo and Russian Karelia (Table 10) are also of this age (Sokolov and Heiskanen, 1985; Gaál, 1990; Silvennoinen, 1991). At Kuusamo there is evidence that the earliest mafic and intermediate volcanics were erupted onto dry land.

Rifting continued episodically along a predominantly NW-SE trend throughout the entire duration of Karelian sedimentation. At Reittiö near Nilsiä, Sariola-type conglomerates were deposited within NW-trending fault-controlled


2.60-2.30 Ga
Sea-level
SW
Siilinjärvi
Carbonatite
NE


2.30-2.06 Ga
Fe-tholeiites (2.1 Ga)
Karjalites (2.2 Ga)


2.06-
Pyhäsalmi
Island Arc
(≥1.92 Ga)
Kainuu-Outokumpu
Back Arc (≥1.96 Ga)


-1.80 Ga
Raahe-Ladoga
Geosuture
Obduction
Subduction
Front

depressions (Paavola, 1984). Jatulian diabase dykes also generally followed the same NW-NNW trends and frequent reactivation of faults is demonstrated by observations of mylonites both cutting and being truncated by dykes (Paavola, 1988). The present distribution of Jatulian sediments in Russian Karelia is also evidently the result of NW-trending structural controls on deposition. Piirainen (1991) however interpreted the widespread 2.2 Ga karjalitic layered intrusions cutting the lower and middle Jatulian sediments as having been emplaced during regional compression rather than extension, and related this to a postulated distant subduction event. The karjalites represent a single-stage differentiated Al-poor tholeiitic magma enriched in Fe and REE (Vuollo et al., 1992). The next major phase of magmatism in the Kiihtelysvaara-Tohmajärvi was around 2.1 Ga (Pekkarinen and Lukkarinen, 1991), where both subalkaline and tholeiitic lavas occur, consistent with an intraplate continental (WPB) setting and possibly coeval with the Fe-tholeiitic dykes intruding the Jatulian quartzites in the Koli area. The Tohmajärvi complex represents a more primitive source magma, with less evidence of crustal contamination (Nykänen, 1992) and probably reflecting significant amounts of crustal extension and thinning.

The 2.1 Ga diabase dykes in North Karelia are continental tholeiites whereas the pyroxene-bearing and uralitized hornblende diabases intruding the Archean basement at Varpaisjärvi and Sonkajärvi are in contrast, oceanic subalkaline tholeiites in character (Toivola, 1988). This may be indicative of an advanced stage of rifting. The younger, 1.97 Ga tholeiitic diabases in North Karelia share features in common with island arc tholeiites, formed in suprasubduction zone environments, even though they were intruded into the Archean craton (Vuollo et al., 1992). As with the earlier dykes, these too trend dominantly NW-SE.

Westwards from Kuopio the crust would have been thinner and subsidence more rapid (Fig. 42), with predominantly deep marine volcanism and sedimentation. The Salo Metavolcanite has low-K island arc affinities and represents a plate margin setting. The transitional Marine Jatulian stage (2.08-1.97 Ga) extended at least across the marginal zone as far as the Svecofennian domain. This represents the break-up stage and commencement of compressional deformation. The 2050 Ma carbonate-black schist sequence associated with mafic volcanism at Kiihtelysvaara (Pekkarinen, 1979) also corresponds to this transition from continental to more marine conditions. Correlatives of this sequence in the Kuopio-Siilinjärvi district include the mafic volcanics dated at 2062 Ma (Kousa, 1986), which were presumably erupted in a submarine cratonic rift or rifted continental margin, which would be appropriate given their oceanic (MORB or OFB)

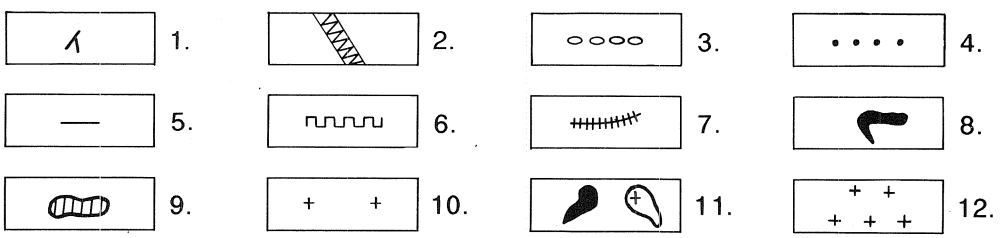


Fig. 42. Plate tectonic model for the Svecofennian Orogeny including extensional evolution and multiple subduction towards the NNE; 1 = Archean basement; 2 = Carbonatite; 3 and 4 = Continental sediments; 5 = Oceanic sediments; 6 = Shelf sediments; 7 = Black schist; 8 = Volcanism; 9 = Ocean floor; 10 = Svecofennian precursor; 11 = Synorogenic ultramafic-and porphyry-type intrusions; 12 = Svecofennian granitoids.

and within-plate characteristics (Lukkarinen, 1990). The lowermost units within the bimodal association at Siilinjärvi include spilitized Fe-tholeiitic lavas, which are followed by rhyolitic pyroclastic rocks, a variety of tuffaceous lithologies and further Fe-tholeiites. In the Otanmäki district Jatulian magmatism included layered mafic intrusions and mafic volcanics dated at 2060 Ma (Talvitie and Paarma, 1980). The Pudozhgorkan and Koikari gabbro-diabase dykes hosting Fe-Ti-V deposits in Russian Karelia may well represent the same processes, particularly in view of the Sm-Nd age of 2.0 Ga (Turchenko, 1992).

The marine transgression was synchronous with the initiation of crustal rifting beneath the Kainuu-Outokumpu zone, forming an intracratonic shallow water basin. The Outokumpu assemblage has some ophiolitic affinities and according to the age of the Horsmanaho gabbro was formed at 1972±18 Ma (Huhma, 1986). Gabbro cutting serpentinites in the Jormua ophiolite complex in Kainuu is of approximately the same age (1960 Ma) and probably records the same short-lived spreading event (Kontinen, 1987). The Jormua ophiolite does however show a more complete and better preserved oceanic crustal section than at Outokumpu, with ultramafic cumulates at the base, gabbros, sheeted dikes and pillow lavas, overlain by pelagic and flysch sediments.

Geological and geochemical studies indicate that the Jormua ophiolitic sequence formed in an incipient oceanic (MORB) setting (Kontinen, 1987). The Jormua metagabbros are slightly REE-depleted in comparison with the mafic rocks of the Outokumpu assemblage, which are more consistent with a back arc or island arc setting (Park, 1988), or a suprasubduction zone environment (Vuollo et al., 1992). The geochemical differences between the Outokumpu assemblage and the Jormua ophiolite reflects their differing geotectonic histories. The volcanic rocks of the Outokumpu area are of low-TiO_2 type (Rehtijärvi and Saastamoinen, 1985), which represents a slow-spreading environment, possibly as a result of subduction of oceanic crust. The high-TiO_2 parental magmas to the Jormua ophiolite together with the sheeted dikes suggest a relatively fast-spreading environment which may have been less favourable for hydrothermal activity and mineralization. The Troodos ophiolite in Cyprus also shows a mixture of MORB-like and island arc characteristics and may therefore represent marginal sea crust (Pearce, 1975; Thy and Moores, 1988). The relatively weak amount of spreading recorded in the Outokumpu assemblage is consistent with a back arc setting throughout the zone.

Gáal (1990) proposed that young oceanic crust was subducted westwards followed by the formation of an oceanic island arc and Ni-deposits. Subsequent collision and crustal shortening resulted in the emplacement of the Kainuu-Outokumpu oceanic crust back onto the Archean foreland between 1.93-1.90 Ga. Following collision, the sense of subduction reversed and a continental margin arc developed. However, the Svecofennian stratabound Zn-Cu deposits clearly formed before the Ni-deposits of the Raahe-Ladoga Zone and more evidence for collision at 1.93-1.90 Ga might be anticipated.

An alternative interpretation involves prolonged, possibly multiple phases of subduction in a N-NE direction towards the continental margin, with back arc spreading, as earlier suggested by Gaál (1986) (Fig. 42). The initiation of subduction is taken as coeval with Marine Jatulian sedimentation and volcanism (2.08-1.97 Ga), causing extensional rifting, vertical faulting and thermal uplift in the back arc region. This coupling of the Kainuu-Outokumpu back-arc with the Pyhäsalmi Island Arc (PIA) is analogous to the extensional opening of the Japan Sea which opened at about 40 Ma when steepening of the subducting slab took place (Ohmoto, 1983). Back arc spreading continued until 25-30 Ma, followed by gradual subsidence of the arc terrane. Bimodal volcanism with Kuroko-type deposits took place at

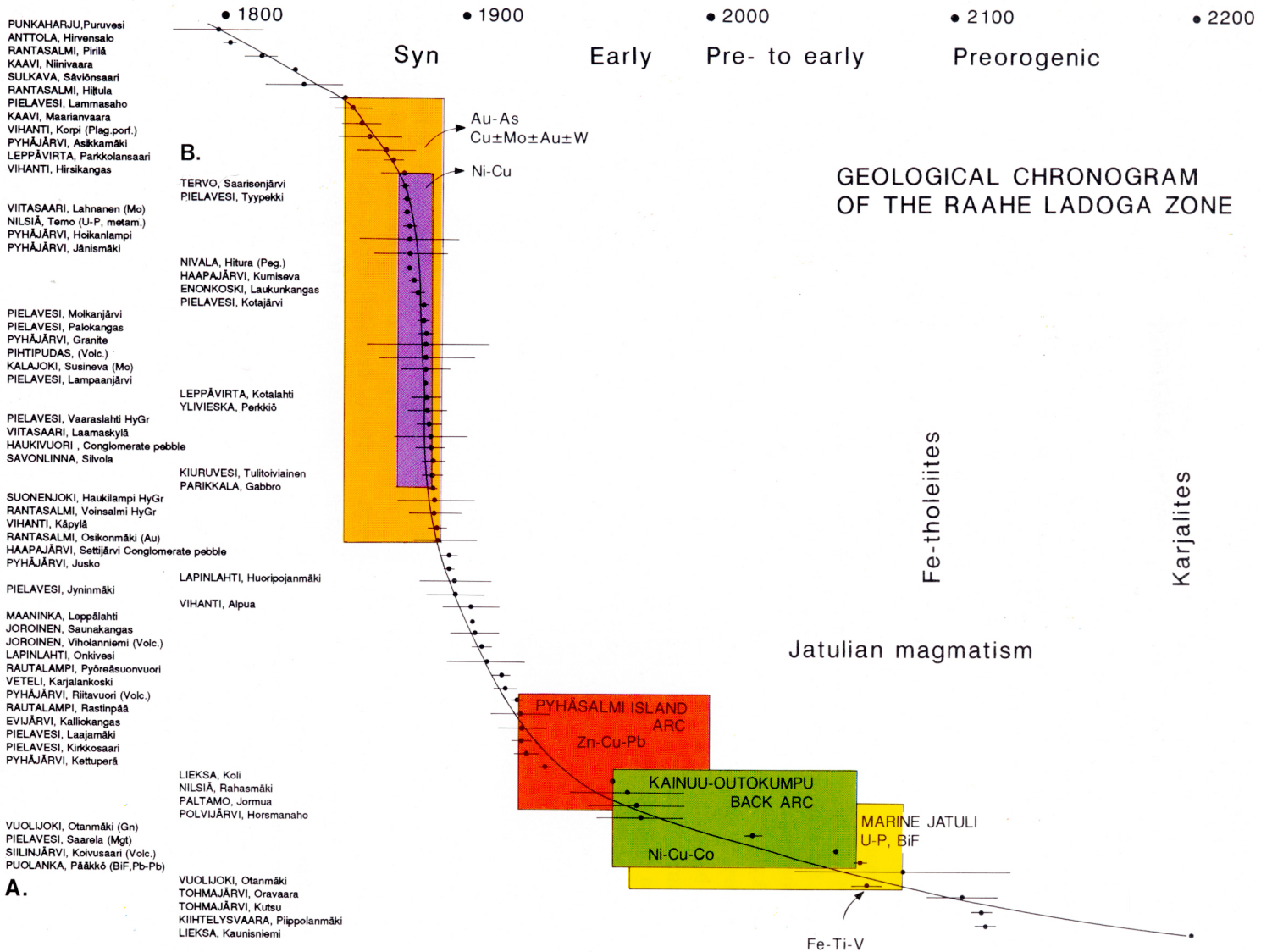
11-16 Ma during the transition from extensional to compressive deformation.

The time interval between the opening of the Japan Sea and formation of the Kuroko deposits is comparable to that between the Horsmanaho gabbro in the Outokumpu assemblage (1970 Ma) and the Riitavuori quartz porphyry (1921 Ma) at Pyhäsalmi, which has been correlated with the Ruotanen bimodal volcanism. However, since the porphyry overlies the Kettuperä gneiss, whose age of 1932 Ma has recently been regarded as indicating a metamorphic event (Marttila, 1993), it is possible that 1921 Ma is a minimum age estimate for PIA volcanism. Subduction would have been initiated long before the eruption of the Riitavuori volcanics. The bimodal Säviä Suite is underlain by the Kalliokylä Conglomerate within the shelf-sedimentary Savijärvi Suite, indicating a distinct hiatus before the onset of island arc volcanism, in contrast to the Jatulian magmatism of the Kuopio district. The tectonomagmatic character and stratigraphic setting of the volcanics, together with the whole metallogenesis of the Raahe-Ladoga Zone suggest that the Pyhäsalmi Island Arc resembles an immature continental margin arc than an oceanic island arc. Lead isotopic studies of volcanics hosting the ore deposits indicate that magmas were derived from lower crust but with a strong mantle influence (Vaasjoki and Sakko, 1988).

The trace element distributions of the northern Ostrobothnia metalavas have MORB-like characteristics (Honkamo, 1989), indicating back-arc spreading or an incipient intracontinental spreading system. In the orogenic model, coeval formations were possibly deposited in an arc setting or subduction zone in central Ostrobothnia and northern Savo (Honkamo, 1985). In southern Ostrobothnia these arc-like characteristics are replaced by basalts having within-plate (WPB) or E-MORB and N-MORB affinities (Vaarma, 1990).

On a large scale the evolution of the Raahe-Ladoga Zone can be compared to a Wilson cycle, as is indicated by the chronogram based on 70 zircon ages determined by the GSF. The chronogram forms a gentle Z-shaped curve (Fig. 43) with two distinct breaks, at 2.06-1.97 Ga and 1.97-1.93 Ga. The chronogram indicates that the total duration of the orogenic processes was at least 150 Ma. Orogenesis can be considered to have commenced at about 1.97 Ga, as mentioned by Koistinen (1981) in connection with the initiation of tectonism in the Outokumpu region. Bimodal arc volcanism also represents the early orogenic stage, whereas the synorogenic stage is characterized by tholeiitic to calc-alkaline magmatism between 1.90-1.87 Ga close to the active continental margin. The I-type characterisitics of the granitoids can be accounted for by appealing to Andino-type plate tectonics (Nurmi and Haapala, 1986). Subduction-related granitoids are more or less coeval with the Svecofennian volcanism in southwestern Finland. Deformational and metamorphic events culminated between 1.89-1.88 Ga with the generation of migmatites and pyroxene granites. Synorogenic granitoids are known from Vihanti (Käpylä, Hirsikangas) to Rantasalmi (Tuusmäki, Osikonmäki). Post-tectonic (1.80 Ga) granitoids are represented by the Puruvesi and Hirvensalo granodiorites in southern Savo (Vaasjoki and Sakko, 1988). In the Rantasalmi-Sulkava area migmatization also took place between 1.83-1.80 Ga, which indicates that tectonomagmatic activity persisted around the Sulkava thermal dome for about 50 Ma after it had ceased elsewhere. In contrast, at Pielavesi the age of the Lammasaho granodiorite indicates that tectonic activity had ceased in that area by about 1.85 Ga and in the Vihanti area at about 1.86 Ga (Vaasjoki and Sakko, 1988).

Sm-Nd results obtained by Huhma (1986) indicate a significant older crustal component in granitoids intruded into the Karelian domain, such as the Maarianvaara granodiorite (1857±8 Ma) and the Puruvesi granite (1797±19 Ma), consistent with anatexis of Archean crust at depth. This contrasts with ε_{Nd} data from Svecofennian granitoids such as that at Molkan-


GEOLOGICAL CHRONOGRAM
OF THE RAAHE LADOGA ZONE
1800
1900
2000
2100
2200
Syn
Early
Pre- to early
Preorogenic
Au-As
Cu±Mo±Au±W
Ni-Cu
B.
PUNKAHARJU,Puruvesi
ANTTOLA, Hirvensalo
RANTASALMI, Pirilä
KAAVI, Niinivaara
SULKAVA, Säviönsaari
RANTASALMI, Hiltula
PIELAVESI, Lammasaho
KAAVI, Maarianvaara
VIHANTI, Korpi (Plag.porf.)
PYHÄJÄRVI, Asikkamäki
LEPPÄVIRTA, Parkkolansaari
VIHANTI, Hirsikangas
TERVO, Saarisenjärvi
PIELAVESI, Tyypekki
VIITASAARI, Lahnanen (Mo)
NILSIÄ, Temo (U-P, metam.)
PYHÄJÄRVI, Hoikanlampi
PYHÄJÄRVI, Jänismäki
NIVALA, Hitura (Peg.)
HAAPAJÄRVI, Kumiseva
ENONKOSKI, Laukunkangas
PIELAVESI, Kotajärvi
PIELAVESI, Molkanjärvi
PIELAVESI, Palokangas
PYHÄJÄRVI, Granite
PIHTIPUDAS, (Volc.)
KALAJOKI, Susineva (Mo)
PIELAVESI, Lampaanjärvi
LEPPÄVIRTA, Kotalahti
YLIVIESKA, Perkkiö
PIELAVESI, Vaaraslahti HyGr
VIITASAARI, Laamaskylä
HAUKIVUORI , Conglomerate pebble
SAVONLINNA, Silvola
KIURUVESI, Tulitoiviainen
PARIKKALA, Gabbro
SUONENJOKI, Haukilampi HyGr
RANTASALMI, Voinsalmi HyGr
VIHANTI, Käpylä
RANTASALMI, Osikonmäki (Au)
HAAPAJÄRVI, Settijärvi Conglomerate pebble
PYHÄJÄRVI, Jusko
LAPINLAHTI, Huoripojanmäki
PIELAVESI, Jyninmäki
VIHANTI, Alpua
MAANINKA, Leppälahti
JOROINEN, Saunakangas
JOROINEN, Viholanniemi (Volc.)
LAPINLAHTI, Onkivesi
RAUTALAMPI, Pyöreäsuonvuori
VETELI, Karjalankoski
PYHÄJÄRVI, Riitavuori (Volc.)
RAUTALAMPI, Rastinpää
EVIJÄRVI, Kalliokangas
PIELAVESI, Laajamäki
PIELAVESI, Kirkkosaari
PYHÄJÄRVI, Kettuperä
LIEKSA, Koli
NILSIÄ, Rahasmäki
PALTAMO, Jormua
POLVIJÄRVI, Horsmanaho
VUOLIJOKI, Otanmäki (Gn)
PIELAVESI, Saarela (Mgt)
SIILINJÄRVI, Koivusaari (Volc.)
PUOLANKA, Pääkkö (BiF,Pb-Pb)
A.
VUOLIJOKI, Otanmäki
TOHMAJÄRVI, Oravaara
TOHMAJÄRVI, Kutsu
KIIHTELYSVAARA, Piippolanmäki
LIEKSA, Kaunisniemi
Fe-tholeiites
Karjalites
Jatulian magmatism
PYHÄSALMI ISLAND ARC
Zn-Cu-Pb
KAINUU-OUTOKUMPU BACK ARC
Ni-Cu-Co
MARINE JATULI
U-P, BiF
Fe-Ti-V

järvi at Pielavesi (U-Pb zircon age 1882±22 Ma), which appear to contain a substantial input from early Proterozoic mantle. There is also little evidence of an Archean precursor to the extensive granitoid complex in central Finland. Similarly the pyroxene granites throughout the Raahe-Ladoga Zone record relatively short crustal residence times before emplacement (Huhma, 1986). These results demonstrate that significant volumes of juvenile crust were added to the continental margin during the Svecofennian Orogeny, with a general tendency for magmatism and metamorphism to young southwards.

The metallogenic implications of the isotopic chronogram with regard to the Raahe-Ladoga Zone will be discussed later in this study.

Tectonics of the Raahe-Ladoga Zone

Although the Raahe-Ladoga Zone (RLZ) is primarily a synorogenic feature, it also contains various elements recording earlier events, including extension, rifting and break-up, and partly involving the reactivation of older basement structures. The Marine Jatulian depositional phase is taken to mark the commencement of compressional activity.

The RLZ is characterized by dextral faults and shear zones that were generated by prolonged plate convergence along a S-SW vector. The effects of this deformation are apparent for a great distance into the foreland. For instance Archean structures have been completely reworked over wide areas near Nunnanlahti, principally in proximity to major shear zones (Kohonen et al., 1991). On the other hand between the zones of reworking, the basement remained very little affected (Luukkonen, 1985). Dextral NW-trending wrench faults with some antithetic conjugate NE-trending faults and dextral N-trending zones dominate the structural pattern of the Archean craton over an area at least 150 km wide (Fig. 44).

Conjugate fracture systems have also controlled volcanism in the PIA zone, with volcanic zones and complexes defining an en echélon chain (Fig. 45 a). Subduction resulted in D_1 recumbent folds and N-NE directed thrusting which were progressively steepened during D_2. This progressive deformational event was associated with amphibolite and granulite facies metamorphism in the marginal zone which, in the Pielavesi area, climaxed prior to the emplacement of the pyroxene granites at 1884 Ma. The Pielavesi area forms part of the Kiuruvesi-Haukivesi complex (which also includes the Rautalampi and Virtasalmi regions), in which deformation and metamorphism both commenced and ceased earlier than in the area surrounding the Sulkava thermal dome (Korsman et al., 1988). Continued convergence resulted in the formation of dextral shear zones along the continental margin with thrusting of the tightly folded trench sequence towards the NW and NNW (Fig. 45 b). This deformation resulted in the formation of the prominent S_3 transverse foliation and SE-plunging L_3 mineral elongation lineation in the Pielavesi area (Figs. 31 and 32) and is also evident within the rigid Archean basement, recorded by block movements in the area covered by the Lapin-

Fig. 43. Descriptive chronogram of the Raahe-Ladoga Zone; compiled from about 70 U-Pb zircon data (y-axis) A = granitoids (exceptions in parentheses) and B = mafic lithologies. Main mineralization epochs are indicated in different colors.

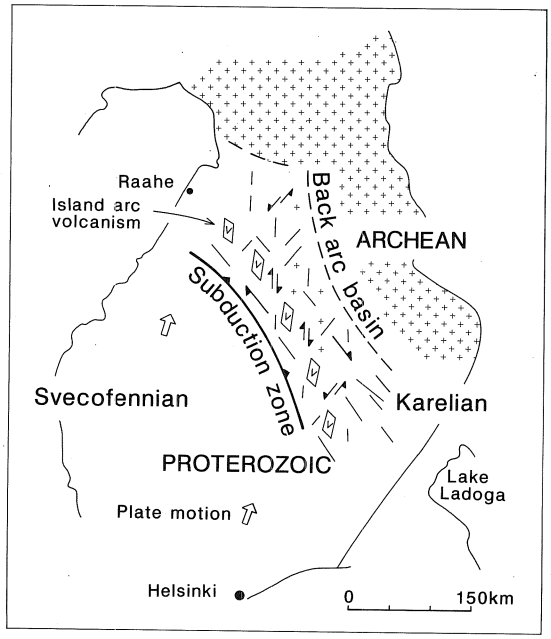


Fig. 44. Plate tectonic interpretation of the evolution of the Raahe-Ladoga Zone.

lahti 1 : 100 000 map sheet (Paavola, 1988). Structures overturned towards the NW or N can be observed along the marginal zone as far as Suonenjoki in the south and Kärsämäki in the northwest. The orientation of ore deposits throughout the marginal zone was largely determined by this D_3 phase of deformation, which must have been a prolonged event, continuing until the intrusion of the late orogenic granites. D_3 deformation also influenced the geometry of metamorphic domains in the Kiuruvesi-Haukivesi complex (Korsman et al., 1988). Late stage movement along these structures is demonstrated by brittle displacements within the Kumiseva gabbro. The NW- NE- and N-trending fault zones that were initiated in the crust during subduction were frequently reactivated resulting in vertical differential block movements during late stages of deformation.

Long-lived wrench faults with conjugate shear and fault zones also repeatedly deformed the PIA zone. The most prominent sinistral deformation zones are apparent in the Auho, Lamujärvi and Säviä zones, although it is also discernible in the Outokumpu-Koli area. The Suvasvesi Shear Zone branches and curves east of Pielavesi into an intensely deformed N-trending zone, whereas the Haukivesi Shear Zone is partly overprinted by brittle deformation along the Kinturi Fault. N-trending deformation zones are charateristic of the whole Savo-Kainuu region.

The Nunnanlahti-Holinmäki Shear Zone in North Karelia was interpreted as broadly coeval with the subparallel Haukivesi and Suvasvesi Shear Zones (Kohonen et al., 1991) although it is sinistral rather than dextral. Northwards it converges with the Laakajärvi Fault at the southern end of the Kainuu Schist Belt. The Närhiniemi stratabound mineralization at Sotkamo has been remobilized into the SSE-trending D_3 lineation (Juhanen, 1989) and a similar structural control was evident at the Hammaslahti Cu-Zn deposit (Loukola-Ruskeeniemi et al., 1990).

Probably as a consequence of prolonged subduction, the Kainuu-Outokumpu back arc was emplaced onto the continental foreland some time after 1.97 Ga (Fig. 42), with the vergence of the nappes coinciding with the direction of subduction (Ward, 1987). The strongly developed SW-plunging lineation associated with this thrusting event in both Archean basement and Proterozoic cover has been described by numerous authors (Wegmann, 1928; Väyrynen, 1933, 1954; Koistinen, 1981; Paavola, 1984, 1988; Park, 1988 a). The syngenetic ore deposits of the Outokumpu region were also polydeformed and are tectonically aligned within the NE-SW thrusting direction. Thrusts are best developed to the N of the Suvasvesi Shear Zone, although deformed remnants of the thrusting and its associated lineation are found further southwest near Juva and Leppävirta. Similarly, N-NE directed thrusting in Kainuu is recorded by thrust-nappes within the Kainuu Schist Belt itself, as well as within Archean and Proterozoic rocks further southwest, virtually

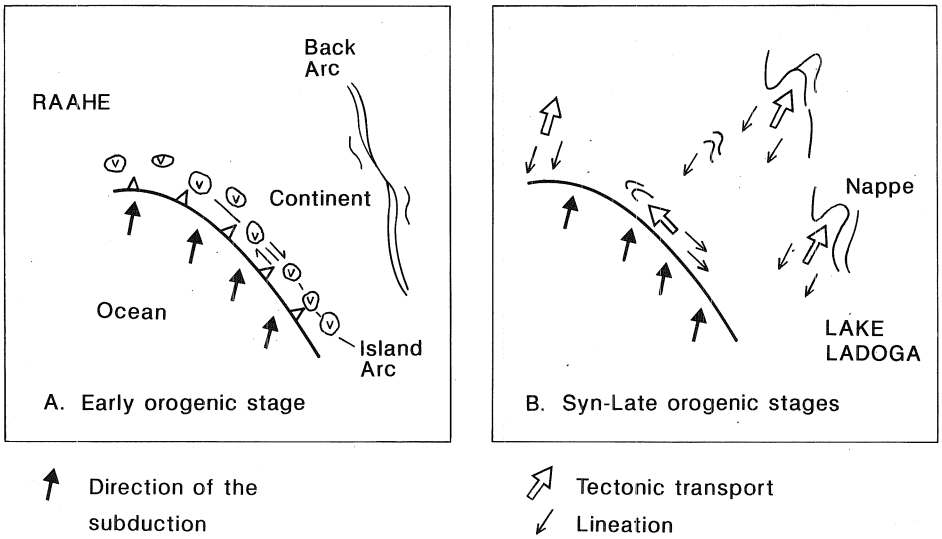


Fig. 45. Schematic illustration of tectonic events in the Raahe-Ladoga Zone.

as far as the PIA zone.

Approaching the Vihanti area, the significance of NW-trending shearing and faults diminished, as the angle of plate convergence increased (Fig.44). The Lampinsaari ore is overthrust towards the NE and accordingly aligned along a SW-plunging trend (Rouhunkoski, 1968). Prolonged subduction is also evident in the NE-directed thrusting in the schists of northern Ostrobothnia. The D_3 deformation along the Haukivesi and Suvasvesi shear zones took place within the D_2-D_3 interval defined by Koistinen (1981) for the Outokumpu district. Deformation throughout the RLZ has been a progressive, continuous process resulting from protracted compression. Volcanism, ore formation and later reorientation and remobilization of ore deposits has been controlled principally by N-NW-trending synthetic and N-NE-trending antithetic shear and fault zones. Brittle deformation during the final stages of orogeny was also controlled by reactivation of the same structures.

Recognition of the actual trace of the subduction zone is difficult. A gravity trough coincident partly with the Kinturi and partly with the Haukivesi faults could mark the site of a primary trench. This zone also coincides with the boundary between the granulite facies Pielavesi block to the east and the lower grade metamorphism in the Korppinen and the Pihtipudas blocks to the southwest (Hölttä, 1988). The SVEKA seismic profile also indicates that there is a disruption to the Moho in this region (Luosto et al., 1984).

METALLOGENESIS OF THE RAAHE-LADOGA ZONE

Definition of metallogenic epochs.

The terms Raahe-Ladoga Zone (RLZ) and Main Sulphide Ore Belt were adopted in the 1960's and 1970's with reference to that part of the early Proterozoic orogenic zone in which there was believed to be a strong srike-slip fault control on the distribution of mineral deposits (Paarma and Marmo, 1961; Härme, 1961; Talvitie, 1971; Gaál, 1972; Kahma, 1973). Kahma (1973) defined the Main Sulphide Ore Belt more specifically as a zone 140-150 km wide and at least 400 km long, situated between a major gravity minimum and the Archean basement complex to the NE. The Belt thus included the Outokumpu Ore District, the Kotalahti Nickel-Copper Ore Zone and the Vihanti Zink Ore Zone.

Piirainen (1975) subsequently presented a plate tectonic context for this mineralization which has since been developed and refined by many workers (cf. Mäkelä, 1980; Koistinen, 1981; Gaál, 1982, 1986, 1990; Park et al., 1984; Park, 1985). The RLZ should be examined as the culmination of a complete spectrum of geological processes, from early extension through to subduction, metamorphism and uplift, since many of the major orogenic structures may have been inherited or reactivated extensional features. Tectonically speaking the RLZ is at least 150 km wide and can be considered to extend as far as the Kainuu Schist Belt in the northeast (Fig.46). The most significant mineralization epochs appear to have been associated with magmatism immediately prior to and during orogenesis. With the exception of the iron formations and the U-P horizon within the Marine Jatulian association, there does not appear to have been substantial ore formation during the earliest stages of Proterozoic sedimentation and rifting, although some mineralization related to Jatulian volcanism has been reported. Within the context of the plate-tectonic orogenic model presented above, the most important tectonic elements of the RLZ from the viewpoint of known and potential mineralization are the Kainuu-Outokumpu Back Arc, Pyhäsalmi Island Arc and the synorogenic intrusions hosting syngenetic and epigenetic mineralization.

The appended general geological and metallogenic map of the Raahe-Ladoga Zone is a compilation based on both published material and unpublished data housed in the archives of the GSF and Outokumpu Co. Where published information on a particular mineralization is available, it has generally been referred to in the text. Special attention has been given to genetic affinities amongst ore deposits and occurrences when compiling the map, so that tectonic features have accordingly been simplified, with emphasis on subduction-related fault and shear movements.

Five principal epochs of mineralization and metallogenic provinces can be recognized in the development of the RLZ (Fig. 46).

1) Preorogenic stage
   - Karjalites (2.2 Ga)
   - Fe-tholeiites (2.1 Ga)

2) Pre- to early orogenic stage
   - Stratiform U-P occurrences and iron formations within the Marine Jatulian association (ca. 2.08-1.97 Ga)
   - Layered intrusions containing Ti-V-Fe deposits (ca. 2.06 Ga)

3) Early orogenic stage
   - Kainuu-Outokumpu Back Arc with Cyprus-type deposits (≥1.96 Ga)
   - Pyhäsalmi Island Arc with Kuroko-type deposits (≥1.92 Ga)

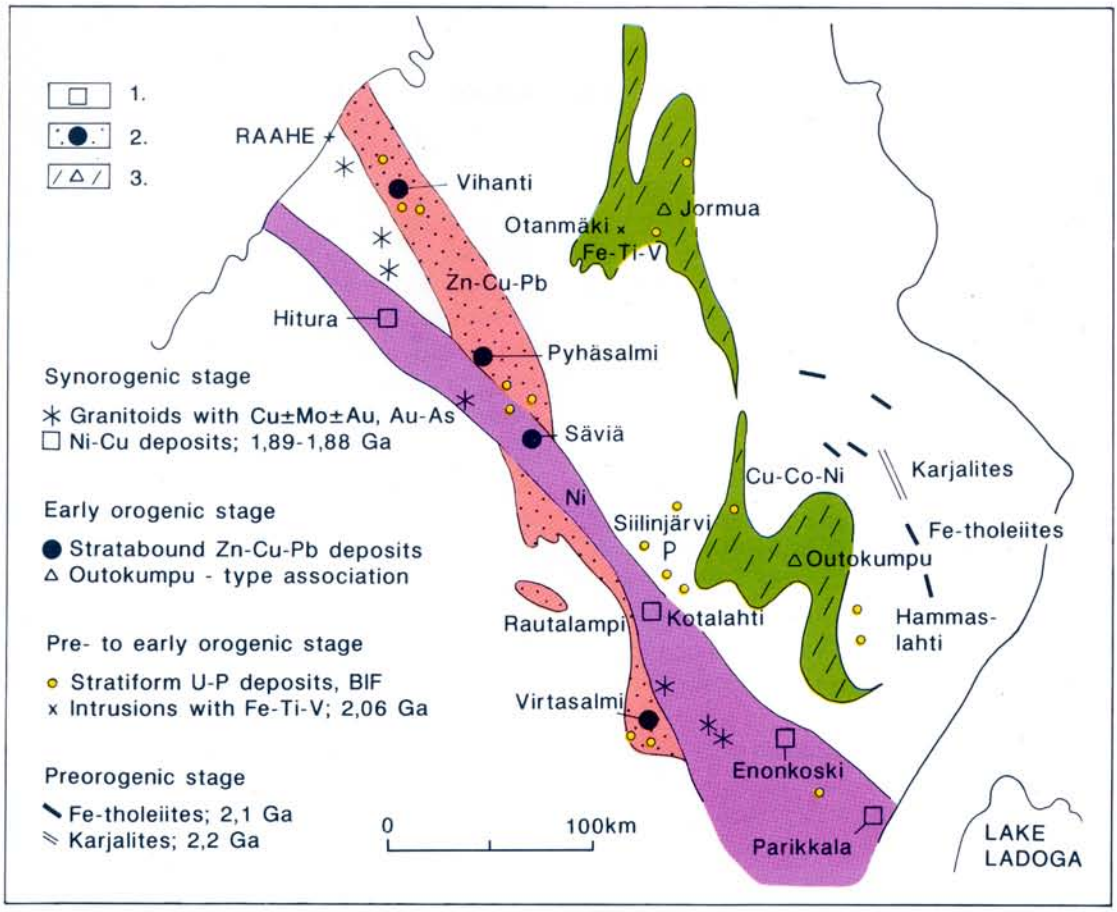


Fig. 46. Metallogenic provinces and epochs within the Raahe-Ladoga Zone. 1. Kotalahti Ni-Belt; 2. Pyhäsalmi Island Arc; 3. Kainuu-Outokumpu Back Arc. U-P deposits are marked with small yellow circles.

4) Synorogenic stage
   - Kotalahti Ni-Belt (ca. 1.89-1.88 Ga)
   - Porphyry type Cu±Mo±Au±W occurrences (≤1.88 Ga)

5) Syn-late orogenic stage
   - Epigenetic Au-As occurrences (ca. 1.88-1.85 Ga)

Preorogenic stage

Karjalites (2.2 Ga)

The Koli layered sill, which has been studied in detail in a profile at Kaunisniemi (Vuollo, 1988), is representative of the widespread karjalitic (or gabbro-wehrlitic) magmatism throughout the Karelian domain. The Koli intrusion represents fractional crystallization from a single pulse of Al-poor tholeiitic magma intruded into both the Archean basement and arkosites of the overlying Jatulian cover sequence at about 2.2 Ga (Vuollo et al., 1992). The parent magma was also fluid-rich and characterized by high FeO and TiO_2 contents and strong LREE enrichments (Vuollo, 1988). The abundance of Pt in the parent magma was evidently limited, so that it is rather unlikely that the karjalitic intrusions host significant PGE deposits. However, fluid circulation connected

with this magmatism resulted in vein-hosted Cu-Au mineralization in a number of places, notably at Kivimaa, near Tervola (Rouhunkoski and Isokangas, 1974). The predominantly continental conditions were probably responsible for the absence of significant sediment-hosted deposits, although the potential for mineralization of this age may be greater in Lapland.

Fe-tholeiites (2.1 Ga)

The majority of diabase dykes intruding the Archean craton and Jatulian sequences in Savo and Karelia appear to belong to the 2.1 Ga Fe-tholeiitic group, which are distinguished from the younger, 1.97 Ga tholeiites by their distinctly higher TiO_2, K_2O, P_2O_5 and $Fe_2O_{3(total)}$ abundances. The Fe-tholeiites represent 10-20% partial melting of the mantle and a highly fractionated source magma (Vuollo et al., 1992).

Piirainen (1991) considered this phase of magmatism to have mineralization potential because, along with Ward (1987, 1988), he regarded the Fe tholeiites as correlative with the Marine Jatulian and the formation of the Hammaslahti Cu-Zn deposit, which has been interpreted as intermediate between Besshi and sediment-hosted types (Loukola-Ruskeeniemi et al., 1990). Because there is evidence that the graywackes hosting the deposit are interdigitated with mafic turbidites related to the Tohmajärvi volcanic complex, Ward (1987, 1988) proposed that mineralization was a result of distal volcanic exhalative processes or else diagenetic hydrothermal activity in the subsiding basin after the cessation of volcanism. Thus, it may be that the Hammaslahti ore is older than Marine Jatulian since the Tohmajärvi complex has been dated as 2.1 Ga (Huhma, 1986), which contrasts with the 2.05 Ga age for the Ottola Formation volcanics within the Marine Jatulian section nearby at Kiihtelysvaara (Pekkarinen and Lukkarinen, 1991). On the other hand, at Tikkala, between Hammaslahti and Tohmajärvi, some turbidites contain pink dolomite clasts that may have been derived from the Marine Jatulian shelf (Ward, 1988), while Nykänen (1968) described gray dolomites, black schists and quartzites from the same area. Because of the structural complexity of the region, primary stratigraphic relationships are uncertain, but it is clearly possible that the Hammaslahti mineralization occurs within this association and, by implication, is Marine Jatulian after all. A connexion with the 1.97 Ga magmatism has also been considerd (Nykänen, 1992). The only other ore showings in Karelia that can be ascribed to the Fe-tholeiitic magmatism are chalcopyrite-bearing veins and uranium enrichment in proximity to dikes intruded into quartzites in the Eno district.

Pre- to early orogenic stage

Iron formations and U-P horizon

The Marine Jatulian association, comprising dolomites/ carbonates, calc-silicate rocks, felsic volcanics, cherts, U-P layers, banded iron formations and black schists represents shallow marine depositional environments. Organically derived carbonate, phosphate and carbon compounds were precipitated in association with sporadic volcanism and chemical sedimentation on the shelf and within continental rifts.

In the Kainuu district extensive Superior-type iron formations were deposited; elsewhere, at Nilsiä, Kuopio and throughout North Karelia, such deposits are present, but not common. This may be a consequence of a more open depositional basin. Changes in conditions of deposition are evident vertically as well as laterally. Both the iron formations and the closely associated U-P horizon have considerable significance as regional stratigraphic markers throughout the whole of the RLZ. In the Karelian domain they preceded Upper Jatulian volcanism and in the PIA area they underlie the island arc volcanics. Therefore, they represent the only phase of mineralization that

can be found in both the Svecofennian and Karelian domains.

None of these deposits have proved to be economically important, with Fe contents at most around 20-30% and forming relatively thin units. The mean concentration of U in the U-P horizon is about 0.04% and P_2O_5 generally remains less than 5%. A total of 42 U-P occurrences have nevertheless been found but not all have been adequately examined.

This lithostratigraphic sequence is also present far beyond the RLZ. As well as being present in central and northern Ostrobothnia (Äikäs, 1988c; Lindmark, 1977; Vanne, 1978), iron formations and U-P layers have been identified in the upper part of the Jatuli sequence in the Peräpohja district (Äikäs, 1980) and can be traced from there westwards across the Swedish border to as far as Kalix. These also contain stromatolitic dolomites and include the Pålänge U-P occurrence (Äikäs, 1988b). Studies in the Pielavesi area support the interpretation of these lithologies within the transitional Marine Jatulian association as an important link in determining the relationship between the Karelian and Svecofennian domains during the pre- to early orogenic stage (2.08-1.97 Ga). The wide distribution of these horizons implies the existence of an extensive marine shelf with conditions conductive to the precipitation of phosphate (cf, Sokolov and Heiskanen, 1985).

Layered intrusions hosting Fe-V-Ti deposits

The arcuate zone of schists in the Otanmäki district follows the trend of a distinctive anorthosite-gabbro-amphibolite association. The metamorphosed Otanmäki layered intrusion was associated with Jatulian magmatism (2060±4 Ma; Talvitie and Paarma, 1980) and is thus effectively the same age as volcanism in the Kuopio-Siilinjärvi district (2062 Ma; Kousa, 1986). The magma was relatively anhydrous with plagioclase as the dominant cumulus phase and sulfides were removed with a melt phase due to liquid immiscibility, leaving oxides to crystallize with late-stage enrichment of Fe, Ti and V (Piirainen, 1975). The emplacement of the Otanmäki gabbro appears to have been controlled by NW- and W-trending fracture systems. According to McLelland (1986), such intrusions are characteristic of anorogenic magmatism which may be a prelude to crustal attenuation and rifting. The Otanmäki deposit and the host intrusions show many similarities with the anorthosites and gabbros associated with Fe-Ti-V mineralization at Karhujupukka in the Kolari area in western Lapland (Karvinen et al., 1988).

The amphibolites, leucogabbros and serpentinites at Koutaniemi, NW of Kajaani have been correlated with the Otanmäki gabbro (Kataikko, 1987), though in some respects the volcanic and sedimentary formations are somewhat reminiscent of those associated with the Vihanti and Pyhäsalmi ore deposits. Moreover, the association consists of carbonates, calc-silicate rocks, cherts, felsic pyroclastics, quartz-feldspar schists with some U-P enrichment. A tremolite-diopside skarn immediately to the east of these contains Fe-Cu-Ni-Zn mineralization. On the Ni-Co diagram presented for the Outokumpu assemblage, the Koutaniemi skarns fall within the mica gneiss and black schist field, while the tremolite-sulphide schists plot in the Ni-mineralization field and the serpentinites plot very close to the Outokumpu serpentinites (Kataikko, 1987). The Koutaniemi mineralization shows a pronounced NW-SE tectonic control.

The Otanmäki-Koutaniemi area is clearly of key importance in trying to better understand the nature of magmatism during the poorly studied 2.06-1.93 Ga time interval. The area shows similarities with both the Outokumpu and the Pyhäsalmi types of mineralization. Studies are complicated by the later intrusion of the Otanmäki alkali gneiss which has a U-Pb zircon age of 1964±6 Ma (Laajoki, 1992), which is remarkably close to the age of the Jormua ophiolite.

Early orogenic stage

Kainuu-Outokumpu Back Arc with Cyprus-type deposits

Serpeninites and black schists outcropping discontinuously for almost 300 km between Outokumpu and Kainuu represent a single tectonostratigraphic unit recording rifting and seafloor spreading (Talvitie et al., 1980; Koistinen, 1981; Kontinen, 1987; Loukola-Ruskeeniemi et al., 1991). The ophiolitic nature of the northern part of this zone has been substantiated by the finding of sheeted dike swarms within the Jormua complex, as well as its overall chemistry (Kontinen, 1987), while recent evidence for the presence of podiform chromite mineralization within the Outokumpu assemblage strengthens the argument that the allochthonous Outokumpu assemblage also contains ophiolitic material (Vuollo, 1992). The zone contains the worked out Keretti, Vuonos and Luikonlahti ore deposits as well as an additional 30 smaller mineralizations. The total in situ metal budget has been estimated at 1.350 kt Cu, 440 kt Zn, 65 kt Ni, 75 kt Co and 20 t Au. The estimates for Cu, Zn and Ni increase substantially if the very large but rather low-grade Talvivaara deposit, which occurs within a more distal black-schist association, is included. On the basis of the eight largest deposits, mean metal abundances have been calculated as 2.7% Cu, 1.0% Zn, 0.2% Co, 0.1% Ni, 0.1 g/t Au, 26% Fe and 21% S (Gaál, 1990). The ores are strongly deficient in Pb and Outokumpu in particular shows very primitive, early Proterozoic (ca. 2.1 Ga) mantle characteristics (Vaasjoki, 1981).The Hammaslahti lead is somewhat exceptional however, with a model age of about 2.3 Ga, which probably records significant Archean crustal component (Vaasjoki, 1981).

Outokumpu-type ore bodies occur within a distinctive association that includes serpentinite, dolomite, calc-silicate skarns and cherts enclosed within black schists (Gaál et al., 1975). The calc-silicate rocks consist of Cr-diopside and Cr-tremolite with some uvarovite and the whole assemblage is overlain by, and thrusted on top of Kalevian mica schists. The ore bodies are stratabound and consist of massive pyrite, pyrrhotite, chalcopyrite and sphalerite, which has often been tectonically separated from more diffuse stockwork mineralization (Koistinen, 1981). The stockwork zones with disseminated Cu-Co-Zn-Ni mineralization and the presence of cordierite-anthophyllite rocks indicate the presence of permeable zones facilitating hydrothermal fluid flow (Treloar et al., 1981). The ore bodies have a general conical form, with an upwards increase in the above-mentioned metals. Cobalt is found in both pyrite and pentlandite. An ophiolitic setting would also be consistent given the similarities between the Outokumpu deposits and those of Cyprus, where hydrothermal activity has produced massive ore bodies in the upper parts of volcanic sequences closely associated with cherts. The Outokumpu ores are also closely connected with calc-silicate lithologies and pelagic sediments were also enriched in various metals due to seafloor hydrothermal activity.

Black schists within the Kainuu-Outokumpu zone can be subdivided into several distinct groups on the basis of their geochemistry (Loukola-Ruskeeniemi, 1992). The Talvivaara-type black schists between Kainuu and Losomäki have distinctly higher Ni, Cu and Zn abundances than those associated with the Outokumpu ores which, on the other hand, are closely associated with calc-silicate and chert horizons as far north as Alanen near Sotkamo. Manganese contents are elevated only in the area between Paltamo and Sotkamo but black schists throughout the whole Kainuu-Outokumpu zone are characterized by relatively high C contents. According to Loukola-Ruskeeniemi (1992) these geochemical differences may reflect differences in spreading regimes, which is in agreement with observed differences in the geochemistry of mafic rocks at Outokumpu and Jormua (Kontinen, 1987). Thus, Jormua may represent somewhat later and more rapid

spreading than recorded at Outokumpu. Hydrothermal activity must also have occurred in the Kainuu region, in order to explain the high metal enrichments in the black schists, even though depositional environments differed from those at Outokumpu (Loukola-Ruskeeniemi, 1992). At Tiittola near Paltamo vein-type Cu mineralization occurs in serpentinite in contact with chlorite schists and mafic volcanics (Ervamaa, 1972; Heino, 1986), which may indicate potential for finding Cyprus-type mineralization in Kainuu as well. Layered intrusions (metadiabases, metagabbros) cut both volcanic rocks and serpentinites in this area.

Petrographical studies indicate that the serpentinites consisted originally of dunites, harzburgites, lherzolites and wehrlites representing residual mantle peridotites and overlying olivine cumulates (Vuollo and Piirainen, 1990). Serpentinization was predominantly isochemical except for the addition of water. This implies that Cr- and Ni-bearing carbonates, skarns and cherts surrounding the serpentinites are not genetically related to serpentinization (cf. Huhma and Huhma, 1970; Gaál et al., 1975) and accordingly an alternative explanation must be sought.

The stratigraphic position and depositional environment of the Outokumpu assemblage is still rather poorly understood. Following Simonen et al., (1978), it has usually been regarded as belonging to the Kalevian but although enclosed by Kalevian schists as a result of thrusting and folding, the ore horizons are separated by a tectonic dislocation from the underlying sediments (Koistinen, 1981). Both the Jormua complex and the Outokumpu assemblage have been interpreted as forming in a shallow basin dominated by carbonate and chemical sedimentation prior to rapid subsidence and deposition of Kalevian sediments (Kontinen, 1987; Park, 1988 b).

The Outokumpu assemblage is in many places found in close proximity to Jatulian sequences and particularly in the northern part of the Kainuu-Outokumpu zone, a useful exploration guide has been the spatial association of serpentinites, black schists and Marine Jatulian iron formations (Mäkelä, 1981). These so-called Outokumpu-type black schists have been variously assigned to the Marine Jatulian (Ervamaa and Heino, 1983) or Lower Kalevian (Loukola-Ruskeeniemi, 1991, 1992) and may also be correlated with volcanic exhalative stratabound mineralization at Närhiniemi, near Sotkamo (Juhanen, 1989), which has Pb charateristics that closely resemble those of Talvivaara (Vaasjoki, 1986). Kopperoinen (1970) classified the serpentinites and soapstones occurring with the Talvivaara black schists as Jatulian.

At both Riihilahti and Huutjärvi mineralization occurs within banded amphibolites and diopside amphibolites representing Jatulian volcanics (Koistinen, 1984 a), while at Miihkali ore is associated with both volcanics and serpentinites. The association serpentinite - black schist - dolomite - skarn and quartz rock is also present at Pieni Vehkalahti near Kaavi, within a zone trending SE from Losomäki to Suvasvesi, while at nearby Vehkalahti serpentinites occur in close proximity to Jatulian volcanics and calc-silicate rocks (Koistinen, 1984 a). At Viitaniemi, the U-P horizon has also been identified, occurring within calc-silicate rocks between Archean basement and Kalevian mica gneisses (Äikäs, 1976). This occurrence at Viitaniemi belongs to a sequence containing Marine Jatulian sediments and mafic volcanics that can be traced via Soisalo and Puutosmäki across the Suvasvesi Shear Zone to Neulamäki near Kuopio.

The Outokumpu assemblage is not restricted to the northeast side of the Suvasvesi Shear Zone but is also present further southwest, as at Soisalo in NE Leppävirta (Koistinen, 1984 a), where quartz rocks occur, and Kangaslampi, within the Kotalahti Nickel Belt, where the complete assemblage is present (Koistinen, 1987). East of Haukivesi the Outokumpu assemblage can be traced through Polvijärvi to Savonranta and from there in a northeasterly

direction towards Rääkkylä.

It is important to distinguish the ophiolitic elements of the Outokumpu assemblage from the pelagic and chemogenic sediments. Sedimentation may have been similar to, or even coeval with the Marine Jatulian, although at present the relationship, if any between Marine Jatulian volcanism (ca. 2.06 Ga) and Outokumpu magmatism is unknown. Both may record progressive extension and ocean floor formation between about 2.06-1.97 Ga, in which case it is meaningless to distinguish between Marine Jatulian and Lower Kalevian. Nevertheless, it is still clear that the Outokumpu complex (ophiolitic sequence) formed before the main phase of Kalevian sedimentation. Although 1.97 Ga albite diabase dikes cut Jatulian quartzites at Nilsiä and have zircons resembling those of Jatulian dikes, none of the 1.97 Ga dykes in the Koli area have been observed to cut the Kalevian sediments (Piirainen, 1991).

Pyhäsalmi Island Arc with Kuroko-type deposits

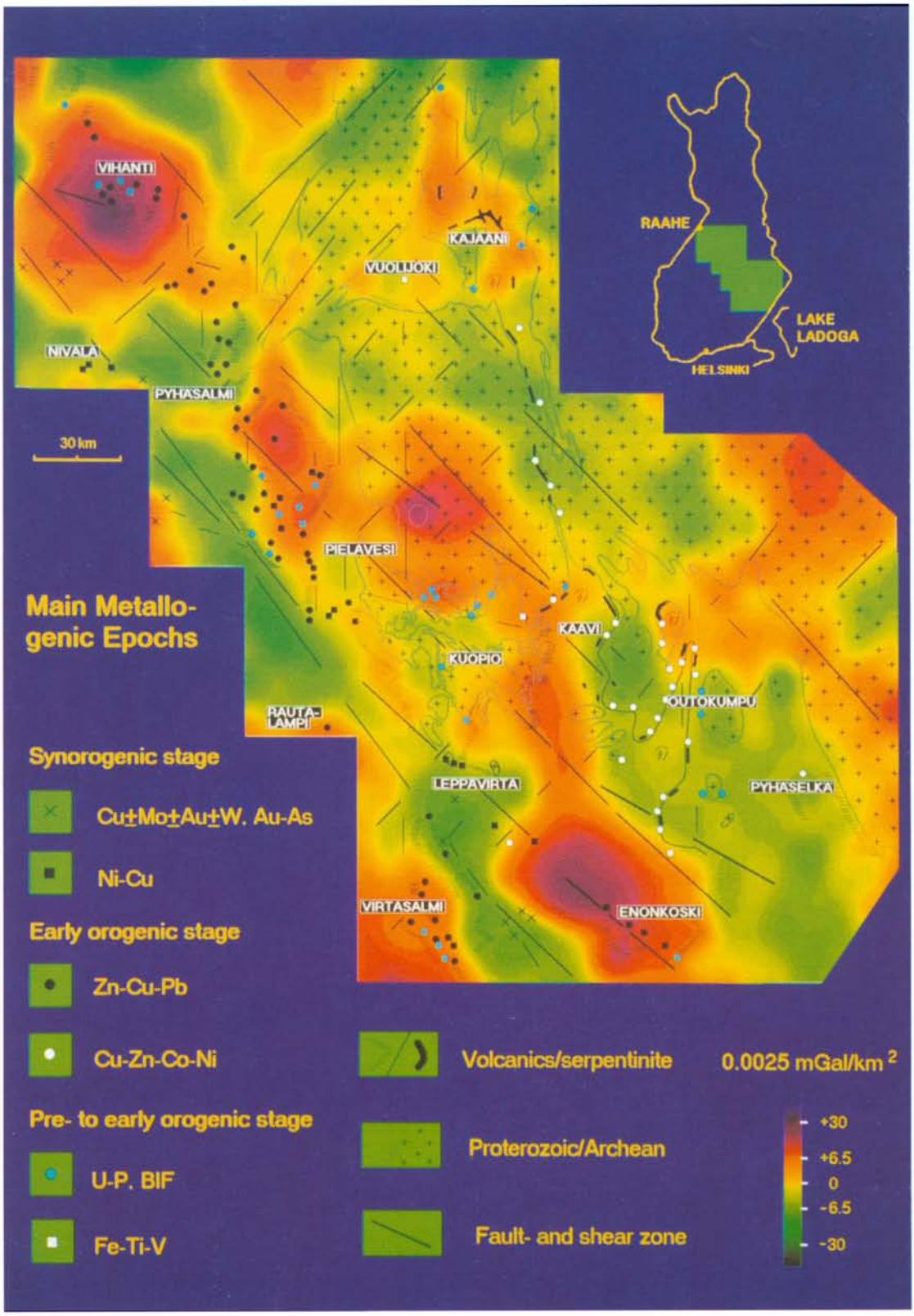
The Pyhäsalmi Island Arc (PIA) zone is about 10-30 km wide and at least 350 km in length. Hydrothermal activity has been confined to volcanic zones and complexes which define a discontinuous deformed belt between Vihanti and Virtasalmi (Fig. 47). In its northern part the zone is characterized by positive Bouguer anomalies on the northeast side of a major gravity trough, while the converse applies in the southern part of the zone.

The Vihanti positive gravity anomaly is some 50-60 km in diameter and presumably reflects the presence of an extensive, predominantly mafic or high-grade metamorphic block in the deep crust. In addition to the Lampinsaari deposit, significant mineralization is present at Nevasaari, Kuuhkamo and Vilminko. Numerous Zn-Cu prospects associated with altered volcanics, calc-silicates, carbonaceous rocks and black schists have been located at Piippola and Kärsämäki. The pyritic Vuohtojoki Zn-Cu occurrence at Kärsämäki is hosted by cordierite gneisses.

The Pielavesi-Pyhäsalmi ore zone is an oval shaped complex some 40 km wide and 100 km in length characterized by bimodal volcanism, hydrothermal alteration and stratabound Zn-Cu mineralization (Figs. 23 and 47). Mineralization tends to occur in close proximity to known major ores, such as Ruotanen, Kangasjärvi and Säviä, with numerous, with clusters of smaller occurrences occupying more distal locations. The stratabound Pukkiharju Zn-Cu mineralization at Rautalampi is very similar to the Pyhäsalmi-Pielavesi deposits on the basis of both geological setting and alteration patterns (Lahtinen, 1988).

There is also a fundamental stratigraphic control on Zn-Cu mineralization in the Virtasalmi-Juva volcanic complex. At Hällinmäki for example, there is no evidence for any intrusive rocks having been responsible for mineralization (Koistinen, 1978), as might have been expected if it were a contact metamorphic skarn deposit. Instead the sulfide mineralization is associated with carbonate rocks, black schists and iron formations probably related to hydrothermal fluid activity and island arc volcanism (Suvanto, 1983). According to Lawrie (1992) the amphibolites of Hällinmäki have a within plate signature indicating rifting of mature arc complexes and do not contain any evidence to suggest the existence of calc-alkaline arc complex. Many deposits in the zone, however, are

Fig. 47. Metallogenic and tectonic features of the Raahe-Ladoga Zone superimposed on a second vertical derivative Bouguer anomaly map.


VIHANTI
KAJAANI
VUOLIJOKI
NIVALA
PYHASALMI
30 km
PIELAVESI
KAAVI
KUOPIO
OUTOKUMPU
RAUTA-
LAMPI
LEPPAVIRTA
PYHASELKÄ
VIRTASALMI
ENONKOSKI
RAAHE
LAKE
LADOGA
HELSINKI
Main Metallo-
genic Epochs
Synorogenic stage
Cu±Mo±Au±W. Au-As
Ni-Cu
Early orogenic stage
Zn-Cu-Pb
Cu-Zn-Co-Ni
Pre- to early orogenic stage
U-P. BIF
Fe-Ti-V
Volcanics/serpentinite
Proterozoic/Archean
Fault- and shear zone
0.0025 mGal/km 2
+30
+6.5
0
-6.5
-30

stratabound in nature and hydrothermal alteration is often present. Occurrences in which Cu is dominant tend to be associated with mafic volcanics whereas felsic and often carbonaceous volcanics preferentially host Zn-dominated mineralization.

The lead isotopic compositions of the PIA-zone sulphide deposits differ clearly from the pure mantle lead of the Outokumpu deposits. While predominantly of mantle origin, there is nevertheless a detectable component that may be derived either from sediments or the adjacent Archean craton (ca. 1925-1975 Ma; Simonen et al., 1978). According to Vaasjoki (1981) the lead age of the Pyhäsalmi-Vihanti ores is about 1.9-2.0 Ga and the data suggest that island arc volcanism was generated from a homogenized crust influenced by mantle-derived magmas. What kind of crust was this and what was its age? In the Bergslagen district of central Sweden, where lead isotopic compositions also indicate a crustal source, a 2.4-2.1 Ga felsic to intermediate precursor for the Svecofennian granitoids and volcanics has been proposed (Baker et al., 1988). Leads from the Mustalampi occurrence at Virtasalmi and the Viholanniemi prospect near Joroinen are similar to one another and have characteristics that fall midway between those of the Vihanti-Pyhäsalmi province and the granitoid complex of central Finland. In general the mantle signature in leads decreases in southern Savo with respect to the Pyhäsalmi region, which is also consistent with the southeastwards younging in age throughout the PIA zone, as shown by a U-Pb date of 1906 Ma for rhyolite at Viholanniemi (Vaasjoki and Sakko, 1988). The metamorphic history of the Virtasalmi-Juva area is similar to that elsewhere in the Haukivesi-Kiuruvesi complex, as is that of the Pukkiharju mineralization at Rautalampi which, on the basis of its isotopic composition, has affinities with the Pyhäsalmi ore type.

The PIA zone is known to contain some 55 mineralizations in addition to numerous economical ore deposits, including Lampinsaari, Ruotanen, Mullikkoräme, Kalliokylä, Kangasjärvi and Hällinmäki. The total in situ metal budget of the PIA zone is estimated as 2400 kt Zn, 500 kt Cu, 100 kt Pb and 15 t Au and mean abundances, excluding the high-grade Hällinmäki Cu ore, are 3.4% Zn, 0.65% Cu, 0.15% Pb, 0.3 g/t Au and 20g/t Ag (Gaál, 1990).

Investigations into the nature and setting of ore deposits in the PIA zone reveals the following general characteristics:

- the presence of volcanic complexes and zones that exhibit structural control by conjugate N- and NE-trending fracture systems. These complexes are associated with gravimetric highs or sharp gradients and medium- to high-grade metamorphism.
- tholeiitic and calc-alkaline submarine volcanism with island arc affinities
- stratabound morphology
- dolomite/carbonate - calc-silicate rocks - felsic volcanics - chert - U-P horizon - black schist association is present in the footwall to ore bodies
- vertical or lateral hydrothermal alteration, sericitization, silicification, chloritization, and the presence of garnet-cordierite-anthophyllite rocks
- Mg, Fe, H_2O and S abundances increase upwards as the ore is approached (with a corresponding increase in MgO/CaO ratio in mafic volcanics). Conversely, Si, Ca, Na and K abundances decrease, as does the Na_2O/K_2O ratio in felsic volcanics.
- ore deposits are pyrite-dominated and display distinct zoning in addition to stratabound form:

Massive ore	FeS_2, ZnS, PbS, Ag
	FeS_2, (FeS), ZnS,
	$CuFeS_2$, Au, Ag, Ba
Stringer ore	FeS, (FeS_2), $CuFeS_2$, Au

- intense tectonic control on geometry of ore deposits

While the PIA zone ore deposits broadly qualify as Kuroko-type deposits, there are nevertheless some distinctive peculiarities. The

overall extensional context of bimodal volcanism appears to have been within an island arc, which is similar to that of the caldera complexes in Honshu where the major faults are parallel to the arc trend (Scott, 1980). The thicknesses of volcanic units within the PIA zone nevertheless appear somewhat trivial compared with the vast thicknesses present in Phanerozoic volcanic sequences. Subduction-related magmatism in the PIA zone would have enabled large scale convective hydrothermal systems to develop and the reducing conditions in shallow marine basins would have provided a favourable setting for sulphide precipitation. No iron formations have been found to occur stratigraphically above the ore deposits of the PIA zone. Island arc volcanism is underlain by the shelf-type association, which is not very typical of Kuroko-type environments. Volcanic-hosted mineralization is situated within the felsic upper part of bimodal volcanic sequences, more reminiscent of the Archean Abitibi Belt in Canada, which has also been interpreted as representing Kuroko-type metallogenesis (Scott, 1988). Deposits in Canada are also clustered into groups which may be partly of tectonic origin and partly inherited from primary caldera complexes.

So far no major felsic volcanic edifices, rhyolitic lava domes or caldera structures have been recognized within the PIA zone. Instead volcanism may represent fissure eruptions. The garnet-cordierite-anthophyllite units may also form laterally persistent horizons. The Na-rich volcanic sequence at Riitavuori, to the north of the Ruotanen ore body somewhat resembles an intrusive rock in appearance and may represent a rhyolitic lava dome such as occur in association with Kuroko deposits. The pyritic Mullikkoräme Zn ore body is situated close by, within altered felsic volcanics that are underlain by mafic pillowed lavas and overlain by a polymodal volcanic sequence (Luukas, 1992). Rhyolite domes associated with Kuroko deposits are believed to form either during or after mineralization (Ohmoto, 1983) and if this relationship is extended to Pyhäsalmi, then the age of the Riitavuori quartz porphyry (1921 Ma) corresponds to the minimum age of the Pyhäsalmi-type mineralization. The paleosome of the Saarela granodioritic gneiss at Pielavesi has been dated at around 2.0 Ga and may represent metamorphosed sediments derived from the Salonsaari Suite, which preceded the mineralized Säviä Suite, thus providing a lower age constraint on the latter. It is also significant that the Laajamäki quartz diorite (1922 Ma), which is approximately the same age as the Kettuperä gneiss and the Kirkkosaari orthogneisses, clearly truncates metasediments (Vaasjoki and Sakko, 1988).

Synorogenic stage

Kotalahti Nickel Belt

As Gaál (1985) noted, the Kotalahti Nickel Belt (KNB) lies within the Raahe-Ladoga Zone and is more specifically associated with the Haukivesi Shear Zone, and the region between the Haukivesi and Suvasvesi shear zones. The Ni-bearing intrusions are located within the regional gravity highs which also coincide with zones of higher metamorphic grade (Fig. 47). The country rocks surrounding the intrusions are typically migmatitic turbidites and schollen migmatites.

The emplacement of these differentiated or multiphase intrusions is characteristically controlled by NW-dextral shear zones and faults with associated drag folds and conjugate or younger sinistral structures (Gaál, 1972, 1980). Major clusters of intrusions occur at Ylivieska, Nivala, Haapajärvi, Pielavesi-Tervo, Leppävirta, Varkaus, Juva, Enonkoski and Parikkala. Individual intrusions in the KNB to be oriented as N- to NE-trending lenses or define en echélon arrays that represent progressive dilation to accommodate intruding magma. For instance the Hitura ultramafic body is but one member of a NE-trending suite of intrusions that, on the basis of geophysical anomaly pat-

terns, appear nevertheless to represent the disruption of what was originally a single magma layer. Most intrusions tend to occur as sills or swarms of sills which are often layered, such as at Perkkiö, Kumiseva, Talluskanava, Laukunkangas and Parikkala. Where mineralization is present, it has apparently formed in the ultramafic basal parts of single- or two-stage intrusions, as documented from intrusions at Pielavesi, Juva (Makkonen, 1992) and Enonkoski (Grundström, 1985). The ore mineralogy of the deposits is dominated by pyrrhotite, pentlandite and chalcopyrite with PGE abundances being rather low.

A close analogue for the KNB is the Thompson Nickel Belt (TNB) in the central part of the Canadian Shield at the boundary between the juvenile early Proterozoic (1.9-1.7 Ga) Churchill and Archean (>2.5 Ga) Provinces. The Ni deposits of the TNB are associated with ultramafic sills within the early Proterozoic platform cover sequence which is underlain by polymetamorphic orthogneisses and paragneisses believed to represent reworked Archean basement (Bleeker, 1990). The intrusions within the KNB are also in close proximity to Archean basement, which may have caused contamination of magma during ascent at higher pressures than in the Vammala zone in southwest Finland, where the crust was somewhat thinner (Mäkinen, 1987). An external source for sulfur is surmised for the Thompson belt deposits while either lowering of temperature or contamination by and assimilation of country rocks were the most significant factors in causing the separation of sulphide from melt in the KNB (Makkonen, 1992). The generation of the tholeiitic parent magma was attributed by Gaál (1985) to water-induced melting of the mantle wedge overlying a subduction zone, at depths of 70-100 km. On the basis of the forsterite composition of olivine from the ultramafic part of the Talluskanava intrusion, it has been calculated that the parent basaltic magma would have had an MgO content of about 10%. The Sm/Nd ratios of synorogenic mafic intrusions indicate LREE enrichment due to lithospheric contamination, although fractionated compositions could also represent pure derivates from mantle sources (Huhma, 1986).

The Ni-critical intrusions in the Kotalahti belt, including Hitura, Kotalahti, Enonkoski and Parikkala, represent a relatively well constrained phase of magmatism between 1.89-1.88 Ga. The Perkkiö gabbro massif at Ylivieska and the intrusions at Tervo also belong to this Ni-critical age group. The Molson dikes which are considered to be coeval with Ni deposits in the TNB are remarkably similar in age (1883 Ma; Bleeker, 1990) to the Kotalahti belt.

The Kotalahti Nickel Belt is distinctly younger than the PIA zone and also contains even younger 1.88-1.87 Ga gabbros, such as Saarisenjärvi, Tyypekki and Kumiseva, which do not appear to be prospective for Ni. Relative timing is demonstrated by their truncating schollen migmatites and quartz diorites that contain D_2 structures and their association with hypersthene granites. The earliest (ca. 1.9 Ga) mafic intrusions, including Alpua and Lapinlahti on the NE side of the Kotalahti belt also appear to be devoid of mineralization.

Total estimated reserves in the Kotalahti Nickel Belt are 400 kt Ni and 150 kt Cu, with mean concentrations of 0.78% Ni and 0.28% Cu (Papunen and Vorma, 1985). These figures are appreciably less than those for the analogous Thompson Collisional Nickel Belt and individual intrusions in the Kotalahti belt also tend to be smaller.

Porphyry type Cu±Mo±Au±W and epigenetic Au-As occurrences (syn-late orogenic)

Gold mineralization within the RLZ is related to the development of a calc-alkaline volcano-plutonic arc in an active continental margin environment that is reminiscent of Phanerozoic porphyry-type Cu-Au deposit settings. Gold occurrences may be further classified as either porphyry-type and epigenetic mesothermal de-

posits, the former being represented by Kopsa at Haapajärvi, Susineva near Kalajoki and Lahnanen at Viitasaari. The Kopsa Cu-Au occurrence is associated with a porphyritic tonalite stock (Gaál and Isohanni, 1979) while Mo-Cu-Au mineralization at Susineva occurs at the margin of a porphyritic granodiorite and the Lahnanen Mo prospect is hosted by pyroxene granodiorite. All three mineralizations are of stockwork type in which fracture zones with or without quartz control the distribution of sulfides (Nurmi, 1984). Chalcopyrite, arsenopyrite and pyrrhotite are the major ore minerals, while loellingite, cubanite, pyrite, sphalerite, native gold, scheelite and molybdenite occur in smaller amounts. Hydrothermal alteration is indicated by silicification, potassic alteration, sericitization and propylitization. Porphyry-type mineralization is hosted by late phases of synkinematic (ca. 1.88 Ga) granitoids that have been dated at Susineva (1883±10 Ma; Huhma, 1986) and Lahnanen (1879±4 Ma; Nironen and Front, 1992). Five distinct ore-element associations have been recognized amongst Svecofennian porphyry deposits, namely Mo-, Cu-, Mo-Cu-Au- and Cu-W- types (Nurmi and Haapala, 1986). Within the RLZ both Mo-bearing and Au-bearing porphyry types have been found, so that the zone resembles an intermediate type between continental and oceanic island arc subclasses (Kesler, 1973).

The Raahe-Ladoga geosuture provided opportunities for gold mineralization during syn- and late orogenic deformation by structurally controlled fluid flow and changing physicochemical gradients. In this respect both felsic magmatism and metamorphism were important and deposits tend to coincide with relatively low metamorphic grades and gravity troughs (Fig. 47). Brittle-ductile or ductile transcurrent shear zones provided the appropriate tectonic setting for late orogenic gold mineralization. Epigenetic gold mineralization may be genetically associated with the porphyry-type mineralization, just as proposed by Gaál and Sundblad (1990) in their classification of porphyry-type and shear-hosted deposits. Mesothermal vein and shear types are clearly younger than the porphyry types but have nevertheless formed prior to 1.85 Ga.

In the area between Ylivieska and Oulainen (Sipilä, 1988) and at Laivakangas near Raahe (Mäkelä and Sundberg, 1985), there exist a number of structurally controlled Au-As occurrences in tonalitic intrusions and surrounding intermediate volcanics. Mineralization is associated with vertical NE- and NW-trending shear zones containing quartz vein systems, with a total width varying between several centimeters and 1.5 m. Gold is present as inclusions in arsenopyrite, loellingite and quartz and small amounts of pyrrhotite, chalcopyrite, scheelite and molybdenite are also found. Metasomatic alteration is confined to weak epidotization, silicification and potassium enrichment. According to Mäkelä and Sandberg (1985), the Laivakangas occurrence is genetically related to late stage magmatic activity associated with intrusion of the quartz diorite magma, combined with the influence of metamorphic fluids and deformation.

The combined Au content of the Ostrobothnian gold deposits (Laivakangas, Kopsa, Vesiperä, Ängesneva, Kiimala) is estimated at about 520 kg. Of this, the Laivankangas deposit contains 0.4 Mt at 4 ppm Au, Kopsa has 1.1 Mt at 2 ppm Au and 0.81% As, with the remainder containing about 0.5 Mt at 2-3 ppm and 0.7-1.0% As (Kontoniemi, 1991).

The epigenetic Osikonmäki Au occurrence near Rantasalmi is hosted by a tonalite intrusion dated at 1887±5 Ma (Vaasjoki and Kontoniemi, 1991) and thus represents the same magmatic event as the porphyry-type intrusions. The main ore minerals present are pyrrhotite, arsenopyrite, loellingite, chalcopyrite, while accessory phases include Bi-Te-B minerals, sphalerite, galena and molybdenite (Kontoniemi, et al., 1991). Gold is present both in native form and as electrum inclusions and although the crystallization of ore minerals

commenced at around 500 degrees C, gold an associated Sb-B minerals formed from fluorine-rich hydrothermal fluids at lower temperatures. Alteration is restricted to narrow sheared and mineralized zones and is characterized by biotitization, silicification and general potassium enrichment. The W-trending Osikonmäki Shear Zone is situated between major NW-trending wrench faults and mineralization took place during D_3 deformation after the formation of F_3 synforms that plunge about 20 degrees SE (Kontoniemi and Ekdahl, 1990). Some quartz veins containing gold and arsenopyrite also formed within the NW-trending F_3 axial plane. According to Korsman et al., (1988) the effects of D_3 deformation are also apparent in the area of the Sulkava thermal dome, which is 1.83-1.80 Ga in age. However, mineralization at Osikonmäki is believed to be older than the emplacement of the nearby Hiltula granodiorite which is dated as 1850±7 Ma (Vaasjoki and Kontoniemi, 1991). In the NE part of the RLZ mesothermal Au-As veins presumably formed somewhat earlier, since tectonic activity appears to have more or less ceased by 1.85-1.86 Ga (Vaasjoki and Sakko, 1988).

The Pirilä quartz-vein hosted Au-mineralization also near Rantasalmi occurs within a volcanic-sedimentary environment overlain by graywackes with some carbonaceous and calc-silicate rocks (Makkonen and Ekdahl, 1988). Sulphide- and gold-bearing quartz veins and lenses occur in a transitional zone between volcanic and sedimentary units, the latter including a narrow silicate, oxide and sulphide facies iron formation. Quartz veins developed during both D_2 and D_3 deformation. Gold was probably ultimately derived from the volcanics of the Pirilä belt and the Tuusmäki-Osikonmäki tonalites.

The gold deposits of southern Savo have been calculated to contain about 7700 kg Au, with Osikonmäki having about 2 Mt of ore grading 3 ppm Au and 0.77% As. The Pirilä occurrence contains about 150 kt of ore at 8 ppm Au (Kontoniemi, 1991).

So far none of the Au occurrences in the Raahe-Ladoga Zone have proven economically viable and it is conceivable that the whole magmatic zone has been eroded to such a depth that any epithermal or porphyry-type deposits that might have existed have been removed by erosion. Porphyry-type mineralization in the RLZ took place towards the end of magmatic evolution in the zone and is reminiscent of Andinotype occurrences. The most interesting mineralizations are those that represent low grade late metamorphic shear-hosted Au-As deposits.

Discussion: Svecofennian Orogeny and metallogenesis

Metallogeny is a potential indicator of tectonic setting. The formation of the Raahe-Ladoga Zone (RLZ) at the margin of the Archean craton is the product of complex processes during the Svecofennian Orogeny. Within the Fennoscandian Shield there is a pattern of younging towards the west and southwest, from the Svecofennian to the Sveconorwegian to the Caledonian. Within the Svecofennian Orogen itself there have also been attempts to identify successive N- to NE-directed subduction zones (Park et al., 1984; Gaál, 1986; Brewer and Pharaoh, 1990), although it is difficult to identify any significant younging trend on the basis of presently available isotopic data. There are nevertheless numerous features, including metallogenesis, which favour interpretation of the Svecofennian Orogen as the product of a number of amalgamated terranes.

According to weight ratios of Cu, Pb and Zn, the deposits of the Pyhäsalmi Island Arc (PIA) zone differ from those of the Skellefte Field

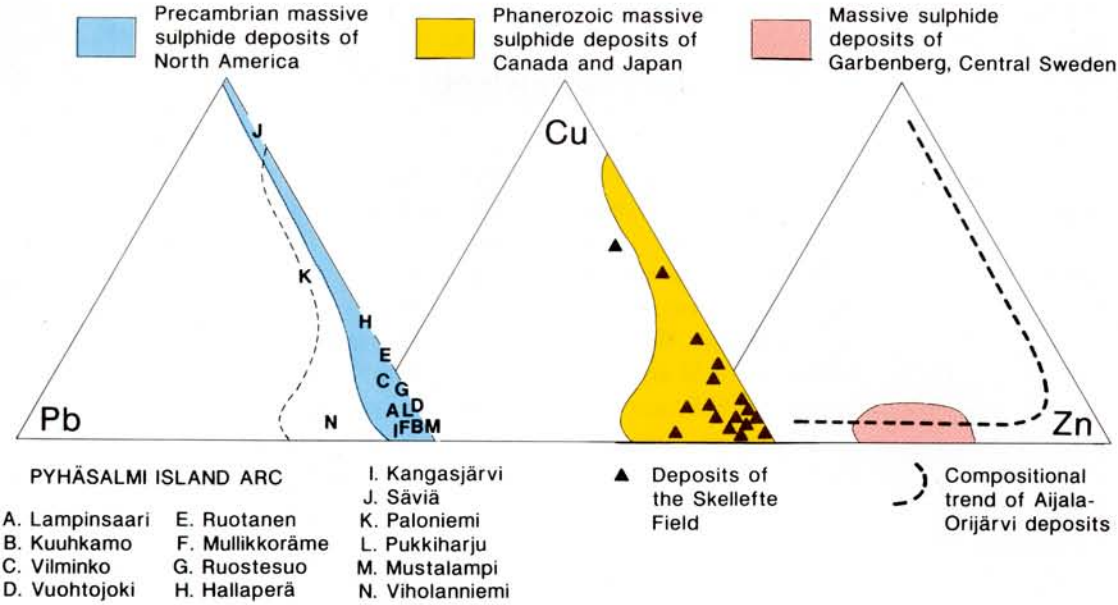


Fig. 48. Cu-Zn-Pb weight ratios of ore deposits in the Pyhäsalmi Island Arc, Skellefte Field and Orijärvi-Bergslagen area. References given in text.

and Bergslagen Province but show a closer resemblance to Precambrian massive sulfides of North America, particularly those of Archean age which are considerably lower in lead (and also in Ba, Au and Ag) than their Phanerozoic counterparts (Fig. 48).

The massive sulphide deposits of the Skellefte Field (Rickard, 1986; Claesson, 1986; Jonsson and Larsson, 1986; Svensson and Willden, 1986) are more clearly of Kuroko type and typically have higher Pb contents than in the PIA zone. However both the PIA zone and the Skellefte deposits lie toward the higher Zn end of the spectrum, in contrast to the Garpenberg deposits in the Bergslagen district of central Sweden (Vivallo, 1984) which are clearly Zn-Pb types (Fig. 48). The sulphide deposits of the Aijala-Orijärvi district in southwestern Finland (Mäkelä, 1989) resemble both Kuroko and Pb-rich Garperberg types. Gold and Ag abundances also increase to the south and west, consistent with a more mature arc setting. The high Au abundances in the Skellefte Field are possibly the result of a later epigenetic overprint, rather than related to early sulphide mineralization (Gaál, 1990; Weihed et al., 1992).

The development of lead from the Outokumpu-type pure mantle composition to the more mature orogenic lead of the sulphide deposits which young to the southwest is an important fact in characterization of metallogenetic provinces. Lead isotopic studies do not favour a common origin for the Skellefte Field and PIA zone deposits, since the Pb isotopic compositions of the Skellefte ores differ significantly from those in the Vihanti-Pyhäsalmi district (Vaasjoki and Vivallo, 1988). The Skellefte volcanism has been considered to be somewhat older than or coeval with the 1.89 Ga Jörn granitoids (Rickard, 1986; Wilson et al., 1987) so that it represents a later phase of magmatism than that in the PIA zone.

Skellefte volcanic activity has been attributed to the formation of a magmatic arc over a N-dipping subduction zone, with ores being of Kuroko type (Rickard, 1986; Weihed et al.,

1992). Eruptions took place from numerous vents possibly as caldera structures. Stratiform and stratabound ores occur predominantly towards the top of the volcanic sequence before the transition to the overlying volcanics. To the south, progressively deeper water environments prevailed, and sedimentation commenced before the onset of volcanism.

Deep seismic reflection profiling along the Gulf of Bothnia has revealed a lower crustal reflector dipping gently northwards that has been interpreted as the relict trace of an early Proterozoic subduction zone (BABEL working group, 1990). This lends support to extrapolations of the Skellefte Field across into Finland via southern Bothnia to the Tampere Schist Belt, as proposed by Korja (1990) on the basis of magnetotelluric investigations. Korja attributed high lower crustal conductivities to remnants of oceanic material or graphitic metasediments within the middle crust. The volcanic rocks of the Tampere district have distinct arc affinities and moreover, appear to represent a relatively mature, rather than primitive oceanic arc (Kähkönen, 1989, 1990). A conductive lower crustal anomaly can be traced from Tampere through southern Savo towards Lake Ladoga (Korja, 1990). Although no Skellefte-type massive sulphide deposits are known from the Tampere Schist Belt, there are a few volcanic-hosted Zn-Cu-Pb-Ag occurrences such as Kankaanpää and Hämeenkyrö, with the most southeasterly mineralization in Finland being the Salo-Issakka prospect at Imatra (Kahma, 1973).

Other metallogenic features also support the correlation between the Tampere Schist Belt and the Skellefte zone. The apparent ring of Ni-deposits around the central Finland granitic batholith complex can be reinterpreted so that its southwestern margin, the Swedish western Bothnian occurrences, the Oravainen Ni-province, Vammala Ni-belt and possibly the deposits in southern Savo define a deformed synorogenic Ni-belt subparallel to the Tampere-Skellefte zone, just as the Kotalahti Nickel Belt is associated with the Pyhäsalmi Island Arc (Fig. 49). The geochemistry of the Swedish west Bothnian deposits is also quite similar to that of the Vammala-type intrusions (Mäkinen, 1987). The Ni-belt postulated to follow the Tampere-Skellefte zone intersects the Kotalahti Nickel Belt in southern Savo, where the Ni occurrences share features of both Kotalahti and Vammala types (Makkonen, 1992). The northern part of the Savo Schist Belt (Kiuruvesi-Haukivesi complex) differs in its tectonic and metamorphic history from the southern part, in the Rantasalmi-Sulkava area, and this has been interpreted to indicate an intervening geosuture (Korsman et al., 1988); this can also be regarded as representing the boundary between the Pyhäsalmi Island Arc and the eastern part of the Skellefte-Tampere Arc.

The postulated subduction zone to the south of the Skellefte-Tampere zone also contains porphyry-type (e.g. Granberg, Tallberg, Tienpää, Ylöjärvi) and epigenetic gold deposits (e.g. Björkdal, Åkerberg, Peräseinäjoki, Pirkkala, Kutemajärvi) which appear to lie along a continuous linear trend from southwestern Finland into northern Sweden (Nurmi et al., 1991). The porphyry-associate deposits of the Skellefte Field appear to be more enriched in gold than their counterparts in the PIA zone and are probably closely related to massive sulphide deposits (Weihed et al., 1992). Epigenetic gold deposits in Sweden developed late in the overall tectonic history (1.87-1.82 Ga) and the final phase of metallogenesis were associated with Li-Be-Sn-Nb-Ta pegmatites and minor post-orogenic Mo-W-bearing pegmatites (Weihed et al., 1992; Teertstra et al., 1992). The Viitaniemi pegmatite in the Tampere Schist Belt (1794±15 Ma; Teertstra et al., 1992) is coeval with the Puruvesi granite in southeastern Finland and a swarm of complex pegmatites in the Kitee district is dated at 1.77 Ga.

The Orijärvi zone in southwest Finland and the central Swedish ores have been interpreted as Kuroko-type (Hietanen, 1975; Lundqvist,

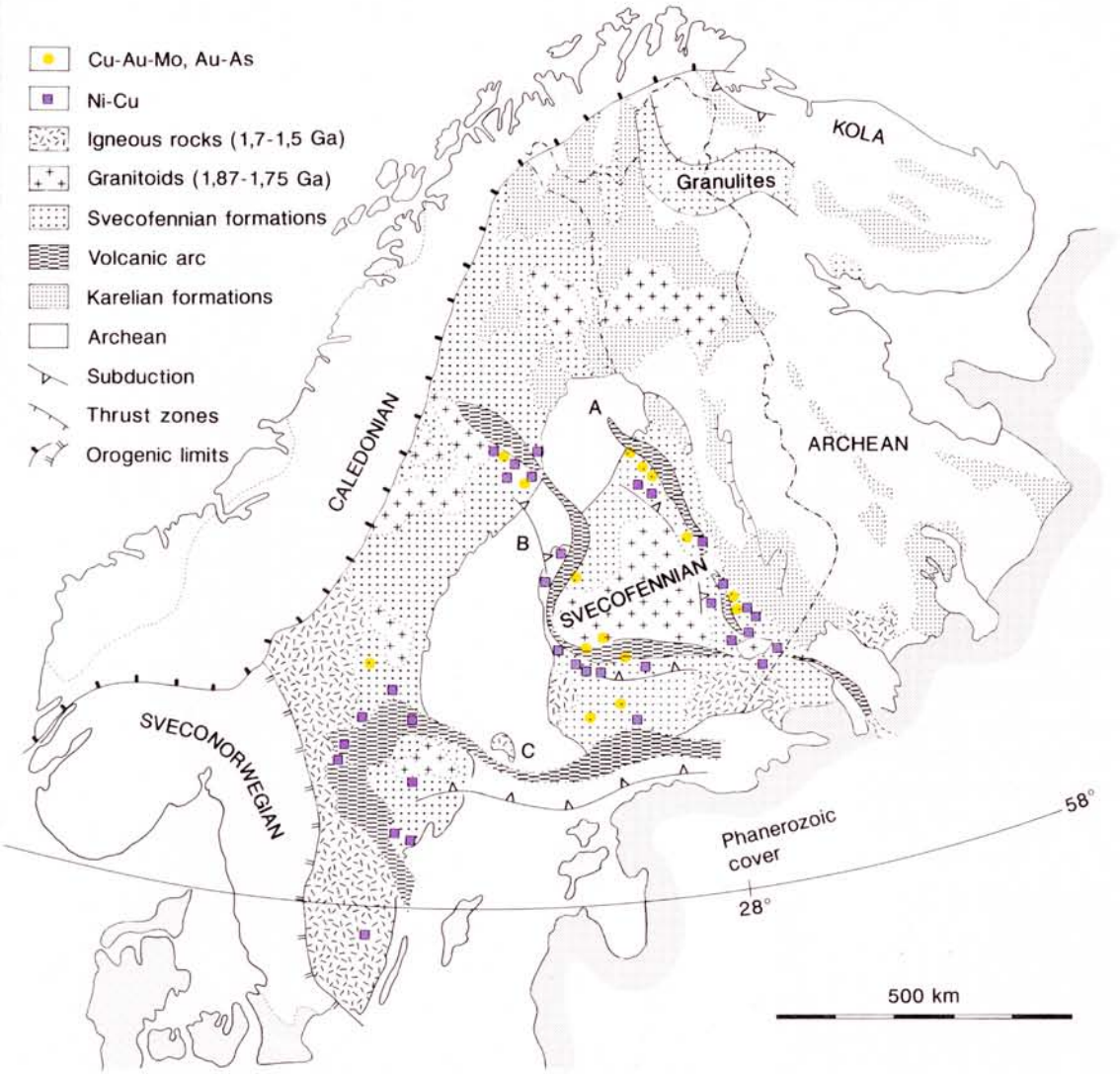


Fig. 49. (A) Pyhäsalmi-, (B) Skellefte-Tampere- and (C) Orijärvi-Bergslagen Arcs and related metallogenesis representing three separate stages within the Svecofennian Orogeny.

1975; Löfgren, 1979; Lundberg, 1980; Loberg, 1980) or as having formed in a continental rift environment (Oen et al., 1982; Vivallo and Rickard, 1984; Baker, 1985; de Groot et al., 1988). Volcanism at Bergslagen has been related to 1.89-1.84 Ga anatexis and reworking of crust formed by subduction processes between 2.4-2.0 Ga (Baker et al., 1988). Gaál and Gorbatschev (1987) and Gaál (1990) interpreted the volcanic-hosted massive sulphides of the Skellefte, Pyhäsalmi, Orijärvi and Bergslagen districts as defining a single U-shaped subduction zone with E-directed subduction. This is difficult to reconcile with all of the features observed in the RLZ. Furthermore, subduction in southern Finland has also been interpreted as

towards the north (Edelman and Jaanus-Järkkälä, 1983; Park et al., 1984; Ward, 1987). Ore elements in deposits of the Orijärvi-Bergslagen belt also suggest more mature characteristics than those of the PIA zone and the Skellefte Field.

The Orijärvi-Bergslagen Arc may represent the third phase of the Svecofennian Orogen, in which the early orogenic stage containing stratabound sulphide deposits is again followed by synorogenic Ni-deposits, particularly in the west (Nilsson, 1985), with only a few small Ni prospects having been identified in the Finnish part of the Arc. Late orogenic granites from Lake Ladoga to Bergslagen host minor Au occurrences (Gaál and Sundblad, 1990), although most of the gold in the Arc is associated with stratabound deposits.

The northeastern part of the Fennoscandian Shield is analogous to the Svecofennian/Karelian Domains in that the Lapland Granulite Belt has been regarded as representing deep water equivalents to the epicontinental Jatulian deposits along the NE margin of the Karelian craton (Barbey et al., 1984). Marker (1985) proposed that subduction towards the SW resulted in the closure of the Kola ocean and the formation of a collisional suture. In this model the protoliths to the granulites and the tholeiitic volcanics would have formed in a back arc basin which was subsequently thrust back southwards onto the foreland. The collision could have taken place as early as 1.98 Ga, with initial rifting having commenced as long ago as 2.4 Ga during early Jatulian deposition. Isotopic studies constrain the age of the granulite protoliths and their metamorphism to about 2.0-1.9 Ga (Bernard-Griffiths et al., 1984), consistent with 2.15-1.90 U-Pb zircon ages from the granulites (Kesola, 1991). This corresponds broadly with the age proposed here for Svecofennian sedimentation. A further point of interest is that the Ni-bearing Pechenga gabbros and ferropicrites have an age of 1970±5 ma (Hanski et al., 1990) which is effectively indistinguishable from that of the Outokumpu and Jormua rocks. Likewise, both of these units have similar Pb compositions. Is it possible that both the SW-thrusting of the granulites and the NE-directed thrusting of the Kainuu-Outokumpu basin both record the closing stages of a major Wilson cycle affecting the whole shield?

SUMMARY AND CONCLUSIONS

Studies of the Savo schists in the Pielavesi district indicate that they can be regarded as Svecofennian. The sequence commenced with the eruption of low-K arc tholeiites in a submarine setting, on a basement consisting of ortho- and paragneisses that are considered to be somewhat older than the 1.93-1.92 Ga U-Pb zircon ages obtained from them. The earliest stages of sedimentation were characterized by thick graywacke sequences, now represented by migmatitic gneisses with sporadic graphitic schists and volcanic intercalations and was gradually replaced by a shallow marine association comprising dolomite/limestone, felsic volcanics, chert, U-P sediments, banded iron formations and black schists. This was in turn followed by extensive bimodal island arc volcanism and Kuroko-type ore formation in its more felsic upper parts. A locally developed conglomerate horizon represents some kind of depositional break prior to this volcanism. A later phase of flysch-type sedimentation is recorded by widespread graded graywackes with locally developed volcanics, carbonates and graphitic schists near the base of the sequence. A further massive volcanic unit, the Kotajärvi

Metalava is younger than at least some of these metagraywackes and is believed to correspond in age to volcanism in the Pihtipudas area, immediately to the southwest (1883±20 Ma; Aho, 1979). Still younger Svecofennian units include the polymodal volcanics and clastic deposits underlying the Settijärvi Conglomerate in central Ostrobothnia.

A similar history can be reconstructed along the whole marginal zone bordering the Archean craton between Vihanti and Virtasalmi, including the Haukivesi basin. The uppermost turbidites represent flysch-type deposits correlated with the metaturbidites of the Outokumpu area and elsewhere in the Savo province, and the Upper Kalevian of Kainuu. The Karelian North Ostrobothnian Schist Belt is also correlative with the Savo schists and contains reliable evidence of deep water deposition during Jatuli time.

The overall early Proterozoic evolution can be interpreted in terms of a Wilson cycle that commenced with the Pt- and Cr-rich 2.44 Ga layered intrusions and Sumian volcanics. Further extensional phases were marked by the intracontinental 2.2 Ga karjalites and 2.1 Ga Fe-tholeiites but this Jatulian magmatism was in general not favourable for mineralization. During the subsequent Marine Jatulian interval Superior-type iron formations were deposited over wide areas, along with a distinctive U-P horizon. The Marine Jatulian corresponded to the break-up stage and transition to compressional deformation. The shelf association of the marginal zone would have been coeval with marine Jatulian deposition in the Karelian domain (2.08-1.97 Ga). From the point of view of mineralization, this Marine Jatulian phase and the immediately succeeding volcanism prior to flysch-type sedimentation are the most significant in the Raahe-Ladoga Zone. The Otanmäki Fe-Ti-V-bearing layered intrusions are approximately the same age (2.06 Ga) as the Jatulian volcanics in the Kuopio district. According to the plate tectonic model presented here the Kainuu-Outokumpu back arc basin formed as a result of extension during the initial phase of prolonged N-NE-directed subduction. The absence of isotopic dates between 2.06-1.97 Ga leaves as problematical the possibility of a relationship between the Kainuu-Outokumpu Back Arc and the Jatulian volcanism. Opening of the marginal basin and Cyprus-type ore formation preceded the bimodal volcanism and associated Kuroko-type magmatism of the Pyhäsalmi Island Arc.

Protracted subduction and consequent thrusting and compression from the S-SW resulted in the development of NW-dextral shear zones and faults, with antithetic conjugate NE trends and synthetic conjugate N trends in both the marginal zone and further within the cratonic foreland. Continued transpression caused NW-directed rotation and overthrusting and the development of a prominent SE-trending elongation lineation. The Kainuu-Outokumpu Back Arc was thrust back onto the foreland into the Kalevian sediments. The same thrusting trend is also evident in the Vihanti area where the effects of the NW-dextral shearing are less marked. Likewise the North Ostrobothnian Schist Belt has been overthrust towards the NE.

Synorogenic metallogenesis in the Raahe-Ladoga Zone is represented by the Ni-prospective ultramafic and mafic intrusions, while porphyry-type Cu±Mo±Au±W occurrences belong to the late stage. The synorogenic stage (1.90-1.87 Ga) post-dated island arc volcanism. Epigenetic Au-As occurrences represent late orogenic mineralization and formed before 1.85 Ga.

The stratabound sulphide ore deposits of the Raahe-Ladoga Zone have been modified by F_1 and also F_2 deformation, with the final significant deformation being D_3. The nickel deposits formed prior to D_2 and are often overturned in D_3 structures. Gold-arsenic occurrences are often controlled by conjugate fracture systems.

Karelian and Svecofennian formations overlap in time. If pre-Svecofennian crust formed prior to 1.93 Ga (Kähkönen, 1987, 1990; Baker et al., 1988; Huhma, 1990), then this implies

that Jatulian sedimentation took place in an oceanic setting. By Marine Jatulian time Svecofennian crust in some areas may have been rather thick and possibly represents multiple subduction events producing island arcs between 2.4-2.0 Ga, as proposed by Baker et al., (1988).

The Svecofennian Orogen is considered here to comprise three successive orogenic events, each of which contains the same broad subduction-related style of metallogenesis. The early orogenic stage is characterized by volcanic-hosted stratabound Zn-Cu-Pb deposits within the Pyhäsalmi Island Arc (1.92 Ga), the Skellefte-Tampere Arc (1.90-1.89 Ga) and the Bergslagen-Orijärvi Arc (1.90-1.86 Ga). The succeeding synorogenic stages are characterized by mafic to ultramafic magmatism with Ni-deposits and the late phases by calc-alkaline I-type granitoids with porphyry-type Cu±Mo±Au±W mineralization and epigenetic Au-As deposits. The Skellefte-Tampere Arc also contains post-orogenic (1.80-1.77 Ga) Li-Be-Sn-Nb-Ta pegmatites.

The Raahe-Ladoga Zone is thus a major geological and tectonic unit at the margin of the Karelian craton which records the whole complexity of Svecofennian orogenic processes, including early extensional events as well as prolonged N-NE-directed subduction. The Archean craton was also actively involved in the Svecofennian Orogeny.

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Appendix 1. Geochemical compositions of dome gneisses, volcanogenic rocks of the Pielavesi sequence and some infracrustal rocks. Sampling sites are presented at the end of App. 1.

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	70.48	74.45	65.85	71.67	67.64	69.70	42.64	45.37	46.54	47.19	49.22	52.22
TiO_2	0.13	0.28	0.48	0.37	0.26	0.34	1.44	1.51	0.74	0.59	1.29	0.83
Al_2O_3	17.57	13.65	14.57	13.14	17.39	14.54	16.90	20.25	16.21	18.01	17.85	15.93
Fe_2O_3tot	1.08	2.59	6.15	4.63	2.50	4.23	16.23	13.59	12.23	12.28	13.02	11.38
MnO	0.04	0.06	0.10	0.09	0.07	0.06	0.26	0.34	0.21	0.21	0.22	0.18
MgO	0.45	0.45	1.48	0.85	0.76	1.03	8.22	3.96	7.91	5.76	4.64	5.58
CaO	4.12	2.31	4.96	3.78	4.97	4.13	10.13	11.38	11.31	11.60	8.93	9.02
Na_2O	4.91	3.98	3.70	3.97	5.39	3.79	2.32	2.46	2.27	2.55	3.74	3.06
K_2O	0.71	1.81	1.93	0.63	0.60	1.02	1.12	0.49	0.95	0.93	0.61	0.71
P_2O_5	0.05	0.06	0.15	0.07	0.08	0.09	0.32	0.35	0.17	0.21	0.22	0.18
S	0.00	0.00	0.01	0.01	0.01	0.23	0.09	0.01	0.02	0.03	0.02	0.12
Total	99.54	99.64	99.38	99.22	99.67	99.17	99.67	99.71	98.56	99.36	99.76	99.21
Cr	27	29	30	28	37	31	133	50	167	74	28	142
Rb	16	23	17	8	0	25	18	4	11	14	1	4
Sr	380	280	432	138	727	247	364	736	504	681	227	224
Y	0	8	31	35	4	23	29	27	16	14	11	18
Zr	580	156	161	159	73	120	79	104	42	37	30	49
Nb	0	7	5	3	3	8	4	5	0	2	0	0
Ba	582	1039	690	229	553	446	283	170	234	233	141	130
Ce	0	0	50	33	17	51	45	42	23	25	18	13
Hf	11	13	0	0	0	0	14	13	17	6	15	11
Ta	0	0	0	0	0	0	0	1	0	0	5	1
Th	1	1	0	0	0	8	0	5	1	0	1	1

	13	14	15	16	17	18	19	20	21	22	23	24
SiO_2	69.51	47.50	48.94	49.24	50.83	52.87	58.56	63.13	68.03	75.80	47.00	47.74
TiO_2	0.57	1.06	2.09	0.93	0.86	0.87	1.04	0.41	0.34	0.14	2.58	1.01
Al_2O_3	14.00	17.47	14.31	15.55	21.13	19.68	15.82	15.02	13.95	13.17	14.28	15.65
Fe_2O_3tot	5.34	5.83	12.77	10.06	8.00	7.48	11.27	4.45	2.75	1.80	18.50	12.53
MnO	0.05	0.11	0.24	0.36	0.13	0.08	0.26	0.15	0.12	0.05	0.18	0.21
MgO	2.14	7.06	5.53	6.19	3.01	3.24	2.55	3.88	1.96	1.21	6.51	7.54
CaO	3.11	15.99	9.88	11.51	7.52	8.04	5.72	7.03	6.57	2.98	7.79	10.61
Na_2O	3.22	2.57	2.67	3.40	3.03	4.93	3.63	3.54	3.18	3.50	2.28	2.31
K_2O	1.39	0.61	0.87	0.53	4.55	1.66	0.35	1.28	2.61	0.96	0.38	1.30
P_2O_5	0.08	0.18	0.31	0.07	0.27	0.35	0.47	0.15	0.13	0.03	0.15	0.22
S	0.75	1.22	2.56	1.72	0.00	1.03	0.20	1.40	0.00	0.12	0.15	0.01
Total	100.16	99.60	100.17	99.56	99.33	100.23	99.87	100.44	99.64	99.76	99.79	99.13
Cr	43	699	94	174	158	31	27	33	31	25	152	390
Rb	28	10	34	1	108	46	0	12	30	14	0	55
Sr	75	279	164	215	731	450	197	270	310	306	101	301
Y	37	21	42	18	18	18	21	17	22	47	29	22
Zr	250	89	220	30	169	154	42	145	142	222	118	79
Nb	4	10	15	2	0	8	6	8	9	8	10	2
Ba	305	317	212	131	900	495	268	378	669	482	157	337
Ce	15	29	51	21	56	40	9	0	0	46	36	28
Hf	14	12	13	9	12	14	11	16	11	14	12	12
Ta	0	4	0	3	3	0	3	0	1	0	1	5
Th	3	1	1	1	10	4	0	1	3	9	0	2

	25	26	27	28	29	30	31	32	33	34	35	36
SiO_2	48.20	48.51	48.88	50.87	50.94	52.16	52.49	52.67	53.28	54.34	55.08	55.71
TiO_2	1.89	0.97	1.30	0.86	1.00	0.97	1.24	1.35	1.35	1.16	0.76	1.10
Al_2O_3	14.67	14.18	15.98	13.87	20.05	15.24	18.97	15.59	17.48	16.83	15.83	16.37
Fe_2O_3tot	15.02	12.71	14.15	13.26	12.00	11.8	10.52	13.02	11.94	10.94	11.06	9.29
MnO	0.26	0.21	0.22	0.21	0.26	0.18	0.16	0.23	0.19	0.18	0.15	0.14
MgO	6.01	8.60	5.89	8.00	4.53	6.98	4.68	4.86	4.17	3.65	4.98	2.67
CaO	10.94	10.49	9.54	8.97	8.34	8.90	7.77	7.14	6.89	6.96	8.34	6.14
Na_2O	2.22	2.70	3.12	2.37	2.18	2.54	3.25	2.37	3.62	2.90	2.79	3.61
K_2O	0.33	0.41	0.46	0.33	0.10	0.47	0.43	1.44	0.53	2.21	0.38	2.10
P_2O_5	0.16	0.07	0.15	0.24	0.29	0.12	0.17	0.33	0.14	0.39	0.10	0.40
S	0.05	0.00	0.02	0.02	0.12	0.04	0.18	0.15	0.32	0.02	0.44	3.36
Total	99.75	98.85	99.71	99.00	99.81	99.40	99.86	99.15	99.91	99.58	99.91	100.89
Cr	124	315	106	545	29	286	65	68	69	55	52	49
Rb	2	12	11	3	0	7	0	52	8	57	4	72
Sr	189	128	278	258	360	175	151	707	197	821	149	794
Y	29	17	25	21	17	16	36	23	21	22	16	20
Zr	105	44	88	69	71	87	115	92	91	111	65	105
Nb	7	0	3	3	6	2	2	8	4	10	0	9
Ba	53	62	148	110	228	133	58	606	83	710	79	847
Ce	20	10	21	36	15	19	25	28	21	18	6	28
Hf	14	12	16	17	14	8	15	9	13	9	14	16
Ta	6	7	2	0	3	0	4	0	2	0	0	4
Th	0	0	1	0	0	0	2	3	0	5	1	3

	37	38	39	40	41	42	43	44	45	46	47	48
SiO_2	56.1	61.35	61.38	64.19	66.68	69.79	71.28	71.47	71.65	73.84	74.95	43.99
TiO_2	1.09	0.76	0.75	1.01	0.48	0.50	0.47	0.17	0.37	0.39	0.20	3.63
Al_2O_3	16.62	15.11	17.63	15.16	16.15	14.63	12.64	15.63	15.17	13.21	13.73	15.91
Fe_2O_3tot	9.28	13.30	5.98	10.14	4.50	5.37	6.87	2.26	2.23	3.55	2.24	17.66
MnO	0.13	0.16	0.11	0.26	0.09	0.11	0.15	0.09	0.05	0.05	0.06	0.27
MgO	2.46	2.85	1.55	2.54	0.66	0.90	2.06	0.53	0.34	1.28	0.60	4.03
CaO	6.43	1.68	4.39	2.00	1.95	4.36	1.55	2.65	1.93	1.83	2.21	8.88
Na_2O	3.46	3.20	3.35	3.06	3.49	3.43	3.14	4.01	3.21	3.51	3.01	2.96
K_2O	2.05	1.07	4.11	1.12	5.49	0.46	1.27	2.80	4.63	1.85	2.63	0.36
P_2O_5	0.43	0.15	0.42	0.12	0.12	0.15	0.05	0.08	0.11	0.12	0.04	1.97
S	3.13	0.00	0.06	0.03	0.00	0.11	0.01	0.00	0.00	0.00	0.00	0.16
Total	101.19	99.63	99.73	99.63	99.61	99.81	99.49	99.69	99.69	99.63	99.67	99.82
Cr	47	30	38	31	38	28	57	27	36	70	34	25
Rb	71	23	139	5	171	7	21	110	101	86	86	4
Sr	839	143	517	197	423	190	153	327	290	162	311	400
Y	23	30	22	36	29	33	28	13	14	19	25	34
Zr	106	107	202	121	226	175	144	87	193	174	203	48
Nb	10	4	10	8	16	4	3	7	3	9	10	42
Ba	724	390	879	541	1042	150	330	777	972	304	518	222
Ce	24	7	61	1	31	0	0	0	14	3	2	53
Hf	11	9	14	12	13	15	11	11	18	13	19	9
Ta	5	0	0	0	3	0	0	0	2	0	5	3
Th	6	1	9	0	10	0	3	2	6	9	12	0

	49	50	51	52	53	54	55	56	57	58	59	60
SiO_2	49.04	49.08	51.54	53.03	53.38	54.54	55.09	55.63	56.17	57.04	57.43	57.48
TiO_2	0.77	1.26	1.59	0.73	0.95	0.88	1.27	1.26	1.10	1.11	1.10	1.07
Al_2O_3	19.92	18.02	18.20	21.07	18.28	16.51	15.39	15.32	15.51	16.66	15.9	15.47
Fe_2O_3 tot	11.26	12.92	11.51	9.17	10.09	8.66	8.32	7.93	7.69	7.52	7.44	7.86
MnO	0.21	0.21	0.19	0.16	0.20	0.14	0.12	0.11	0.11	0.11	0.10	0.12
MgO	4.91	4.67	4.07	2.67	4.06	4.96	5.06	4.95	4.89	4.04	4.06	4.38
CaO	10.33	8.90	7.31	7.82	8.22	7.11	6.08	6.34	6.45	5.41	5.27	5.77
Na_2O	2.56	3.18	3.28	4.05	2.97	3.81	3.88	4.15	4.10	4.43	4.32	3.98
K_2O	0.66	1.00	1.28	0.80	1.18	2.30	3.54	2.90	2.64	2.57	3.27	2.74
P_2O_5	0.07	0.36	0.71	0.17	0.32	0.43	0.64	0.75	0.73	0.58	0.58	0.56
S	0.01	0.04	0.02	0.08	0.03	0.01	0.02	0.01	0.05	0.01	0.02	0.02
Total	99.75	99.64	99.70	99.75	99.68	99.35	99.41	99.35	99.44	99.48	99.49	99.45
Cr	46	94	55	37	37	158	175	172	185	114	135	142
Rb	15	28	30	11	33	48	94	34	47	39	69	49
Sr	359	506	427	511	488	952	1331	1821	1640	1669	1373	1476
Y	19	27	19	20	30	22	22	17	16	14	21	17
Zr	34	111	89	44	137	204	262	146	182	147	226	181
Nb	1	9	8	4	4	9	26	9	3	5	9	5
Ba	156	598	418	260	397	1270	1286	1510	1204	1287	1283	1226
Ce	0	24	22	23	18	96	99	92	124	88	111	102
Hf	13	10	9	14	16	9	15	9	8	11	15	15
Ta	4	4	0	4	2	0	6	0	0	4	2	0
Th	1	0	0	1	6	1	3	2	0	0	1	6

	61	62	63	64	65	66	67	68	69	70	71	72
SiO_2	60.31	60.44	61.35	41.86	46.67	47.66	49.36	47.24	50.32	50.59	51.12	57.67
TiO_2	0.95	0.94	0.76	2.47	0.42	0.56	0.33	1.02	0.31	0.31	0.25	0.96
Al_2O_3	16.12	16.15	16.39	14.33	10.82	10.59	5.76	18.15	20.37	18.61	18.60	16.82
Fe_2O_3tot	6.51	6.09	7.09	16.17	9.64	9.47	7.25	10.45	8.73	7.38	6.91	9.22
MnO	0.09	0.09	0.12	0.20	0.17	0.17	0.15	0.15	0.13	0.14	0.14	0.14
MgO	3.26	3.18	3.15	9.10	11.67	13.93	14.71	10.78	7.80	8.31	8.09	3.16
CaO	4.94	4.78	5.69	9.44	17.13	13.61	19.58	8.60	8.96	11.76	12.07	6.00
Na_2O	4.18	4.59	2.96	2.17	0.88	1.40	0.80	2.28	2.62	2.40	2.41	3.49
K_2O	2.79	2.72	1.98	0.96	0.69	0.23	0.31	0.64	0.31	0.20	0.14	1.92
P_2O_5	0.40	0.43	0.21	0.10	0.10	0.19	0.02	0.17	0.01	0.01	0.01	0.31
S	0.02	0.02	0.00	0.06	0.05	0.01	0.02	0.09	0.37	0.20	0.18	0.01
Total	99.57	99.43	99.70	96.86	98.24	97.82	98.29	99.57	99.93	99.91	99.92	99.70
Cr	103	110	101	88	687	809	1094	164	210	128	109	40
Rb	57	37	62	16	3	0	6	17	0	2	0	67
Sr	1244	1583	417	289	520	242	76	399	286	316	306	434
Y	13	12	23	27	12	17	10	17	6	9	5	29
Zr	175	179	134	74	21	28	14	97	10	15	9	182
Nb	5	8	8	7	0	0	0	0	1	0	0	11
Ba	964	1201	569	359	135	60	84	162	119	95	80	495
Ce	52	43	27	40	6	11	0	25	0	2	6	88
Hf	15	9	16	16	12	12	13	11	11	9	14	13
Ta	0	3	4	4	1	1	4	3	0	0	3	0
Th	4	3	6	0	0	1	2	1	0	0	0	9

	73	74	75	76	77	78	79	80	81
SiO_2	59.54	61.98	62.11	63.18	63.40	63.85	63.43	64.98	75.68
TiO_2	0.76	0.79	0.89	0.84	0.66	0.56	0.67	0.57	0.21
Al_2O_3	17.20	16.92	16.44	16.93	17.50	16.34	17.12	17.02	13.21
Fe_2O_3tot	7.51	6.32	6.14	7.39	5.78	5.63	5.47	4.21	2.30
MnO	0.13	0.09	0.09	0.06	0.08	0.12	0.08	0.07	0.06
MgO	3.06	2.27	2.30	3.45	1.71	2.20	0.98	0.75	0.31
CaO	6.14	5.34	4.17	1.88	4.21	5.31	2.62	1.71	1.69
Na_2O	3.62	3.84	3.32	2.52	3.64	3.49	3.38	3.06	4.58
K_2O	1.47	1.86	3.83	3.16	2.38	1.89	5.31	6.66	1.64
P_2O_5	0.21	0.25	0.25	0.07	0.24	0.15	0.24	0.33	0.04
S	0.03	0.01	0.01	0.22	0.02	0.00	0.04	0.00	0.00
Total	99.67	99.67	99.55	99.70	99.62	99.54	99.34	99.36	99.72
Cr	59	58	79	159	46	56	42	31	28
Rb	36	58	122	136	79	42	116	100	14
Sr	624	600	692	239	540	533	235	197	192
Y	18	19	20	21	9	18	25	23	5
Zr	139	193	242	185	146	117	403	516	161
Nb	6	12	15	15	6	7	12	9	4
Ba	546	659	929	633	604	858	1576	2223	561
Ce	22	0	56	45	10	0	75	81	0
Hf	8	16	16	15	9	13	17	20	14
Ta	0	0	4	0	0	0	0	0	1
Th	4	1	4	12	3	1	6	8	2

	Numbers
Dome gneisses	1 - 6
Salonsaari Suite	7 - 13
Savijärvi Suite	14 - 22
Säviä Suite	23 - 47
Koivujoki Suite	48 - 63
Infracrustal rocks	64 - 81

Sample No	Rock type	Location	Map sheet	X/Y Coordinates
1	Orthogneiss	Salonsaari	3314 05	7028.320/476.400
2	Orthogneiss	Kylmälahti	3313 09	7007.700/482.000
3	Paragneiss	Kirkkosaari	3314 07	7016.000/483.480
4	Orthogneiss	Sikoniemi	3314 07	7016.740/482.500
5	Paragneiss	Joutsenniemi	3314 06	7030.850/477.300
6	Orthogneiss	Leväniemi	3313 09	7007.500/482.180
7	Mafic metalava	Sulkava	3314 09	7038.320/484.840
8	Mafic metalava	Kaitamo	3314 05	7023.400/476.600
9	Mafic metatuffite	Salonsaari	3314 05	7028.220/476.300
10	Mafic metalava	Salonsaari	3314 05	7028.220/476.320
11	Mafic metatuffite	Sikoniemi	3314 07	7016.440/482.400
12	Mafic metatuffite	Keski-Ukko	3314 05	7022.850/479.080
13	Interm. metatuffite	Kokkosaari	3314.07	7018.240/481.620
14	Calc-silicate rock	Kukkomäki	3314 08	7029.800/481.720
15	Mafic metatuffite	Kirkkosaari	3314 07	7016.160/484.090
16	Mafic metatuffite	Rusala	3314 04	7020.000/477.800
17	Calc-silicate rock	Pangansalo	3314 07	7015.820/486.840
18	Interm. metatuffite	Sulkava	3314 09	7038.500/486.050
19	Interm. metatuffite	Kalliokylä	3314 06	7030.800/479.840
20	Felsic metatuffite	Uivero	3314 08	7025.000/481.380
21	Felsic metatuffite	Heinämäki	3314 06	7031.280/475.170
22	Felsic metatuffite	Kotajärvi	3314 09	7037.900/483.100
23	Mafic metavolcanite	Pyöreinen	3314 09	7035.300/487.080
24	Uralite porphyrite	Haapakoski	3314 07	7018.600/486.820
25	Mafic metatuffite	Kirkkosaari	3314 07	7016.160/484.080
26	Mafic metalava	Kokkosaari	3314 07	7018.220/481.520
27	Mafic metalava	Valkeisjärvi	3314 09	7036.850/488.050
28	Mafic metalava	Sulkava	3314 09	7038.050/485.080
29	Cordierite-anthophyllite gneiss	Uivero	3314 08	7023.840/480.800
30	Uralite porphyrite	Jylhä	3314 04	7018.000/477.050
31	Mafic metatuffite	Karkonsaari	3314 08	7020.700/481.260
32	Mafic metalava	Haapajärvi	3314 08	7020.560/487.300
33	Interm. metatuffite	Jylhä	3314 04	7018.000/477.050
34	Interm. metavolcanite	Haapakoski	3314 07	7018.800/486.200
35	Interm. metavolcanite	Saarela	3314 05	7025.960/471.000
36	Intermediate metatuffite	Haapajärvi	3314 08	7020.560/487.300
37	Intermediate metatuffite	Haapajärvi	3314 08	7020.560/487.300
38	Cordierite-anthophyllite gneiss	Sulkava	3314 09	7038.440/485.250
39	Felsic metavolcanite	Lampuri	3314 08	7021.600/487.620
40	Cordierite-anthophyllite gneiss	Uivero	3314 08	7023.840/480.800
41	Felsic metavolcanite	Haapakoski	3314 07	7018.850/486.620
42	Felsic metavolcanite	Kovero	3314 08	7022.120/480.700
43	Garnet-cordierite-gneiss	Suuri Suvisuo	3314 09	7034.520/483.500
44	Felsic metavolcanite	Saarela	3314 05	7026.060/471.280
45	Felsic metavolcanite	Haapakoski	3314 07	7018.020/586.700
46	Felsic metavolcanite	Haapajärvi	3314 08	7020.780/437.400
47	Felsic metavolcanite	Pangansalo	3314 07	7015.680/486.600
48	Gabbro	Tommonmäki	3314 06	7031.500/479.700
49	Gabbro	Kalliokylä	3314 06	7031.500/479.700
50	Gabbro	Kotajärvi	3314 06	7036.240/474.820
51	Gabbro	Joutsenniemi	3314 08	7028.300/480.100
52	Gabbro	Uivero	3314 08	7023.120/480.800
53	Kotajärvi Metalava	Tommonmäki	3314 09	7031.780/481.650
54	Kotajärvi Metalava	Juurikka	3314 09	7039.300/482.400
55	Kotajärvi Metalava	Tommonmäki	3314 09	7031.650/481.860
56	Kotajärvi Metalava	Ruokokangas	3314 06	7031.020/472.680
57	Kotajärvi Metalava	Koivujoki	3314 06	7030.260/471.200
58	Kotajärvi Metalava	Kettuvuori	3314 06	7032.740/472.560
59	Kotajärvi Metalava	Sulkava	3314 09	7038.500/484.850

Sample No	Rock type	Location	Map sheet	X/Y Coordinates
60	Kotajärvi Metalava	Myhkyri	3314 05	7021.800/479.700
61	Kotajärvi Metalava	Koivujoki	3314 06	7030.260/471.200
62	Kotajärvi Metalava	Heinämäki	3314 06	7031.200/475.200
63	Kotajärvi Metalava	Kaitamo	3314 05	7023.300/476.460
64	Peridotite	Kivimäki	3314 09	7033.200/484.200
65	Mafic inclusion in orthogneiss	Salonsaari	3314 05	7025.400/477.500
66	Peridotite	Salonsaari	3314 05	7026.880/477.900
67	Peridotite	Salonsaari	3314 05	7026.880/477.900
68	Peridotite	Saarela	3314 05	7029.180/472.600
69	Gabbro	Tyypekki	3313 09	7008.700/484.400
70	Gabbro	Tiirinluoto	3314 05	7028.200/472.180
71	Gabbro	Saarela	3314 05	7026.220/473.060
72	Quartz diorite	Kivimäki	3314 09	7033.320/483.800
73	Quartz diorite	Saarela	3314 05	7025.480/470.900
74	Quartz diorite	Pielavesi	3314 07	7017.000/488.000
75	Granodiorite	Säviä	3313 09	7008.850/483.300
76	Granodiorite	Vaaraslahti	3314 08	7028.400/484.500
77	Granodiorite	Ruokokangas	3314 06	7031.020/472.720
78	Granodiorite	Kolujärvi	3314 03	7035.500/465.400
79	Porphyry granite	Mäntysaari	3314 05	7029.080/472.800
80	Porphyry granite	Hirvivuori	3314 09	7031.700/485.200
81	Aplite	Lapinsaari	3314 08	7023.400/481.900

Appendix 2. U-Pb analytical results

2 A: U-Pb zircon data / A 291 - Pielavesi, Kirkkosaari; orthogneiss.

Sample no.	Fraction (g/cm)	Concentration µg/g ^{238}U	Concentration µg/g ^{206}Pb	measured $^{206}Pb/^{204}Ph$	Isotopic composition of lead $^{206}Pb=100$ 204	207	208	Atom rations and radiometric ages, Ma $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A297A	d>4.6	232.0	62.05	23559	.003115	11.741	6.183	.3091	4.986	.11699
								±19	±31	±5
								1736	1816	1910
B	d>4.6	174.2	49.63	79029	-	11.745	6.477	.32938	5.334	.11745
								±69	±13	±9
								1835	1874	1917
C	4.2<d<4.6	436.3	111.59	20964	.004087	11.736	6.385	.2956	4.761	.11681
								±15	±25	±5
								1669	1778	1908
D	4.0<d<4.2	968.7	238.00	15244	.005407	11.665	6.113	.2840	4.538	.11592
								±15	±24	±8
								1661	1737	1894
A291E	4.2<d4.6	307.7	81.89	12827	.006594	11.751	7.599	.3076	4.945	.116622
								±16	±27	±7
								1728	1809	1905
F	4.0<d<4.2	398.5	104.45	4474	.01884	11.831	8.117	.3029	4.836	.11577
								±19	±32	±19
							1705	1791	1892	
G	d>4.6	191.6	53.12	9742	.00856	11.806	7.984	.3204	5.164	.11690
								±22	±36	±9
								1791	1846	1909
H	d>4.6	194.0	53.12	13451	.005802	11.751	7.800	.3165	5.094	.11673
								±18	±29	±11

2 B. U-Pb zircon data / A1199-Pielavesi, Kirkkosaari; paragneiss

Sample (Zircon, density/size) 238U		^{206}Pb/ (ppm)	Measured1) $^{207}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	1) $^{206}Pb/^{206}Pb$	2) $^{207}Pb/^{238}U$	2) $^{207}Pb/^{235}U$	Age, Ma ^{206}Pb	
A1199A	+4.5/100-200	206	14193	0.1348	0.1280	.36565	6.7571	2151	
1199B	4.3-4.5/+200	421	15360	0.1328	0.1088	.35542	6.4698	2125	
1199C	4.2-4.3/+200		702	12198	0.1323	0.1057	.34785	6.2942	2114

1) Corrected for blank
2) Corrected for blank and initial common lead (206/204=14.7, 207/204=15.8, 208/204=34.3

2 C. U-Pb zircon data / A178 - Pielavesi; Saarela granodioritic gneiss.

Sample no.	Fraction (g/cm)	Concentration µg/g ^{238}U	Concentration µg/g ^{206}Pb	measured $^{206}Pb/^{204}Ph$	Isotopic composition of lead $^{206}Pb=100$ 204	207	208	Atom rations and radiometric ages, Ma $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A178-Saarela, Pielavesi; granodioritic gneiss										
A178aA	total	953.3	265.78	10809	.001242	12.070	6.272	.3222	5.355	.12053
	borax fusion							±22	±41	±36
								1800	1877	1964
aB	total									
	abr 2h	946.2	270.83	15637	.004802	12.204	6.241	.3308	5.537	.12140
								±33	±30	±11
								1842	1906	1977
bC	d>4.3	360.8	105.34	6320	.01348	11.900	9.190	.3374	5.452	.11718
	clear							±18	±29	±10
	abr 3h							1874	1893	1913
bD	d>4.3	457.5	132.75	5105	.01436	11.909	8.317	.3354	5.417	.1172
	Ø>70;abr 3h							±18	±60	±11
	long crystals							1864	1887	1913

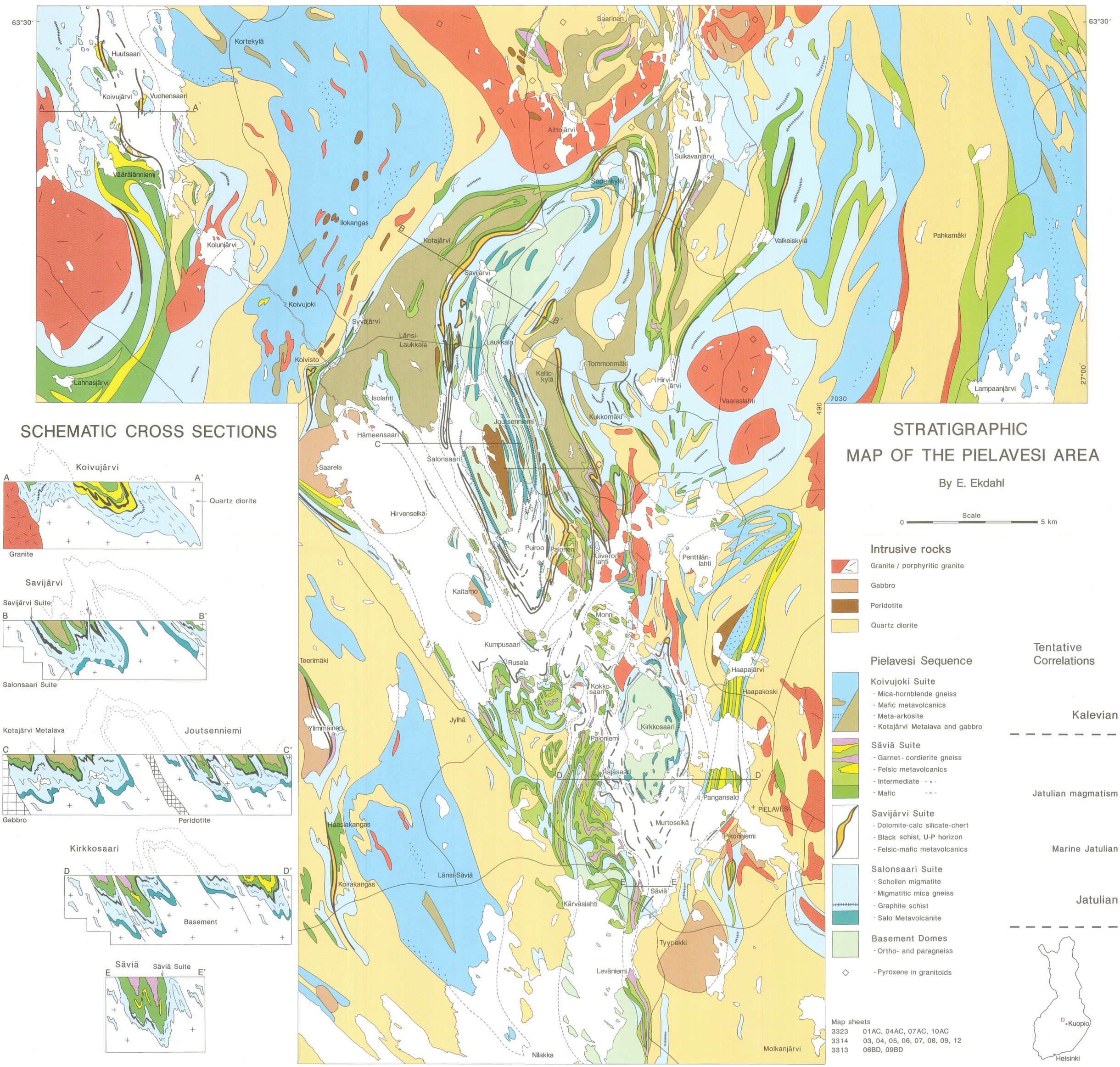
2 D. U-Pb isotopic data on zircons from the Saarisenjärvi gabbro of Tervo and the Tyypekki gabbro of Pielavesi. Decay constants: Jaffey et al. 1971.

Sample no.	Fraction (g/cm)	Concentration µg/g ^{238}U	Concentration µg/g ^{206}Pb	measured $^{206}Pb/^{204}Ph$	Isotopic composition of lead $^{206}Pb=100$ 204	207	208	Atom rations and radiometric ages, Ma $^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
A294A	d>4.6	131.5	38.20	32077	.001213	11.476	14.928	.3358	5.306	.11459
								±17	±28	±8
								1866	1869	1873
B	d>4.6; HF	129.1	37.80	22858	.002620	11.500	14.866	.3386	5.351	.11464
								±18	±29	±7
								1879	1877	1874
C	4.2<d4.6	421.2	121.46	48448	.001447	11.489	10.690	.3333	5.270	.11469
								±17	±28	±3
								1854	1864	1875
D	4.0<d<4.2	892.2	257.23	18381	.003481	11.506	9.282	.3332	5.264	.11459
	abr							±18	±29	±8
								1853	1863	1873
A298-Tyypekinlampi, Pielavesi; gabbro										
A298A	d>4.3;abr	397.1	116.48	11024	.004188	11.527	16.276	.3390	5.362	.11471
								±18	±31	±19
								1881	1878	1875
B	d>3.6;abr	909.1	262.44	12638	.005592	11.548	20.517	.3337	5.277	.11473
								±18	±30	±18
								1856	1865	1875
C	4.0<d<4.2	974.8	282.30	20047	.002878	11.508	17.724	.3347	5.293	.11470
	Ø>70;abr							±19	±30	±6
								1861	1867	1875

abr = grains abraded

2 E. U-Pb zircon data / A 189 - A 853 Haapajärvi, Kumiseva; pyroxene gabbro-gabbro pegmatoid-unakite

Zircon fraction		Concentration		Atom ratios							Acoepted
d = desity Ø = size in um HF= preleached in HF		Ug/g		measu-red	blank corrected			corrected for blank and common lead; age (Ma)			age
		^{238}U	^{206}Pb radio-genic	$^{206}Pb/^{204}Pb$	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{206}Pb$	$^{208}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}U$	Ma
A189-Haapajärvi, Kumiseva; pyroxene gabbro											
A	d>4.55	166.7	48.82	13485	20914	11557	.12002	.3386 ±32 1879	5.365 ±52 1879	.11492 ±15 1878	1879
	4.3<d<4.55	328.6	95.88	2059	2104	.12138	.15730	.3373 ±21 1873	5.346 ±52 1876	.11495 ±75 1879	
C	4.3<d<4.55	324.3	95.21	17761	22432	.11559	.14010	.3393 ±21 1883	5.378 ±36 1881	.11498 ±26 1879	
A852-Haapajärvi, Kumiseva; gabbro pegmatoid											
A	d>4.3 Ø >70	388.9	112.30	16681	21147	.11550	.13580	.3337 ±23 1856	5.285 ±38 1866	.11486 ±10 1877	1879
B	4.2<d<4.3	701.3	203.4	2085	2113	.12139	.15832	.3352 ±22 1863	5.314 ±42 1871	.11499 ±45 1879	
A853-Haapajärvi, Kumiseva; unakite											
A	4.4<d<4.6	334.6	87.86	1652	1689	.12332	.10604	.3035 ±16 1708	4.826 ±26 1789	.11532 ±14 1884	1897
B	4.3<d<4.55 Ø <160	491	111.0	1018	1029	.12716	.121134	.2613 ±14 1496	4.107 ±27 1655	.11401 ±33 1864	
C	4.4<d<4.6 HF; uncrushed	316.0	86.07	1918	1967	.12228	.10319	.3148 ±23 1764	5.009 ±38 1820	.11541 ±19 1886	
D	4.3 d 4.5 Ø <160; abr clear;colourless	416.8	102.02	989	.09866	12.794	12.479	.2829 ±15 1606	4.470 ±25 1725	.11460 ±17 1873	
E	4.3 d 4.5 abr long crystals	503.5	115.28	877	.11208	12.956	12.977	.2646 ±14 1513	4.173 ±24 1668	.11440 ±19 1870	
F	4.3 d 4.5 abr large, short dark crystals	456.5	108.43	967	.10087	12.849	12.056	.2745 ±14 1563	4.347 ±24 1702	.11486 ±11 1877	



STRATIGRAPHIC MAP OF THE PIELAVESI AREA

By E. Ekdahl

Scale 0 — 5 km

Intrusive rocks

- Granite / porphyritic granite
- Gabbro
- Peridotite
- Quartz diorite

Pielavesi Sequence

Koivujoki Suite
- Mica-hornblende gneiss
- Mafic metavolcanics
- Meta-arkosite
- Kotajärvi Metalava and gabbro

Säviä Suite
- Garnet-cordierite gneiss
- Felsic metavolcanics
- Intermediate - " -
- Mafic - " -

Savijärvi Suite
- Dolomite-calc silicate-chert
- Black schist, U-P horizon
- Felsic-mafic metavolcanics

Salonsaari Suite
- Schollen migmatite
- Migmatitic mica gneiss
- Graphite schist
- Salo Metavolcanite

Basement Domes
- Ortho- and paragneiss
- Pyroxene in granitoids

Tentative Correlations

Kalevian

Jatulian magmatism

Marine Jatulian

Jatulian

Map sheets

3323	01AC, 04AC, 07AC, 10AC
3314	03, 04, 05, 06, 07, 08, 09, 12
3313	06BD, 09BD

Kuopio

Helsinki

SCHEMATIC CROSS SECTIONS

Koivujärvi

A — A'

Quartz diorite

Granite

Savijärvi

Savijärvi Suite

B — B'

Salonsaari Suite

Kotajärvi Metalava

Joutsenniemi

C — C'

Gabbro

Peridotite

Kirkkosaari

D — D'

Basement

Säviä

Säviä Suite

E — E'

KARTTAKESKUS, Helsinki 1993

RAAHE - LADOGA ORE ZONE

Synorogenic stage

- Granitoids enriched in Cu±Mo±Au, Au-As (late orog.)
- Ni-Cu deposits

Early orogenic stage

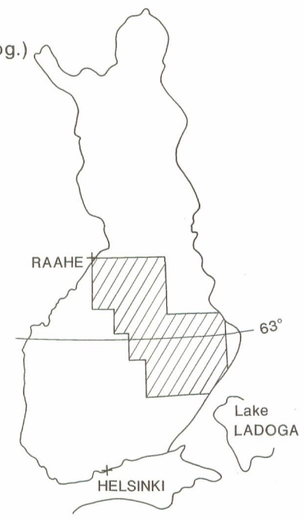
- Stratabound Zn-Cu-Pb-mineralization
- Outokumpu-type association

Pre- to early orogenic stage

- Stratiform U-P deposits, BIF
- Intrusions enriched in Fe-Ti-V

Preorogenic stage

- 2.1 Ga Fe-tholeiites
- 2.2 Ga Karjalites



Scale

0 10 20km

Compiled by E. Ekdahl 1993

Explanation

- Jotnian sediments (Muhos Formation)
- Approximate extent of early orogenic volcanism
- Serpentinite
- Mafic dykes
- Carbonatite
- Proterozoic / Archean
- Trace of overthrust
- Right lateral wrench fault
- Fault and shear zone
- Second order faults

Ore deposits and showings

Cu ± Au, Au - As

1. Laivakangas
2. Kurula
3. Ängesneva
4. Vesiperä
5. Pöhiölä
6. Kopsa
7. Saanijärvi
8. Ritovuori
9. Mäkrä
10. Lahnanen
11. Tuohilahti
12. Pirilä
13. Hakojärvi
14. Luomasenlampi
15. Osikonmäki

Ni - Cu

1. Perkkiö
2. Hitura
3. Makola
4. Kusiaiskallio
5. Ajakainen
6. Nuottijärvi
7. Nuoranen
8. Kumiseva
9. Saarinen
10. Ilokangas
11. Koivujoki
12. Saarela
13. Tuliniemi
14. Saarisenjärvi
15. Talluskanava
16. Kotalahti
17. Koirusvesi
18. Sarkalahti
19. Varkaus
20. Niemilahti
21. Laukunkangas
22. Makkola
23. Hälvälä
24. Kuokkala
25. Kekonen
26. Saarikko

Zn - Cu - Pb

1. Aittola
2. Siliäneva
3. Lampinsaari
4. Nevasaari
5. Kuuhkamo
6. Vilminko
7. Pelkoperä
8. Näsälä
9. Viitasenjärvi
10. Ritokoski
11. Hätämaanperä
12. Piipsanneva
13. Pyhälänmäki
14. Naurisniemi
15. Honkaperä
16. Ristisenoja
17. Saviselkä
18. Vuohtojoki
19. Karkumaa
20. Lehtikangas
21. Kurpas
22. Kaskela
23. Sirviö
24. Korpijoki
25. Mullikkoräme
26. Särkijoki
27. Ruotanen
28. Hallaperä
29. Kalliokylä
30. Ollinniemi
31. Hulanperä
32. Väärälä
33. Lahnasjärvi
34. Kumpusjärvi
35. Kangasjärvi
36. Teerimäki
37. Koirakangas
38. Kärnä
39. Sulkava
40. Puiroo
41. Kärppäsaari
42. Paloniemi
43. Säviä
44. Leväniemi
45. Vauhkola
46. Äyskoski
47. Vehkalampi
48. Pukkiharju
49. Syvänsi
50. Pärnänen
51. Hakuniemi
52. Karankalahti
53. Sahinjoki
54. Hällinmäki
55. Mustalampi
56. Luomanen
57. Lahnalahti

Cu - Co - Zn - Ni

1. Koutaniemi
2. Saviranta
3. Talvivaara
4. Alanen
5. Pappilanmäki
6. Korpimäki
7. Hankamäki
8. Losomäki
9. Vehkalahti
10. Sorsasaari
11. Kokka
12. Luikonlahti
13. Hoikka
14. Riihilahti
15. Usinjärvi
16. Kuusjärvi
17. Keretti
18. Vuonos
19. Perttilahti
20. Marjakuiva
21. Horsmanaho
22. Kylylahti
23. Sola
24. Saramäki
25. Miihkali
26. Joutenlahti
27. Papinniemi
28. Leppälahti
29. Petäjäjärvi
30. Hietajärvi
31. Savonranta
32. Kangaslampi

U - P

1. Relletti
2. Lampinsaari
3. Nevasaari
4. Vilminko
5. Huutsaari
6. Tossavanlahti
7. Koivisto
8. Savijärvi
9. Sulkava
10. Uivero
11. Haasiakangas
12. Toso
13. Vironniemi
14. Kuivasteenmäki
15. Rahvo
16. Temo
17. Tesmo
18. Neulamäki
19. Pitkälahti
20. Näärinki
21. Tutunen
22. Louhi
23. Pääkkö (BIF)
24. Koutaniemi
25. Nuottijärvi (BIF)
26. Tuomivaara
27. Losonvaara
28. Viitaniemi
29. Rukkajärvi
30. Mantilanniemi
31. Louhonsaari
32. Karhunsaari

Cu

1. Hammaslahti

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