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Crustal evolution of the Svecofennian and Karelian domains during 2.1-1.79 Ga, with special emphasis on the geochemistry and origin of 1.93-1.91 Ga gneissic tonalites and associated supracrustal rocks in the Rautalampi area, central Finland

by Raimo Lahtinen

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CRUSTAL EVOLUTION OF THE SVECOFENNIAN AND KARELIAN DOMAINS DURING 2.1 - 1.79 GA, WITH SPECIAL EMPHASIS ON THE GEOCHEMISTRY AND ORIGIN OF 1.93 - 1.91 GA GNEISSIC TONALITES AND ASSOCIATED SUPRACRUSTAL ROCKS IN THE RAUTALAMPI AREA, CENTRAL FINLAND

by

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with 29 figures and 7 tables

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The Rautalampi district is situated in the Savo Schist Belt (SSB) adjacent to the Archean craton and within the transition zone between the Svecofennian and Karelian domains. It contains 1.93-1.91 Ga gneissic tonalites and associated supracrustal rocks that formed within a Paleoproterozoic immature island arc. Gneissic tonalites are cogenetic and probably comagmatic with felsic volcanics and they represent low degrees of melting of low-K tholeiitic island arc basalts with gabbroic to amphibolitic residues. The Pukkiharju Zn-Cu occurrence is a VMSD-type mineralization associated with altered rocks now metamorphosed to cordierite-sillimanite gneisses and garnet±cordierite±orthoamphibole/orthopyroxene (GCO) rocks and gneisses. Cordierite-sillimanite gneisses occur within the mineralization zone and GCO rocks and gneisses, characterized by a lack of sulphides and showing mainly felsic and mafic volcanic protoliths, occur below the mineralized zone and are considered to represent a semi-conformable lower alteration zone. GCO rocks and gneisses in the Toholampi area contain abundant sulphides and were mainly derived from a mixed sedimentary/tuffaceous protolith; they are interpreted as representing the lateral continuations of the mineralized zone.

The Kangasjärvi, Pielavesi, Pyhäsalmi and Venetpalo areas all contain 1.93-1.91 Ga rocks that show geological and geochemical similarities with the gneissic tonalites and supracrustal rocks at Rautalampi. The Kettuperä gneiss (subvolcanic or volcanic rock) near Pyhäsalmi represents a slightly more evolved and younger source but otherwise the Pyhäsalmi region shares many of the general features of 1.93-1.91 Ga evolution. These rocks and associated VMSD occurrences are considered to represent aborted rift stage at about 1.92 Ga ago.

A plate tectonic model is presented which includes break up of a continent at 2.1 Ga and continental evolution in the Karelian domain during 2.1-1.91 Ga. During this time period, the basement to the Svecofennian domain was possibly developing in an exotic location with respect to its present position. Three different collision events (orogenies) at 1.91-1.90, 1.89-1.88 and 1.86-1.84 Ga are distinguished, involving both thin-skinned tectonics and crustal scale thrusting.

Key words (Georef Thesaurus, AGI): schist belt, plutonic rocks,tonalite, metavolcanic rocks, gneisses, cordierite, anthophyllite, geochemistry, genesis, plate tectonics, plate collision, underplating, Paleoproterozoic, Karelian, Svecofennian, Rautalampi, Finland

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INTRODUCTION

The Savo Schist Belt (SSB in Fig. 1) lies adjacent to the Archean craton and contain numerous Zn-Cu and Ni deposits and occurrences, associated with the Main Sulphide Ore Belt of Finland (Kahma 1973) or, as it is sometimes known, the Raahe-Ladoga Zone (RLZ). This also coincides with the boundary zone between the Archean craton and Early Proterozoic juvenile crust (Huhma 1986). Isotope data (Vaasjoki 1981, Huhma 1986, Lahtinen and Huhma in prep.) indicate a rather primitive origin for sulphides in the Zn-Cu occurrences and for the associated host rocks. The SSB shows a pronounced metamorphic block structure (Korsman et al. 1984) with abundant pyroxene granitoids surrounded by contact metamorphic aureoles (Hölttä 1988). The presence of highly deformed orthogneisses in the SSB has generated controversies concerning the "basement" and correlations of supracrustal rocks with cover sequences in the craton have been also proposed (Ekdahl 1993). Altogether, these rocks occupy a critical position when considering Paleoproterozoic crustal evolution and collisional tectonics. The Svecofennian as a whole in Finland is characterized by abundant granitoids (Fig. 1) having rather similar ages and lacking clear regional age zonation trends (1.89-1.88 Ga, e.g. Huhma 1986). The 1.93-1.91 Ga ages obtained from the SSB represent the main exceptions to these generalizations, and consequently form a major part of this study. This indicates very rapid crustal growth and the general importance of granitoids in Svecofennian crustal evolution. The magmatic, structural and metamorphic overprinting during 1.89-1.88 Ga events has obscured or destroyed many older features and thus hinders the interpretation of older events.

The Paleoproterozoic Svecokarelian granitoids have traditionally been classified as synkinematic (synorogenic) and late-kinematic (late-orogenic) granitoids (eg., Eskola 1932, 1960; Saksela 1936; Wahl 1936). Sederholm (e.g. 1932) classified granitoids into four groups of which group I can be correlated with synkinematic and group II with late-kinematic granitoids. Postorogenic granitoids belong to his third group and rapakivi granites to his fourth group. Simonen (1960, 1980b) divided the granitoids of the Svecofennidic area into different petrographic provinces according to the most acid end members as follows: granodiorite, trondhjemite, charnockite, granite and microcline granite provinces. The first four of these were considered to represent synkinematic (synorogenic) granitoids, which are anatectic crustal magmas derived from parental magmas of slightly varying composition (quartz dioritic-granodioritic). The late-orogenic microcline granites were considered metasomatic and anatectic in origin (Simonen, 1960,1980b). Small postkinematic (postorogenic) stocks of massive granitoids (Sederholm 1932) and post-Svecokarelian rapakivi granites are the youngest granitoids (Simonen 1980b). Nurmi and Haapala (1986) divided the Proterozoic granitoids on the same basis into four main groups: synkinematic (1860-1930 Ma, most commonly 1870-1890), late-kinematic (1800-1850) and postkinematic (ca. 1800 Ma) Svecokarelian granitoids and anorogenic rapakivi granites (1540-1700 Ma). The occurrence of 1.93-1.91 Ga gneissic tonalites (Helovuori 1979; Korsman et al. 1984; Vaasjoki and Sakko 1988, Ekdahl 1993) has been problematic. For example, the Kettuperä granite gneiss in Pyhäsalmi area and

orthogneisses in Pielavesi area has been considered to be basement to the volcanics (Helovuori 1979, Ekdahl 1993). The synkinematic granitoids have been further subdivided to granitoids in schist belts and the granitoid complex of central Finland (Front and Nurmi 1987). The granitoids intruding schist belts are predominantly granodiorites, tonalites and trondhjemites, and voluminous granites are absent. The granitoid complex of central Finland is composed of intrusions ranging in composition from gabbro to more voluminous granite. The most marked differences between these groups are higher FeO/MgO and K_2O/Na_2O ratios for similar SiO_2 contents in the granitoid complex (ibid.).

Based on the calc-alkaline and I-type characteristics of synkinematic granitoids, a partial melting model of igneous material was proposed by Front and Nurmi (1987). They invoked partial melting of a basaltic source for the origin of tonalitic rocks based on small mafic inclusions interpreted as restite material. The granodiorites and granites are considered to represent different magma pulses and melting of more crustal meta-igneous material and/or differentiation lower in the crust. The difference between granitoids in schist belts and granitoid complex was attributed to a lower degree of melting of thicker crust in the granitoid complex. Nironen (1989a) studied five Svecofennian granitoid plutons and interpreted some of the mafic inclusions as microgranitoid enclaves indicating magma mingling and possibly also magma mixing. Repeated mixing and mingling between crustal and mantle-derived magma, or magma derived by partial melting of a subducted oceanic slab combined with partial melting and/or assimilation of crustal rocks, was proposed for generation different magmas. The Svecofennian granitoids studied so far have $\varepsilon_{Nd}(1.9)$ from -1 to +3 and show no evidence for any significant Archean crustal contribution, indicating instead that they consist largely of juvenile mantle derived material with a minor admixture of older continental crust (Huhma 1986).

The geochemistry of gneissic tonalites (especially 1.93-1.91 Ga group) and related rocks in the Rautalampi area and in the reference areas have been studied to reveal their origin. The geochemistry and tectono-magmatic affinity of metavolcanics in the Rautalampi area are also considered. The garnet±cordierite±orthoamphibole/orthopyroxene (hereafter GCO) rocks and gneisses are an important rock assemblage that occur in association with Zn-Cu occurrences. Their geochemistry and origin is also studied with special emphasis on their relation to ore forming processes. The geochemistry of synkinematic granitoids and associated mafic rocks have also been studied to assess differences in source components and associated mantle type. A tentative plate tectonic model (2.1-1.79 Ga) is presented where the evolution of different schist belts and Svecofennian granitoids and crust as whole is evaluated in the light of the present data.

MATERIAL AND ANALYTICAL METHODS

Most of the material for this study was collected by the author between 1985-1988. Some drill core samples from the Pukkiharju area (Lahtinen 1988) are also included and a few samples have been contributed by other people. Altogether 263 samples from the Rautalampi area and 21 samples from reference areas are included in the study. The samples were taken from outcrops by a mini-drill or hammer, except for those from Kettuperä which were selected from diamond drill cores. The Kettuperä samples were from depths 55-80 m and 100-125 m from drill core PYS-27, which had previously been sampled (80-100 m) for U-Pb

zircon dating (Helovuori 1979). Thin sections were prepared from 173 samples.

Major and most trace element analyses of 141 samples were determined by ICP-AES techniques after $LiBO_2$-fusion at the Geological Survey of Finland. International standards and recommended values were used for calibration during major and trace element determinations. To compensate for instrumental drift international standard W-2 was analyzed as every fifth sample. Reanalysis, as well as five additional refusions of W-2 has been done to evaluate the analytical precision and accuracy. Accuracy is very good and analytical precision is good (1-3 %) for most of the elements considered here. Sodium, potassium, phosphorus, chromium and nickel have rather poor analytical precision (5-15 %) at the concentrations typical in this study. Five duplicate gneissic tonalite samples were sampled separately and analyzed to obtain the total variation, including sample heterogeneity. The results are in agreement with the refusion results and do not indicate any large sampling error. So far, only XRF-analyses from the Outokumpu Co laboratories are available for 122 samples of GCO rocks and gneisses. The oxide sums are normally low (about 90%) and so, these results are only used in crude comparisons to evaluate possible trends.

Rare-earth elements and some trace elements (Rb, Ta, Th, U and Hf) were determined from selected samples by the neutron activation method at the Technical Research Centre of Finland. The precision and accuracy of the method is presented by Rosenberg et al. (1982). The general precision is ± 2-10 %, but can be significantly worse near detection limits. The Rb, Ta, Th and U values in most mafic amphibolites are near or below the detection limits which thus hinders the use of results for these elements.

GENERAL GEOLOGICAL SETTING

The major Precambrian units in southern and central Finland are the Archean basement and the Palaeoproterozoic Svecokarelian terrains (Fig. 1). The latter has traditionally been divided into Karelides and Svecofennides based mainly on differences in supracrustal rocks (eg. Eskola 1963, Simonen 1980b, see also Gaál and Gorbatschev 1987). In northern and eastern Finland the Karelides are characterized by epicontinental sediments (Sariolan and Jatulian) deposited on the Archean crust between 2.45-1.9 Ga, as well as by metavolcanics and mafic dykes of different ages. The upper Karelian schists include marine sediments (Kalevian) deposited, at least partly on or at the edge of the Archean craton. The Karelides also contain the 1.97 Ga old Outokumpu ophiolite assemblage, which has been interpreted as an allochthonous nappe (Koistinen 1981). The Svecofennian comprise of ca. 1.9 Ga old orogenic terrains in southern and western Finland, comprising volcanic-sedimentary belts and migmatitic gneiss belts. The main schist belts are the Savo Schist Belt (SSB), Bothnian Schist Belt (BSB), Tampere Schist Belt (TSB) and Hämeenlinna Schist Belt (HSB) (Fig. 1). The Kemiö-Mäntsälä Belt (KMB) is composed of several smaller schist belts including the Orijärvi area. One major feature within the Svecofennian crust is the large Central Finland Granitoid Complex (CFGC).

A major suture zone has been delineated along the boundary of the SSB (Koistinen 1981) marking, together with the NW-SE-trending Raahe-Ladoga zone (RLZ) the approximate boundary zone between the Karelides and Svecofennides. The RLZ has been interpreted as a major strike-slip fault system (Gaál 1972, Talvitie 1975) with additional vertical components, resulting in distinct block-like domains

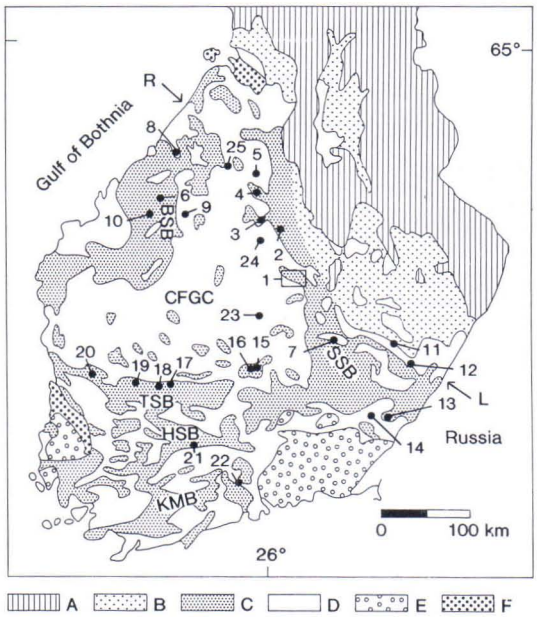

Fig. 1. Simplified geological map of southern Finland, after Simonen (1980a). A. Archaean rocks; B. Karelian schists; C. Svecofennian schists, gneisses and migmatites; D. Svecofennian plutonic rocks; E. rapakivi granites; F. Jotnian sedimentary rocks. CFGC - central Finland granitoid complex; SSB - Savo schist belt; BSB - Bothnian schist belt; TSB - Tampere schist belt; HSB - Hämeenlinna schist belt; KMB - Kemiö-Mäntsälä belt; R-L - Raahe-Ladoga zone. Study areas (1-7, this study; 8-25, Nurmi 1984): 1 - Rautalampi (see Fig. 2); 2 - Leväniemi, Pielavesi; 3 - Kangasjärvi; 4 - Pyhäsalmi; 5 - Venetpalo; 6 - Veteli; 7 - Saunakangas; 8 - Rautio; 9 - Halsua; 10 - Evijärvi; 11 - Varparanta; 12 - Silvola; 13 - Äitsaari ja Salosaari; 14 - Käkövesi; 15 - Sydänmaa; 16 - Hernemäki; 17 - Pukala; 18 - Värmälä; 19 - Hämeenkyrö; 20 - Kankaanpää; 21 - Aulanko; 22 - Mäntsälä; 23 - Palokka; 24 - Mäntylä; 25 - Kopsa.

of varying metamorphic grade (Korsman et al. 1984, Luosto et al. 1984). The Main Sulphide Ore Belt (Kahma 1973) with distinctive galena Pb isotopic compositions (Kouvo and Kulp 1961, Vaasjoki 1981) also coincides with the RLZ. The Sm-Nd isotopic data for plutonic rocks from northeast of the interpreted suture confirms the existence of Archean basement below the Proterozoic cover rocks (Huhma 1986).

As noted above, the Precambrian granitoids in Finland have been classified as synkinematic, late-kinematic and post-kinematic Svecokarelian granitoids and the rapakivi granites (eg. Eskola 1932, Simonen 1960, 1980b). The zircon U-Pb-ages of Svecofennian synkinematic granitoids range mainly from 1.90-1.87 Ga (Huhma 1986, Patchett and Kouvo 1986) and apart from the occurrence of 1.93-1.91 Ga aged gneissic tonalites (Helovuori 1979, Korsman et al. 1984, Vaasjoki and Sakko 1988, Ekdahl 1993) in the SSB they do not show any clear age zonation. The late-kinematic microcline granites of southern Finland occur in association with 1.84-1.81 Ga migmatites (Korsman et al. 1984, Hölttä 1986, Vaasjoki and Sakko 1988, Suominen 1991). Post-kinematic granitoids around 1.80 Ga (Vaasjoki 1977, Welin et al. 1983, Korsman et al. 1984, Huhma 1986) occur as small plutons. The rapakivi granites occur as batholiths and plutons and have isotopic ages from 1.65 to 1.54 Ga (Vaasjoki et al. 1991; see also Vaasjoki 1977).

GEOLOGY OF THE RAUTALAMPI AREA

The Rautalampi area is situated in the SSB adjacent to the Archean craton (Fig. 1) and is characterized by metamorphic blocks with stepwise differences in the metamorphic conditions between blocks (Korsman et al. 1984). The Rautalampi area forms a metamorphic block that has locally attained granulite facies conditions.

A simplified geological map of Rautalampi area is shown in Figure 2. The northern part of the map area contain only a few outcrops and is compiled mainly from the aerogeophysical data. The Rautalampi area contains volcanogenic amphibolites and felsic gneisses, gneissic tonalites and intermediate to felsic gneisses of unknown origin. Locally, migmatitic mica gneisses and rocks composed of quartz ±plagioclase, garnet, cordierite and orthoamphibole/

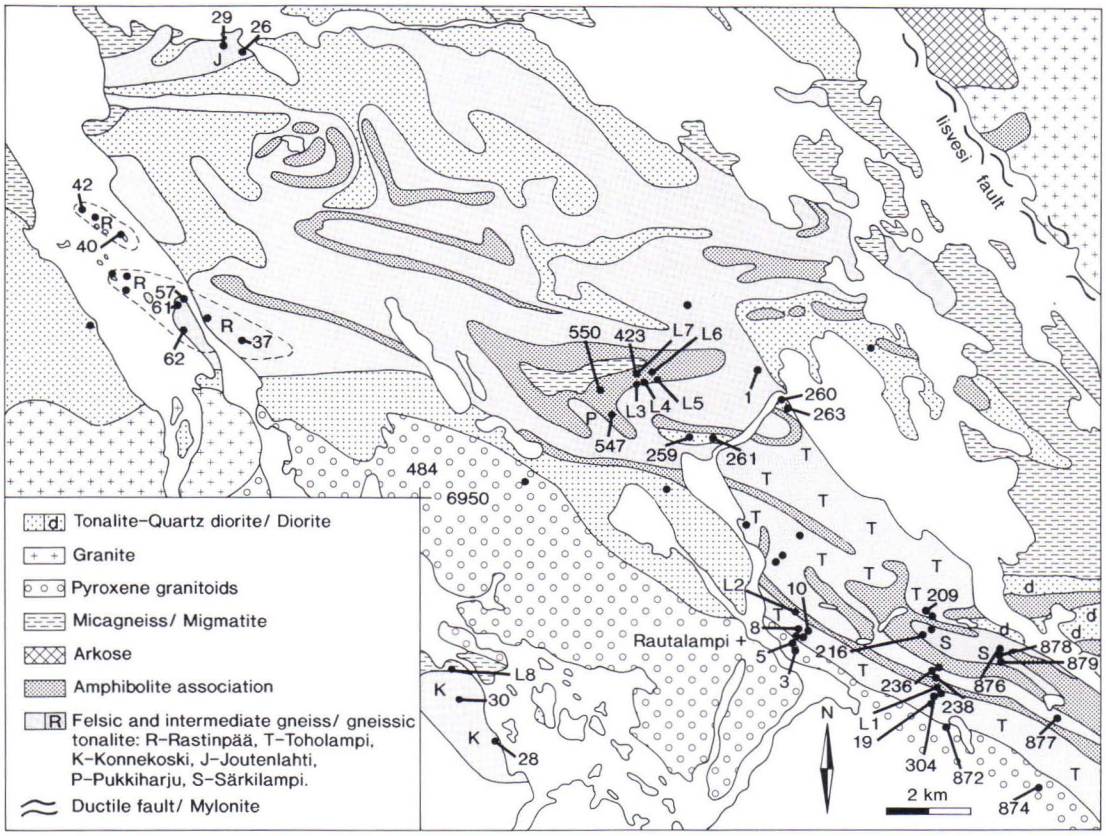

Fig. 2. Simplified geological map of the Rautalampi area based on the map by Pääjärvi (1991) and the mapping by the author and Äikäs (1977). The numbers refer to samples in Tables 1-3 and L1-L8 refer to sampled profiles of GCO rocks and gneisses (see text).

orthopyroxene are found. The main metallogenic feature is Pukkiharju Zn-Cu prospect (Lahtinen 1988, 1989). The inferred age of these rocks is 1.93-1.91 Ga, based on the age of the gneissic tonalites at Rastinpää (1922±12 Ma, Korsman et al. 1984) and Toholampi (1914±4 Ma, Pyöreänsuonvuori granitoid in Vaasjoki and Sakko 1988). The other tonalites and quartz diorites, locally also gneissic, are interpreted to be of synkinematic age (1.90-1.88 Ga).

The Rautalampi area is bounded to the north and west by the rocks of the CFGC and to the south by the hypersthene granitoid complex. The zircon age from the hypersthene granite (1887±16 Ma) (Korsman et al. 1984) constrains the lower age limit for the hypersthene granitoid complex. The Rautalampi block and the Kuopio block to the east, metamorphosed in the stability field of muscovite (Korsman et al. 1984), are separated by the Iisvesi fault (Fig. 2), which is characterized by locally protomylonitic to ultramylonitic porphyritic microcline granite.

Rastinpää gneissic tonalite

The Rastinpää gneissic tonalite is an oval shaped complex and has a satellite body in the

northwest (Fig. 2). In the aeromagnetic pixel map (not shown) the main complex can be subdivided into a lesser magnetic interior and to more magnetic outer rim, which in turn is partly divisible into two different parts. Because of the poor exposure, the contacts of the complex with the surrounding rocks are mainly obscure. Gneissic tonalites in the satellite body occur with migmatites, composed of biotite±garnet±hornblende gneisses with boudinaged amphibolite layers. The leucosome is tonalitic to granodioritic in composition. The western contact of the main body contains migmatites, granodiorites and gneissic tonalites that locally contain garnet and have abundant pegmatitic dykes. This area is strongly deformed and all rock types occur as rolled brecciated fragments in a migmatitic matrix. This is interpreted as a tectonic zone dividing the Rastinpää tonalite and associated paragneisses from the granitoid area to the west (Fig. 2). Fragmented mafic dykes with a within plate basalt (WPB)-affinity (discussed later) in the migmatite are also strongly deformed in this zone and exhibit a strong mineral lineation.

The Rastinpää gneissic tonalite is normally grey on weathered surfaces with quartz ribbons. It is partly granitized and migmatitic in character with mobilized tonalitic veins. Migmatization is strongest in the satellite body and in the western part of the main body. Boudinaged amphibolite dikes and xenoliths and fine-grained felsic dikes are locally present. Amphibolite dikes are 0.05-2 m wide and occur as flame-like structures or broken fragments and they can contain gneissic tonalite enclaves; younger pegmatitic tonalite and granite veins also crosscut them. Mafic enclaves interpreted as country rock xenoliths are found near the complex margins and are more intermediate in composition and banded in character compared to amphibolite dikes. Felsic 0.05-2 m wide fine grained dikes also occur as disrupted fragments in the gneissic tonalite.

Gneissic tonalites are fine to medium-grained and they normally have a granoblastic texture with coarser subidioblastic plagioclase grains suggestive of a blastohypidiomorphic texture. Quartz exhibits deformation bands and strong undulose extinction and also occurs as mobilized grain aggregates. The major minerals are plagioclase (40-50 %), quartz (40-50 %) and biotite (10-15 %). Hornblende occurs sporadically as a minor constituent. Accessory potassium feldspar occurs mainly as antiperthite and epidote, magnetite, apatite, zircon, carbonate and sericite constitute ubiquitous accessory phases. Mafic dykes in gneissic tonalite are fine to medium-grained rocks and granoblastic. Plagioclase (40-50%), hornblende (30-40%), clinopyroxene (10-20%), and occasionally biotite are major minerals. Clino-pyroxene occur both as small grains and as large grain aggregates. Biotite±sphene±apatite±opaques are found as accessories. Mafic xenoliths differ mineralogically from the dykes in having quartz but lacking clinopyroxene. The normal modal composition is plagioclase (50-60%), hornblende (30-40%), quartz (2-10%) and biotite (2-10%) with sphene, apatite and opaques as accessories. The WPB-affinity mafic dykes are fine-grained rocks with a lepidoblastic texture. In addition to plagioclase, hornblende and biotite, they contain abundant sphene, apatite, zircon and opaque.

Geology and petrography of the Rautalampi-Tyyrinmäki region

Rautalampi-Tyyrinmäki region is situated east and southeast of the town of Rautalampi (Fig. 2) and has been previously studied by Äikäs (1977). Supracrustal gneisses, amphibolites and different types of gneissic tonalites and trondhjemites predominate. To the south is a large pyroxene granitoid complex with pyroxene-bearing quartz diorites, tonalites and granodiorites at the margins. These are cut by coarse-grained granites and medium-grained quartz

monzodioritic dikes. The contact zone is characterized by breccias and migmatites with stromatic, agmatic and schlieren structures. Between the pyroxene granitoid complex and the Toholampi gneissic tonalite there is a 200-400 m wide sequence consisting of felsic gneisses (partly orthogneisses), hornblende gneisses, amphibolites and a 30-50 m wide zone veined garnet-biotite-orthopyroxene gneiss (migmatite) with clinopyroxene-rich boudins. The Toholampi gneissic tonalite is a sill-like body about 150 m wide that can be traced for at least 5 km. North of Toholampi gneissic tonalite occur a zone of amphibolite association rocks including a 10-20 m thick association of GCO rocks and gneisses. These rocks are found in several places (e.g. L1-L2 in Fig. 2) and seem to form a continuous stratiform horizon.

The volcanogenic amphibolites, felsic gneisses and garnet-rich felsic gneisses of the amphibolite association occur as zones from 100-300 m trending at 315° (Fig. 2). Garnet-cordierite-sillimanite gneisses containing graphite and pyrrhotite are present locally in the margins of the amphibolite associations (Äikäs 1977). The felsic gneisses are locally homogenized and mobilized. Coarse-grained tonalitic patches and segregations indicating incipient melting and quartz veins and lenses are common. Tonalitic patches are also found in amphibolites. Supracrustal associations are surrounded by tonalitic/trondhjemitic gneisses of varying origin. Some of them have the same characteristics as the Toholampi gneissic tonalite and could be interpreted as gneissic tonalites (marked also with T in Fig. 2), but others are gneisses with no indication of their origin. In the middle of the Rautalampi-Tyyrinmäki region is a body of gneissic tonalite at least 2 km long and 100-300 m wide that has been termed the Särkilampi gneissic tonalite. Crosscutting pegmatite granites are the youngest rocks in the area.

Pyroxene-bearing quartz diorites, tonalites and granodiorites in the pyroxene granitoid complex consist of fine to medium grained rocks that have locally a distinct foliation. The texture is hypidiomorphic and the dominant major minerals are plagioclase, quartz, biotite and, in granodiorite, K-feldspar. Hornblende, ortho- and clinopyroxene are also very abundant. K-feldspar occurs in quartz diorites and tonalites mainly as antiperthite and filling the interstices. Apatite, epidote and opaque occur as accessory minerals. Slightly foliated quartz monzodioritic dikes are medium-grained and the texture is hypidiomorphic, with the major minerals being plagioclase, quartz and K-feldspar. Hornblende and magnetite occur as minor constituents. Metamict allanite, pyroxene, zircon and epidote are accessories. Antiperthite and myrmekite are very common and perthite is locally present in antiperthite patches.

Felsic gneisses associated with the pyroxene granitoids and the Toholampi gneissic tonalite are trondhjemitic to tonalitic in modal compositions with the major minerals being plagioclase, quartz and biotite. Biotite however occurs only as a minor constituent (< 5%) in the most felsic gneisses. Other minor components are K-feldspar, hornblende, garnet and orthopyroxene. The gneisses form distinctive and partly ghost-like layers 0.05-3 m thick and locally have more leucocratic fragments and oval shaped fragments that are rich in epidote and anorthitic plagioclase. Felsic gneisses have been partly mobilized and seem to migmatize the amphibolites. These mobilized gneisses are rich in orthopyroxene compared with biotite. The origin of the felsic gneisses is problematic but as the texture is always granoblastic and they form layers with locally occurring fragments, they are interpreted as supracrustal in origin. Some of the more homogenous thicker layers may also be of tonalitic sill origin. Veined gneiss (migmatite) has plagioclase +quartz±K-feldspar veins alternating with biotite-rich veins with abundant garnet and pyroxenes. Amphibolite associated with veined gneiss consists of hornblende-plagioclase-clinopyroxene with abundant biotite.

The Toholampi gneissic tonalite resembles the Rastinpää gneissic tonalite. Bluish quartz

ribbons are very common as also are thin amphibolite dykes. Fragmental mafic enclaves occur near the contact and are interpreted as xenoliths. The main minerals in the gneissic tonalite are plagioclase (44-53 %), quartz (38-43 %) and biotite (4-10 %). Orthopyroxene occurs as a minor constituent in samples where biotite content is low. K-feldspar is also an abundant minor constituent, especially in granitized samples. Hornblende, apatite, magnetite and zircon are the accessory minerals. The texture is granoblastic and, in the best preserved samples, blastohypidiomorphic. Mobilized strongly undulose quartz aggregates and antiperthite are very common. The Toholampi gneissic tonalite differs from the Rastinpää gneissic tonalite in having lower biotite contents, abundant orthopyroxene and that it is lacking epidote. One mafic dyke contains elongated plagioclase and roundish hornblende aggregates in a fine-grained groundmass of plagioclase-hornblende-quartz with accessory pyroxene and opaque minerals showing blastoporphyritic texture. Veined mafic xenoliths near the northern contact are heterogenous with quartz and tonalitic segregations and are orthopyroxene-bearing (10%) with plagioclase, hornblende, biotite and quartz.

The Särkilampi gneissic tonalite is lighter in colour than other gneissic tonalites in the Rautalampi-Tyyrinmäki region. Major minerals include plagioclase (56-68 %), quartz (21-29 %), biotite (4-10 %) and locally hornblende (0-6 %). K-feldspar, apatite and opaque occur as accessory minerals. Chlorite occurs as an alteration product of hornblende. The texture varies from blastohypidiomorfic to granoblastic and mylonitic features are also locally observed. The Särkilampi gneissic tonalite differs from the other gneissic tonalites in having higher plagioclase and hornblende contents, orthopyroxene is also absent.

Amphibolite associations (Fig. 2) are composed of amphibolites, hornblende gneisses, felsic gneisses and garnet-bearing felsic gneisses. Amphibolite-felsic gneiss layering is locally seen as alternating layers of amphibolite and felsic gneiss 3 mm- 50 mm thick. Amphibolites also occur as 1-8 m thick layers that are locally cut by light coloured dykes (0.3-0.5 m). Amphibolites are often fragmentary rocks, which reflects their volcanic origin. One amphibolite association about 10 m thick contain ghost-like banded amphibolite with amphibole-rich patches (0.5x1.5 cm) of possible lapilli-origin. Although a volcanic origin is postulated for most of the amphibolites, a hypabyssal origin for the some of massive amphibolites is also probable. The occurrence of calc-silicate-rich bands and patches show that carbonate deposition possibly occurred intermittently during volcanism. Amphibolites often contain, abundant clinopyroxene in addition to hornblende. Orthopyroxene is not uncommon, and garnet and quartz-rich intermediate amphibolites are also present. Addition of carbonates is seen as accessory carbonate in some samples and clinopyroxene+plagioclase patches are found locally. The anorthite content of plagioclase (bytownite) in some samples is also very high. Retrograde chlorite and bluegreen amphibole are found sporadically. Garnet-bearing felsic gneisses are found as 10-20 m thick units alternating with thin amphibolite-micagneiss layers. Garnet can be also very abundant but normally these are quartz-rich rocks with plagioclase, biotite (10%) and variable amount of garnet.

North of the Toholampi gneissic tonalite is amphibolite association with GCO rocks and gneisses near to the contact. These have been studied in detail in two places (L1-L2 in Fig. 2). The southern location (L1) is situated north of sample 304 and here the contact with the Toholampi gneissic tonalite is covered by overburden. These rocks occur as 10-150 cm thick layers with 5-30 cm thick amphibolite layers in the southern part. The abundance of amphibolite layers increases northwards and the northern part is characterized by both homogenous and banded amphibolites (20-150 cm) with only sporadic occurrences of garnet-orthopy-

roxene/orthoamphibole gneisses. Orthopyroxene-bearing felsic gneiss layers are also present. The thickness of the association containing GCO rocks and gneisses is about 20 m. Locally these are coarse-grained rocks that exhibit evidence of having been mobilized e.g. to the necks of boudinaged amphibolites. Layers are normally rather distinct, but some very coarse-grained and heterogenous layers can be of composite origin. The first five meters of the southern part is composed of orthopyroxene/orthoamphibole-cordierite rocks with substantial amount of plagioclase occurring in only one sample. These are granoblastic to porphyroblastic rocks with quartz and phlogopitic mica as the major minerals. The most common accessories are green spinel, sulphides and other opaques. Garnet, zircon, apatite and both rutile and sphene are also occasionally found. Zircon shows a variety of morphologies; small zircon inclusions in cordierite are associated with coarse-grained (0.05-0.2 mm), clear and almost euhedral zircons, but metamict variants are not uncommon. Orthoamphibole (anthophyllite) surrounds orthopyroxene (hypersthene) and is also intergrown with biotite. The rocks to the north contain substantial amounts of plagioclase and are thus called gneisses. Otherwise the mineralogical composition is similar, but poikiloblastic (opaques+quartz +plagioclase) garnet is abundant in places. Hypersthene and plagioclase are also seen replacing garnet.

The other location (L2) studied in detail occur about 2 km northwest from L1 (Fig. 2). The contact with the Toholampi gneissic tonalite is exposed here and is apparently tectonic, with hypersthene-anthophyllite gneiss being strongly mylonitized. It shows a retrograde mineralogy where hypersthene and anthophyllite have been altered to chlorite, amphiboles (both hornblende- and cummingtonite-like) and unidentified alteration products. Phlogopitic mica, sericite and carbonate-rich seams are very common. Here the GCO association is about 22 m thick. Layering is similar to L1 with abundant amphibolite layers, but 10-20 cm thick felsic gneiss layers are more abundant. Mineralogically the samples are similar to the GCO gneisses of L1 and only one sample almost devoid of plagioclase is found. Phlogopitic mica has overgrown gneissic banding in one sample indicating late potassium activity.

Geology and petrography of the Pukkiharju area

A metamorphosed Zn-Cu-occurrence interpreted as volcanic-exhalative in origin is present in the Pukkiharju area (Lahtinen 1988, 1989) (Fig. 2). Volcanogenic amphibolites and felsic gneisses and calc-silicate rich rocks, mica gneisses, garnet-sillimanite gneisses and graphite-bearing rocks are also common in the area. Locally cordierite-sillimanite gneisses and GCO rocks and gneisses also occur. The petrography of the Pukkiharju area has been described in detail by Lahtinen (1988) and only certain features are briefly discussed here. Layered felsic gneiss (quartz porphyry) forms a lenticular body at most 50 m in thickness (Fig 2, sample 550) and has 1-3 mm "quartz eyes" in a fine-grained granoblastic groundmass which are considered to be recrystallized quartz phenocrysts. The main minerals present are plagioclase and quartz and with phlogopite being a minor constituent. In calc-silicate rich parts the rock also contains tremolitic amphibole, carbonate and zoisite. One gneissic tonalite crosscuts amphibolites and has the same predominant foliation as the surrounding rocks (Fig. 2, sample 547); major minerals are plagioclase (54 %), quartz (24 %) and biotite (11 %). Orthopyroxene, garnet, hornblende, apatite, opaque and zircon occur as accessories. About 200 m east of the quartz porphyry (sample 550 in Fig. 2) is a 70-100 m thick tonalitic gneiss

which also outcrops about 400m southeast. It has a blastoporphyric-hypidiomorphic like texture with plagioclase grains in a fine and even-grained granoblastic groundmass. Plagioclase grains are sometimes wrapped by the groundmass reflecting a porphyritic origin and in other instances occur as poikiloblastic porphyroblasts. Plagioclase anorthite content is low (An < 20) and it shows albitic twinning. Biotite is a major mineral, and garnet and orthopyroxene are abundant minor minerals. K-feldspar (antiperthite), green mica, apatite, opaques and zircon are typical accessories. This rock has been interpreted as an altered hypabyssal intrusion (discussed later) and cogenetic with volcanics.

Amphibolites have been divided into homogeous massive amphibolites, clinopyroxene amphibolites and banded amphibolites (Lahtinen 1988). Massive amphibolites have hornblende and plagioclase as major minerals and quartz, opaques and apatite as accessory phases. Clinopyroxene amphibolites are fragmentary or patchy rocks with clinopyroxene-plagioclase rich parts. Phlogopitic mica, tremolitic amphibole, quartz, carbonate, opaques, zoisite, sphene and apatite are accessory phases. Banded amphibolites contain amphibole-rich bands normally 0.2-5 cm thick that alternate with clinopyroxene-, quartz-feldspar- and mica-rich bands. Massive amphibolites are considered to have been lavas, but some of them may have been dykes. Clinopyroxene amphibolites show volcaniclastic features and are considered to represent brecciated lavas and/or tuff breccias and possibly also tuffs. Banded amphibolites originated as tuffs or tuffaceous deposits. The occurrence of carbonate and abundant calc-silicates indicate the activity of carbonate during deposition or later. Thin plagioclase phyric dykes (<1m) with WPB-affinity crosscut amphibolites and also pegmatites, and are the youngest rocks found from Pukkiharju area. They are blastoporphyritic rocks with 1-5 mm long plagioclase phenocrysts in a fine and even-grained granoblastic-lepidoblastic ground mass. Plagioclase, biotite, hornblende and quartz are the major phases and abundant apatite, sphene and zircon occur as accessory minerals with opaques, carbonate, clinopyroxene and K-feldspar (mainly as antiperthite).

Cordierite-sillimanite gneisses are found within and near to the mineralized horizons and contain abundant sulphides (Lahtinen 1988). They are characterized by quartz, K-feldspar, phlogopitic mica, sillimanite and cordierite. Plagioclase is uncommon (< 10%) and is absent in some cases. Stratabound GCO gneisses and rocks are found at several localities (L3-L7 in Fig. 2) in the Pukkiharju area. Three locations (L3-L5) form a continuous zone for at least 300 m. The thickness of this association is 50-100 m, but it is partly covered by overburden and so the sampling is not continuous between different locations. The most common rock type is a felsic gneiss with variable amounts of phlogopitic mica, orthoamphibole, garnet and cordierite. Orthopyroxene is found in small amounts sometimes inside orthoamphibole flakes. Bluish cordierite is often absent and cummingtonite is sometimes found. Accessory opaques, apatite, spinel and zircon are found in variable amounts, but zircon is lacking from many samples. Some garnet-bearing felsic gneisses have zircon both as small inclusions in biotite and as rounded crystals. Samples containing only small inclusions are also common. Rutile and hematized magnetite are also found. Garnet-orthoamphibole-cordierite-rich gneisses and rocks (devoid of plagioclase) are also present. Coarse alternating garnet- and orthoamphibole-rich bands are very common in these rocks. Garnet is sometimes poikiloblastic with fine-grained quartz, opaques and plagioclase. Rocks of this association occur as 10-150 cm thick layers alternating with 5-100 cm thick amphibolite layers. Both homogeneous and banded amphibolites occur and are sometimes rich in garnets, especially near contacts with the garnet-orthoamphibole-cordierite gneisses. Orthopyroxene and cummingtonite are also found in some amphibolites. Mobilization of coarse-grained garnet-orthoamphibole-cordier-

ite gneisses is not uncommon and anthophyllite-rich "pegmatites" also occur.

The other association studied (L6-L7) occurs north of the above association (Fig. 2), but it is uncertain whether this association is equivalent to that described above. The association at L7 is 18 m thick and occurs between an amphibolite 2.5 m thick to the south and another at least 6 m thick to the north. Its southern continuation is open, but the northern contact is well defined as only amphibolites and quartz-feldspar gneisses are found going further north. About 1.5 m north from the southern amphibolite is a garnet-bearing felsic gneiss with phlogopitic mica. The abundance of orthoamphibole increases northwards. The main association contains anthophyllite±garnet-bearing gneiss layers (normally 10-40 cm) that alternate with amphibolites (5-25 cm) and garnet-bearing felsic gneisses (10-20 cm). Anthophyllite is abundant and occurs as long flakes or as radiating aggregates. Garnet is abundant and garnet-orthoamphibole-cordierite rich variants are found near the northern contact. Amphibolite layers are often rich in garnet and garnet±orthoamphibole "pegmatites" are not uncommon.

The original local stratigraphy remains obscure due to the absence of younging observations but if the alteration and mineralization patterns are used, a tentative younging direction can be obtained (Lahtinen 1988). The tentative local stratigraphy is as follows. Mica gneisses (garnet-bearing felsic gneisses) and metavolcanics are lowermost followed by stratabound GCO gneisses. Above these is the amphibolite association with quartz-feldspar gneisses (metavolcanics and tuffaceous rocks). The proportion of felsic metavolcanics and subvolcanic sills (dome ?) is greatest near the Pukkiharju Zn-Cu mineralization (close to sample 550 in Fig. 2). The abundance of calc-silicate lithologies steadily increases in mafic and felsic metavolcanics, which are followed by impure calcitic dolomites and carbonate-rich calc-silicate gneisses. These are followed by graphite-bearing lithologies and very graphite-rich black schists (up to 22% C) are also found. Mica gneisses and garnet-sillimanite gneisses are the uppermost rocks. The present total thickness of this formation is less than 1 km.

Geology and petrography of the other areas

Rastinpää-type gneissic tonalites have been sampled in the Konnekoski and Joutenlahti areas (Fig. 2). South of Konnekoski, in the pyroxene granitoid complex, there is a "dome-like" occurrence of gneissic tonalite that is surrounded by migmatitic GCO gneiss about 10 m in thickness (L8 in Fig. 2). Contacts with surrounding rocks are poorly constrained. This association contains 10-40 cm thick layers of heterogenous and coarse-grained garnet-orthopyroxene-cordierite gneisses with thin (5-30 cm) layers of quartz-feldspar gneiss and garnet-bearing amphibolites. Boudinage of amphibolites and mobilization of garnet-orthopyroxene-cordierite gneiss is common. In the Joutenlahti area gneissic and locally banded tonalite has been sampled. To the northwest of Joutenlahti are banded felsic gneisses that have been interpreted as supracrustal in origin (Fig 2, sample 29). The modal composition of gneissic tonalites and banded felsic gneiss is trondhjemitic and their texture is typically granoblastic, although blastohypidiomorphic variants occur locally in the gneissic tonalites. Mineralogy differs from that of the Rastinpää gneissic tonalite in the lower biotite content (4-6 %). Minor retrograde bluish green amphibole also occurs.

Samples have also been taken from layered felsic gneiss of probable volcanic origin (Fig. 2, sampling site 1). The texture is granoblastic and the modal composition is tonalitic. Sam-

ples of gneissic tonalite and quartz diorite from SE of Pukkiharju (Fig. 2 samples 259 and 261) and from gneissic tonalite and granodiorite in Virtaniemi (Fig 2. samples 263 and 260), along with analytical data for granitoids in the Rautalampi area and adjacent areas to the north, have been provided by A. Pääjärvi.

Metamorphism and structural features

The metamorphism in the Rautalampi area has been studied by Korsman et al. (1984) and the main features are as follows. Granulite facies conditions (660°-700°C and 5.1-5.5 kb) were attained and this was related in time and space to the proximity of hypersthene-bearing granitoids. This is reflected in the presence of hypersthene in the Toholampi gneissic tonalite and in mobilized gneisses, and tonalitic melt patches in amphibolites near the hypersthene-bearing granitoids. The other gneissic tonalites are devoid of hypersthene and are more indicative of amphibolite facies conditions.

The structural features of the Rautalampi area have not been studied in detail and this combined with poor exposure in many areas leaves the structural evolution of the area open at this moment. In the Rautalampi-Tyyrinmäki area (see also Äikäs 1977) magnetic anomalies and pronounced schistosities have a predominant 290 strike. Schistosities are mostly subvertical but further north have more gentle dips. Lineations such as small-scale fold axes and mineral lineations are usually very strong and have normally gentle plunges to about 290 and 110 and also record a younger folding phase. The asymmetrical fold necks and boudins found locally indicate a strong shear component. Younger semi-brittle to ductile faults with trends of 310°-330° and 20°-60° are very common, and are accompanied by pegmatites and the mobilization of tonalitic material. These are also locally sheared and may well be correlated with the 320°-trending Iisvesi fault zone and its conjugates. If the almost undeformed tonalitic melt patches in amphibolites are correlated with the 1887 Ma hypersthene granite (Korsman et al. 1984), this would indicate the termination of the main deformation events (excluding the Iisvesi fault zone and related structures) at about the same time.

GEOLOGY OF THE REFERENCE AREAS

Rastinpää-type gneissic tonalites are very common in the SSB near Archean craton and some have been sampled as reference material. The Veteli granodiorite has also been sampled in the BSB and there are additional samples from the Saunakangas gneissic trondhjemite-granodiorite in the SSB (Huhma 1986).

Kangasjärvi gneissic granodiorite

At Kangasjärvi area (Fig. 1) there is a strata-bound massive sphalerite-pyrite deposit hosted by felsic volcanics with calc-alkaline character, underlain by tholeiitic basic volcanics and calc-alkaline intermediate to felsic volcanics (Rasilainen 1991). These are structurally underlain by a gneissic granodiorite (Rehtijärvi 1984). The granodiorite is strongly foliated and quartz ribbons are very abundant. Fine-grained granit-

ic dikes occur commonly and the gneissic granodiorite is locally granitized. Textures are mylonitic and K-feldspar content is variable, while muscovite and chlorite are very common. Modal composition is granodioritic, but original composition was probably tonalitic.

Leväniemi gneiss

The Säviä schist belt in the Pielavesi area hosts a Cu-Zn occurrence within garnet-cordierite-anthophyllite rocks. Gneissic tonalite at Leväniemi has been interpreted as basement to the Säviä schists and named accordingly basement gneiss (Makkonen 1981, Ekdahl 1993). Amphibolitic enclaves are very common. The major minerals are plagioclase, quartz and hornblende, while biotite, opaques, epidote, sphene, apatite and zircon occur as accessory phases (Makkonen 1981). This gneiss is part of the Leväniemi dome and comparable with other dome orthogneisses (Ekdahl 1993). The age of similar orthogneisses at Kirkkosaari is 1925±5 Ma (ibid.).

Kettuperä gneiss

The Pyhäsalmi zinc-copper-pyrite deposit is situated in Ruotanen schist belt, which is bordered to the east by younger quartz diorite (Helovuori 1979, Mäki 1986). Between the volcanic association and the quartz diorite is a strongly deformed granite gneiss (1932±15 Ma), which has been interpreted as basement (Helovuori 1979). Samples from this gneiss have been taken from a drill core. The gneiss is reddish to greyish in colour, heterogenous and locally strongly mylonitic. Textures vary from mobilized augen gneiss with strongly undulating quartz aggregates to blastoporphyritic, with zoned plagioclase grains up to 3 mm sized in fine-grained granoblastic groundmass. These are subidioblastic and strongly altered (sericite+saussurite) with recrystallized albitic rims. Major minerals are plagioclase (30-50 %), quartz (20-45 %), biotite (2-12 %) and hornblende (0-15 %). K-feldspar is also found as a major mineral in strongly mobilized and mylonitic parts of the Kettuperä granite gneiss. Modal composition is tonalitic to granodioritic, but the original composition has probably been mainly tonalitic. The textural evidence is suggestive of a very strongly deformed porphyritic subvolcanic (sill) or even a volcanic origin.

Venetpalo area

Hautala (1968) divided the rocks of the Venetpalo area into five main units: two gneiss domes, schists, mafic dykes, granites and sulphide occurrences. Oligoclase gneisses are the main rock type found in the gneiss domes. Locally occurring K-feldspar in these rocks is not of primary origin and is attributed to the effect of younger granites. The contacts between oligoclase gneisses and the surrounding schists have been studied from two drilled sections: at Kaskela there is a gradual change to a finer-grained variant followed by sillimanite-cordierite gneiss and locally occurring cordierite-anthophyllite rock, while in the eastern part of the Venetpalo dome intercalations of mica gneisses, quartz feldspar gneisses and amphibolites occur (ibid.). The Kaskela sulphide occurrence is hosted by cordierite-anthophyllite rocks. There are two samples from the area, one of which is plagioclase porphyritic epidote-

rich felsic gneiss with a tonalitic modal composition and blastoporphyritic and mylonitic texture. The other sample is gneissic granodiorite with a fine-grained granoblastic groundmass and is slightly mylonitic with mobilized quartz aggregates and 1-3 mm plagioclase grains. According to Hautala (1968), the oligoclase gneiss normally has a tonalitic to trondhjemitic modal composition, but in some parts it is granodioritic. Gneissic granodiorite resembles his oligoclase gneiss.

Veteli granodiorite

The Veteli granodiorite is medium-grained and slightly foliated granodiorite that contains oval-shaped fine-grained plagioclase porphyritic, intermediate enclaves ranging from a few centimetres to some tens of centimetres in length. The texture is granoblastic rather than igneous, but subhedral amphibole, magnetite and apatite needles suggest an igneous precursor. These enclaves are regarded as microgranitoid enclaves (Vernon 1983). The granodiorite has an intrusive contact into the surrounding amphibolites and mica schists and it gradationally changes to a porphyritic variant (P. Pietikäinen 1992, oral communication). Sampled rocks are hypidiomorphic to slightly porphyritic with plagioclase (40-50 %), quartz (30 %) and K-feldspar (10-20 %) as major minerals, the latter mainly filling the interstitial spaces. The most abundant minor constituents are biotite (4-5 %) and its alteration product chlorite. Muscovite and epidote are also common. Opaque, sphene, apatite, carbonate and zircon occurs as accessory minerals.

GEOCHEMISTRY OF THE RAUTALAMPI AREA

Geochemistry of gneissic tonalites, felsic gneisses and granitoids

Rastinpää gneissic tonalite

Representative analyses of Rastinpää gneissic tonalite are presented in Table 1. Strongly altered samples have been excluded. Variations in major and trace elements are expressed on Harker diagrams (Fig. 3). A striking feature is the variation in Y abundances, which effectively divides the data into low-Y and high-Y groups. High-Y samples are found in the satellite body and from the interior of the main Rastinpää body, whereas low-Y samples are concentrated at the margins of the main body. These groups are also distinguishable on the other diagrams. SiO_2 values partly overlap, but the low-Y group tends to have more high SiO_2 values. Although both groups have decreasing trends on Al_2O_3 and MgO diagrams, the high-Y group has lower values. On the TiO_2, FeO, CaO, Ba, Sc and Sr diagrams these two groups form overlapping trends where the high-Y group has higher levels of TiO_2, FeO, and in some cases CaO, Sc and Sr, but lower levels of Ba. Clear differences in abundances are observed for MnO, Ta, Th, and Y (Table 1). The level of Na_2O, P_2O_5, and Rb contents are rather uniform, although the high-Y group has an increasing trend in Na_2O and Hf (not shown) compared to the low-Y group which shows no clear trend. The high-Y group has a distinct increasing trend with K_2O and Zr (not shown) and decreasing trend with V, in contrast to the low-Y group, which has no trend with K_2O and Zr and only a slightly increasing trend in V.

Table 1. Selected major (anhydrous) and trace element data and CIPW norms for Rastinpää gneissic tonalites.

Sample	R40	R57	R62	R37	R42	R61
		Low-Y group			High-Y group	
SiO_2	73.44	73.40	71.65	71.98	73.65	72.66
TiO_2	0.29	0.28	0.33	0.31	0.31	0.29
Al_2O_3	13.76	13.41	14.38	13.98	13.51	13.45
FeO*	3.43	3.65	3.46	3.83	3.64	3.29
MnO	0.09	0.05	0.07	0.09	0.12	0.07
MgO	0.59	1.20	1.26	0.98	0.70	0.76
CaO	2.59	2.80	3.26	3.43	2.78	3.26
Na_2O	4.34	3.97	4.10	4.21	3.82	4.58
K_2O	1.45	1.16	1.40	1.12	1.35	1.52
P_2O_5	0.02	0.08	0.09	0.07	0.12	0.12
Ba	561	480	273	585	345	519
Sc	8.1	10.3	9.3	12.1	10.0	8.9
Sr	163	180	172	183	182	160
V	20	37	18	35	19	28
Y	21.9	14.7	10.5	34.9	39.1	29.7
Zr	199	205	142	141	193	112
Rb	18.4	18.9	33.0	19.6	26.0	26.0
Ta	0.35	0.22	0.30	0.42	0.43	0.37
Th	2.9	1.7	1.1	3.8	2.9	2.6
U	0.46	0.65	0.39	1.41	0.20	0.20
Hf	6.2	3.5	4.9	3.8	5.9	4.8
La	18.7	13.0	8.9	13.5	17.3	16.1
Ce	39.0	25.0	16.5	23.0	34.0	27.0
Nd	21.0	14.9	9.4	14.3	21.0	16.3
Sm	4.4	3.3	1.9	4.4	6.1	5.1
Eu	1.15	0.88	0.89	0.69	1.21	0.64
Tb	0.54	0.37	0.27	0.54	0.71	0.50
Yb	1.7	1.4	1.4	2.9	6.0	2.8
Lu	0.20	0.16	0.16	0.36	0.72	0.35
$^{a}A/CNK$	1.03	1.05	1.02	0.97	1.06	0.89
$(La/Yb)_{CH}$	7.4	6.3	4.3	3.2	1.9	3.9
$^{b}Eu/Eu^{*}$	0.83	0.85	1.44	0.50	0.66	0.42
Q	33.56	35.42	31.13	31.68	36.69	30.63
Or	8.57	6.86	8.27	6.62	7.98	8.98
Ab	36.72	33.59	34.69	35.62	32.32	38.76
An	12.72	13.37	15.59	15.94	13.01	11.65
C	0.39	0.72	0.41	-	1.00	-
Di	-	-	-	0.53	-	3.31
Hy	7.46	9.32	9.08	8.86	8.14	5.89
Il	0.55	0.53	0.63	0.59	0.59	0.55
Ap	0.05	0.19	0.21	0.16	0.28	0.23

Note.- Major elements (wt%) and Ba,Sc,Sr,V,Y and Zr analyzed at the GSF Laboratory by ICP-AES; other trace elements and REE by neutron activation at the Technical Research Centre of Finland. FeO^{*}: total Fe as FeO.
$^{a}A/CNK$ = Molecular ratio of $Al_2O_3/(CaO+Na_2O+K_2O)$.
$^{b}Eu/Eu^{*}$ = Observed Eu value/value obtained by interpolation between Sm and Tb.

These groups are also discriminated by chondrite-normalized REE curves (Fig. 4) where the low-Y group is characterized by higher La/Yb ratios and a weak or positive europium anomaly. The high-Y group has lower La/Yb ratios and negative europium anomalies. Light REE's (LREE) are of the same order of magnitude in both groups and increase with increasing SiO_2 content. These groups also plot separately on the FeO/MgO-SiO_2 (Fig. 3) and MgO-FeO (not shown) diagrams where the high-Y group has a slightly tholeiitic character and the low-Y group is more calc-alkaline. The low-Y group is slightly corundum normative compared to the generally diopside-normative high-Y group (Table 1).

The difference between these two groups is difficult to explain solely by alteration or metamorphic processes although the scatter of MgO, CaO, K_2O, Sr and Ba indicates some degree of element mobility. The higher K_2O and Ba values and lower CaO values compared to SiO_2 in some samples (data not shown) indicate local destruction of plagioclase but the negative Eu anomaly and the higher CaO and Sr contents in high-Y group can not be explained by metasomatic effects. Similarities in the abundances of the mobile components K_2O, Rb and the LREE rules out the possibility of contamination being the reason for the difference between the groups. The same elements are generally enriched in first forming melts, so the high-Y group can not be the residue of the low-Y group or vice versa. It is concluded that the observed difference between these groups is mainly primary in nature.

Variations in the chemical composition of granitoids can be caused by crystal fractionation, contamination/assimilation, unmixing of liquids, magma mixing, restite unmixing or a combination of these. There is also the possibility that different phases within the same body to have different origins. There are great differences in some element abundances and some elements have totally differing trends which can not be explained by magma mixing or restite unmixing, both of which should form linear trends in Harker diagrams.

Hildreth (1981) considers diffusive differentiation in zoned magma chamber to be a important process, especially in high-silica rhyolites. For example, the rhyolitic Bishop Tuff is divided to an early and late phase with a large compositional gap in chemical composition. The mechanism proposed by Hildreth (1981) could partly explain the variation in the Rastinpää gneissic tonalite. The high-Y group resembles a roof zone phase and is enriched in Sm, heavy REE (HREE), Sc, Mn, Y, Ta and Th and depleted in Mg, Al, Ba and Eu. The main differences compared with enrichment factors of Hildreth (1981) are the lack of enrichment of Na, Rb and U and depletion of P, Ca, Ti, V, Fe, Sr and Zr. Also the possible enrichment factors found here are very low compared to those observed by Hildreth (1981). The Bishop Tuff is a K-rich rhyolite and is therefore compositionally different from the low-K Rastinpää tonalite. According to Hildreth (1981) the size of the magma chamber, magma composition and especially the composition and amount of volatiles changes the behaviour of elements in diffusion processes and their enrichment factors. His examples are eruptive rocks, but the same features are found also in some plutonic rocks, even though posteruptive changes in the composition of unerupted magma are likely to be more significant.

Crystal fractionation of plagioclase could explain to some extent the differences in the Eu anomaly and HREE, but at the same time Ca and Sr contents should also decrease. This is however, opposite to the observed trend. The difference in trends and totally separate groups defined by some elements means that crystal fractionation is a very improbable mechanism to account for the differences between these two groups. Normal crystal settling is also generally considered an inadequate mechanism in felsic magmas with high viscosities. On the other hand, according to Sparks et al. (1984) sidewall crystallization with convective frac-

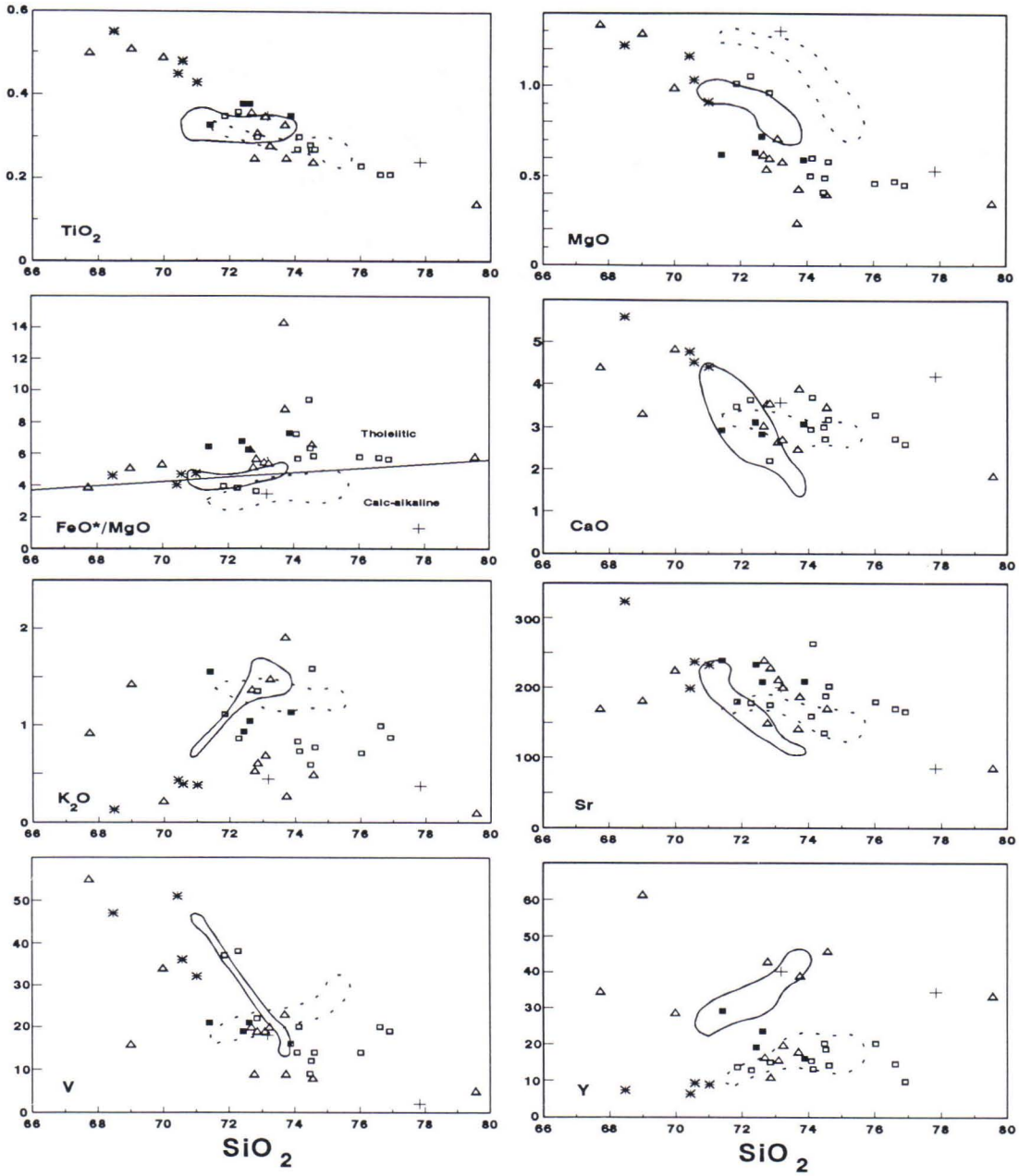

Fig. 3. Harker-type selected major element (%) and trace element (ppm) variation diagrams for 1.93-1.91 Ga Rastinpää gneissic tonalites and related rocks in the Rautalampi area. Oxide totals were recalculated to 100 %. Data from this study. Solid line outlines the Rastinpää high-Y group and dotted line the low-Y group. Filled squares - Toholampi gneissic tonalites; open squares - other gneissic tonalites; open triangles - supracrustal gneisses; crosses - volcanogenic gneisses; asterixes - mobilized gneisses.

tionation (Rice 1981) can produce evolved magmas rapidly.

The Rastinpää high-Y group shows a clear decreasing trend for Al, Ca and Sr and a strong increasing trend for K, Ba, Y and REE's, and somewhat weaker increasing trend for Na. These features could be due to fractional crystallization of plagioclase (not necessarily crystal settling). Magnesium, Sc and to some extent Fe have decreasing trends and reflects the presence of a ferromagnesium phase, but the strong increase in K nevertheless excludes biotite, leaving amphiboles and pyroxenes as possibilities. Vanadium shows very strong depletion at higher SiO_2 values (Fig. 3) and indicates the possibility of a magnetite as a crystallizing phase. The incompatible trace elements Ba, Y and Zr, with bulk distribution coefficients of the order of 0.1-0.3 (approximation from the distribution coefficients for plagioclase, amphibole and pyroxenes; Hanski 1983) have been used to calculate the extent of possible fractional crystallization. Barium (400-200 ppm) gives 54-63 %, Y (22-46 ppm) gives 56-65 % and Zr (125-250 ppm) gives 54-63 % fractional crystallization. Considering K as an incompatible trace element with a bulk distribution coefficient of 0.3, we get 66 % fractional crystallisation (0.8-1.7 % K). Potassium and Ba are considered as mobile elements but their

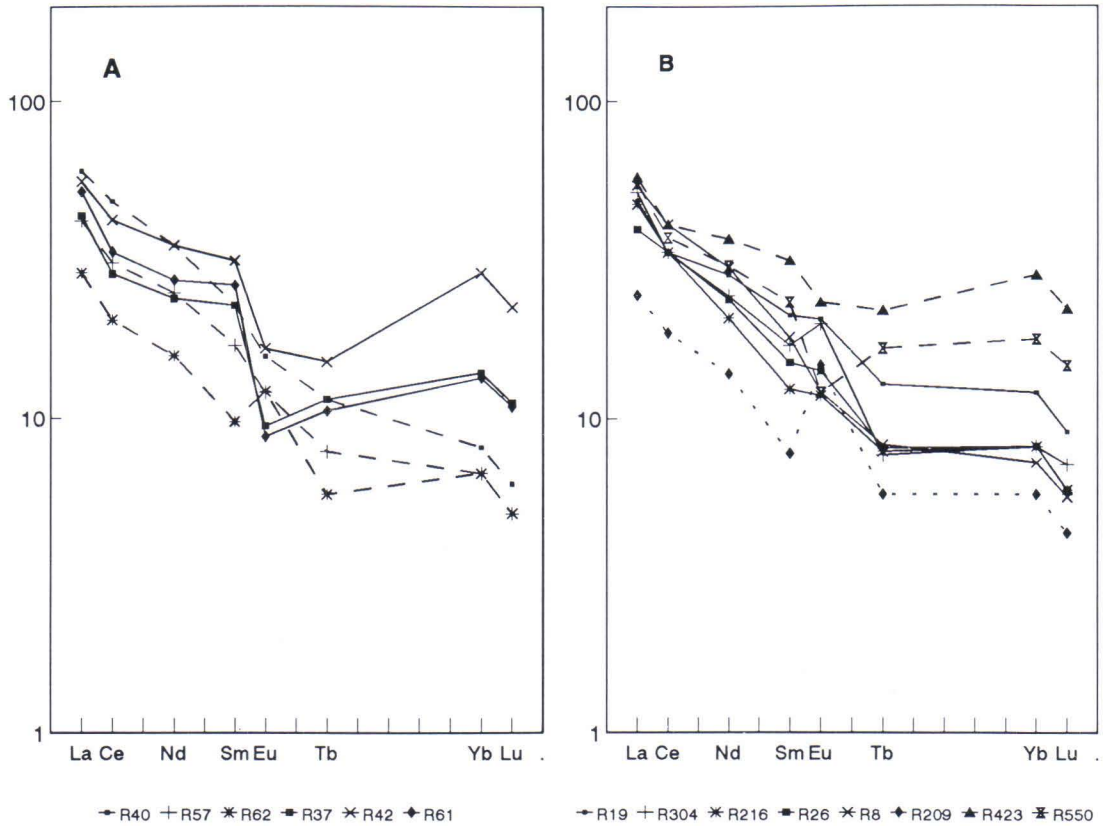

Fig. 4. Chondrite normalized rare-earth element patterns for 1.93-1.91 Ga Rastinpää gneissic tonalites (A) and related rocks in the Rautalampi area (B). Normalizing values are recommended chondrite values from Boynton (1984). Sample numbers refer to samples in Tables 1 and 2. A: solid line - high-Y group; dashed line - low-Y group. B: solid line - Toholampi type gneissic tonalites; dashed line - supracrustal gneisses; dotted line - mobilized gneiss.

similar behaviour in these calculations with respect to the normally immobile elements Y and Zr indicates that K and Ba have acted as immobile elements in these samples. Taking 60 % percent as the extent of fractional crystallisation, we obtain bulk distribution coefficients of 2.5 for V (48-12 ppm) and 1.5 for Sc (14-9 ppm). The behaviour of V is critical to the oxygen fugacity, but the twofold increase in the bulk distribution coefficient compared to Sc indicates the presence of magnetite or at least amphibole as a possible crystallizing phase. The normal distribution coefficients for V are typically of the same order of magnitude or lower for pyroxenes and amphiboles compared to Sc. This favours the occurrence of magnetite. There is in fact a difference in magnetite content between rocks with high V contents and abundant accessory magnetite and rocks with low V abundances and lacking magnetite but the metamorphic nature of samples should nevertheless be kept in mind.

Plagioclase fractionation often produces a very distinct Eu anomaly because this element has a higher distribution coefficient for plagioclase than do other REE's. The data in Figure 4 do not show any evidence for this. The occurrence of magnetite would indicate higher oxygen fugacities and the change in Eu^{2+}/Eu^{3+} redox ratios could explain the behaviour of Eu as a mainly trivalent cation.

The low-Y group shows a decreasing trend for Ti, Al and Mg and increasing trends in Ba and V contents. The other elements do not form any noticeable trends and the variation cannot be explained by fractional crystallisation. One interpretation is that the low-Y group rocks reflect side-wall crystal accumulation and trapped evolved magma where plagioclase and some ferromagnesium mineral have acted as cumulate phases. The positive Eu anomaly and low LREE contents of the sample from the margin (Fig. 4, sample 62 in Table 1) as well as the high variation in LREE compared to HREE would also favour a mixture of cumulate and evolved magma.

The Rastinpää gneissic tonalite is a strongly deformed and partly migmatized rock and the sampling density is low due to poor outcrop density. These features and the uncertainty of the actual distribution coefficients of minerals in magmas of these compositions hinder the interpretation of the data. Although there are uncertainties, the low-Y group is interpreted as resulting from side-wall accumulation (plagioclase+ferromagnesium mineral) with trapped evolved magma (Sparks et al. 1984). Most of the migrated interstitial melt has either erupted or become concentrated in other parts of the magma chamber. There are indications that these gneissic tonalites have been sill-like bodies, so that a flow differentiation processes and enrichment of evolved melt in the middle of sill would also have been possible. The variation in chemical composition of the high-Y group is due to fractional crystallization of plagioclase, the presence of ferromagnesium mineral, and to some extent also magnetite.

Toholampi gneissic tonalite

The sill-like Toholampi gneissic tonalite is situated near the pyroxene granitoid complex and it contains orthopyroxene, indicating that it has experienced granulite facies conditions. The variation in K_2O, Na_2O and Ba values is due to orthopyroxene formation and local destruction of plagioclase. Compositionally it resembles the Rastinpää gneissic tonalites but has somewhat higher TiO_2 and Sr values and lower MgO, Ta and Th values (Table 2). Yttrium (Fig. 3) and zircon contents (not shown) show a decreasing trend from the high-Y to low-Y group. The high-Y sample is from the centre and it has only a weak positive Eu-anomaly (Fig. 4). The sample having a strong positive Eu-anomaly and low Y content is from the margin of the sill. These features are interpreted as side-wall crystallization combined with flow differentiation.

Other Rastinpää-type gneissic tonalites

Variation exists in the chemical composition of these gneissic tonalites especially in trace element composition. They resemble Rastinpää and Toholampi types, but some have significantly lower K_2O, Rb, Ta and Th values (Table 2 and Fig. 3). According to Y and REE contents (Figs. 3 and 4) they belong to low Y-group. Some of them have a significant positive Eu-anomaly.

Table 2. Selected major (anhydrous) and trace element data for gneissic tonalites and felsic gneisses from the Rautalampi area.

Sample	R19	R304	R216	R236	R26	R28	R30
Type[a]	a	a	b	b	c	c	c
SiO_2	71.40	73.87	71.85	72.27	74.08	74.13	74.51
TiO_2	0.33	0.35	0.35	0.36	0.27	0.30	0.27
Al_2O_3	14.48	13.02	14.10	13.60	13.32	13.34	13.19
FeO*	4.01	4.31	3.97	4.02	3.63	3.41	3.12
MnO	0.10	0.10	0.08	0.09	0.05	0.08	0.10
MgO	0.62	0.59	1.01	1.05	0.50	0.60	0.49
CaO	2.93	3.08	3.48	3.64	2.95	3.70	2.72
Na_2O	4.49	3.46	3.95	4.02	4.35	3.69	4.00
K_2O	1.55	1.13	1.11	0.86	0.83	0.73	1.58
P_2O_5	0.09	0.09	0.10	0.09	0.02	0.02	0.02
Ba	962	312	290	423	310	353	631
Sc	12.0	6.3	12.1	10.3	7.1	8.8	3.9
Sr	239	209	180	178	159	263	188
V	21	16	37	38	14	20	12
Y	29.1	16.1	13.7	12.9	15.4	13.2	18.6
Zr	327	222	204	190	172	135	148
Rb	17.7	6.1	24.0	6.5	17.8	4.3	12.9
Ta	0.18	0.15	0.35	0.23	0.29	0.09	0.16
Th	0.2	0.5	1.5	0.3	1.7	0.6	0.2
U	0.13	-	0.25	0.26	0.39	0.24	0.34
Hf	6.6	5.5	4.6	4.2	4.7	3.8	4.8
La	15.1	16.0	14.7	12.0	12.2	13.4	12.3
Ce	27.0	27.0	27.0	19.3	27.0	26.0	22.0
Nd	17.0	14.6	12.4	9.4	14.2	15.4	12.7
Sm	4.1	3.3	2.4	2.2	2.9	2.3	2.7
Eu	1.50	1.45	0.86	0.85	1.03	1.08	0.87
Tb	0.60	0.36	0.37	0.34	0.38	0.32	0.33
Yb	2.5	1.7	1.7	1.8	1.7	1.7	2.4
Lu	0.29	0.23	0.19	0.23	0.19	0.21	0.32
[b]A/CNK	1.01	1.04	1.00	0.98	0.99	0.98	0.99
$(La/Yb)_{CH}$	4.1	6.3	5.8	4.5	4.8	5.3	3.5
[c]Eu/Eu*	1.12	1.41	1.08	1.17	1.12	1.43	1.02

Note.- Major elements (wt%) and Ba,Sc,Sr,V,Y and Zr analyzed at the GSF Laboratory by ICP-AES; other trace elements and REE by neutron activation at the Technical Research Centre of Finland. FeO*: total Fe as FeO.
[a]a- Toholampi gneissic tonalite; b- Toholampi-type gneissic tonalite; c- gneissic tonalite; d- felsic gneiss of propable igneous origin; e- mobilized gneiss; f- supracrustal gneiss; g- volcanogenic felsic gneiss.
[b]A/CNK = Molecular ratio of $Al_2O_3/(CaO+Na_2O+K_2O)$.
[c]Eu/Eu* = Observed Eu value/value obtained by interpolation between Sm and Tb.

Supracrustal gneisses

Supracrustal gneisses are either volcanic (tuffaceous) and/or sedimentary in origin. They too have been subdivided into high-Y and low-Y groups (Fig. 3 and Table 2.). In the low-Y group, Toholampi gneisses have compositions almost identical with those of gneissic tonalites, except for their lower MgO. Others are characterized by lower abundances of K_2O, MgO and Rb and slightly higher CaO and Sr contents. The high-Y group contains rocks with both high and low K_2O values. Volcanogenic gneisses (felsic volcanics) are also low in K_2O,

Table 2 (continued).

Sample Type[a]	R238 d	R877 d	R8 d	R209 e	R423 f	R550 g	R29 g
SiO_2	76.02	76.61	73.69	70.57	69.00	77.83	72.76
TiO_2	0.23	0.21	0.33	0.48	0.51	0.24	0.25
Al_2O_3	12.85	12.47	13.76	14.33	13.84	12.11	14.16
FeO*	2.67	2.71	3.44	4.84	6.61	0.69	2.79
MnO	0.05	0.07	0.07	0.08	0.13	0.03	0.08
MgO	0.46	0.47	0.24	1.03	1.29	0.53	0.54
CaO	3.29	2.73	2.49	4.52	3.32	4.19	3.56
Na_2O	3.67	3.68	3.97	3.63	3.66	3.95	5.28
K_2O	0.71	0.99	1.91	0.39	1.43	0.37	0.53
P_2O_5	0.05	0.06	0.10	0.13	0.21	0.06	0.05
Ba	320	179	697	252	242	123	163
Sc	7.9	7.6	9.1	14.0	18.6	7.7	9.7
Sr	180	170	142	237	182	85	150
V	14	20	23	36	16	2	9
Y	20.2	14.6	18.0	9.3	61.5	34.4	43.0
Zr	187	189	211	229	193	165	146
Rb	1.4	7.9	25.0	2.0	24.4	2.0	1.5
Ta	0.36	<0.02	0.18	0.12	0.46	0.69	0.62
Th	2.2	2.3	1.4	0.1	2.2	2.7	3.0
U	0.56	0.35	0.30	0.14	-	-	0.95
Hf	3.7	4.8	4.9	4.2	5.0	5.1	5.3
La	14.0	18.2	16.9	7.6	17.9	17.1	20.0
Ce	32.0	31.2	33.0	15.0	33.0	30.0	36.0
Nd	19.3	16.4	17.9	8.3	22.0	18.1	19.8
Sm	3.4	3.7	3.5	1.5	6.1	4.5	6.2
Eu	0.90	0.80	0.88	1.07	1.70	0.88	1.05
Tb	0.51	0.40	0.39	0.27	1.03	0.78	0.82
Yb	2.6	1.3	1.5	1.2	5.9	3.7	4.4
Lu	0.31	0.19	0.18	0.14	0.71	0.47	0.50
[b]A/CNK	1.00	1.03	1.05	0.98	1.02	0.83	0.90
$(La/Yb)_{CH}$	3.6	9.4	7.6	4.3	2.0	3.1	3.1
[c]Eu/Eu*	0.81	0.71	0.81	2.08	0.40	0.57	0.53

Ba and Rb and they are characterized by low La/Yb-ratios and negative Eu-anomalies as are other high-Y group gneisses as well (Fig. 4). Sorting, winnowing, welding, postemplacement crystallization and synvolcanic alteration all can affect the geochemical composition of volcanic rocks. Therefore, it is not clear if the low contents of K_2O, Ba and Rb are primary features in the rocks interpreted as volcanogenic in origin. Other supracrustal gneisses with low K_2O and Rb are rich in plagioclase and they could be interpreted to have had an arkositic protolith, with loss of e.g. clays (K,Rb) during sorting and deposition. The rocks devoid of K_2O, Rb and Th could have also lost these components to either a fluid phase or melt during granulite facies metamorphism (see below).

Mobilized gneisses

Mobilized gneisses have lower contents of SiO_2, K_2O, Rb, Th, U and Y than most other gneisses and are also characterized by very low REE contents and a very strong positive Eu-anomaly (Figs. 3-4 and Table 2.). Mineralogically they differ from other felsic gneisses in having low contents of biotite and abundant orthopyroxene, which favours a small degree of dehydration-melting induced by deformation to produce a limited amount of granitic melt and residual mobilized gneisses with abundant plagioclase and orthopyroxene. Dehydration without melting is more difficult to study and so it is possible that some samples low in K_2O, Rb, Th, and U have lost a portion of these elements to the fluid phase.

Särkilampi gneissic tonalite and related felsic gneisses

The Särkilampi gneissic tonalite and geochemically related gneisses differ from the above described Rastinpää-type gneissic tonalites in having lower SiO_2, TiO_2, FeO, Sc, Y, Zr, Ta, Th and U and higher Al_2O_3 and Sr values (Table 3). Na_2O, K_2O, P_2O_5, Ba and Rb abundances fluctuate but are rather similar. MgO, CaO and V contents are higher, but they follow a typical "differentiation trend" on Harker-diagrams. These rocks have major element composition comparable, except for their high CaO values, to the synkinematic tonalites of the schist belts (Fig. 5). Ba, Sc and Sr contents are also comparable, but the Särkilampi gneissic tonalite and related gneisses have slightly lower Rb and anomalously low contents of La, Zr, Ta, Th and U compared to the synkinematic tonalites (data not shown). The REE abundances are also low and the rocks are characterized by a moderate La/Yb-ratio (Fig. 6).

Pyroxene granitoids

Pyroxene quartz diorites and tonalites are lower in SiO_2 than the median contents of typical tonalites in schist belts and on major element diagrams they form a "differentiation trend" (Fig. 5 and Table 3). The exceptions are higher TiO_2 and slightly lower MgO contents in pyroxene quartz diorites and tonalites. The Sr, Ba and La contents are high (Fig. 5) and Th contents are very low (Table 3). They have REE distribution (Fig. 6) with moderately negative and positive Eu-anomalies (Table 3). The pyroxene quartz leuco monzodiorite has anomalously high contents of Ba, Zr, Th, Y, U and LREE (Table 3). Also Al_2O_3, CaO, Na_2O and Sr contents are high for such a K_2O rich rock.

Other granitoids

The Pukkiharju gneissic tonalite (sample R547 in Table 3) has high TiO_2, FeO and Sc and low Al_2O_3 and Sr contents, and thus resembles the Rastinpää-type gneissic tonalites except for its lower SiO_2 contents. The other quartz diorites and tonalites can be divided into two groups, where one (eg. samples R259 and R263), apart higher CaO contents, follows the typical synkinematic major element "differentiation trends". The other group (eg. samples

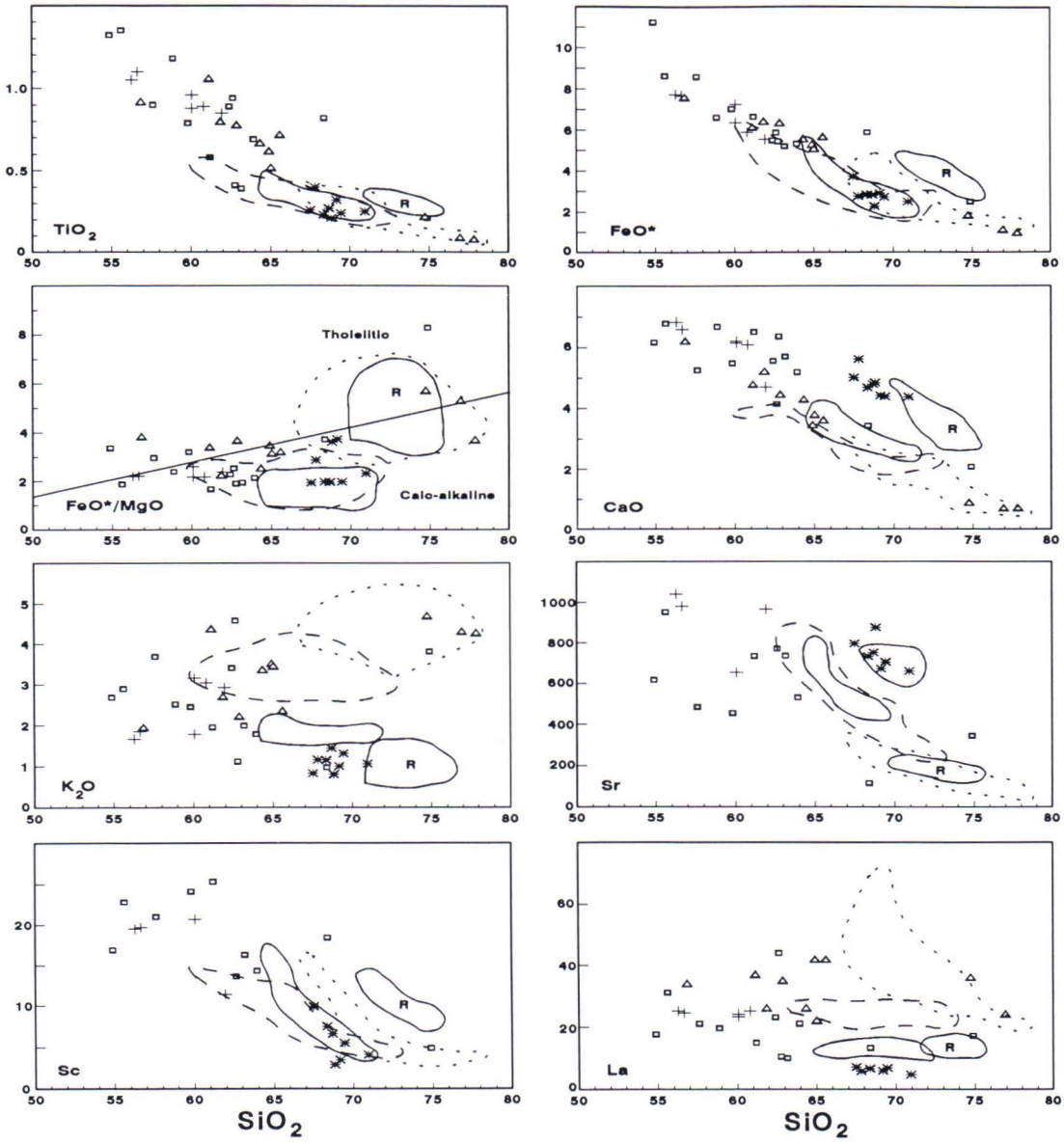

Fig. 5. Harker-type selected major element (%) and trace element (ppm) variation diagrams for 1.91-1.88 Ga granitoids, gneissic tonalites and felsic gneisses from Rautalampi area and the area to the NW from Rautalampi. Oxide totals were recalculated to 100 %. Data from this study except for syntectonic granitoids (Nurmi 1984) and for granitoids from NW of Rautalampi (Pääjärvi, unpublished data). Solid line with R - main field of Rastinpää-type rocks; solid line without R - main field of the median contents of syntectonic tonalites; dashed line - main field of the median contents of syntectonic granodiorites; dotted line - main field of the median contents of syntectonic granites; crosses - pyroxene quartz diorite/tonalite (ca. 1.890-1.885 Ga) ; open squares - tonalites and granodiorites in the Rautalampi area (1.89-1.88 Ga ?); asterixes - Särkilampi gneissic tonalites and related gneisses (1.91-1.88 Ga ?); open triangles - granitoids from NW of Rautalampi (1.89-1.88 Ga).

Table 3. Selected major (anhydrous) and trace element data for granitoids, gneissic tonalites and felsic gneisses from the Rautalampi area.

Sample	R5	R872	R874	R3	R260	R261	R10
Type[a]	a	a	a	b	c	d	e
SiO_2	56.28	60.07	56.65	63.34	62.61	55.67	63.91
TiO_2	1.05	0.88	1.10	0.38	0.94	1.35	0.69
Al_2O_3	18.14	17.24	18.02	18.35	15.66	16.40	16.42
FeO*	7.75	7.26	7.69	5.74	5.90	8.64	5.34
MnO	0.13	0.12	0.12	0.09	0.14	0.08	0.10
MgO	3.53	2.79	3.46	0.24	2.33	4.62	2.50
CaO	6.83	6.19	6.59	3.40	4.12	6.78	5.18
Na_2O	4.25	3.37	4.18	5.41	3.31	2.90	3.85
K_2O	1.66	1.78	1.85	2.94	4.58	2.89	1.78
P_2O_5	0.38	0.30	0.34	0.11	0.41	0.67	0.23
Ba	746	1055	1077	3080	893	878	651
Sc	19.5	20.7	19.7	28.7	13.6	22.8	14.3
Sr	1040	652	979	457	770	951	529
V	161	136	157	5	116	212	105
Y	17.0	26.1	22.6	40.4	23.2	21.4	14.2
Zr	200	170	222	849	395	179	183
Rb	43.0	57.4	53.7	32.0	122	117	54
Ta	0.40	0.55	0.43	0.43	0.78	0.60	0.29
Th	0.2	0.3	0.2	17.6	4.4	0.8	1.8
U	0.15	0.47	0.20	1.70	1.24	1.08	0.14
Hf	3.6	4.8	4.3	13.7	8.1	4.3	4.0
La	25.0	23.2	24.4	202	44.0	31.0	21.0
Ce	51.0	42.9	44.1	291	88.0	59.0	38.0
Nd	28.0	24.6	27.5	123	51.0	37.0	18.6
Sm	4.8	6.6	5.9	17.1	9.2	7.9	3.3
Eu	1.69	1.38	1.35	2.80	1.65	1.92	1.15
Tb	0.53	0.71	0.66	1.34	0.76	0.65	0.40
Yb	1.5	2.0	1.7	2.9	2.2	1.8	1.2
Lu	0.14	0.27	0.23	0.29	0.24	0.22	0.12
[b]A/CNK	0.85	0.92	0.86	1.00	0.88	0.81	0.93
$(La/Yb)_{CH}$	11.2	7.8	9.7	50.0	13.5	11.6	11.8
[c]Eu/Eu*	1.14	0.68	0.75	0.56	0.62	0.83	1.12

Note.- Major elements (wt%) and Ba,Sc,Sr,V,Y and Zr analyzed at the GSF Laboratory by ICP-AES; other trace elements and REE by neutron activation at the Technical Research Centre of Finland. FeO*: total Fe as FeO.
[a]a- Pyroxene quartz diorite/tonalite; b- Pyroxene quartz leuco monzodiorite; c- gneissic granodiorite; d- gneissic quartz diorite; e- gneissic tonalite; f- Särkilampi gneissic tonalite; g- felsic gneiss of probable volcanic origin.
[b]A/CNK = Molecular ratio of $Al_2O_3/(CaO+Na_2O+K_2O)$.
[c]Eu/Eu* = Observed Eu value/value obtained by interpolation between Sm and Tb.

R260 and R261), which also includes granodiorites is heterogenous and characterized by high K_2O, TiO_2 and La contents and in part, a tholeiitic character (Fig. 5). Granitoid samples collected from the N or NW of the Rautalampi area belong to the CFGC and for these, only major element and REE data are available. The granites resemble the reference granites, but the quartz diorites and granodiorites differ from typical synkinematic granitoids in having higher TiO_2 contents and higher FeO/MgO-ratio (Fig 5). The REE distribution is presented in Figure 6.

Table 3 (continued).

Sample Type[a]	R259 e	R263 e	R547 e	R876 f	R878 f	R879 f	R1 g
SiO_2	61.18	63.17	68.38	70.95	69.15	67.50	69.45
TiO_2	0.58	0.39	0.82	0.25	0.32	0.26	0.24
Al_2O_3	15.49	17.02	14.13	15.88	16.98	16.85	16.61
FeO*	6.67	5.23	5.89	2.50	2.92	3.74	2.74
MnO	0.10	0.10	0.16	0.05	0.04	0.08	0.07
MgO	4.00	2.70	1.58	1.07	0.78	1.92	1.38
CaO	6.50	5.69	3.41	4.36	4.40	5.00	4.37
Na_2O	3.26	3.59	4.40	3.78	4.28	3.70	3.74
K_2O	1.95	1.99	0.97	1.06	1.00	0.83	1.31
P_2O_5	0.27	0.12	0.26	0.10	0.13	0.12	0.09
Ba	684	532	389	383	437	472	349
Sc	25.3	16.3	18.4	4.1	3.5	9.9	5.5
Sr	734	735	115	660	671	795	705
V	192	120	38	38	36	78	53
Y	15.9	10.7	24.2	6.0	7.4	7.7	6.3
Zr	92	84	149	61	96	61	60
Rb	22.0	29.0	17.9	45.1	21.0	14.2	28.0
Ta	0.21	0.24	0.44	0.14	0.13	0.07	0.23
Th	0.6	1.6	1.1	0.5	0.2	0.1	1.1
U	0.50	0.48	-	0.27	0.13	0.14	0.30
Hf	2.3	2.2	4.0	1.9	2.6	1.5	1.6
La	14.9	10.0	13.3	4.7	5.9	7.1	6.8
Ce	23.0	16.5	24.0	7.8	10.1	10.8	11.9
Nd	14.0	9.5	14.1	4.0	5.6	5.7	6.1
Sm	3.4	2.5	3.3	1.0	1.4	1.3	1.1
Eu	0.71	0.54	1.12	0.35	0.43	0.50	0.38
Tb	0.33	0.23	0.67	0.13	0.16	0.16	0.13
Yb	1.4	1.0	2.5	0.4	0.5	0.5	0.6
Lu	0.22	0.15	0.35	0.05	0.07	0.07	0.07
[b]A/CNK	0.80	0.92	0.98	1.04	1.05	1.05	1.07
$(La/Yb)_{CH}$	14.1	11.1	3.6	7.9	8.0	9.6	7.6
[c]Eu/Eu*	0.69	0.73	0.95	1.11	1.00	1.17	1.12

Geochemistry of mafic and intermediate volcanics, dykes and xenoliths

Pukkiharju volcanics

Compositionally the Pukkiharju mafic and intermediate volcanics are tholeiitic low-K basalts to basaltic andesites (Table 4, Figs. 7-8). Potassium mobility is seen in the irregular behaviour of this element in Figure 8, where it does not follow the general differentiation trend of the more immobile elements in Figure 9 (e.g. TiO_2 and Zr). Although scattered, the generally low K values in mafic samples indicate a low-K character. An overall tholeiitic character is postulated for these rocks (Fig. 8). The most mafic volcanics show rather primitive compositions, indicating only a small amount (10%) of olivine crystallization from mantle derived melts for the least evolved samples (Lahtinen 1988). The deduced crystallization sequence for the Pukkiharju volcanics is olivine(+Cr-spinel)-clinopyroxene-plagioclase, which differs from the normal olivine-plagioclase crystallization order found in Mid-

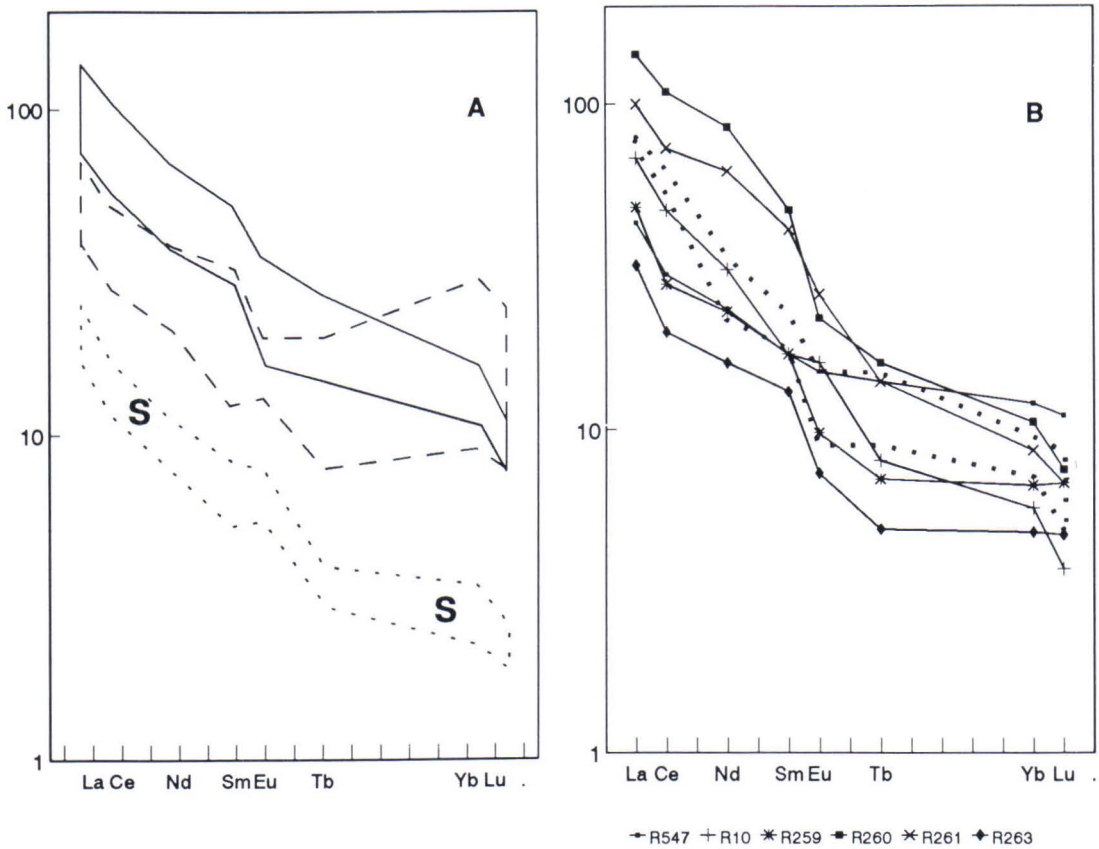

Fig. 6. Chondrite normalized rare-earth element patterns for 1.91-1.88 Ga granitoids, gneissic tonalites and felsic gneisses from Rautalampi and the area to the NW. A: solid line (Pääjärvi, unpublished data) - quartz diorites and granodiorites from NW of Rautalampi (8 samples); dashed line - main field of 1.93-1.91 Ga gneissic tonalites and related rocks; dotted line with S - Särkilampi-type gneissic tonalites and related rocks (5 samples). B: Sample numbers refer to samples in Table 3; dotted line - pyroxene quartz diorites and tonalites (5 samples).

Ocean Ridge Basalts (MORB) (e.g. Perfit et al. 1980). The crystallization of clinopyroxene before plagioclase indicates a relatively higher water content in the magma (e.g. Sakuyama 1983), due to the lower liquidus temperatures of plagioclase in water-rich magmas (Yoder 1969). The REE distributions and MORB-normalized trace element patterns (Figs. 10-11) are compatible with a slightly LREE-enriched island arc tholeiite (IAT) origin (Pearce 1982,1983). The Rb, Th and Ta values are below or near the detection limit and so the patterns concerning these elements in Figure11 are only indicative, but the behaviour of analytically more reliable Ba and Ce is nevertheless similar, suggesting that the pattern is, in general sense, correct. The low level of Zr-Yb in Figure 11 favours a rather high degree (30%) of mantle melting (Pearce 1982). All these features indicate the occurrence of a depleted mantle (DM) source for the melt, with an additional subduction component and to a rather immature, oceanic island arc origin with rather thin crust. Although the occurrence of thicker mature crust can not be excluded (especially in an extensional environment), the occurrence of DM-type mantle is inevitable.

Toholampi volcanics

The Toholampi volcanics are geochemically similar to the Pukkiharju volcanics (Figs. 8-11), but some differences are apparent. The Toholampi volcanics seem to have slightly lower SiO_2 values (Fig. 8) and higher CaO values (Fig. 9), which can be partly attributed to the occurrence of more calc-silicates (primarily carbonates) in the Toholampi samples. Three samples are deviant, with higher TiO_2 and Y values, indicating a difference either in parent magma or in crystallization history. Sample R595 is a rather primitive basalt with 118 ppm Ni and 10.5% MgO (Table 4). Although, the REE distribution curve does show some scatter (Fig. 10), it is rather similar to Pukkiharju volcanics, albeit at a lower level. In particular, the negative Ce-anomaly is considered to be of analytical origin. The MORB-normalized trace element pattern also shows a pattern similar to that of the Pukkiharju volcanics, but the K and Th abundances indicate higher levels of these incompatible elements in the parent magma. Sample R233 has small positive Eu (Fig. 10) and P and Ti (Fig. 11) anomalies, but otherwise is comparable with Pukkiharju samples. Sample R240 differs from the Pukkiharju and other Toholampi samples due to its medium-K nature (Fig. 8) and because of its higher Sr and lower V values (Fig. 9). The REE distribution curve show a higher La/Yb ratio (Fig. 10 and Table 4) and the MORB-normalized trace element pattern shows a distinctly "humped" form, with negative Ta, Zr, Hf and positive Sm anomalies (Fig. 11).

Although there are some slight geochemical differences between Pukkiharju and main part of Toholampi volcanics, they are in a general sense similar to one another and therefore inferentially comparable in origin with the Pukkiharju volcanics. Sample R240 clearly differs geochemically and although a tholeiitic in character according to Figure 9, it has a trace element pattern more like that found in oceanic calc-alkaline basalts (Pearce 1983). The mantle component has been of DM type as with the Pukkiharju volcanics, but the proportion of subduction component has probably been higher. A noticeable feature is that this sample is from an association occurring between pyroxene granitoid complex and the Toholampi gneissic tonalite and is thus from a stratigraphically different position to the other Toholampi volcanics.

Xenoliths in gneissic tonalites

One sample was collected from the Toholampi gneissic tonalite and four samples from the Rastinpää gneissic tonalite. One corresponds to basaltic andesite, three are basaltic andesites near the boundary with basaltic trachyandesite and one is trachyandesitic in composition according to Figure 7. They are medium-K rocks with both tholeiitic and calc-alkaline affinities (Fig. 8). They are geochemically scattered which is, at least partly, attributed to heterogenous origins and possibly also due to interaction with the tonalite host. They differ from the Pukkiharju and Toholampi volcanics in their higher SiO_2, K_2O, P_2O_5, Sr and generally lower TiO_2 and V (Figs. 8-9), with the latter in particularly showing a decreasing trend, which possibly reflects the crystallization of magnetite during their crystallization history. Samples R586 and R592 show higher La/Yb ratios (Fig. 10 and Table 4) and although, scattered, a more convex "humped" trace element pattern (Fig. 11) than the Pukkiharju and Toholampi volcanics. These xenoliths are more comparable with sample R240 than with the normal volcanics of the Rautalampi district.

Dykes in gneissic tonalites

There is little textural evidence of comingling and thus it is unclear whether or not these dykes are of similar age or distinctly younger than their hosts. The occurrence of a tonalite inclusion in one Rastinpää dyke indicates that the tonalite in this case was either totally or

Table 4. Selected major (anhydrous) and trace element data for amphibolites (volcanics), mafic dykes and mafic inclusions in gneissic tonalites from the Rautalampi area.

Sample	R204	R987	R405	R595	R233	R240	R586
Type[a]	a	a	a	a	a	a	b
Location[b]	P	P	P	T	T	T	R
SiO_2	51.19	51.59	55.16	49.84	52.62	48.23	59.46
TiO_2	0.66	0.72	1.11	0.53	1.56	0.62	0.37
Al_2O_3	16.31	16.23	14.37	11.11	15.27	18.29	17.81
FeO*	9.95	10.44	13.52	10.23	13.16	10.57	5.84
MnO	0.16	0.19	0.27	0.21	0.21	0.17	0.16
MgO	8.64	7.00	4.77	10.54	4.37	7.99	2.58
CaO	10.87	11.16	6.36	15.50	8.81	11.83	7.17
Na_2O	1.84	2.20	4.02	1.55	3.50	1.65	5.33
K_2O	0.28	0.40	0.30	0.44	0.27	0.52	1.13
P_2O_5	0.10	0.07	0.12	0.05	0.23	0.13	0.15
Ba	52	123	56	111	70	148	461
Co	45	48	30	35	44	50	16
Cr	330	110	<10	335	-	-	<10
Cu	66	79	24	8	50	58	10
Ni	78	48	11	118	22	55	15
Sc	45.0	41.3	43.2	51.2	41.4	32.0	16.2
Sr	199	203	96	149	227	641	544
V	272	283	358	265	396	202	108
Y	14.0	14.5	26.8	9.2	21.9	13.8	14.3
Zn	89	96	134	82	135	88	90
Zr	40	40	79	<30	65	<30	60
Rb	<6	<6	<6	<8	<8	<7	13
Ta	0.15	0.10	0.31	<0.10	0.17	0.12	0.21
Th	0.3	0.4	0.3	0.5	<0.2	0.5	0.8
Hf	1.1	1.1	2.3	0.6	1.4	0.8	1.5
La	4.3	4.4	6.8	2.7	5.6	6.1	6.1
Ce	9.2	9.0	15.9	4.7	12.3	11.8	12.0
Sm	1.8	2.0	3.0	1.3	2.9	2.2	2.2
Eu	0.56	0.70	1.00	0.52	1.17	0.84	0.67
Tb	0.33	0.35	0.62	0.28	0.61	0.35	0.31
Yb	1.33	1.59	2.8	0.93	2.3	1.10	1.46
Lu	0.20	0.22	0.35	0.09	0.23	0.14	0.20
$(La/Yb)_{CH}$	2.2	1.9	1.5	2.0	1.6	3.9	2.8

Note.- Major elements (wt%) and Ba,Co,Cr,Cu,Ni,Sc,Sr,V,Y,Zn and Zr analyzed at the GSF Laboratory by ICP-AES; other trace elements and REE by neutron activation at the Technical Research Centre of Finland. FeO*: total Fe as FeO.

[a]a- amphibolite/metavolcanic; b- mafic inclusion in gneissic tonalite; c- mafic dyke in gneissic tonalite; d- dykes having WPB-affinity; e- younger mafic dyke.

[b]P- Pukkiharju; R- Rastinpää; T- Toholampi; K- Konnekoski.

almost totally crystallized before the intrusion of the mafic magma. If the tonalite hosts were only partly crystalline, there could have been interaction of mafic magma with evolved magma or late stage magmatic fluids. This would have changed the composition of the mafic magma and led in particular to an increase in the level of incompatible elements such as K, Rb, Ba and Th (Holden et al. 1991). There are five dyke samples from the Rastinpää

Table 4 (continued).

Sample	R592	R576	R589	R593	R548	R590	R218
Type[a]	b	c	c	c	d	d	e
Location[b]	R	K	R	R	P	R	T
SiO_2	53.08	43.09	48.15	48.79	56.26	52.65	52.45
TiO_2	0.61	1.78	0.84	0.57	2.00	2.04	0.74
Al_2O_3	18.51	18.45	10.83	12.21	15.88	15.53	19.17
FeO*	9.67	15.14	10.66	10.01	9.34	10.12	7.44
MnO	0.20	0.26	0.21	0.20	0.15	0.15	0.18
MgO	5.59	5.79	15.33	11.97	3.83	4.95	5.91
CaO	7.13	11.32	10.94	13.71	6.40	7.97	10.08
Na_2O	4.34	2.90	1.09	1.93	2.91	3.19	3.41
K_2O	0.63	0.67	1.78	0.49	2.45	2.38	0.41
P_2O_5	0.24	0.60	0.17	0.12	0.78	1.02	0.21
Ba	62	202	182	55	898	707	163
Co	27	33	63	48	32	30	33
Cr	55	<10	1360	773	52	120	-
Cu	25	37	26	39	72	37	65
Ni	20	<10	436	152	27	62	18
Sc	30.1	37.8	39.0	51.3	20.2	24.0	35.7
Sr	343	672	25	184	735	631	476
V	238	329	225	245	149	165	250
Y	16.1	35.9	14.7	14.7	24.9	31.5	14.7
Zn	113	160	171	98	145	167	102
Zr	<30	67	45	30	220	200	48
Rb	<6	-	13	<10	71	62	<6
Ta	0.14	-	0.22	<0.11	1.1	1.25	0.08
Th	1.1	-	<0.4	<0.4	4.5	4.0	0.2
Hf	1.1	-	1.6	0.7	5.7	5.8	0.8
La	7.1	-	5.8	3.2	41.0	50.0	7.0
Ce	14.7	-	11.9	6.8	83.0	95.0	15.3
Sm	2.4	-	2.0	1.8	8.7	12.0	2.2
Eu	0.81	-	0.68	0.57	2.02	2.60	0.87
Tb	0.39	-	0.30	0.37	0.83	0.98	0.39
Yb	1.7	-	1.52	1.3	1.7	2.4	1.5
Lu	0.21	-	0.20	0.14	0.23	0.26	0.16
$(La/Yb)_{CH}$	2.9	-	2.6	1.7	16.3	14.1	3.1

gneissic tonalites that vary between medium-K basalt to high-K (nearly very high-K) trachybasalt (Figs. 7-8). Two samples are calc-alkaline (samples R589 and R593) and three are tholeiitic. Sample R589 has high Cr (1360 ppm), Ni (436 ppm) and MgO (15.3%), but also high K_2O (1.8%, Table 4). The high K with high Ti could indicate the stabilization of biotite during interaction with evolved melt. Whether or not these features are due to a primitive melt or cumulate origin is unclear, but the low Sr (25 ppm) favours a cumulate origin. Altogether the high K, combined with low Th and primitive nature is not characteristic of either DM or EM-type origin and suggests a disturbed geochemistry. The Tb minimum in the REE distribution curve

is probably analytical in origin and is not considered diagnostic (Fig. 10). The trace element pattern is scattered, but is rather similar to that of the volcanics (Fig. 11). Sample R593 is also a primitive medium-K basalt (Table 4) and has a low La/Yb ratio (Fig. 10). The trace element data show "humped" pattern (Fig. 11), but otherwise are comparable with most primitive volcanics (sample at 12% MgO in Fig. 9). The other three dyke samples from Rastinpää show variable K and Ba, and it is not clear whether these features are of magmatic origin. These samples have, apart K and Ba, a similar geochemistry to that of the volcanics (Fig. 9) which favours a comparable origin.

Dyke R576 (Table 4) from Konnekoski area is a medium-K tholeiitic basalt in composition (Figs. 7-8) and is characterized by high TiO_2, P_2O_5, Sr and Y (Fig. 9), thus resembling to some degree the WPB-dykes discussed below. The low Zr contrasts with to the evolved nature indicated by low Ni (<10 ppm) and high Al_2O_3 (18.5%) and indicates a different origin when compared with these WPB-dykes. The low SiO_2 and high Al_2O_3 and Sr (Table 4) could suggest a cumulate origin (plagioclase), but petrographical studies do not support this idea. The lack of REE and more complete trace el-

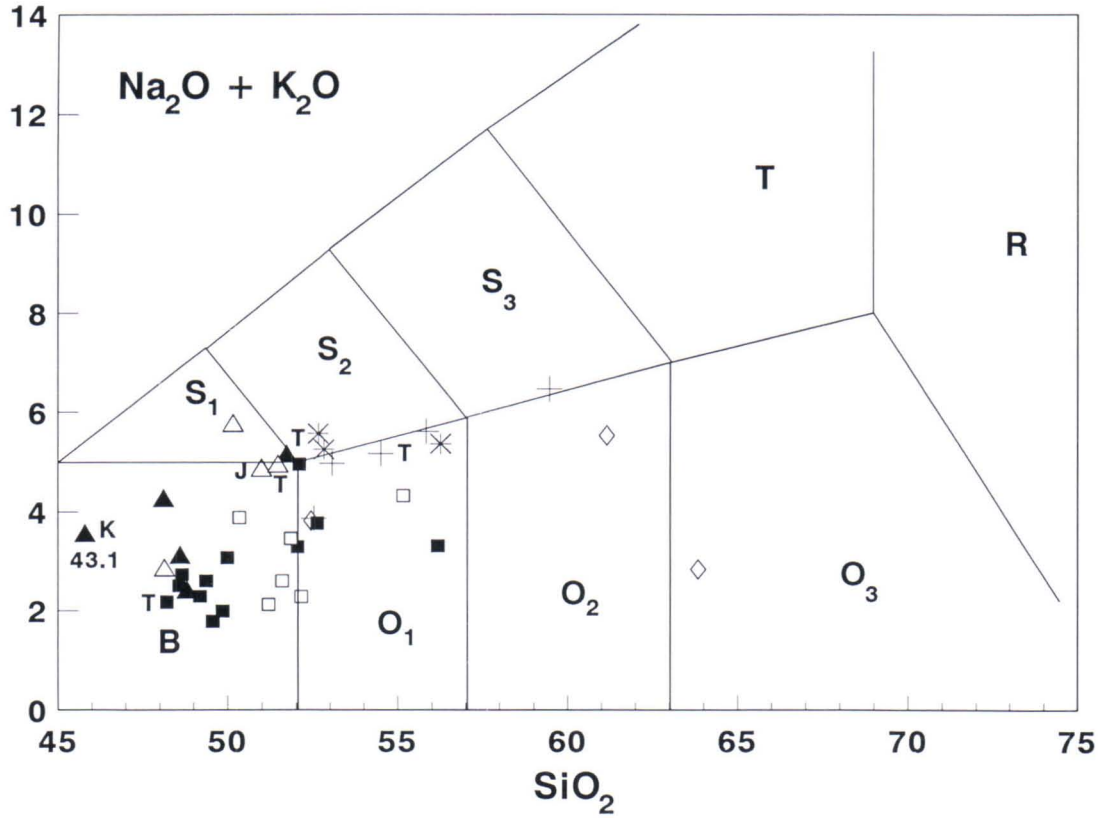

Fig. 7. Total alkali silica (TAS) diagram (Le Bas et al. 1986) for mafic volcanics, inclusions and dykes in the Rautalampi area. B - basalt, O_1 - basaltic andesite, O_2 - Andesite, O_3 - dacite, R - rhyolite, S_1 - trachybasalt, S_2 - basaltic trachyandesite, S_3 - trachyandesite and T - trachyte and trachydacite. Open squares - Pukkiharju volcanics; filled squares - Toholampi volcanics; crosses - mafic inclusions in gneissic tonalites; open triangles - mafic dykes in gneissic tonalites; asterixes - younger WPB-affinity dykes; open diamonds - mafic dykes in amphibolite associations. T - Toholampi and sample R240, K - Konnekoski, J - Joutenniemi.

ement data prevents more detailed interpretation, but a WP affinity can not be excluded. Joutenniemi and Toholampi dykes are medium-K basalts near to trachybasalts in composition (Figs. 7-8) and the Toholampi dykes are tholeiitic and Joutenniemi dyke calc-alkaline according to Figure 8. The lack of REE and more complete trace element data hinders interpretation. They have geochemical similarities with the volcanics (Fig. 9), but the higher Sr, Al_2O_3 (19%) and low Zr in the other Toholampi dyke is comparable with sample R240 from Toholampi.

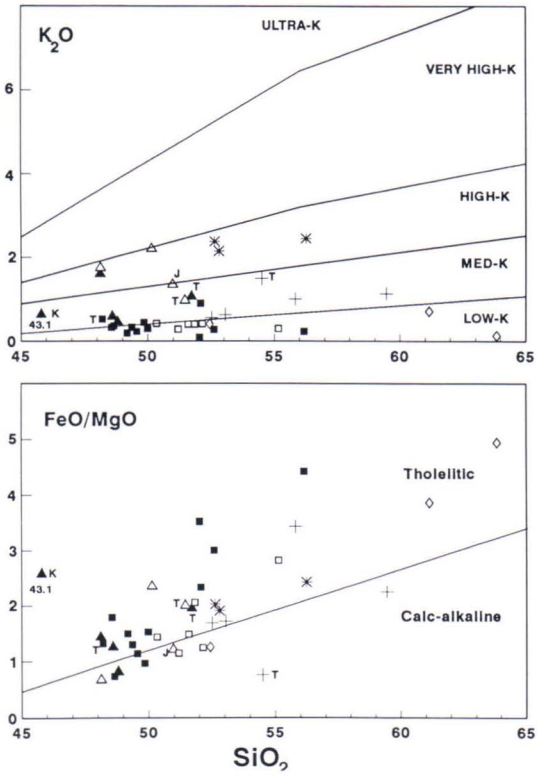

Fig. 8. K_2O-SiO_2 and FeO/MgO-SiO_2 diagrams for mafic volcanics, inclusions and dykes in the Rautalampi area. Boundary lines for K_2O-SiO_2 diagram are from Kähkönen (1989). The division between tholeiitic and calc-alkaline fields on the FeO/MgO-SiO_2 diagram is from Miyashiro (1974). See Figure 7 for symbols.

Dykes having WPB-affinity

These dykes are among the youngest rocks in the Rautalampi area and have crosscutting relationships with the volcanics. In the Pukkiharju area one dyke also cut pegmatite. They are high-K tholeiitic basaltic trachyandesites in composition (Figs. 7-8) and are characterized by high TiO_2, P_2O_5, Sr and Zr (Fig. 9 and Table 4). The REE distribution shows a LREE enriched pattern with a high La/Yb ratio (Fig. 10). The MORB-normalized trace element pattern (Fig. 11) shows an enriched pattern characteristic of WPB (Pearce 1983). There is a small negative Ta-anomaly which could indicate the occurrence of a small subduction component or crustal contamination. On the other hand these rocks are rather evolved and the occurrence of a Ti-bearing phase during the crystallization history is possible. The low Sc and V could favour this (Table 4). These rocks have an alkaline affinity, as seen in their high TiO_2 and P_2O_5, and they for instance plot in the alkali basalt field in the TiO_2-Zr/(P_2O_5 x 10000) diagram (not shown, Winchester and Floyd 1976). Although, these dykes are evolved rocks, they show clear alkaline WPB affinity and point to an EM-type mantle.

Dykes within the Toholampi amphibolite association

The orthopyroxene-bearing dyke R218 (Table 4) is a low-K calc-alkaline basaltic andesite in composition (Figs. 7-8). The low-K nature is questionable, due to the negative K-Rb anomaly in the trace element pattern (Fig. 11). The REE distribution is comparable with that of the volcanics, but the trace element pattern is "humped" in nature and is thus more comparable with sample R240. The other two more evolved samples are also low-K rocks, but they show clear tholeiitic affinities (Fig. 8). In particular, the sample at SiO_2 63.8% in Figure 8, has a very low K_2O value (0.1%). They show

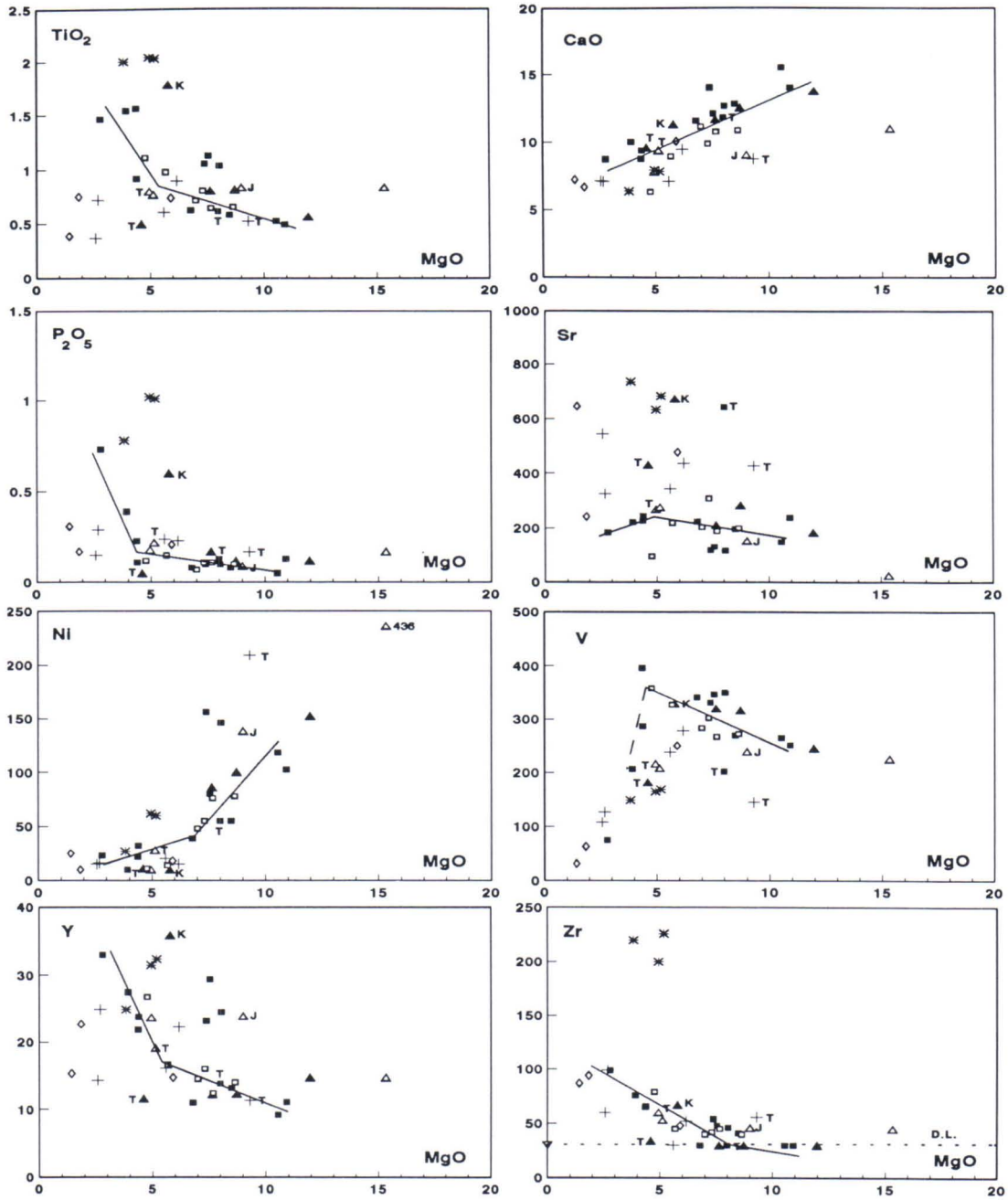

Fig. 9. Harker-type selected major element (%) and trace element (ppm) variation diagrams for mafic volcanics, inclusions and dykes from the Rautalampi area. Oxide totals were recalculated to 100 %. See Figure 7 for symbols. D.L. is the detection limit of Zr.

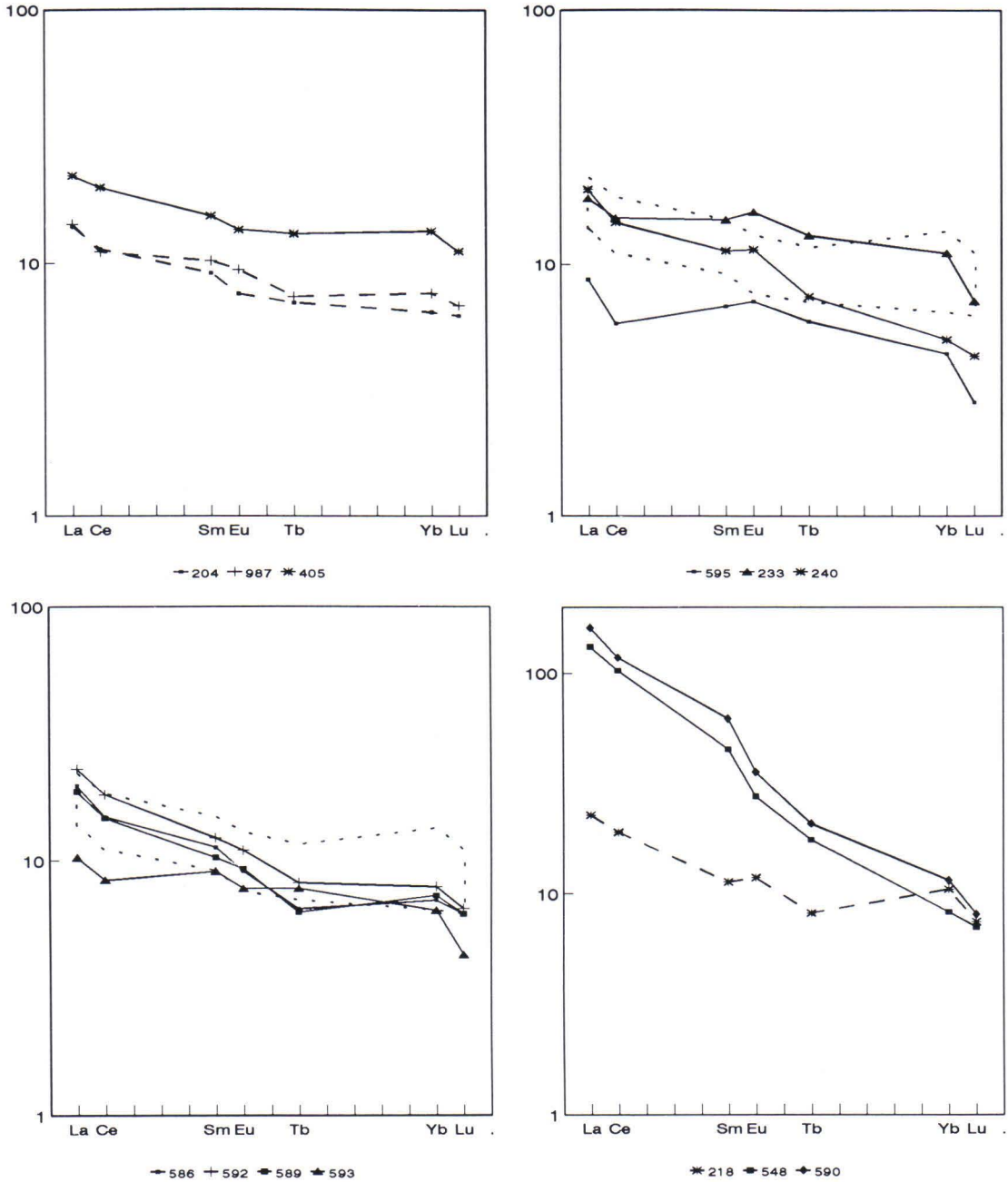

Fig. 10. Chondrite normalized rare-earth element patterns for Pukkiharju volcanics (upper left), Toholampi volcanics (upper right), mafic inclusions and dykes in gneissic tonalites (lower left) and younger mafic dykes (lower right). Sample numbers refer to Table 4. Dotted line outlines the variation of Pukkiharju volcanics in upper right and lower left figures. Lower right figure: solid line - WPB-affinity dykes; dashed line - dyke from amphibolite association.

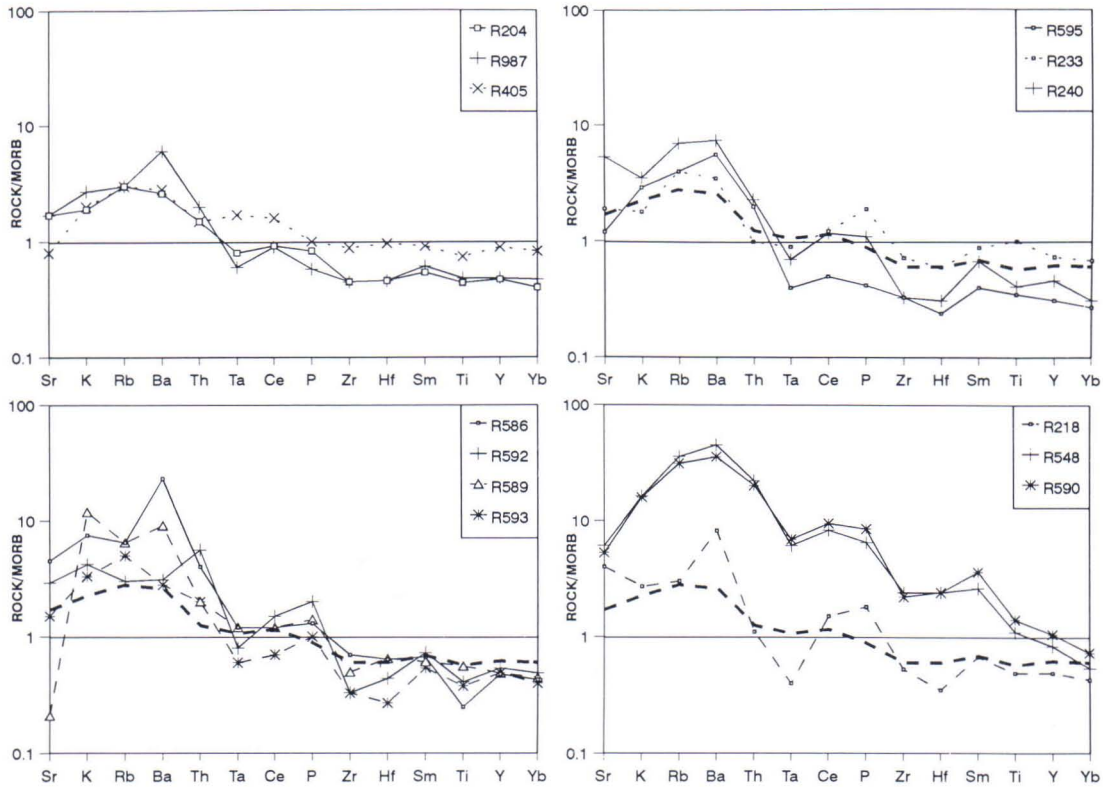

Fig. 11. Mid-ocean ridge basalt-normalized trace element patterns for Pukkiharju volcanics (upper left), Toholampi volcanics (upper right), mafic inclusions and dykes in gneissic tonalites (lower left) and younger mafic dykes (lower right). Sample numbers refer to Table 4. Normalizing values are from Pearce (1982). The dashed line in three of the figures is the Pukkiharju volcanic average. Notice that the Rb, Ta and Th values are often near detection limits in most primitive volcanics.

variable geochemistry (Fig. 9), but the lack of other trace element data hinders interpretation. The low-K nature of these samples could be due to loss of some mobile elements during granulite facies metamorphism, but more data is needed to verify this possibility.

Geochemistry and origin of altered rocks

Cordierite-sillimanite gneisses and tonalitic gneiss in the Pukkiharju area

Cordierite-sillimanite gneisses are found associated with mineralized zones at the Pukkiharju Zn-Cu occurrence and they show enrichment of magnesium and potassium and depletion of sodium, calcium and strontium (Lahtinen 1988). Analyses of two samples are given in Table 5 and their REE distribution in Figure 12. The REE patterns and TiO_2, Al_2O_3 and Zr values are comparable with those for felsic gneisses of probable volcanic origin (see samples R550 and R29 in Table 2) and accordingly a similar protolith is assumed for these cordierite-sillimanite gneisses. Magnesium and K, especially in sample R305, show enrichment

and Ca, Na and Sr show depletion when compared to proposed protoliths. Barium follows K and is also enriched, such that Ba-rich zones are found associated with mineralization and indicated by the presence of barium feldspar (hyalophane) in some samples (Lahtinen and Johanson 1987). Iron shows rather strong variation in the Rautalampi felsic gneisses (Fig. 3) so that it is not clear whether it is enriched or not. The increase of iron due to sulphide precipation is seen in sample R165, but whether there has been a corresponding increase of Fe in the silicate phase is open. The same applies to elements like V and Sc. The chemical index of alteration (CIA) has been used to quantify the effects of weathering, especially feldspars (Nesbitt and Young 1982). Fresh igneous felsic rocks have CIA values under 55 and often ≤ 50. Increases in CIA values (Table 5) reflect the destruction of plagioclase and sample R165 show also slightly more negative Eu minimum when compared to samples R550 and R29 in Table 2, which probably indicates slight depletion of Eu. The present mineralogy is due to regional metamorphism and previously these rocks have been rich in chlorite and sericite/muscovite. All these features are similar to those of hydrothermal alteration zones found in many volcanic-associated massive sulphide deposits (VMSD, e.g. Franklin et al. 1981) and a similar origin is proposed for cordierite-sillimanite gneisses in Pukkiharju Zn-Cu prospect.

About 200-300 m away from the mineralized rocks (and possibly lower in the stratigraphy) is a tonalitic gneiss that is characterized by high Na_2O and low CaO, as can be seen from one representative sample (R540 in Table 5). Four other samples have been taken from this unit and they all show the same geochemical characteristics as sample R540. The REE pattern (for sample R540 in Fig. 12) and SiO_2, TiO_2, Al_2O_3, Y and Zr values are comparable with those of the high-Y felsic gneisses and volcanics in the Rautalampi area and consanguineous origin is proposed. Apart from enrichment of Na and depletion of Ca, there is a slight increase in Fe, possibly also in Mg and depletion in Sr (data not shown). The CIA value of 50 does not indicate the presence of severe alteration and the increase in Na indicates "spilite" type alteration. A subvolcanic sill/dome origin is proposed, but a volcanic protolith is also possible. The slight enrichment of Na and Fe are comparable with background alteration ("spilitization") and lower conformable alteration zones found associated with some VMSD (Franklin et al. 1981, Walford and Franklin 1982, Parry and Hutchinson 1981).

Garnet±cordierite±orthoamphibole/orthopyroxene (GCO) rocks and gneisses

GCO rocks and gneisses are found at several localities in the Rautalampi area. Eight profiles across these rocks have been sampled (130 samples) and five of these (L1, L4, L5, L7 and L8) will be discussed in more detail. So far, there are only XRF-analyses from the Outokumpu Co laboratories available for most of these samples. The oxide sums are normally low (about 90%) indicating that the complex and coarse-grained mineralogy with, for example abundant garnet, have caused analytical problems. Therefore, these analyses are not particularly accurate, but can be correlated with each other and give an approximation of chemical composition. Sulphur analyses are only indicative and variable, but tend overall to be too high, especially in lower values. Some representative data analyzed by other methods are shown in Table 5. Another problem with these rocks is also the possibility of mass and volume change. Variation in possible protolith compositions, lack of detailed trace element data for the main part of the data and analytical uncertainties makes it impossible to calculate the extent of these changes. Zirconium, Ti and Al are normally considered to be rather immobile during alteration as demonstrated above in the case of the cordierite-sillimanite gneisses, but Al show overlapping values with some

Table 5. Selected major (anhydrous) and trace element data for tonalitic gneiss, cordierite-sillimanite gneisses, GCO rocks and gneisses, and garnet-bearing felsic gneisses and mica gneisses from the Rautalampi area.

Sample	R540	R219	R165	R305	R327	R332	R341
Type[a]	a	b	b	c	d	c	e
Location[b]	P	P	P	L1	L2	L2	L3
SiO_2	74.78	76.83	70.25	55.80	69.97	55.86	52.80
TiO_2	0.31	0.27	0.33	1.03	0.69	1.42	0.65
Al_2O_3	12.21	12.47	12.52	16.28	12.13	13.92	17.60
FeO*	4.43	3.02	3.54	11.27	7.20	16.76	11.98
MnO	0.09	0.03	0.09	0.20	0.17	0.24	0.48
MgO	0.58	2.56	7.58	13.63	4.11	9.89	9.76
CaO	0.65	1.43	0.35	0.46	2.34	0.93	2.83
Na_2O	6.17	2.34	0.43	0.24	3.28	0.32	3.69
K_2O	0.72	1.00	4.83	0.66	0.00	0.12	0.06
P_2O_5	0.06	0.05	0.08	0.43	0.11	0.54	0.15
FeS_2	<0.2	<0.2	3.1	<0.2	4.0	<0.2	<0.2
Ba	256	371	684	227	112	111	68
Co	5	6	12	24	18	33	35
Cu	<1	17	26	38	148	59	17
Ni	1	16	17	21	22	11	103
Sc	11.8	12.2	12.9	27.9	21.1	30.8	45.3
Sr	65	132	32	16	157	9	109
V	6	5	21	82	75	87	313
Y	41.4	39.5	41.0	44.9	50.7	55.7	14.5
Zn	67	88	136	157	1630	425	170
Zr	168	187	205	245	308	117	<30
La	18.3	18.8	21.0	16.2	17.0	7.0	3.5
Ce	36.0	40.0	42.0	36.0	37.0	19.0	6.8
Sm	4.7	5.9	7.4	7.4	5.8	5.1	1.7
Eu	1.10	1.49	1.08	0.53	1.40	0.45	0.75
Tb	0.89	1.09	1.04	0.96	0.96	1.00	0.33
Yb	4.5	5.1	4.7	3.7	5.4	5.7	1.2
Lu	0.63	0.54	0.62	0.48	0.68	0.70	0.15
[c]CIA	50.0	62.2	65.8	89.6	55.6	85.6	60.9
$(La/Yb)_{CH}$	2.7	2.5	3.0	2.9	2.1	0.8	2.0
[d]Eu/Eu*	0.67	0.73	0.45	0.22	0.70	0.25	1.26

Note.- Major elements (wt%) and Ba,Co,Cu,Ni,Sc,Sr,V,Y and Zr analyzed at the GSF Laboratory by ICP-AES; REE by neutron activation at the Technical Research Centre of Finland. FeO*: total Fe as FeO.Sulphur calculated as pyrite and Fe in pyrite is subtracted from total FeO.

[a]a- tonalitic gneiss; b- cordierite-sillimanite gneiss; c- orthopyroxene/orthoamphibole-cordierite rocks; d- orthopyroxene/orthoamphibole ± cordierite gneiss; e- orthopyroxene/orthoamphibole-garnet ± cordierite gneiss; f- garnet-bearing quartzfeldspar gneiss; g- hornblende-biotite gneiss (migmatite); h- garnet-pyroxene mica gneiss (migmatite).

[b]P- Pukkiharju; T- Toholampi; R- Rastinpää; L1-L6 refer to Fig. 2.

[c]CIA = Molecular ratio of $(Al_2O_3/(Al_2O_3+CaO+Na_2O+K_2O))*100$ (Nesbit and Young 1982).

[d]Eu/Eu* = Observed Eu value/value obtained by interpolation between Sm and Tb.

mafic volcanics and felsic gneisses. Normally however, felsic gneisses and volcanics show Al_2O_3 values less than 14% and mafic volcanics values over 14%. The Zr/TiO_2 ratio for felsic gneisses and volcanics is mainly over 450 and for mafic volcanics less than 150 (normally < 100). Felsic mica gneisses also show often values near to 450, but normally they have

Table 5 (continued).

Sample	R346	R368	R374	R402	R546	R563	R242
Type[a]	e	e	f	e	g	f	h
Location[b]	L3	L4	L5	L6	R	T	T
SiO_2	71.66	54.88	67.72	72.26	62.07	76.04	66.46
TiO_2	0.27	0.48	0.56	0.48	0.51	0.26	0.71
Al_2O_3	14.03	14.53	14.12	11.54	15.35	11.68	14.56
FeO*	4.62	16.49	7.58	7.88	7.57	5.39	6.08
MnO	0.10	0.11	0.15	0.07	0.13	0.14	0.09
MgO	3.97	8.95	1.95	5.37	3.91	2.08	2.98
CaO	0.81	1.69	2.82	0.46	5.27	0.81	3.21
Na_2O	3.62	1.56	3.20	1.75	2.83	3.16	3.24
K_2O	0.91	0.73	1.71	0.19	2.11	0.39	2.48
P_2O_5	0.01	0.58	0.19	<0.01	0.25	0.05	0.19
FeS_2	<0.2	<0.2	0.5	<0.2	<0.2	<0.2	<0.2
Ba	250	121	718	95	351	181	588
Co	5	19	10	13	23	<1	23
Cu	28	15	55	17	31	38	38
Ni	38	27	3	14	59	<10	52
Sc	15.0	36.7	21.1	15.8	17.7	6.9	17.5
Sr	72	25	110	34	283	51	285
V	29	193	22	35	125	5	124
Y	40.4	34.6	54.3	36.3	21.1	52.3	17.3
Zn	1090	57	142	111	91	125	137
Zr	139	79	191	210	119	233	218
La	18.0	9.0	20.0	28.0	-	-	-
Ce	33.0	23.0	38.0	51.0	-	-	-
Sm	4.7	6.8	6.1	8.2	-	-	-
Eu	1.20	0.95	1.60	0.82	-	-	-
Tb	0.72	1.10	0.96	0.83	-	-	-
Yb	4.4	2.9	5.8	3.8	-	-	-
Lu	0.62	0.39	0.78	0.60	-	-	-
[c]CIA	62.7	67.0	53.5	75.0	48.1	62.5	51.4
$(La/Yb)_{CH}$	2.8	2.1	2.3	4.9	-	-	-
[d]Eu/Eu*	0.78	0.42	0.80	0.47	-	-	-

lower values (400-350) that further decrease with more pelitic nature (Lahtinen 1988 and unpublished data by the author). Garnet-sillimanite-bearing gneisses can show rather low values, partly overlapping with mafic volcanics, but normally they have Zr values over 100 ppm, which is atypical for mafic volcanics of the study area. The mica gneisses and garnet-sillimanite-bearing gneisses normally have low CIA values (< 60) pointing to rather immature sediments, but the possibility of Ca enrichment due to the common occurrence of carbonates should be kept in mind. Hornblende-biotite gneiss (migmatite) R546 (Table 5) is probably of tuffaceous origin as is also seen in the low CIA value. Garnet-pyroxene mica gneiss (migmatite) R242 also has a low CIA value. They have Zr/TiO_2 ratios 233 and 307 respectively and thus differ clearly from mafic and felsic volcanics.

Although there are many uncertainties, as discussed above, an approach has been tried using the Zr/TiO_2 ratio to discriminate the possible protoliths of GCO rocks and gneisses.

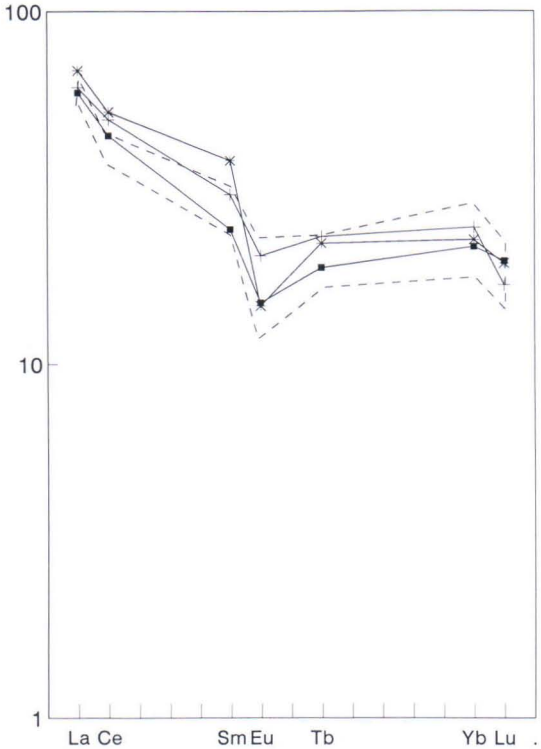

Fig. 12. Chondrite normalized rare-earth element patterns for tonalitic gneiss and cordierite-sillimanite gneisses from the Pukkiharju area. Sample numbers refer to Table 5. Dashed line outlines the normal variation of high-Y group gneissic tonalites and volcanogenic gneisses.

Both of these elements are considered to be rather immobile so that the use of this ratio should not be affected by mass and volume changes. Samples that have high Zr, rather low TiO_2 and a Zr/TiO_2 ratio over 450 are considered to have had protoliths comparable with felsic volcanics. Samples with low Zr, higher TiO_2 and a Zr/TiO_2 ratio less than 150 (normally 100) are considered to have mafic volcanic protolith. Samples with Zr/TiO_2 ratios between 150-450 are considered to have had a sedimentary or tuffaceous origin with higher ratios indicating a more felsic source and lower ratios a more mafic source. A sedimentary origin with an exotic source can not be distinguished from tuffaceous rocks with variable mixture of felsic and mafic material of local origin. Intermediate volcanics with Zr/TiO_2 ratios between 150-450 and SiO_2 values between 58-68% are, based on the available data, generally absent from the Rautalampi area and thus the possibility of intermediate volcanic protoliths is excluded.

Toholampi area

There are two sampling profiles in the Toholampi area (L1-L2 in Fig. 2). L1 is considered in detail here and only a few references will be made to L2. The chemical variation seen in L1 is shown in Figure 13. The southern part of the profile (A) is composed of cordierite-hypersthene/anthophyllite rocks of variable thickness (25-100 cm). These rocks are normally devoid of or contain only minute amounts of plagioclase, as seen in the extremely low Ca and low Na and Sr values. Potassium shows variable but low values with only one high value reflected in the abundance of phlogopitic mica (later?) in this sample. Zirconium, TiO_2 and the Zr/TiO_2 ratio fluctuate indicating both mafic and sedimentary type protoliths. Low SiO_2 and elevated Al_2O_3 and Cr correlate with samples having low Zr and Zr/TiO_2 corresponding with a mafic protolith. These samples also have FeO and MgO peaks which is partly primary in origin, but the level is higher than that found normally in mafic volcanics, indicating later enrichment of Fe and Mg. Samples that plot in the suggested sedimentary field have lower FeO and MgO, but these levels are also high for such silica-rich rocks. Part of the increase in Fe in these rocks is attributed to the increased sulphide content.

The southernmost sample in Figure 13 is R305 in Table 5 and Figure 14. The REE distribution curve show similarities with felsic volcanics and gneisses but the extremely deep Eu minimum indicates severe depletion. High Zr and REE data could favour felsic origin, but the high TiO_2 and Al_2O_3 favour a mixed or

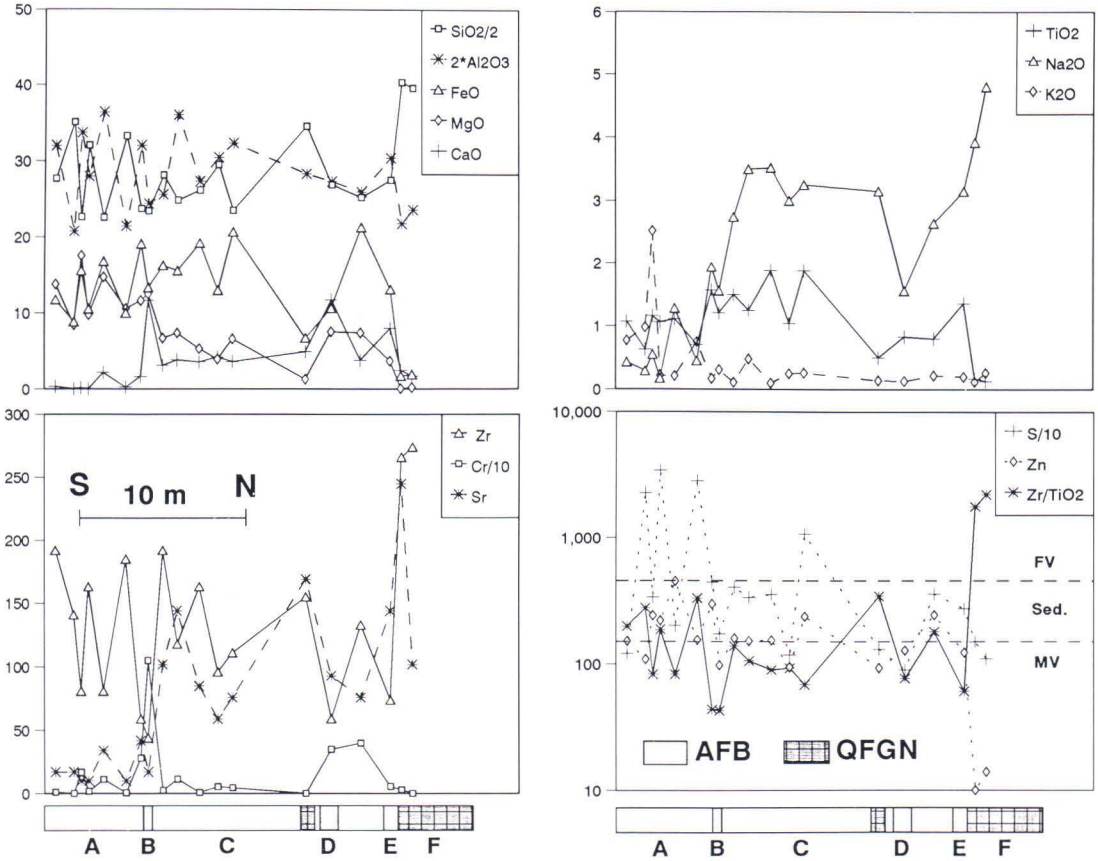

Fig. 13. Variation of selected elements in GCO rocks and gneisses and related rocks in profile L1 (Fig. 2). AFB - amphibolite (metavolcanic); QFGN - quartz-feldspar gneiss. Capitals are referred in the text.

sedimentary origin as does also the high P_2O_5 (Table 5). The almost total loss of alkalies is also reflected in the high CIA value. Chlorite has a CIA value near 100 and the original mineral composition was possibly dominated by chlorite+quartz.

The amphibolite layer B (Fig. 13) is rich in Cr and MgO and show normal geochemistry with high CaO comparable with the mafic volcanics in the Rautalampi area. Cordierite±hypersthene/anthophyllite lithologies north of this (C in Fig. 13) contain plagioclase and locally abundant garnet. This is manifest in an increase of CaO and especially Na_2O and Sr. The FeO contents show similar variation or even higher values, but MgO shows lower values, and these rocks are almost devoid of K_2O when compared to the southern part. Zr/TiO_2 ratios, as well as high Cr in some samples, indicate mafic protolith but the rather high Zr values indicate either loss in mass or a tuffaceous/sedimentary protolith for most samples. A conspicuous feature is the very high level of TiO_2. One orthopyroxene/anthophyllite-bearing quartz-feldspar gneiss showed an only slightly altered composition as also did amphibolites at the northern end (D and E in Fig. 13). Quartz-feldspar gneiss (F) in the northern end probably has a near primary composition.

The highest sulphur values are found at the

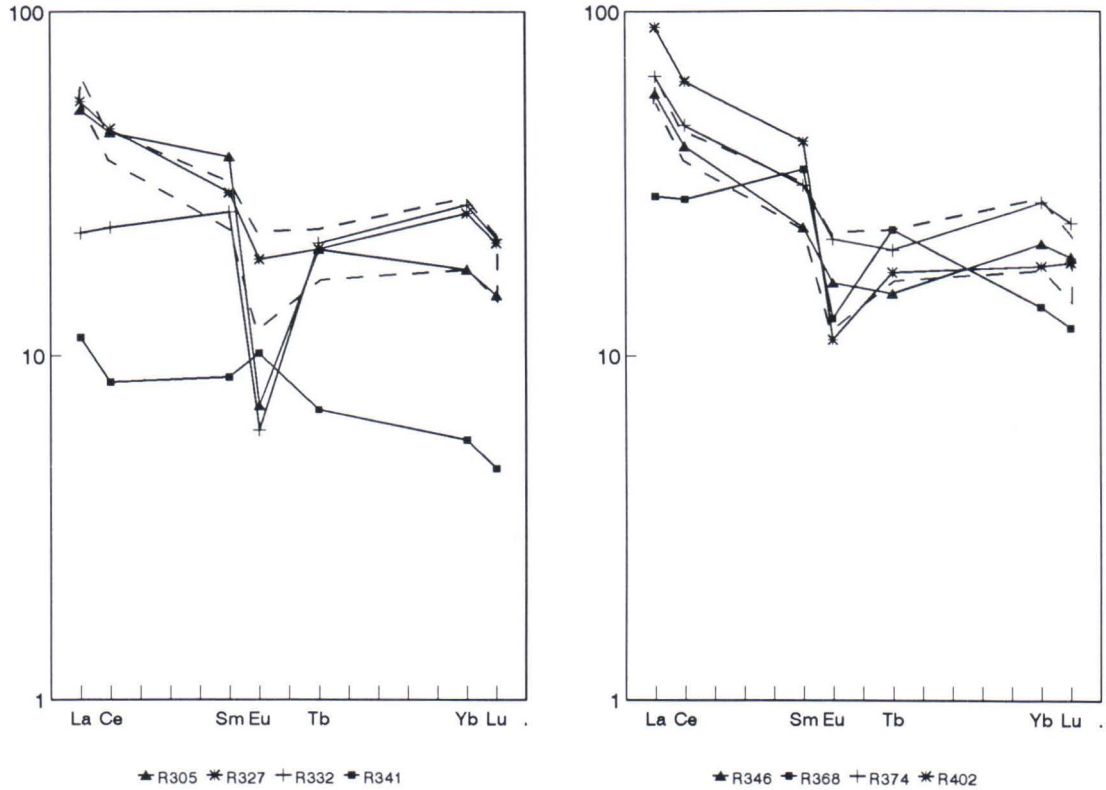

Fig. 14. Chondrite normalized rare-earth element patterns for GCO rocks and gneisses in the Rautalampi area. Sample numbers refer to Table 5. Dashed line outlines the normal variation of high-Y group gneissic tonalites and volcanogenic gneisses.

southern end (C) from samples having an inferred sedimentary origin. Three samples also show high Cu values (360-960 ppm, data not shown). Some Cu values in the order of 100-200 ppm are found in the northern part of the profile. The normal variation of Zn in the mafic volcanics is 80-130 ppm and felsic rocks 40-80 ppm. Elevated Zn values often correlate with elevated Cu, but the highest value of 450 ppm is found in a sample devoid of Cu. Some high Ni values are also found (data not shown), but they correlate with high Cr and it is not clear if there are substantial amounts of Ni in the sulphide phase.

Although there are differences between these profiles, these rocks have been found in approximately the same position in several places indicating a general stratabound origin. The lack of complete sequences hampers the correlation of these profiles. L2 has only one sample (R332, Table 5) that is almost totally devoid of plagioclase. Low Zr, high TiO_2 and a Zr/TiO_2 ratio of 82 in this sample indicate a mafic volcanic protolith, but the REE distribution is more indicative of a felsic origin (Fig. 14). The low LREE level could indicate depletion that possibly also affected Zr. This sample has similarities with sample R305 from L1 and has high CIA, deep Eu minimum, high P_2O_5 and high TiO_2.

Other samples from L2 contain abundant plagioclase and are similar to the northern part of L1. They are also characterized by higher FeO when compared with MgO and very low K_2O. Late activity of K is seen in one sample wherein phlogopitic mica overgrows gneissic

banding. These samples differ further in having higher Zr/TiO_2 ratios (normally 400-900) when compared to normal values (< 200) found in the northern part of L1. Only when going north of sample 332 there is a comparable sample (Zr/TiO_2 ratio of 120) Sample R327 is a representative for the high Zr/TiO_2 ratio samples (Table 5) and has a REE distribution comparable with that of felsic volcanics and gneisses (Fig. 14). The CIA value is rather low and indicates only slight alteration, consistent with the high Sr. This sample is totally devoid of K and has high Zn and elevated Cu. The high TiO_2 compared to low Al_2O_3 could indicate Ti enrichment, but a primary origin (tuffaceous) or increase due to mass loss should also be considered. Zinc values are always anomalous in L2 (200-1600 ppm). Copper values are also elevated, but they are less than 150 ppm. The Ni values in sulphide-rich samples are < 20 ppm, indicating very low Ni values in sulphides. There are two amphibolite and one quartz-feldspar gneiss samples from L2 and these show normal compositions without any distinct change in composition.

Garnet-bearing quartz-feldspar gneisses are found as 10-20 m thick (at least) units within the main amphibolite association in Toholampi area (between samples 216 and 236 in Fig. 2). These rocks have been found at several localities but it is uncertain whether they form a single stratabound horizon. Sample R563 in Table 5 is from this association. The SiO_2, TiO_2, Al_2O_2, Y and Zr values are comparable with high-Y felsic volcanics and gneisses in the Rautalampi area, the main differences being higher FeO, MgO and lower CaO and Sr. Copper and Zn also show elevated values. The low CaO, possibly also lower Na_2O, associated with the elevated CIA value could indicate either a sedimentary (weathering) or altered origin with a felsic protolith. The general geochemical similarity with felsic volcanics favours the latter volcanic origin.

Pukkiharju area

There are five profiles (L3-L7 in Fig. 2) in the Pukkiharju area and three of them (L4, L5, L7) are considered below in more detail. Profiles L3-5 are probably from the same stratabound horizon. The observed northern contact of L4 (A in Fig. 15) is composed of homogenous and banded amphibolites. Homogenous amphibolite (northernmost sample in Fig. 15) has chemical composition comparable with metavolcanics. Banded amphibolite (next sample) shows an increase in FeO and S depicting sulphide enrichment. Notice that there is no Zn enrichment associated with the sulphides. These two samples have rather high Na_2O values. A thin quartz-feldspar gneiss B (20 cm layer) has a composition comparable to that of felsic volcanics and gneisses, except its high Sr and Na_2O, which could be due to alteration, particularly considering the high Na_2O value in garnet-anthophyllite gneisses occurring to the south. GCO gneisses and rocks with low SiO_2, Zr and high Al_2O_3 values in Fig. 15 (C and E) are interpreted as having had mafic protoliths. These values and low Zr/TiO_2 ratios (< 100) are comparable with the thin amphibolite layer D. Samples from proposed felsic protoliths have high SiO_2, Zr and low Al_2O_3 values with high Zr/TiO_2 ratios. TiO_2 values in felsic protoliths vary from 0.14-0.60%, but there is no correlation with FeO (2-11%) and the sample with the lowest TiO_2 has high FeO (9.6%). One sample (fifth from south in E) deviates from these general trends and has a Zr/TiO_2 ratio of 165 (sample R368 in Table 5). The disturbed REE distribution (Fig. 14) and chemical composition, especially high P_2O_5, indicate a mixed and sedimentogenic protolith for this sample.

CaO is depleted and is normally < 1%, but three samples with a mafic protolith north of D have higher values. FeO and MgO are normally enriched, but the three above considered samples show only small changes in these elements. Although scattered there is a decrease in Na_2O from north to south followed by a concomitant increase in K_2O. Note also the high level of Na_2O in the three samples with mafic protoliths occurring north of D. Two samples

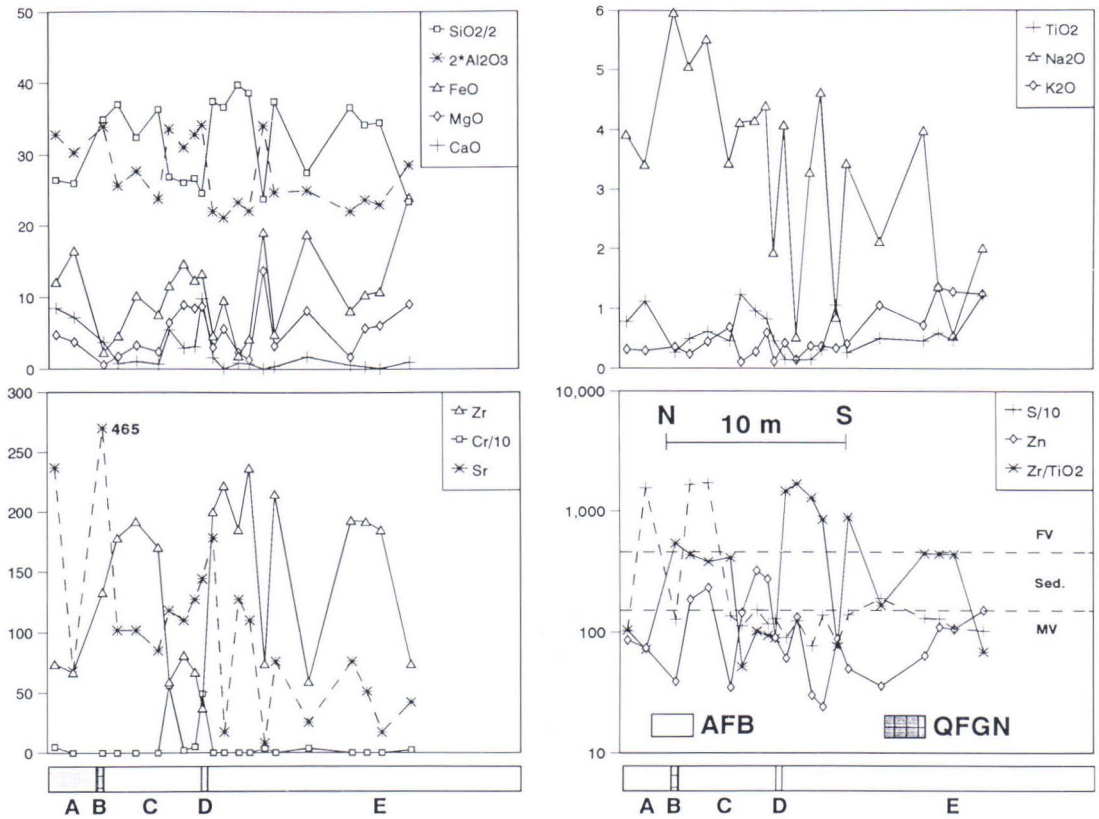

Fig. 15. Variation of selected elements in GCO rocks and gneisses and related rocks in profile L4 (Fig. 2). AFB - amphibolite (metavolcanic); QFGN - quartz-feldspar gneiss. Capitals are referred in the text.

with very low Na_2O values have also extremely low CaO and Sr, which is in accordance with their plagioclase-free mineralogy. High sulphur contents are found in two samples in the north associated with elevated Zn, but high Zn values are also found from rocks devoid of sulphides. Slightly elevated Cu (70-80 ppm) is found from the two sulphide-bearing samples, but other samples show only low Cu values (data not shown).

L3 is located 100 m SW of L4 and is quite comparable with it. The northern contact is a banded amphibolite which is followed with three samples having Zr/TiO_2 ratios (400-500) comparable with L4 samples in similar positions. These are followed by three samples with low Zr/TiO_2 ratios. Sample R341 in Table 5 is a representative sample of these rocks. Low CaO and K_2O, slightly elevated FeO and MgO and clearly enriched Na_2O are the main differences with respect to comparable mafic volcanics. The REE pattern (Fig. 14) is, apart from its positive Eu anomaly, also comparable with that of the mafic volcanics. These rocks are followed by samples with high Zr/TiO_2 ratios (500-1080) and the situation is thus similar with that for L4. Sample R346 (Table 5 and Fig. 14) has high MgO and low CaO, but is otherwise comparable with felsic volcanics. Notice the high Zn value associated with low S (< 0.1%). Some other elevated Zn values are also found but Cu values are low (< 30 ppm). Nickel

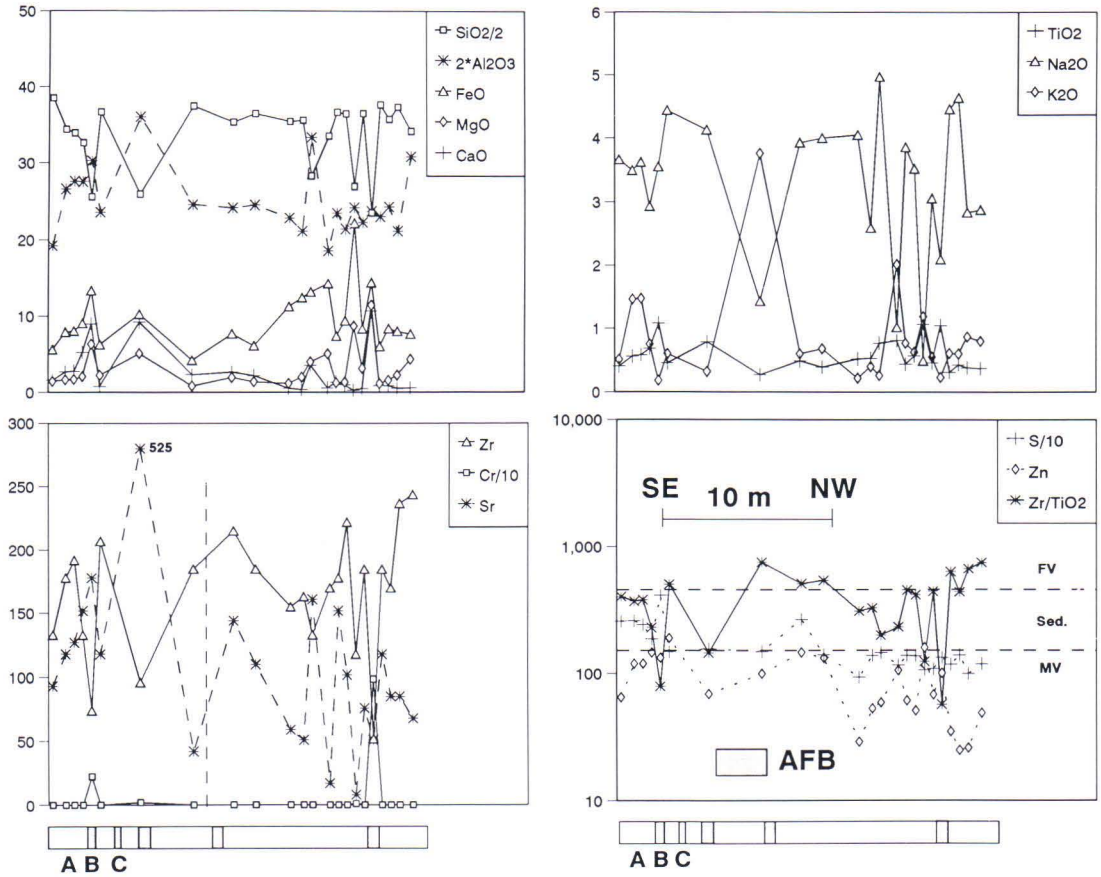

Fig. 16. Variation of selected elements in GCO rocks and gneisses and related rocks in profile L5 (Fig. 2). AFB - amphibolite (metavolcanic); QFGN - quartz-feldspar gneiss. Dashed line in lower left diagram correlates with a break in the profile. Capitals are referred in the text.

show some higher values associated with high Cr indicating a primary origin within a silicate phase.

Profile L5 is situated about 200 m northeast from L4 and is divided to two parts separated by 13 m section covered by overburden. The SE part of profile (A in Fig. 16) contains garnet-bearing quartz-feldspar gneisses that have abundant biotite (10-15%) and locally also cummingtonite; these rocks could also be called biotite-plagioclase gneisses. They contain abundant CaO and elevated K_2O and their Zr/TiO_2 ratios plot in the sedimentary field. A representative sample, R374 in Table 5, has a composition comparable with sample R423 (Table 2) and is probably of sedimentary origin. The REE distribution curve is also comparable with those of the high-Y felsic volcanic and gneisses (Fig. 14). The occurrence of zircon both as inclusions in biotite and as small rounded grains are also consistent with a sedimentary origin. The slightly higher FeO and MgO and lower Sr in sample R374 could be either primary in origin or represent small change in composition. An amphibolite layer (B in Fig. 16) has a near primary chemical composition, but the Na_2O level seems to be somewhat high for such a Cr-rich (220 ppm)

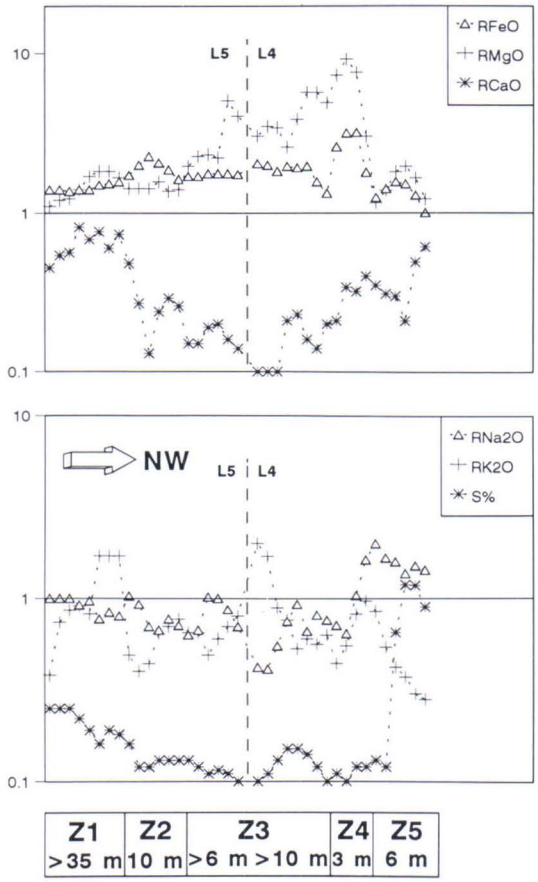

Fig. 17. Variation of calculated enrichment/depletion ratios (R) for selected elements in GCO rocks and gneisses in combined profile L4+L5. Moving average based on three samples has been used and the thickness of the sampled unit has been omitted. Different zones have been considered in the text.

volcanic. Garnet-anthophyllite-bearing felsic gneiss NW of this amphibolite has low CaO and high Na_2O (Fig. 16). The amphibolite C differs from normal mafic volcanics in having high Sr. Garnet-bearing felsic gneiss exposed just before the break in the profile has a Zr/TiO_2 ratio comparable with that of felsic volcanics and apart from its high K_2O, low Na_2O and possibly low CaO, it has a chemical composition in accord with a felsic volcanic origin.

The rocks in the NW part of this profile have both felsic and tuffaceous/sedimentary protoliths. The one sample with a low Zr/TiO_2 ratio (122) has rather high Zr (118 ppm), favouring a sedimentary origin. FeO and MgO values, although scattered, show an increase when going to NW associated with a corresponding decrease in CaO. K_2O increases and Na_2O shows fluctuating values while sulphur values are highest in the SE part of the profile and lowest values in the NW part. Zinc does not follow this trend, but there are many low values in the NW part of the line. Copper and Ni are low everywhere.

L5 seems to show the southern and L4 the northern contact of this association which thus appears to have a thickness of about 50-100 m. To see the possible change in enrichment and depletion of some elements the sample values were divided by the assumed protolith values to obtain enrichment/depletion ratio R. Protolith composition was estimated by using the subdividing lithologies into felsic volcanics and gneisses, sedimentary rocks and mafic volcanics. Abundances of TiO_2 and Zr were used to find the most suitable unaltered sample for comparison using data from this study and Lahtinen (1988, unpublished). Chromium values were also used for samples having mafic protoliths. This is a very crude method and it is further limited by analytical uncertainty. Thus these ratios are only indicative and while they may show possible trends, the absolute values are not accurate. Ratios were calculated for L5 and L4 samples. Moving averages of three samples were used to smooth the scatter seen in Figs. 15-16. The result is shown in Fig. 17, where the sample length is not taken in account and profiles L5 and L4 are combined.

The results have been divided to five overlapping zones. Zone 1 could be divided into two parts where the four first samples show R values near to predicted protoliths. RMgO shows increasing values and RCaO rather constant values in Zone 1. The K_2O increase in three samples is due to one sample that has high K_2O value and it is not clear whether this is diagnostic for this part of Zone 1 or not. Zones

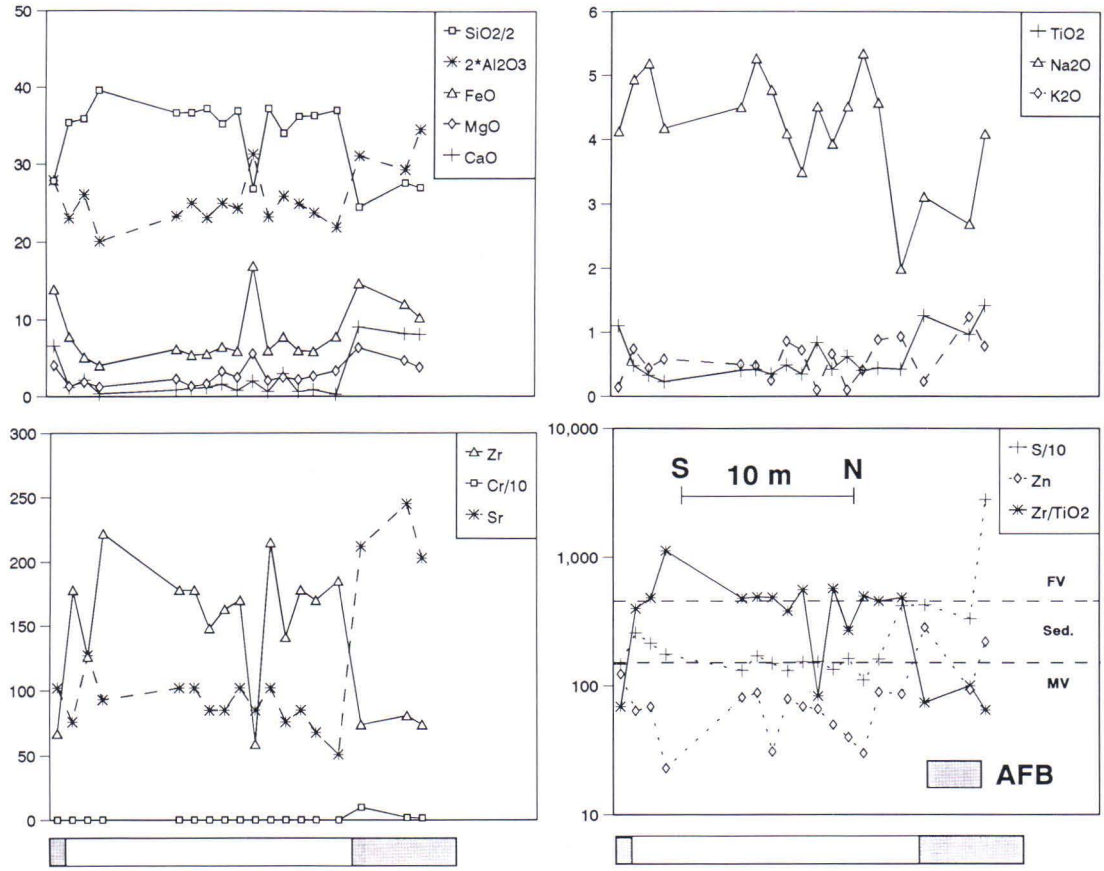

Fig. 18. Variation of selected elements in GCO rocks and gneisses and related rocks in profile L7 (Fig. 2). AFB - amphibolite (metavolcanic).

2-5 show low RCaO values with lowest values in Zone 3. RK_2O values are scattered and low in these zones, but high values are found in the middle of Zone 3. RNa_2O values are also variable, but are normally low in Zones 2-4. Zone 2 is enriched in FeO over MgO and differs thus from Zones 3-5. The thickness of Zone 3 is not known, but L5 and L4 may partly overlap (Fig. 17) indicating a thickness of 10-15 m. Zone 3 shows nearly constant RFeO values with increasing RMgO values. Zone 4 shows the highest RFeO and RMgO values with slightly increasing RCaO values. Zone 5 is characterized by high RNa_2O with increasing RCaO. Sulphur decreases from Zone 1 to Zone 2. Zones 2-4 show low values, but the northwestern end of Zone 5 is enriched in S. RZn (not shown) shows enriched values in Zones 1 and 4-5. Zones 2-3 are normally depleted in Zn, but in the middle of Zone 3 is an enriched horizon (subzone) associated with enriched RK_2O. Notice that this subzone is devoid of sulphides and also that some high Zn values in Zone 4 occur in samples devoid of sulphides (see Fig. 16).

Profile L7 (Fig. 2) can be followed for 200 m to the east (L6). One sample in L7 had a mafic protolith and one had a sedimentary protolith, while others clearly show a felsic origin. Variation between samples is small. They show low CaO and K_2O and high FeO and MgO (Fig.

18). Na_2O values are variable, but normally they are enriched when compared to possible protoliths. Sulphur values are normally low and increase towards both ends as also does Zn. Enrichment/depletion values have been calculated as above for this line also (data not shown) and they show rather constant ratios with high RFeO, RMgO, RNa_2O and low RCaO and RK_2O. This association is not directly comparable with the zones in Figure 17, but the level of RFeO, RMgO, RCaO and RK_2O and also the higher level of RMgO when compared to RFeO are comparable with Zones 3-4. The main exception is clearly enriched nature of Na_2O which is comparable with Zone 5 in Figure 17. This association has similarities with the Na-rich samples in Zone 4.

Profile L6 is comparable with L7 and shows similar characteristics and seems to be a continuation of this association towards the east. The Na_2O values are normally high (data not shown) but one sample, which has the highest MgO also has a lower Na_2O value. The chemical composition of this sample (R402) is shown in Table 5 and its REE pattern in Figure 14. It is slightly enriched in LREE, but otherwise it has a chemical composition compatible with felsic protolith.

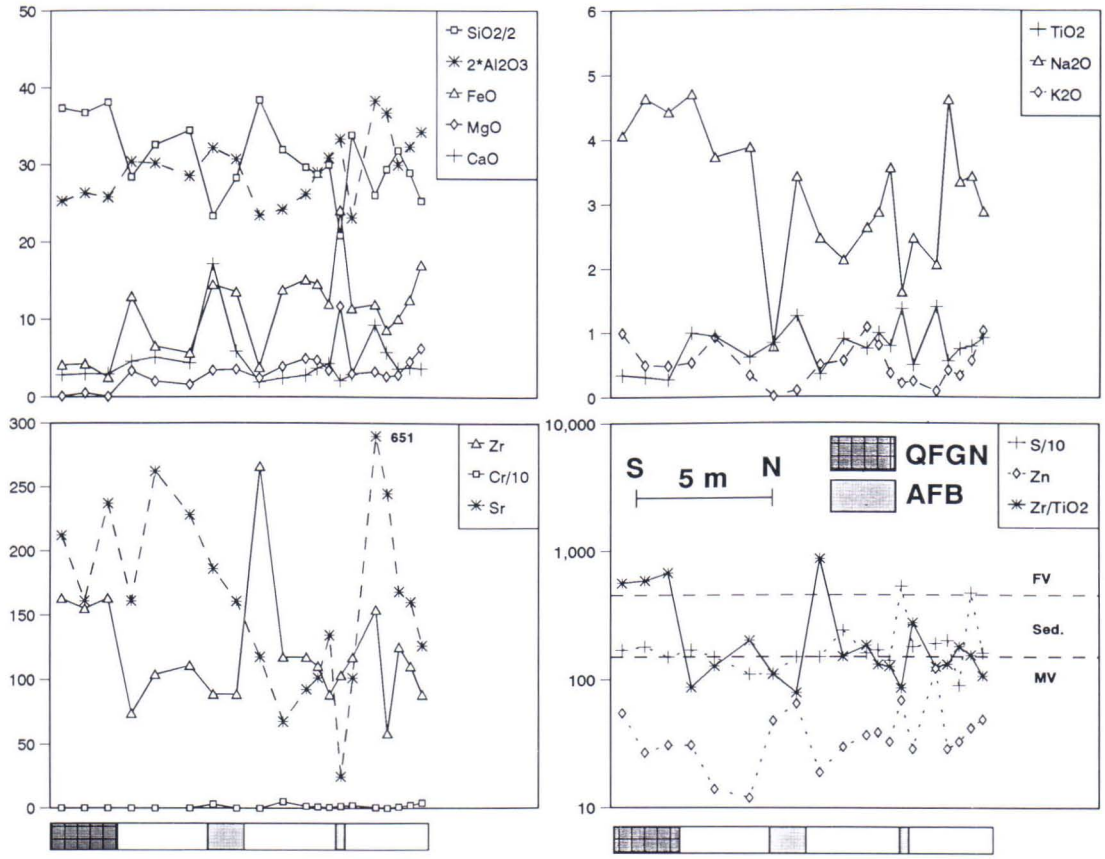

Fig. 19. Variation of selected elements in GCO rocks and gneisses and related rocks in profile L8 (Fig. 2). AFB - amphibolite (metavolcanic); QFGN - quartz-feldspar gneiss.

Konnekoski area

Profile L8 (Fig. 2) is a nearly 10 m thick association of garnet-orthopyroxene-cordierite gneiss layers alternating with amphibolite and quartz-feldspar gneiss layers. High TiO_2 with elevated Zr indicate a sedimentary or mixed protolith for most samples (Fig. 19). Samples that plot in the mafic volcanic field according to their Zr/TiO_2 ratio have both elevated SiO_2 and Al_2O_3 and are more likely to be of sedimentary or tuffaceous origin than direct volcanic origin. One sample with a high Zr/TiO_2 ratio also has high SiO_2 and low Al_2O_3 which is in accordance with a felsic protolith. The difference between this and Pukkiharju occurrences is the generally high CaO level (normally 2-4%) and the predominance of FeO over MgO. Uncertainty concerning the composition of the probable protolith makes comparison difficult. FeO values (normally 10-15%) are probably higher than in sedimentary rocks, but the MgO values (normally 2-5%) are not exceptionally high. Na_2O values are about the same level or slightly higher, but K_2O values are clearly lower than those found in sedimentary rocks (normally 2-3%). Four samples show high P_2O_5 (0.3-0.7%), also indicating sedimentary origin. Sulphur shows sporadic higher values with elevated Cu (140-200 ppm). These features differ from the Pukkiharju area, but are somewhat similar to the L1 plagioclase-bearing samples (Fig. 12), the main difference being the very low Zn values in L8 compared to more elevated Zn values in L1.

Origin of garnet±cordierite±orthoamphibole/orthopyroxene (GCO) rocks and gneisses in the Rautalampi area

Eskola (1914) was the first to undertake detailed studies on cordierite-anthophyllite rocks. He considered those in the Orijärvi area to represent Fe-Mg metasomatism caused by nearby granite. A metasomatic origin was also favoured by Simonen (1948) and Kano (1963). Tuominen and Mikkola (1950) proposed a model where clay-rich material flowed to fold hinges with depletion of alkalies and calcium associated with subsequent metamorphism. A partial melting origin wherein cordierite-orthoamphibole/orthopyroxene assemblages are thought to have formed as restites of certain metasedimentary and metavolcanic rocks after extraction of a granitic melt has been also proposed (e.g. Savolahti 1966, Grant 1968, Hoffer and Grant 1980). Marttila (1976a) considered the cordierite-garnet-anthophyllite rocks in Pielavesi area to represent mafic volcanics that have lost certain elements (Na and Ca) during expulsion of a fluid phase from an anatectic melt as a result of filter-pressing activity during shear movements. Sulphurization during metamorphism is also one possible mechanism for changing the chemical composition of a silicate phase to more magnesium-rich and hence make the crystallization of cordierite and antophyllite possible (Bachinski 1976). Isochemical metamorphism of hydrothermally altered rocks, first proposed by Vallance (1967) is however now the generally accepted model for most cordierite-anthophyllite rocks associated with VMSD (e.g. Franklin et al. 1981). Many VMSD have a distinct and discordant alteration pipe associated with stringer ore. In Cu-Zn type VMSD the alteration pipe is often composed of an inner zone enriched in Mg-chlorite and surrounded by a sericite-rich zone (Franklin et al. 1981). Chlorite+quartz is a typical mineral pair in the inner zone, but it is unstable at higher metamorphic grades (>590 °C, P_{H2O}=2.07 kb) and the pair cordierite+anthophyllite can then form instead (Fleming and Fawcett 1976). One example of this type of cordierite-anthophyllite rock is found at the Millenbach deposit where the regional chlorite-dominated mineralogy become transformed to cordierite-anthophyllite rocks within the thermal aureole of granites (Knuckey et al. 1982). The occurrences of cordierite-anthophyllite rocks associated with VMSD in Finland are also interpreted to have originated as

hydrothermally altered volcanic rocks (Latvalahti 1979, Helovuori 1979, Treloar et al. 1981, Puustjärvi 1981, Makkonen 1981 and Smith et al. 1992). Cordierite-hypersthene rocks in the West Uusimaa Complex are similar to cordierite-anthophyllite rocks in the Orijärvi area and thus represent higher grade variants of similar rocks (Schreurs and Westra 1985).

A sedimentary origin for cordierite-anthophyllite rocks has also been postulated (e.g. Kamineni 1979, Reinhardt 1987). Interlayered calcareous, aluminous and magnesian rocks of the Rosebud Syncline in Queensland, Australia has been interpreted as metamorphosed carbonate-evaporite-pelite sequence wherein magnesian cordierite-anthophyllite rocks are derived from evaporitic magnesian clay minerals (Reinhardt 1987). Weathering of mafic volcanic material and redeposition during warm saline conditions can also favour the development of chlorite-rich assemblages that can be further metamorphosed to cordierite-orthoamphibole, cordierite-orthopyroxene and cordierite-phlogopite gneisses (Moore and Waters 1990). Hydrothermally altered rocks can also form detritus that is transported to some distance either as a fine suspension in a brine or as a sediment gravity flow (Goetz and Froese 1982).

VMSD Cu-Zn deposits show variable types of alteration in which the pipe alteration associated with stringer ore is the most prominent feature. A lower semi-conformable alteration zone is also a very prominent feature in some VMSD (Franklin et al. 1981). Walford and Franklin (1982) describe an alteration pipe from Anderson Lake mine that extends to such a lower semi-conformable alteration zone. This alteration zone includes a lower temperature distal part where Fe was added and Na lost, and an intermediate temperature zone, where Fe and Mg were added and Na lost. A higher temperature zone with enrichment of Mg and Na occurs at the intersection with the alteration pipe. Parry and Hutchinson (1981) describe similar changes that have overprinted earlier regional Na-metasomatism (spilite). Lagerblad and Gorbatschev (1985) describe also extensive alteration in Bergslagen, where Na and Mg are enriched and ore-related elements are depleted. Strata-bound exhalative occurrences of cordierite-anthophyllite rocks associated with Cu-Zn deposits have been reviewed by Beeson (1988). They often occur in association with siliceous chemical sediments and are interpreted to have formed due to hydrothermal alteration of tuffaceous material on the sea floor, or by the chemical precipitation of Mg-rich, Ca-poor minerals. These rocks are considered as stratigraphical marker horizons for Cu-Zn mineralization.

The GCO rocks and gneisses in Pukkiharju area (L3-7) reflect a stratabound nature and can occur as distinct layers with amphibolite layers. These amphibolites show geochemical similarities with mafic volcanics and have a composition close to their expected primary composition. Some of these may be of sill-like origin, but the occurrence of heterogeneous and banded variants indicates a volcanic origin for at least some of these rocks. Samples from homogeneous amphibolites have been taken from their central parts since the occurrence of garnet-rich margins in some amphibolites indicates that a change in composition may have taken place near the margins. GCO rocks and gneisses in the Pukkiharju area show high FeO and MgO and low CaO. Zirconium, TiO_2, SiO_2, Al_2O_3 levels and Zr/TiO_2 ratios are mostly comparable with both felsic and mafic volcanic rocks. Felsic protoliths dominate over mafic and only a few rocks of mixed or sedimentary protolith are found. Titanium is generally considered immobile, but small enrichments of Ti have been reported, for example, in some samples at the Anderson Lake alteration pipe (Walford and Franklin 1982). On the other hand, some samples of this study have low Ti with high Fe, Mg and extremely low CaO, indicating its relative immobility at least in these samples. It is often considered that REE behave as immobile elements, but exceptions are also found (e.g. Humpris 1984). In this case however, REE

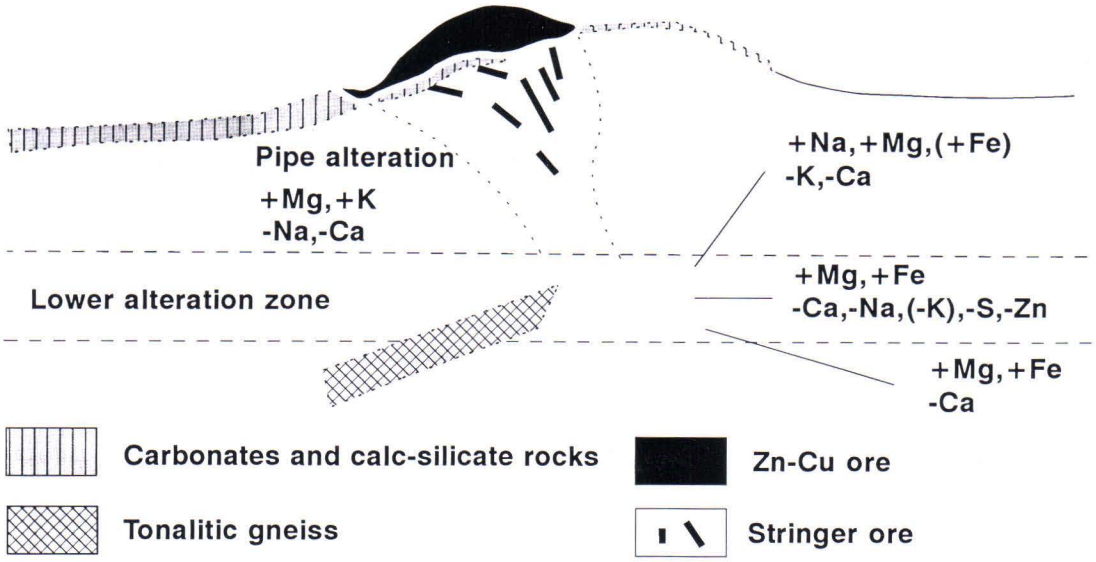

Fig. 20. Hypothetical model for the evolution of alteration zones associated with the Pukkiharju Zn-Cu mineralization. Note that no massive sulphide ore body has been found yet.

patterns are comparable with those of inferred protoliths and indicate relatively immobile behaviour of REE. The main exceptions are the low Eu abundances accompanying low Ca and Na values, that reveal the leaching of Eu during feldspar destruction.

The chemical composition of GCO rocks and gneisses in Pukkiharju area favour an altered volcanic origin. The small proportion of rocks with mixed or sedimentary protoliths do not favour a redeposited origin. Low K could indicate a partly restitic origin for these rocks, with loss of granitic melt during metamorphism. The similar level of REE's and especially LREE do not favour the loss of felsic melt, which is normally enriched in LREE. Sample R368 (Fig. 14) inferred to be of sedimentary or mixed origin has lower abundances of LREE, but at the same time it has high K when compared to other samples. Potassium is normally also enriched in the first forming melts and so the LREE depleted nature in this sample is more probably of primary origin. Although some loss, gain or redistribution of mobile elements such as K and Rb is probable during metamorphism, their generally low K nature is considered to be premetamorphic in origin.

Crosscutting feeder and associated alteration pipe relationships at Pukkiharju can be reoriented into a stratabound position by the partial transposition of a feeder zone into the plane of gneissosity (Friesen et al. 1982). The absence of sulphides, apart from at the margins, and distinct layering do not suggest pipe geometry and an origin comparable with lower semi-conformable alteration zone is proposed instead. Profiles L3-5 seem to be from the same association, which is 50-100 m in thickness. Profiles L6-7 are not directly comparable with L3-5 and lack of outcrops hinders the interpretation. These two associations may represent the same folded horizon along strike or alternatively, the same association with L6-7 higher in the stratigraphy. In the latter case the thickness of this association would be about 400 m. Overlapping zoning has been noticed in the vertical direction, but the lack of full sequences precluded the study of horizontal differences. A

hypothetical model for the origin of GCO rocks and gneisses in Pukkiharju area and their relation to cordierite-sillimanite gneisses and mineralized zone is shown in Fig. 20 and proposed stages are discussed below. At present, no massive ore has been found but Lahtinen (1989) proposed a tentative model where the subvertical alteration zone has been transposed into subparallel position with detachment from massive orebody.

During the first stage, there was a loss of Ca and K, accompanied by an increase in Na and slight enrichment of Mg and Fe. The fluid involved would probably have been modified sea water. A convective cell would have formed at this stage either due to magmatic activity (felsic sills) or due to high heat flow from deeper magma chambers. This alteration was confined to permeable zones, possibly to pyroclastic and tuffaceous rocks, and massive lavas and stocks would show only slight alteration effects. Calcium was deposited as carbonates in porous volcanics near to the discharge vent and on the sea bottom. Continuous redeposition of volcanic material and volcanic activity has formed rocks with thin to massive carbonate layers now seen as calc-silicate rocks and impure carbonate rocks. Carbonate cement in volcanic rocks would have formed an impermeable zone "cap rock" and fluid flow would be confined to discharge vent. Increase in temperature and fluid-rock interactions modified the fluid, now confined to the low alteration zone, and more severe destruction of plagioclase and biotite occurred. Sodium is not depleted in the margins, but there is an increased loss of Ca and increase in Mg and Fe. Continuous leaching in the inner zone locally caused total destruction of plagioclase and biotite, as recorded by extremely low Ca, Na and K values and high Mg and Fe. Chlorite+quartz would have been the major alteration minerals, presently seen as a cordierite+anthophyllite/hypersthene assemblage. Garnet occurs in some samples with higher Fe/Mg ratios. Sulphur, Zn and Cu also seem to have been leached also during this stage, but some higher Zn in silicate phases are locally found. The pipe alteration and sulphide precipation in the discharge vent has produced sulphides and chlorite-sericite alteration seen now as cordierite-sillimanite gneisses. More severe chlorite-type alteration is either yet not found or was inhibited due to the carbonate-rich environment. This proposed hypothetical model is based on models proposed for Cu-Zn type VMSD (Franklin et al. 1981, Walford and Franklin 1982). Lack of complete trace element data makes it impossible to perform mass balance calculations and to take into account the possible mass and volume changes.

GCO rocks and gneisses in the Toholampi area (L1-2) differ from those in Pukkiharju area. Although they also occur as distinct layers and form a stratabound association, they differ in general occurrence of sulphides and high abundances of rocks with inferred sedimentary protoliths. This association can be divided to two groups. Cordierite+hypersthene/anthophyllite rocks (group 1) are almost devoid of plagioclase and have high Mg, Fe and low Ca, Na, Sr. Potassium shows variable, but rather high values. Both mafic and sedimentary protoliths have been found. Sedimentary protoliths show high Ti and Al, but high Zr and REE patterns are more compatible with a felsic origin. La/Yb ratios of felsic volcanics (samples R550 and R29 in Table 2) are 3.1 and only slightly higher than those found in typical mafic volcanics (1.5-2.2). Mixing of mafic volcanic material with felsic material would only slightly depress REE abundances, but would not change the pattern to any great extent. High Ti and Zr accounts for mass loss, if these samples are considered to have mixed sedimentary origin. Titanium can be also a mobile element, but high Al would require it to have been mobile at the same time. The occurrence of zircons as both small inclusions and as coarser grains, as well as also elevated P in some samples, favour a sedimentary origin. Clear coarse grained zircons may also be younger zircons crystallized during granulite

facies metamorphism. Although, there is some uncertainty concerning the origin of these rocks, a mixed sedimentary origin is proposed.

Gneisses that contain substantial amounts of plagioclase (group 2) show elevated CaO, Na_2O and variable but normally low K_2O, while FeO values are high and always enriched over MgO; this is also seen in the abundant occurrence of garnet. All three protolith types are found. Felsic protoliths often have elevated TiO_2 (>0.5%), suggesting a tuffaceous origin. Sulphur normally shows higher values when compared with Pukkiharju association. High S associated with high Cu and Zn are found, but Zn is also high in some samples having low sulphide abundances. These two profiles occur about 2 km apart, but the lack of complete sections hinders interpretation. Profile L2, which occurs to the northwest contains more rocks with felsic tuffaceous protoliths and rocks devoid of plagioclase are lacking. The highest sulphide abundances in L1 occur in plagioclase-free rocks of sedimentary origin (group 1) while in L2 they are in plagioclase-bearing rocks of felsic tuffaceous origin (group 2). Sulphide-rich (S 2.3-3.4%, Cu 360-960 ppm, Zn 110-220 ppm) group 1 samples in L1 show high Cu/Zn ratios of (2.3-6.7). Group 2 samples in L1 with elevated S (0.3-1%, Cu 75-230 ppm, Zn 160-240 ppm) show lower Cu/Zn ratios (0.5-0.9, normally 0.7-0.9). Sulphide-bearing (S 0.7-2.1%, Cu 43-117 ppm, Zn 200-970 ppm) group 2 samples in L2 show low Cu/Zn values (0.1-0.6, normally 0.1-0.2). L1 samples have higher Cu/Zn ratios than those found in L2, a difference also expressed with respect to group 2 samples. A substantial amount of Zn (40-120 ppm) is found in felsic and mafic rocks where it occurs in a silicate phase. The above differences are mainly due to differences in the sulphide phase so that the subtraction of Zn bound in silicates would not change the overall picture. A pyrrhotite-sphalerite vein has been found from a similar association 1.5 km east-southeast from L1 (Marttila 1976b, unpublished report). The vein-like nature probably indicates tectonic mobilization of sulphides. Another occurrence with disseminated sphalerite in a probably similar association has been found about 5 km east-eastsouth from L1 (ibid.).

The proposed model for the origin of GCO rocks and gneisses in the Toholampi area (L1-2) is a stratabound horizon distal to discharge vent. Continuous sedimentation of detritus composed of hydrothermally altered felsic and mafic rocks associated with precipitation of Fe- and Mg-rich minerals and sulphides from brine, alternating with higher rates of sedimentation of volcanic tuffaceous material, could account for the differences between L1 and L2. Less altered rocks with more obvious felsic or mafic volcanic protoliths may have been altered during the flow of hydrothermal fluid on the sea bottom. Amphibolites that have near primary compositions are either lavas or dyke rocks. In this model L1 would have been nearer to the discharge vent than L2. High Ti and Zr in some samples of sedimentogenic or mixed volcanic origin indicate mass loss either due to mass loss in the source area (altered rocks) or due to sedimentary processes.

The Konnekoski association mainly contains samples with sedimentary protoliths and has similarities with plagioclase-bearing group 2 samples in the Toholampi area. The main differences are the low Zn values (normally < 60 ppm) combined with low S (normally < 0.2%). Three samples (S 0.2-0.5%) show elevated Cu (140-200 ppm) associated with low Zn (34-49 ppm). The predominance of Fe over Mg is reflected in the abundant occurrence of garnet. High Ca with high Fe, and also high P in some samples, could indicate a mixed exhalative-sedimentary origin, possibly associated with detritus enriched in altered material. Low Zn values are enigmatic and an altered origin is also possible. More data are needed to verify these possibilities, but overall a sedimentary origin is proposed.

GEOCHEMISTRY OF GRANITOIDS AND FELSIC GNEISSES IN REFERENCE AREAS

Kettuperä gneiss

The Kettuperä tonalitic to granodioritic gneiss (metavolcanic) is heterogeneous lithology with large variations in composition (Fig. 21 and Table 6). The relatively high TiO_2 and low Al_2O_3 and Sr contents, as well as the high FeO/MgO ratio are closer to the Rastinpää-type than to typical synkinematic tonalites. The main differences with respect to the Rastinpää-type are however, lower FeO, MgO, CaO, Sc, V and higher Na_2O, Y, Zr, Ta, Th and U contents. The LREE abundances are also higher (Figs. 4 and 22). The felsic variants of the Kettuperä gneiss have the same chemical composition as the felsic volcanics surrounding Pyhäsalmi orebody (Mäki 1986, Table 1), although the latter have somewhat lower levels of Zr and LREE.

Kangasjärvi gneissic granodiorite

The Kangasjärvi gneissic granodiorite is mylonitic and shows variable K_2O abundances (Fig. 21 and Table 6). The original composition was probably tonalitic/trondhjemitic. The chemical composition resembles the Rastinpää high-Y group having only slightly higher Ta, Th and U contents. The REE level is also slightly higher and the negative Eu-anomaly weaker than in the Rastinpää high-Y group (Figs. 4 and 22).

Leväniemi gneiss and some orthogneisses in the Pielavesi area

The Leväniemi gneiss is characterized by low contents of K_2O, but otherwise it resembles the Rastinpää high-Y group expect for slightly higher Na_2O, Sc and possibly Ta contents (Fig. 21 and Table 6). The REE distribution is characterized by a low La/Yb ratio and a strong negative Eu-anomaly (Fig. 22). Four analyses from orthogneiss domes in the Pielavesi area (samples 1,2,4,6 in Appendix 1, Ekdahl 1993) have been also included in this study. Three samples have geochemical compositions comparable with Rastinpää gneissic tonalites. Samples 4 and 6 are comparable with high-Y group and sample 2 with low-Y group. One orthogneiss (sample 1) differs from other samples in having lower TiO_2, FeO and higher Na_2O and Al_2O_3. High Sr (380 ppm) and Zr (580 ppm) are also significant features that indicate differences in origin.

Venetpalo area

There are only two samples from this area. The gneissic granodiorite (sample V95 in Table 6) is geochemically similar to the Rastinpää high-Y group and Kangasjärvi gneissic tonalites (Figs. 21-22). Plagioclase porphyritic gneiss differs from these in chemical composition and resembles typical synkinematic tonalites, having only slightly higher contents of

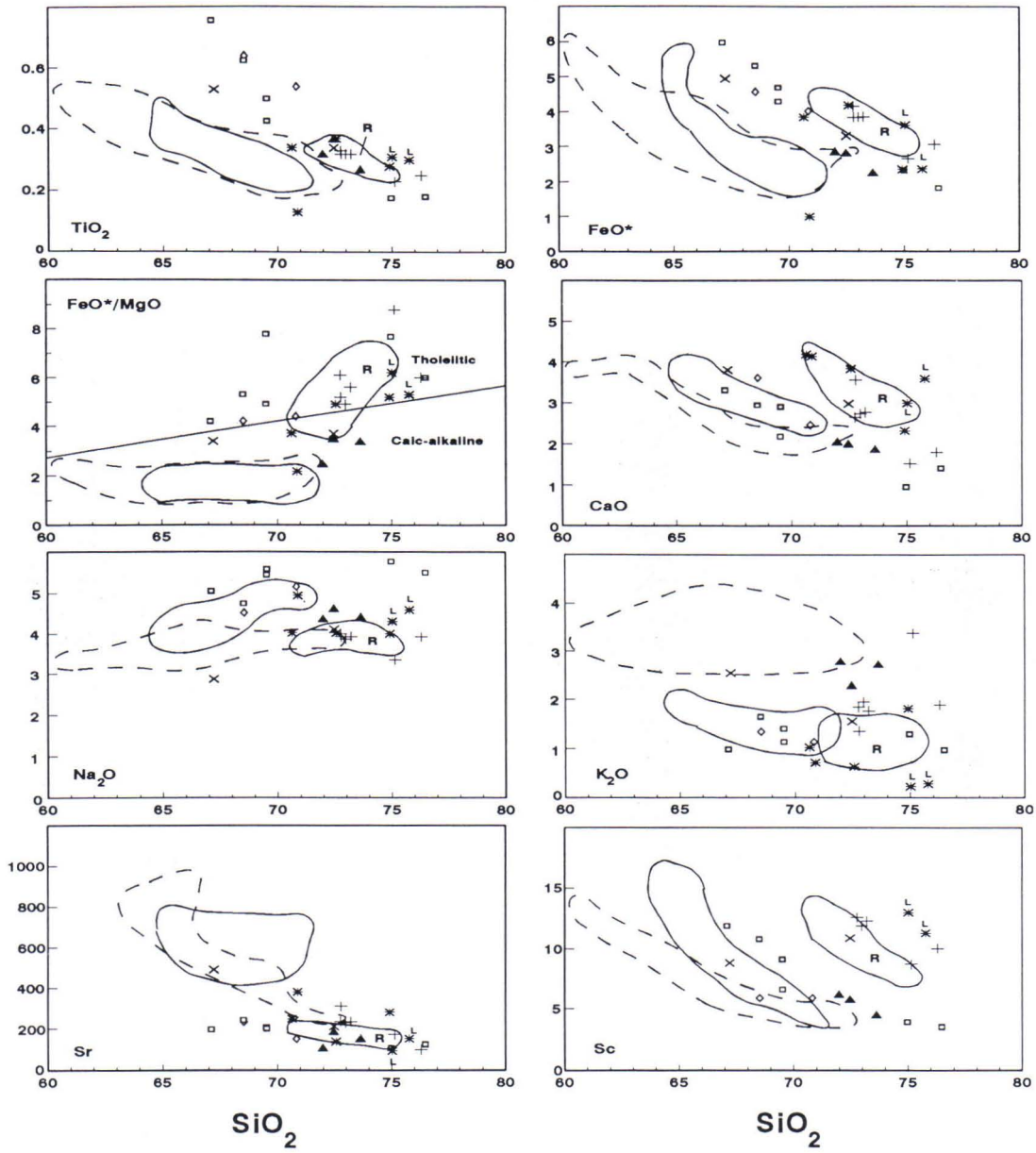

Fig. 21. Harker-type selected major element (%) and trace element (ppm) variation diagrams for gneissic tonalites/granodiorites and gneisses in reference areas. Oxide totals were recalculated to 100 %. Data from this study except for syntectonic granitoids (Nurmi 1984) and for orthogneisses in the Pielavesi area (samples 1,2,4,6 in Appendix 1, Ekdahl 1993). Solid line with R - main field of Rastinpää-type gneissic tonalites and related rocks; solid line without R - main field of the median contents of syntectonic tonalites; dashed line - main field of the median contents of syntectonic granodiorites; crosses - Kangasjärvi gneissic granodiorite; asterixes - orthogneisses in the Pielavesi area and Leväniemi gneiss (L); open squares - Kettuperä gneiss (volcanic); oblique crosses - Venetpalo area; diamonds - Saunakangas tonalite/granodiorite; filled triangles - Veteli granodiorite.

Table 6. Selected major (anhydrous) and trace element data for gneissic tonalites and gneisses from reference areas.

Sample	K246	R249	L3	L23	P66	P69	V94
Type[a]	a	a	b	b	c	c	d
SiO_2	72.81	76.29	75.02	75.77	76.53	67.04	67.22
TiO_2	0.32	0.25	0.31	0.30	0.18	0.76	0.53
Al_2O_3	13.30	12.17	12.86	12.59	13.25	15.09	16.39
FeO*	3.83	3.05	3.60	2.35	1.81	5.97	4.93
MnO	0.07	0.05	0.05	0.03	0.03	0.12	0.09
MgO	0.74	0.51	0.58	0.44	0.30	1.43	1.46
CaO	3.56	1.80	2.99	3.59	1.41	3.31	3.80
Na_2O	3.92	3.92	4.31	4.59	5.50	5.05	2.87
K_2O	1.35	1.89	0.23	0.28	0.97	0.98	2.54
P_2O_5	0.10	0.07	0.05	0.06	0.02	0.25	0.17
Ba	461	304	161	511	448	242	608
Sc	12.6	10.0	13.0	11.3	3.5	11.9	8.8
Sr	310	98	93	152	124	197	491
V	26	16	-	-	7	35	62
Y	31.4	29.1	-	-	33.9	42.4	16.0
Zr	159	144	207	185	210	476	191
Rb	18.4	27.0	-	-	15.0	20.0	98.0
Ta	0.58	0.61	0.42	0.58	1.24	1.32	0.71
Th	2.8	3.3	0.8	1.2	7.6	3.7	8.2
U	1.22	1.34	0.29	0.64	2.70	1.69	1.69
Hf	3.7	3.3	-	-	7.5	9.3	3.8
La	27.0	19.2	11.9	14.6	46.0	31.0	40.0
Ce	51.0	36.0	24.6	32.0	78.0	63.0	53.0
Nd	27.0	22.0	20.4	19.0	36.0	34.0	26.0
Sm	5.6	4.7	4.5	6.2	6.9	6.4	3.6
Eu	1.15	1.05	0.80	0.95	1.22	1.85	0.91
Tb	0.73	0.72	0.61	0.76	0.78	0.99	0.34
Yb	3.6	3.6	3.5	3.3	3.2	4.4	1.2
Lu	0.45	0.43	0.52	0.43	0.36	0.50	0.18
[b]A/CNK	0.92	1.03	1.00	0.87	1.05	0.98	1.14
$(La/Yb)_{CH}$	5.1	3.6	2.3	3.0	9.7	4.8	22.5
[c]Eu/Eu*	0.65	0.68	0.56	0.48	0.57	0.87	0.84

Note.- Major elements (wt%) and Ba,Sc,Sr,V,Y and Zr analyzed at the GSF Laboratory by ICP-AES; other trace elements and REE by neutron activation at the Technical Research Centre of Finland. FeO*: total Fe as FeO.
[a]a- Kangasjärvi gneissic granodiorite; b- Leväniemi gneiss; c- Kettuperä gneiss (volcanic); d- Venetpalo porphyritic felsic gneiss; e- Venetpalo gneissic granodiorite; f- Saunakangas tonalite/granodiorite; g- Veteli granodiorite; h- Veteli microgranitoid enclave.
[b]A/CNK = Molecular ratio of $Al_2O_3/(CaO+Na_2O+K_2O)$.
[c]Eu/Eu* = Observed Eu value/value obtained by interpolation between Sm and Tb.

FeO, TiO_2 and Rb. It has higher LREE abundances and a higher La/Yb ratio than the gneissic granodiorite (Fig. 22).

Veteli granodiorite

The Veteli granodiorite is rich in K_2O and Rb compared to the Rastinpää-type gneissic tonal-

Table 6 (continued).

Sample Type[a]	V95 e	S667 f	S668 f	Ve90 g	Ve91 g	Ve92 g	Ve93 h
SiO_2	72.47	68.53	70.82	73.63	71.90	72.46	60.16
TiO_2	0.34	0.64	0.54	0.27	0.32	0.37	0.91
Al_2O_3	14.22	15.46	14.74	14.03	14.45	14.50	16.31
FeO*	3.29	4.55	4.00	2.25	2.85	2.81	6.90
MnO	0.08	0.07	0.07	0.07	0.06	0.05	0.15
MgO	0.90	1.09	0.91	0.66	1.14	0.81	2.83
CaO	2.98	3.61	2.46	1.88	2.06	2.01	4.63
Na_2O	4.11	4.52	5.16	4.42	4.38	4.63	4.57
K_2O	1.55	1.34	1.13	2.73	2.79	2.29	3.27
P_2O_5	0.06	0.19	0.17	0.06	0.05	0.07	0.27
Ba	633	310	206	461	450	414	561
Sc	10.9	5.9	5.9	4.5	6.2	5.8	16.3
Sr	210	232	151	153	109	187	302
V	39	35	22	17	31	27	134
Y	22.1	33.7	42.4	16.5	17.9	21.1	17.7
Zr	135	437	324	185	199	248	129
Rb	29.0	41.0	-	49.0	44.0	43.0	61.0
Ta	0.56	3.10	-	0.87	0.81	0.88	0.56
Th	4.7	7.7	-	4.0	3.7	3.5	2.3
U	0.84	2.60	-	1.68	1.66	1.69	1.21
Hf	3.8	10.8	-	5.2	5.0	5.7	3.3
La	21.0	80.0	-	22.0	23.0	24.0	17.5
Ce	30.0	120.0	-	35.0	39.0	40.0	30.0
Nd	14.8	43.0	-	13.6	19.7	22.0	14.0
Sm	3.6	7.0	-	3.2	3.3	4.0	3.9
Eu	0.76	1.68	-	0.70	0.70	0.84	1.17
Tb	0.44	0.86	-	0.37	0.36	0.45	0.48
Yb	2.5	3.5	-	1.7	1.5	2.1	1.5
Lu	0.33	0.40	-	0.23	0.22	0.28	0.17
[b]A/CNK	1.03	1.00	1.04	1.03	1.03	1.05	0.83
$(La/Yb)_{CH}$	5.7	15.4	-	8.7	10.3	7.7	7.9
[c]Eu/Eu*	0.67	0.77	-	0.71	0.69	0.68	0.95

ites. Also Ta and U contents are higher and FeO, CaO and Sc contents are lower, and comparable with the Kettuperä gneiss (Fig. 21 and Table 6). The REE pattern has a lower level but otherwise its form is comparable to the pattern for the Kettuperä gneiss (Fig. 22). The microgranitoid enclave corresponds compositionally to high K andesite (Table 6).

Saunakangas tonalite/granodiorite

Only two samples from the Saunakangas tonalite/granodiorite have been analyzed (Table 6). They are lower in Ba and higher in Ta, Th and U than the Kettuperä gneiss, but otherwise they follow the Kettuperä trend (Fig. 21). Saunakangas has also higher levels of LREE and a higher La/Yb ratio than the Kettuperä gneiss (Fig. 22).

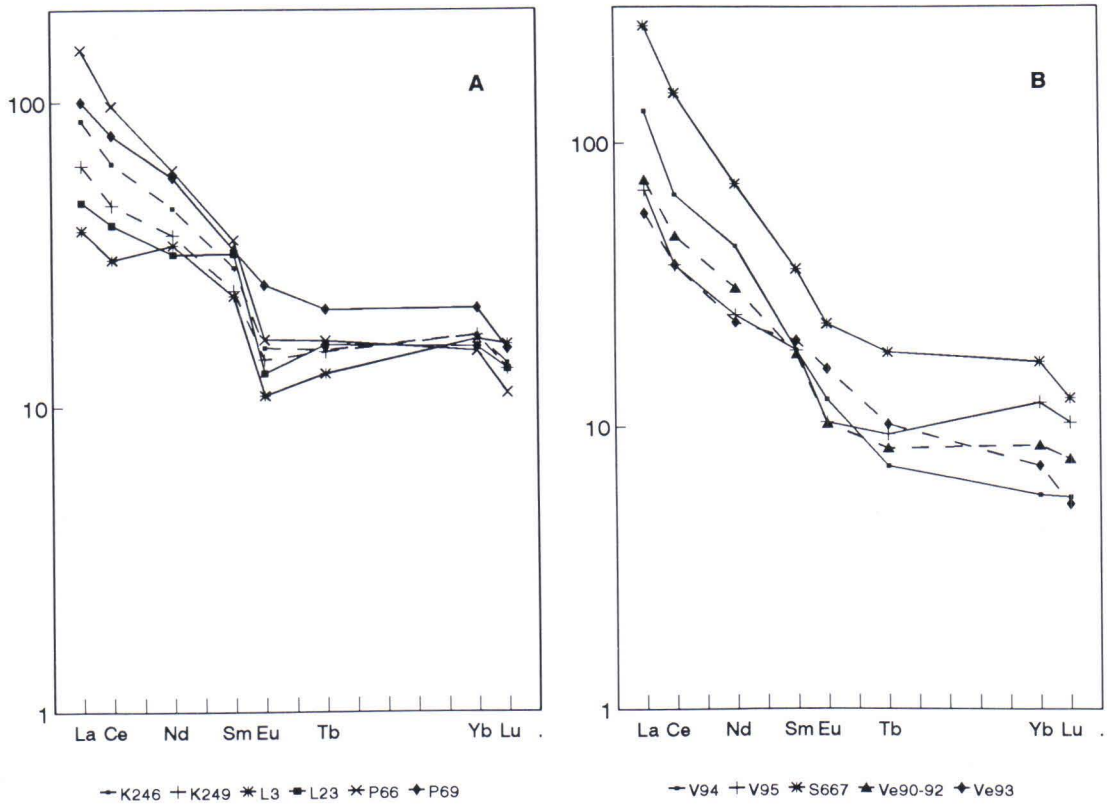

Fig. 22. Chondrite normalized rare-earth element patterns of gneissic tonalites/granodiorites and gneisses in reference areas. Sample numbers refer to samples in Table 6.

ORIGIN OF THE 1.93-1.91 GA GNEISSIC TONALITES AND OTHER GRANITOIDS

The main sites of terrestrial magmatism are at destructive plate margins and in continental within-plate environment. The main contribution to primitive island arc magmatism comes from the depleted mantle derived melts, combined with subduction related components derived from subducted sediments and altered MORB (e.g. Perfit et al. 1980, Gill 1981, Hawkesworth et al. 1991). The proportion of subduction component present and especially its trace element composition is difficult to evaluate and is probably best estimated in situations where the mantle is highly depleted in certain trace elements (Hawkesworth and Ellam 1989). The evolution to more mature island arcs and thicker crust produces more silica-rich rocks and the calc-alkaline to shoshonitic trends due to differences in fractional crystallization trends and/or melting/assimilation of pre-existing rocks (eg. Gill 1981). In continental margin settings the melting of pre-existing crust, including possibly older and partly LIL-depleted granulite, produces more evolved rocks with mixed components. The mantle component can also be heterogenous and the contribution of enriched subcontinental lithos-

pheric mantle (EM) to the genesis of destructive margin rocks is important, especially when magmatism migrates inland to areas under continental crust. The lithospheric mantle component, which was most likely produced in some past subduction episode, is also postulated to occur in some island arc suites (Hawkesworth et al. 1991). For example a mixed origin is proposed for Tecuya rhyolites with assimilation of continental crust by MORB-related magmas and subcontinental lithosphere-derived melts during mid-oceanic ridge-induced magmatism along a continental margin (Sharma et al. 1991).

In a continental within-plate environment the possible mafic components include melts from depleted mantle, subcontinental lithospheric mantle, sublithospheric mantle or combinations of these (see e.g. Zindler and Hart 1986, Carlson 1991 and Zartman et al. 1991). At the same time there can be fusion within the crust of accreted island arc protoliths, continental-crust protoliths, depleted granulite protolith and also remelting of refractory crust.

Different granitoid types have been distinguished on the basis of composition and geotectonic occurrence (see Pitcher 1987, 1993, Leake 1990 for review), and geochemical and petrological characteristics (e.g. Chappell and White 1974, Pearce et al. 1984a, Brown et al. 1984, Whalen et al. 1987, Castro et al. 1991). The importance of depleted mantle, subcontinental lithospheric mantle, melting of different sources in different crust-types, fractional crystallization, assimilation/contamination and mixing of diversity of magmas in granitoid genesis have been discussed in numerous studies (e.g. De Paolo 1981, Rogers and Hawkesworth 1989, Holden et al. 1991, Zorpi et al. 1991 and references above). Experimental petrological studies do not favour the generation of significant quantities of granitic or trondhjemitic magmas by partial melting of mantle (summarized by Wyllie 1984), so the mantle signatures found in many granitoids are attributed to crystal fractionation of basic mantle-derived magmas, melting of newly mantle derived source or mixing between mantle derived melts with crustal source. On the other hand, there is a need for crustal derived component to account for all the geochemical and isotopic signatures found in granitoids, especially with respect to evolved granitoids. There is also an apparent space problem, in the case of voluminous great granite batholiths that do not appear to have incorporated crustal material from either upper, middle or lower crust (Leake 1990).

The origin of granitoids has been attributed to differentiation, partial melting, restite unmixing, magma mixing, unmixing of liquids and metasomatic processes. Fractional crystallization (not necessarily crystal settling) has played a role in the chemical variation of many zoned granitoid plutons. Also in some situations when there is only minor amount of silicic rocks combined with a great proportion of mafic and intermediate rocks, crystallization-differentiation can explain the origin of silicic rocks. In most cases there is a chemical gap between basaltic melts and silicic melts, which militates against the possibility that granitoid melts are derived directly from basaltic melts by fractional crystallization. White and Chappell (1977) proposed the restite unmixing model to explain chemical variation in certain granitoid suites. According to their model, the linear trend on Harker variation diagrams is the result of mechanical separation, or unmixing of SiO_2-poor restite and SiO_2-rich melt fractions. Also crucial to their model is the abundant occurrence of mafic enclaves, which are more abundant in the mafic members of granitoid suites. In many cases the mafic enclaves of granitoids are microgranitoid enclaves and are considered to represent the mingling of mafic and silicic magmas (e.g. Reid et al. 1983, Furman and Spera 1985, Castro et al. 1991). Mixing should also cause linear variation in major and trace element Harker diagrams, although the patterns can be complicated by the mixing of more than two components. The mixing model implies the presence of silicic magma

that has originated by melting. Unmixing of liquids (e.g. Hildreth 1981) can produce great chemical variation during magma evolution, but silicic magmas originate by partial melting. Silicate liquid immiscibility is also one possible mechanism (Roedder 1979). Alkali metasomatism has also been invoked for trondhjemite genesis (Drummond et al. 1986). A melting model with different source materials is the most common explanation for the origin of granitoid magmas. However, situations are normally complex and involve multi-staged models, including melting/crystallization with assimilation, homogenization and storage, volatile transport, magma mixing, and magma unmixing.

Granitoid sources vary from primitive mafic underplates producing Ca-rich metaluminous tonalites to mature sediments producing K-rich peraluminous S-type granitoids. Apart from some S-type granites, mafic magmatism is normally associated with granitoid magmatism and different types of hybrid rocks are not uncommon. It is not always clear whether this is due to mechanical mixing and/or true magma mixing but comingling and possibly also magma mixing are ubiquitous features and they are therefore briefly considered here. The inhibition of mixing of mafic and silicic magmas by chilling of the mafic magma has been predicted by thermal physical modelling (Sparks and Marshall 1986). According to such studies, it is only in situations where the proportion of mafic magma is large that complete hybridization is possible or else where mixing involves two evolved magmas. The importance of the mixing ratio is seen in the Abu volcano group, where in the mafic-dominated mixture, basalt and dacite magmas mixed in the liquid state while in the silicic-dominated mixture, the basalt magma was quenched (Koyaguchi 1986). On the other hand, clear chemical and isotopic evidence for chemical modification of mafic enclaves and mass transfer between mafic enclaves and felsic magmas has been presented (Bloomfield and Arculus 1989, Eberz and Nicholls 1990, Holden et al. 1991, Zorpi et al. 1991). The main exchange processes postulated between mafic and felsic components are the mechanical transfer of crystals (mainly from felsic to mafic) and mass transfer of alkali metals, alkali earths and transition metals. The degree of hybridization of mafic and felsic components is induced by convective stirring of the magma chamber and physical disintegration of the mafic component. This is also seen in the results of Rutter and Wyllie (1989), where the assimilation of mafic rocks by granite melt is very slow in the absence of convective heat transfer and/or mechanical disaggregation of mafic inclusions. The stabilization of biotite during interaction between granitic and basic magmas can result in basic material containing more K_2O than coexisting felsic melt (Johnston and Wyllie 1988). The general occurrence of biotite-rich small mafic patches in granitoids could be partly explained this way. Therefore, there is apparently good evidence for mixing/assimilation occurring between mafic and felsic rocks. At the same time there are many occurrences where only mingling has taken place without great chemical modification of components.

Mixing has been invoked to explain the genesis of calc-alkaline rocks (e.g. Koyaguchi 1986), and by depressing the FeO/MgO ratio of these rocks. The tholeiitic parental liquid can possibly also evolve to calc-alkaline liquid line of descent by reacting with ultramafic wall rock, in combination with crystal fractionation due to falling temperatures in subduction-related magmatic arcs (Keleman 1990). Experimental systems of tonalite-peridotite at 15 kb also shows how tonalitic melts with added peridotite have lower FeO/MgO-ratios and Ca/(Mg+Fe)-ratios close to natural calc-alkaline compositions (Carrol and Wyllie 1989).

Petrogenesis of Rastinpää-type gneissic tonalites

As considered before, the restricted range of SiO_2 and the nonlinear chemical variation of some elements in Rastinpää gneissic tonalites are not consistent with simple restite unmixing or magma mixing as probable mechanism for the origin of these tonalites, while basalt fractionation is inappropriate on geological grounds since there are few rocks of intermediate composition and gabbroic intrusions of similar age are lacking. Barker (1979) presented a summary of models for the genesis of high-Al_2O_3 and low-Al_2O_3 trondhjemitic-tonalitic liquids by differentiation and partial melting of a mafic parent. Partial melting is considered a realistic model with a possible source being either amphibolitic or gabbroic in order to produce low-Al_2O_3 tonalitic liquid. The composition of the source is constrained as being approximately basaltic to basalt-andesitic. Potassium is strongly enriched in the first forming melts and, therefore the low-K_2O nature of Rastinpää gneissic tonalite excludes the possibility of a more intermediate and K_2O-rich source.

Tonalitic composition can be produced by hydrous partial fusion of basalts at pressures less than 10 kb (Green and Ringwood 1968, Holloway and Burnham 1972, Helz 1973,1976, Spulber and Rutherford 1983). Partial melting of basaltic material at high pressure (> 20 kb) can also produce melts of broadly tonalitic chemistry (e.g. Green and Ringwood 1968, Stern and Wyllie 1978). High pressure crystallization experiments (hydrous and anhydrous) nevertheless indicate that high silica compositions can not be the direct partial melts of basaltic source compositions at high pressures (Green and Ringwood, 1968, Stern et al.,1975). These experimental studies show that both low pressure gabbroic or amphibolitic and high pressure eclogitic assemblages are potential source candidates for magmas of tonalite composition.

An unfortunate complicating factor in predicting parent magma composition is that the existing plutons are the "residues" of magma chambers and posteruptive changes in composition of the unerupted magma can be extensive. There is some evidence for crystallization-differentiation and the possibility of different magma batches in the genesis of low-K and low-Al_2O_3 tonalites in Rautalampi area. Also the occurrence of comagmatic volcanics is very probable. Although there are variations in the chemical compositions of these rocks, they are still consistently of low-K and low-Al_2O_3 type with broadly similar chemical characteristics. The major element composition of the least fractionated high-Y samples of Rastinpää gneissic tonalite is here compared to recalculated fluid free analysis of experimentally produced high SiO_2 residual liquid (10% melting) from DO8 oceanic tholeiite with gabbroic residue (Spulber and Rutherford 1983); SiO_2 71-71, TiO_2 0.35-0.8, Al_2O_3 14.5-15, FeO 4-4.7, MnO 0.09-0.06, MgO 1.0-0.9, CaO 4-3.4, Na_2O 3.5-3.6, K_2O 0.8-0.4 and P_2O_5 0.1-0.45 respectively. There are differences in TiO_2, K_2O and P_2O_5 contents, but otherwise the chemistries are in excellent agreement. K_2O contents are critically dependent on to percentage of melting and also upon small differences in source rock composition (DO8 0.06 % K_2O). The silica enrichment during partial melting of a tholeiitic basalt at $P_{fluid} < 2$ kb is controlled by the crystallization of Fe-Ti oxides, which depend on the f_{O2} and fluid phase (Spulber and Rutherford 1983). Thus, small differences in starting compositions, f_{O2}, fluid compositions and pressures could explain the variation in K_2O, TiO_2 and possibly P_2O_5. The presence of apatite in the basalt melt experiments by Helz (1976) indicates that the occurrence of apatite in the melt residue controls the concentration of P_2O_5 in the melt. At higher water pressures the residual assemblage is amphibolitic, which can also produce SiO_2 enriched melts at low degrees of basalt melting (Holloway and Burnham 1972).

Table 7. Compositions of parent basalts and calculated model melt compositions.

Type[a]	MORB				IAT			
Residue[c]	o	a	b	c	o	a	b	c
Melt-%		10	18	30		10	18	30
K_2O	0.20	1.10	0.66	0.49	0.43	2.38	1.42	1.06
TiO_2	1.40	0.55	0.55	0.52	0.84	0.33	0.33	0.31
Rb	2	14	11	7	4.7	32	25	16
Ba	20	79	60	48	60	237	179	143
Zr	90	369	116	101	40	164	51	45
Ta	0.29	0.75	0.57	0.47	0.10	0.26	0.20	0.16
Th	0.26	2.12	1.06	0.72	0.37	3.02	1.51	1.02
Sr	121	89	180	247	231	170	344	472
Y	33	55	19	18	17	29	10	9
Sc	41	12	7	7	40	11	6	7
La	3.72	10.6	6.8	5.8	2.48	7.05	4.5	3.9
Ce	11.0	30.5	14.3	12.8	6.94	19.2	9.0	8.1
Sm	3.26	7.4	1.3	1.2	1.89	4.3	0.73	0.72
Eu	1.21	1.1	0.52	0.54	0.71	0.65	0.31	0.31
Tb	0.76	1.4	0.21	0.20	0.45	0.79	0.12	0.12
Yb	3.22	6.1	1.1	1.0	1.95	3.7	0.64	0.63

Note.- K_2O and TiO_2 in % and others in ppm.
[a] MORB- Mid-Ocean Ridge basalts (Pearce 1982); IAT- Island Arc Basalts (Pearce 1982); RT- Median contents of most primitive tholeiitic volcanics from the Rautalampi area.
[b] The distribution coefficients are approximated from the data for andesitic and dacitic rocks (Hanski 1983).
[c] o- the composition of parent basalts cited above; a- gabbroic residue at 1 kb (45 % pl, 35 % cpx, 15 % ol and 5 % il); b- amphibolitic residue at 5 kb (65 % af, 20 % cpx, 12 % pl and 3 % mt); c- amphibolitic residue at 5 kb (77 % af, 20 % cpx and 3 % mt). Data for a from Spulber and Rutherford (1983) and for b and c from Holloway and Burnham (1972).

Their experiments on olivine tholeiite produced, at 4.9 kb (f_{H2O} about 0.6) and 875 °C ,18 per cent of melt with a broadly similar major element composition to that observed in the Rastinpää type gneissic tonalites. The main differences are the slightly lower level of FeO and MgO. From these experimental studies it is clear that a small degree (about 10-15 %) of partial melting of tholeiite can produce melts with major element compositions comparable to the low-K_2O and low-Al_2O_3 tonalites in Rautalampi area.

An attempt was made to match mathematical melting models with the trace element composition of these tonalites. The models considered below produce tonalite melt by 10 and 18 per cent melting of a basaltic source with different trace element compositions. To be consistent with the experimental data cited above, the residual mineral assemblages are as follows: gabbroic: 45 per cent plagioclase; 35 per cent clinopyroxene; 15 per cent olivine; 5 per cent ilmenite; amphibolitic: 65 per cent amphibole; 20 per cent clinopyroxene; 12 per cent plagioclase; 3 per cent magnetite. Melting under equilibrium conditions is assumed, and calculations follow the simple batch melting equation (Shaw 1970). Three basaltic tholeiites with different trace elements, as well as variable K_2O and TiO_2 characteristics have been studied: average MORB, average IAT and the average of the most primitive tholeiitic volcanics in Pukkiharju area. The major element compositions of these types are broadly similar and comparable. The modelling is highly dependent upon the distribution coefficients selected, which are subject to considerable uncertainty. This is particularly crucial in present situation,

Table 7 (continued).

Type[a]	RT				Distribution coefficients[b]					
Residue[c]	o	a	b	c						
Melt-%		10	18	30	pl	cpx	ol	af	mt	il
K_2O	0.34	1.88	1.12	0.84	0.2	0.0	0.0	0.2	0.0	0.0
TiO_2	0.63	0.25	0.25	0.24	0.0	0.5	0.0	4.0	8.0	50
Rb	5	34	27	17	0.1	0.0	0.0	0.0	0.0	0.0
Ba	100	395	298	239	0.3	0.1	0.0	0.2	0.0	0.0
Zr	40	164	51	45	0.1	0.3	0.0	1.0	0.2	0.2
Ta	0.10	0.26	0.20	0.16	0.0	0.2	0.0	0.5	1.0	5.0
Th	0.36	2.94	1.47	0.99	0.0	0.0	0.0	0.1	0.5	0.5
Sr	195	143	290	400	3.0	0.2	0.0	0.3	0.0	0.0
Y	12	20	7	7	0.0	1.5	0.0	2.5	0.5	0.5
Sc	46	13	7	8	0.0	10	0.5	8	5	5
La	3.8	10.8	6.9	6.0	0.20	0.45	0.0	0.50	0.24	0.65
Ce	7.6	21.0	9.9	8.8	0.17	0.51	0.0	0.90	0.28	0.60
Sm	1.7	3.8	0.65	0.65	0.07	0.95	0.0	4.0	0.40	0.34
Eu	0.59	0.54	0.26	0.26	1.9	0.68	0.0	3.4	0.31	0.20
Tb	0.32	0.56	0.09	0.08	0.05	1.4	0.0	6.1	0.37	0.18
Yb	1.28	2.4	0.42	0.41	0.03	1.3	0.0	4.9	0.22	0.27

when the possible melts are of high-silica and low-K nature. Most of the published distribution coefficients for high silica rocks are from high-K rhyolites and rhyodacites, which are not appropriate in this situation. Therefore, the calculated distribution coefficients determined from andesitic and dacitic melts are used instead (Table 7). Many of these distribution coefficients are, especially in the case of ilmenite, still very poorly calibrated.

The results of the model calculations are shown in Table 7 and the REE modelling results in Figure 23 As a comparison, an amphibolitic residue with 30 % of melt has been also calculated. The most meaningful results are of elements with low bulk distribution coefficients (< 0.2) and the shape of the REE-patterns. The level of Rb, Ba, Zr and Th in the Rastinpää high-Y group are in good agreement with gabbroic residue in island arc tholeiite or the Rautalampi volcanic-type source and exclude MORB as a possible source. Strontium has a similar trend. The levels of TiO_2, Ta and to some extent Sc are strongly controlled by the amount of ilmenite and by the distribution coefficients for ilmenite, which can all vary considerably. Also the possible variance in distribution coefficients of Y and Sc for clinopyroxene and amphibole hinders the use of these elements. The tonalitic model melt with a gabbroic residue in the Rautalampi-type volcanic source yields a REE pattern characterized by modest LREE enrichment and a significant negative Eu anomaly. The level is somewhat lower, but otherwise the essential features are in excellent agreement with the REE patterns of the Rastinpää high-Y group tonalites. The amphibolitic residue is not favoured, because of the different level and pattern of the REE. The amphibolitic model would also indicate lower levels of Zr, Th and Y than what was observed.

These results demonstrate that a basaltic source with gabbroic residue produced in experimental studies is capable of producing tonalitic melt with major element compositions, REE patterns and trace element compositions that correspond closely to the Rastinpää high-Y group. The best approximation to such a basaltic source is a low-K island arc tholeiite (K_2O about 0.2 %) with moderate LREE enrich-

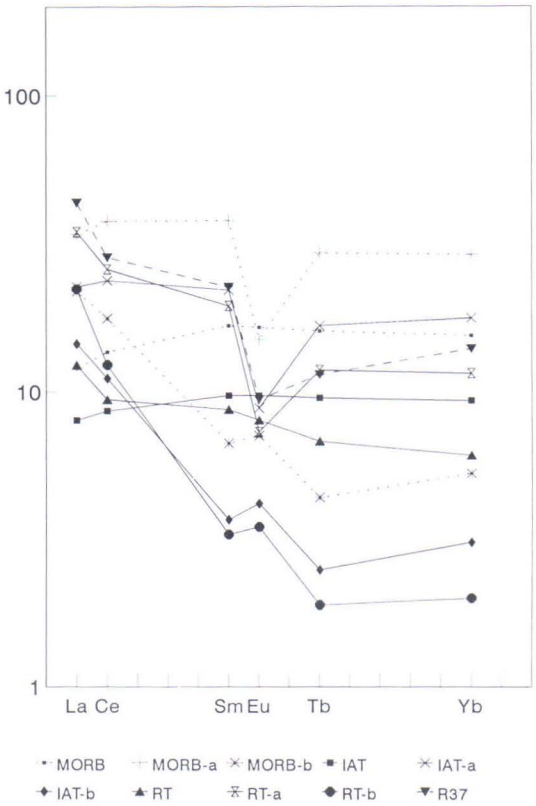

Fig. 23. Chondrite normalized rare-earth element patterns of parent basalts (MORB, IAT, RT) and calculated model melt compositions; MORB-a, IAT-a and RT-a refer to gabbroic residue and, MORB-b, IAT-b and RT-b refer to amphibolitic residue. See Table 7 for data sources and text for explanation. R37 refers to Table 1.

ment. In this modelling the most mafic samples of Rastinpää high-Y group have be assumed to represent the parent melt. This assumption is based on the fact that more mafic parents comparable to these rocks are lacking. The Rastinpää low-Y group has been interpreted as accumulated rocks with trapped inter-crystalline melt and so their possible parent magma is very difficult to evaluate. There is evidence for accumulation of plagioclase and ferromagnesium minerals, but otherwise most of the trace elements are at about the same level than in the high-Y group; this would indicate similar source.

The gneisses of volcanogenic origin resemble the Rastinpää high-Y group and are probably cogenetic with them. The other high silica (> 70 %) gneissic tonalites in the Rautalampi area are difficult to interpret due to small number of samples. Compared to the Rastinpää high-Y group they have somewhat higher Sr and lower abundances of K_2O, Ba, Y, Rb, Ta and Th at equivalent SiO_2 values and often have higher La/Yb ratio with a weak positive Eu anomaly. As considered earlier the lower abundance of some elements may be due to their migration with a vapour phase or a small degree of melting during granulite facies metamorphism, as in the case of mobilized gneisses. Similarities in LREE abundances do not support the idea of melt extraction and many of these samples show no indication of granulite facies mineralogy. High level differentiation, possibly flow differentiation as discussed earlier with respect to the Rastinpää and Toholampi gneissic tonalites, may be responsible for some of these differences. The variations considered above could also be explained by the presence of a small amount of amphibole in the residue (about 10 %), which would depress the level of Tb and Yb as seen in Figure 23, as well as trends in other elements (Table 7). Garnet is not able to account for this since there is no change in Tb/Yb ratio. At the same time, small changes in f_{O2} would drastically decrease the distribution coefficient for Eu. Thus, it is possible to produce melts of low-Y nature from the same basaltic source at higher f_{H2O} and f_{O2}. The heterogenous nature can therefore be attributed to either or both differentiation and release of vapour phase.

On the whole the Rastinpää-type gneissic tonalites and related rocks in the Rautalampi area consistently reflect a rather primitive island arc origin. This also evident in Figure 25, where they clearly plot out of the modern calc-alkaline andesite field and within the more calcic side, indicating more primitive source (Brown 1982). The Ta-Th and Rb-Sr diagrams (Figs. 26 and 28) could be interpreted as indi-

cating differences in melt percentage and source composition, while the Rb-Sr diagram also reflects differences in residue (amount of plagioclase in residue)(see Table 7). The Rastinpää-type gneissic tonalites and related rocks are considered to be the result of a small degree of melting of island arc low-K tholeiitic basalts with a mainly gabbroic residue.

Petrogenesis of other gneissic tonalites and granitoids in the Rautalampi area

The Pukkiharju gneissic tonalite (sample R547 in Table 3) has lower SiO_2 compared to the Rastinpää-type tonalites, but it otherwise resembles them and is probably cogenetic. Younger gneissic tonalites (1.89-1.88 Ga ?) are of the high-Al_2O_3 type and are characterized by high Sr. They have higher CaO-contents (Fig. 5) than normal synkinematic tonalites, indicating a slightly more primitive source, but otherwise they are similar. Samples R259 and R263 have low Tb/Yb-ratios and REE patterns compatible with an amphibole-rich residue after 20-30 percent melting of island arc basalt source (compare Figs. 6 and 23). This would also explain the low abundances of Zr, Th and Ta, but is at variance with the high K_2O and Sr. Altogether, it seems plausible that the source for these rocks would have been more calc-alkaline in nature, with higher abundances of K_2O, Sr and probably also of LREE, compared to the source of the Rastinpää-type tonalites (see also Figs. 26 and 28).

The Särkilampi gneissic tonalites and related gneisses of probable volcanic origin are low in total REE and especially in Yb, and the Tb/Yb-ratio is moderate. The low contents of Ta, Th, U and Zr at such high SiO_2 levels excludes the possibility of these rocks being the products of fractional crystallization of a mafic parent. The fractional crystallization of zircon could explain some of these features, but the zircon saturation level exceeds 100 ppm Zr in most situations (Watson and Harrison 1983). Low degree melting of garnet amphibolite would be in accordance with the observed REE pattern (compare Figs. 6 and 23) and low abundances of Sc, Y, Zr and Hf. The low contents of Th and U indicate a source that has been depleted in these elements. The most probable source is a tholeiitic to calc-alkaline island arc basalt, which has lost Th, U and possibly also some of K_2O, Rb and Ta to a vapour phase during granulite facies metamorphism. The calcic nature seen in Fig. 25 is perhaps slightly modified due to the possible K_2O loss in the source. The Ta-Th and Rb-Sr relations may also have been modified (Figs. 26 and 28) but the high Sr-contents favour the interpretation of a plagioclase-free residue with a Sr-rich source.

Only a few samples from the margin of the pyroxene granitoid complex have been analyzed, so it is impossible to study the relationship between different pyroxene granitoid types and the pyroxene granitoid complex as a whole. The pyroxene quartz diorites and tonalites show enrichment in LREE and high contents of Ba, Sr and K_2O as well as an anomalously low Th/U-ratio (about 1) at SiO_2 values of 56-60% (Table 3 and Fig. 5). These features exclude the possibility of a direct low-K island arc basalt source and points to a more evolved LIL-element enriched origin through differentiation or melting. The Th-Ta-relations suggest a within-plate origin (Fig. 26), but the rather low FeO/MgO-ratio in these rocks does not suggest a strong tholeiitic character (Fig. 5). The low Th-values could be attributed to a source depleted in Th, but the high Rb, K_2O and REE-distribution contradicts this interpretation. The high Zr and medium TiO_2 (Table 3) compared to Th-Ta-relations, slightly alkalic nature of some samples (Fig. 26) and high Sr-contents (Fig. 28) also suggest a within plate origin. Such an origin for these rocks could be envisaged through melting or differentiation of within plate basalts combined with later differ-

entiation and contamination. The pyroxene quartz leuco monzodiorite has very high abundances of Ba, Zr and LREE's (Table 3), as well as high Th-contents. The high Zr is consistent with a fluorine-bearing magma and possibly also with a within plate component.

The gneissic quartz diorites and granodiorites in the Rautalampi area are heterogeneous and the mafic rocks (< 60 %) have high K_2O contents and high FeO/MgO-ratios (Fig. 5). The more siliceous lithologies are of the synkinematic type. Sample R261 has high contents of TiO_2, P_2O_5, Ba, Sr, Zr and Rb (Table 3). These and the Th-Ta-relation (Fig. 26), slightly alkalic nature (Fig. 25) and Rb-Sr relationships are compatible with a within plate character. The REE-distribution is similar to that of the pyroxene granitoids (Fig. 6). Quartz diorites and granodiorites to the N and NW of Rautalampi have normally higher FeO/MgO-ratios and are more mafic than the median contents of synkinematic tonalites and granodiorites (Fig. 5). The REE data indicates cogenetic origin for the quartz diorites, granodiorites and possibly also for granites. The REE-distribution (Fig. 6) is similar to that of the pyroxene granitoids, but there are no trace element data available to verify the similarity.

Petrogenesis of reference areas

The Kangasjärvi gneissic granodiorite, the Leväniemi gneiss and most of the other orthogneisses in the Pielavesi area, and the Venetpalo gneissic granodiorite all have the same geochemical characteristics as the Rastinpää-type gneissic tonalites at Rautalampi. They are therefore considered to have had the same origin by melting of a low-K_2O island arc tholeiitic basalt source. The low K_2O-contents (0.2-0.3 %) of the Leväniemi gneiss and possibly also low Th (Table 6) are nevertheless enigmatic. The occurrence of these samples stratigraphically below and rather near the mineralized zone could indicate the existence of the lower conformable alteration zone with Na-alteration as proposed for Pukkiharju Zn-Cu-mineralization (this study). Altogether, it is not clear whether these samples are from plutonic or homogenized supracrustal rocks. The Venetpalo plagioclase porphyritic gneiss resembles synkinematic granodiorites and tonalites in having rather high K_2O and Rb (Table 4) but the higher Ca indicates a more primitive source (e.g. Fig. 25). This and the REE-distribution (Fig. 22) and Th-Ta-relation (Fig. 26) rather indicate a calc-alkaline origin through melting or differentiation. The Kettuperä gneiss is of low-Al_2O_3 type with low Sr abundances (Fig. 28). These features, together with the high REE and low Tb/Yb ratio (Fig. 22) point to a gabbroic residue with possibly a small amount of amphibole. The high abundances of Zr, Ta, Th, U, LREE and Th-Ta-relation (Fig. 26) in the most mafic sample also indicate a mafic source with a more enriched nature compared to normal island arc tholeiite. Only two samples of Saunakangas tonalite/granodiorite have been analyzed, which hinders interpretation, although they do to some extent follow the Kettuperä trend in having higher Zr, Ta, Th, U and LREE (Table 6). In particular the high Ta content (Table 4) should be verified with more samples. The low Sr-contents (Fig. 28), Eu-minimum and low Tb/Yb-ratio compared to the level of these elements indicate a gabbroic residue.

The Veteli granodiorite has abundant microgranitoid enclaves, which could indicate magma mixing. The low Tb/Yb-ratio (Fig. 22) excludes the presence of garnet while low Sr contents (Fig. 28) and the negative Eu anomaly indicates the occurrence of plagioclase in the source residue. The Th-Ta-relations (Fig. 26) and REE-distribution as well as K_2O and Rb-abundances (Table 6) at high SiO_2 (72-74 %) favours an evolved island arc source and a low

degree of melting with gabbroic residue (small amount of amphibole is also permissible). The microgranitoid enclave has probably interacted with the surrounding magma and gained K_2O and Rb, thus changing its composition. Altogether, these features are characteristic of calc-alkaline andesite.

Petrogenesis of synkinematic granitoids

Published median contents of different granitoid phases in the Svecofennian (Nurmi 1984) and also the original data from the same granitoids (available from the Geological Survey of Finland) have been used as reference analyses. The Ta-values of granitoids from Mäntylä, Silvola and Kaartila are excluded due to apparent contamination, deduced from their high W-values. Limited REE data are also available (Nurmi and Haapala 1986). The locations of sampling sites are shown in Figure 1. The tonalites (tonalites and trondhjemites), granodiorites and granites are considered separately and the main fields of the median contents of different granitoid phases are shown in Figures 5 and 21. The tonalites have greater similarity than granodiorites and form a rather homogenous group (Nurmi 1984). In the Harker variation diagrams, the median contents of most tonalites show high-Al_2O_3 (not shown) and high-Sr characteristics with median SiO_2 contents of 65-70 % (Fig. 21). Exceptions are the high-SiO_2 Aulanko tonalite porphyry and the Kaartila tonalite-monzogranite and these are not included in the main fields. These and other deviations from the "typical synkinematic trends", as well as the nature of individual granitoids are discussed briefly below. Also included are the available data for mafic rocks associated with the granitoids.

The Käkövesi pluton is composed of hornblende tonalite and younger granodiorite, which are considered as separate phases (Nurmi et al. 1984, Heinonen 1987). The pluton intrudes garnet-cordierite gneisses, mafic metavolcanics and sillimanite quartzites (ibid.). The main differences between these two phases are the lower level of Al_2O_3 and Sr in tonalite with a compositional gap between it and the granodiorite (Heinonen 1987). These phases are heterogeneous and partly overlap, as can be seen in the Harker-diagrams (see Fig. 14 in Heinonen 1987). The high Al_2O_3 contents of the granodiorite phase are reflected in the high A/CNK values (1.1-1.3) compared to the tonalite (mostly 0.8-1.1). The high A/CNK values for the granodiorite are characteristic of S-type

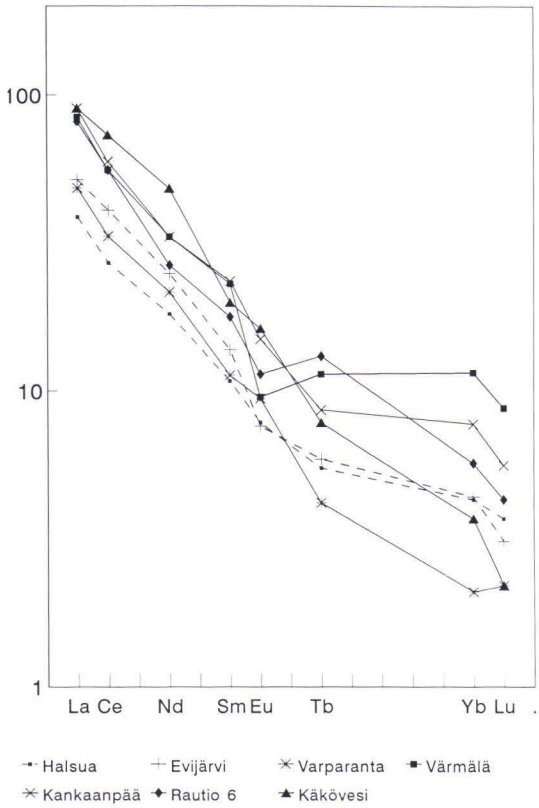

Fig. 24. Chondrite normalized rare-earth element patterns of syntectonic tonalites (dashed lines) and granodiorites (solid lines). Data from Nurmi and Haapala 1986.

granites, but the high Sr contents (800-1200 ppm) and Na_2O-K_2O relations are more typical of I-type intrusions. The surrounding metabasalts have high abundances of K_2O (0.8-1.9 %), P_2O_5 (0.32-0.45 %), Rb (22-91 ppm), Ba (260-500 ppm), Sr (270-650 ppm), La (22-29 ppm), Zr (110-150 ppm), Ta (0.34-0.77 ppm) and Th (3.9-8.4 ppm) (see also Fig. 27). Thus they have calc-alkaline to shoshonitic chemical characteristic which indicates a mature arc environment. The high Tb/Yb-ratio in the REE data for the granodiorite indicates the occurrence of garnet in the residue (Fig. 24). The tonalites are rather mafic rocks (SiO_2 56-64 %) and therefore they would require a higher degree of melting, if a partial melting model is invoked. On the other hand the tonalites also resemble andesites and a cogenetic origin with these is also possible. Both granitoids plot in or near calc-alkaline fields in Figures 25 and 26 and in the high Rb and Sr part of Fig. 28. The heterogeneous nature of the geochemical data points to a complex history for the pluton, with a mixed source combined with varying degrees of assimilation and differentiation. The geochemical data in general indicate a mature arc environment with a genetic link to calc-alkaline/shoshonitic rocks.

The Silvola and Kaartila stocks are surrounded by migmatitic cordierite gneisses, diopside amphibolite, quartz feldspar gneiss, garnet gneiss and metaconglomerate intercalations (Kuusola 1982, Nironen 1989a). According to Nironen (1989a) the Silvola stock is homogenous and composed of tonalites (1886±5 Ma, U-Pb zircon), trondhjemites and granodiorites, and it was emplaced as a diapir during local D_2 deformation. Geochemically it is calcic indicating a more primitive origin compared to modern calc-alkaline arcs and it plots within the S-field in Fig. 25. The low-K (K_2O 0.9-1.2) and high Al_2O_3 (16.8-17.4 %) nature of the most mafic tonalite samples (SiO_2 64-66 %) indicates a low degree of melting of a basaltic source with amphibolitic residue. The FeO (3.4-3.8 %), MgO (1.4-1.6 %), CaO (4.9-5.4 %) and Na_2O (3.6-4.2 %) contents in the most mafic samples are consistent with the experimental data for basalt melting cited previously (p. 63). The Sr (500-600 ppm), Rb (25-35 ppm)(see Fig. 28), La (14-17 ppm), Ba (500-600 ppm), and Th (2.2-3.2 ppm) contents, as well as major elements favour a LREE enriched low-K island arc tholeiitic to calc-alkaline basaltic source.

The Kaartila pluton is generally heterogeneous, except near the centre, with modal compositions ranging from monzogranite to tonalite (Nironen 1989a). The pluton is calcic in nature and concentric, having higher values of K_2O, Ba, Rb and Th in the centre. The heterogeneity of the pluton and abundant mica gneiss and amphibolite xenoliths, as well as relict accessory garnets indicate a mixed crustal source consisting of metagreywackes and amphibolites (Nironen 1989a). The high SiO_2 (71-77 %) and low Al_2O_3 (11.7-13.8 %) and Sr (140-240 ppm) along with high Th (7-19 ppm) and La (42-113 ppm) contents favour a small degree of melting. The low Rb (25-50 ppm) in the tonalites (K_2O 1.2-1.4 %) and also abundances of other elements (e.g. high La and Th) are similar to the Kettuperä and Saunakangas trends and suggest a low degree of melting of slightly evolved island arc rocks with a gabbroic residue. The more evolved rocks are due to assimilation of mica-rich country rocks and/or fractional crystallization.

The Varparanta pluton is a domal structure surrounded by migmatitic cordierite gneisses and diopside amphibolites of volcanic origin (Nurmi et al. 1984, Nironen 1989a). The pluton is composed mainly of tonalite, but trondhjemite, magnetite-rich trondhjemite and quartz diorite banded tonalite are also present. According to Nironen (1989a) wall rock xenoliths occur mainly near the contacts and between the domes. Microgranitoid enclaves occur mainly in the centre of the eastern dome and diorite fragments occur irregularly throughout the pluton, mainly between the two domes. He interpreted the quartz diorite as the fractionated

product of a dioritic/gabbroic magma and considered that the tonalites were generated by mixing with quartz diorite and trondhjemite magma, or alternatively that the trondhjemites were the product of fractional crystallization of a tonalitic magma. The elevated K_2O contents of some tonalite samples (see Fig. 41 in Nironen 1989a) as also higher Th contents indicate small scale contamination. The major element compositions of the tonalites, excluding the higher MgO, are consistent with a low degree (about 20 %) of melting of basaltic source with amphibolitic to eclogitic residue. The low HREE abundances and high Tb/Yb-ratios favour the existence of garnet in the residue (Fig. 24). The low Rb (25-40 ppm) and La (9-15 ppm) in most mafic tonalites and the Th-Ta relation (Ta 0.61 ppm and Th 2.8 ppm in Fig. 26) indicate a rather primitive source. The most probable candidate is therefore a low-K tholeiitic to calc-alkaline island arc basalt. The higher MgO contents with respect to experimental results could indicate interaction with mafic rocks as discussed previously (p. 62). The diorite/gabbro samples are basaltic and the quartz diorite sample is basaltic andesite in composition. The Th-Ta relation (Fig. 27), high P_2O_5 (0.3-0.5 %), La (19-36 ppm), Zr (130-260 ppm), Sr (400-800 ppm) and Ba (230-370 ppm) contents indicate within-plate affinities for these rocks. The TiO_2 contents (1.3-1.4 %) of diorite/gabbro are somewhat lower than those normally found in WPB but on the other hand these samples are rich in MgO (11-12 %), suggesting a rather primitive origin.

There is only one unpublished field report describing the Mäntsälä pluton (Nironen 1982), which is mostly granodioritic except for a granitic central part, and trondhjemite at its northern contact. The pluton cuts supracrustal rocks and contains schist xenoliths. The geochemistry of major and trace elements indicates the presence of two different magma patches; a trondhjemitic and granodioritic to granitic (data not given). The more siliceous granodiorites (SiO_2 > 68 %) are the result of mixing of trondhjemitic magma with granodioritic magma or assimilation of the surrounding schists. The high contents of K_2O (3.5-4.5 %), Ba (800-1000 ppm), Rb (80-110 ppm), and Th (normally 6-13 ppm) and Ta (0.6-1.3 ppm) in granodiorite are best explained by invoking a crustal source or assimilation. The high abundances of Na_2O (3.5-4.1 %) and CaO (2.8-1.9 %) and rather low A/CNK (0.96-1.05) at SiO_2 65-68 % in granodiorite suggest a source that was immature and close to an original igneous protolith in composition, if the schist xenoliths are considered as restites. The major element and trace element data are consistent with the interpretation that the trondhjemite magma might have originated through melting of a basaltic source leaving an amphibolitic to eclogitic residue. The Th-Ta relation and low K_2O (1.5 %), La (9-11 ppm) and Rb (25-33 ppm) contents at SiO_2 66-68 %, for the two most mafic trondhjemites, as well as the low Th/U-ratio (1.6-2.0) indicate a rather primitive island arc basalt to basaltic andesite source. The high contents of Sr (1000-800 ppm) in these two samples also favour plagioclase-free amphibolitic to eclogitic source with high initial Sr-contents.

The Aulanko granodiorite is surrounded by mafic-intermediate metavolcanics and late-orogenic microcline granites (Simonen 1948). The characteristic features of the granodiorite are the occurrence of idiomorphic hornblende and the rounded mafic inclusions with hypidiomorphic textures. The higher and rather irregular contents of Rb (190-220 ppm, normally 85-120 ppm) and Ta (1.6-4.5 ppm, normally 0.3-0.8 ppm) in some samples indicate autometasomatic processes within late-magmatic liquids or the effect of the late-orogenic microcline granites which are enriched in these elements (normally Rb 200-330 ppm and Ta 1-3 ppm). On Harker diagrams (not shown) samples with high Rb have lower Ba and in some samples also lower Sr, but otherwise they follow normal trends. The La (10-13 ppm), Rb (80-90 ppm), Sr (600-750 ppm), Ta (0.3-0.5 ppm) and Th (4-5 ppm) contents of the most

mafic samples (SiO_2 61-63 %) suggest an origin by melting of calc-alkaline basaltic to basaltic andesitic rocks, at least for these samples. The median granodiorite contents plot in the calc-alkaline fields in Figures 25, 26 and 28. The low Th/U-ratio (1.4-2.4) also favours an island arc environment. The occurrence of mafic igneous enclaves and the scatter of data shows that mingling and mixing and/or contamination between different melts has occurred. The marginal variant of the Aulanko granodiorite is trondhjemite porphyry, which has Rb, Th and U contents (Nurmi 1984) comparable to those of the granodiorite. Simonen (1948) interpreted this rock as a metasomatic product of the Aulanko granodiorite caused by albitization. The equivalent level of Na_2O and higher level of Rb than in the granodiorite is however opposite that which would be expected in sodium metasomatism (e.g. Drummond et al. 1986). Altogether, except for the low K_2O, the trondhjemite porphyry resembles granite rather than typical trondhjemites (see point at 75 % SiO_2 in Fig. 25 and point at Rb 140 ppm in Fig. 28).

The Hämeenkyrö pluton and the Värmälä pluton intrude the Tampere schist belt (TSB) and were studied in detail by Nironen (1989a). Both plutons are surrounded by rocks of vol-

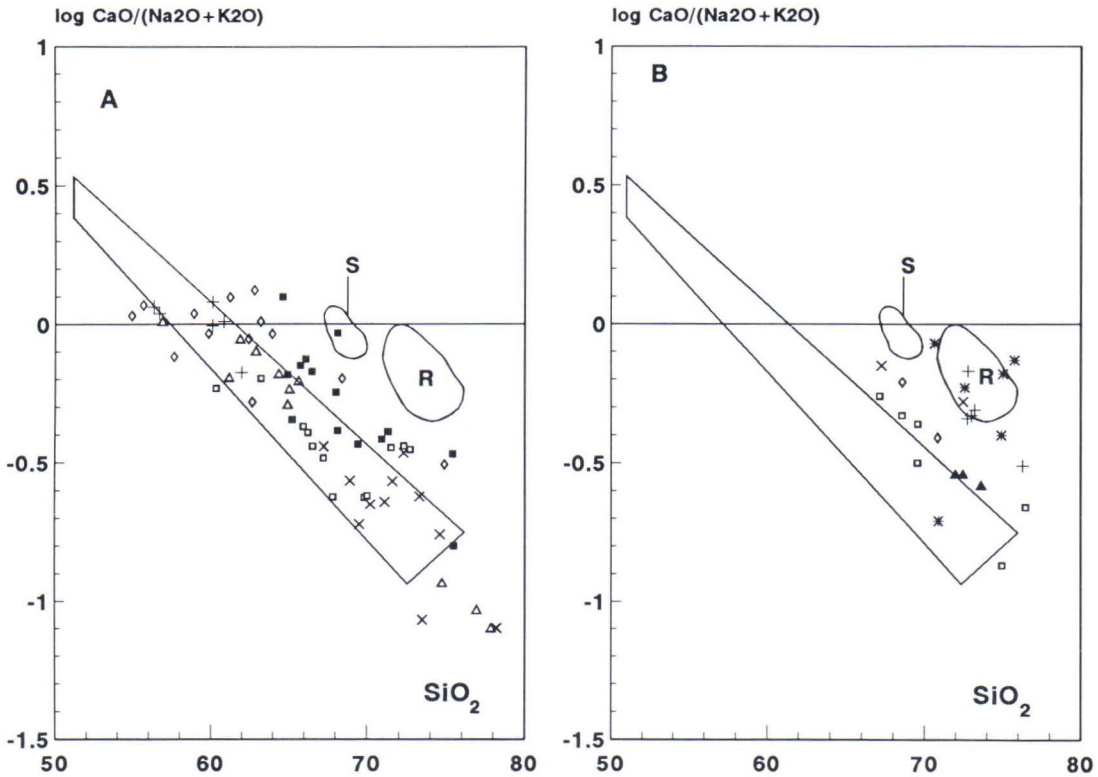

Fig. 25. Calc-alkali-SiO_2 diagram. Outlined area with solid lines - modern calc-alkaline andesites (Brown 1982); field R - Rastinpää-type gneissic tonalites and related rocks; field S - Särkilampi gneissic tonalites and related rocks. A: filled squares - syntectonic tonalites; open squares - syntectonic granodiorites; oblique crosses - syntectonic granites; open triangles - granitoids from NW of Rautalampi; open diamonds - tonalites and granodiorites from Rautalampi area; crosses - pyroxene quartz diorites and tonalites. B: crosses - Kangasjärvi gneissic granodiorite; asterixes - orthogneisses from Pielavesi area and Leväniemi gneiss; open squares - Kettuperä gneiss (volcanic); oblique crosses - Venetpalo area; open diamonds - Saunakangas tonalite/granodiorite; filled triangles - Veteli granodiorite. Data sources as in Figures 5 and 21.

canic origin, with intercalations of metaturbidites. The Hämeenkyrö pluton is composed of multiply intruded rocks, which range from tonalite to monzogranite in modal composition (Front 1981). The main intrusive phases are distinguished by major element ratios as marginal tonalite, prevailing granodiorite (1885±2 Ma, U-Pb zircon) and granite (Nironen 1989a). Mafic enclaves of both xenolithic and microgranitoid origin are present, but neither magma mixing and mingling nor partial melting can explain all the chemical characteristics found in Hämeenkyrö pluton. According to Nironen (1989a), the injection of separate, already differentiated magma batches and mixing and mingling with a basic magma could explain the chemical characteristics and compositional gaps. The Th (6-16 ppm), Ta (0.6-1.0 ppm), La (20-32 ppm) and Rb (70-140 ppm) contents indicate crustal and mixed sources for the rocks in the Hämeenkyrö pluton. The tonalites and granodiorites are rather mafic rocks with SiO_2 values of 53-61 % (see Fig. 11, Nironen 1989a) so that they could represent high-K andesitic magma batches cogenetic with the TSB volcanics. The low Al_2O_3 and Sr contents deviate from those of normal synkinematic granodiorite trends.

The Värmälä pluton is mostly granodioritic (1878±3 Ma, U-Pb zircon) with a granitic margin (Nironen 1989a). The central part consists of porphyritic quartz monzodiorite to monzogranite while the southern contact is with a separate pluton with quartz dioritic to dioritic composition. Both mafic xenoliths and microgranitoid enclaves are found. Nironen (1989a) proposed a model where the diorite was intruded first. A stratified magma chamber formed deeper in the crust and the fluorine-bearing granite ascended to its final level of emplacement before the granodiorite. The quartz monzodioritic cumulus-type magma was generated at the bottom and sides of the magma chamber and was the last to intrude. The separate diorite-quartz diorite pluton (SiO_2 54-58 %) resembles mature arc basaltic andesites, in having high K_2O (1.4-5.0 %), La (17-32 ppm), Ba (300-600 ppm) and Rb (62-120 ppm) and seen also in the Ta-Th-diagram (Fig. 27). The high-K (3.4-5.5 %) and high-Si_2O (67-75 %) nature of the Värmälä pluton, and the absence of intermediate composition favours a partial melting origin for these rocks. The Rb (80-120 ppm) contents and Ta-Th relation (Ta 1.0 ppm and Th 9.1 ppm in Fig. 26) furthermore indicate a crustal source. The Eu-minimum (Fig. 24) shows that plagioclase has been a crystallising phase (see Nironen 1989a) in these samples or that the residue contained plagioclase.

The Pukala pluton, which also intrudes the TSB, was considered by Kähkönen (1989) as a hypabyssal porphyry and is an elongate body (1x12 km^2) with modal compositions ranging from trondhjemite through granodiorite to monzogranite (Nurmi et al. 1984). All the three phases distinguished by Nurmi et al. (1984) have about the same levels and variation in SiO_2 contents (trondhjemite 65-72 %, granodiorite 67-74 % and monzogranite 65-74 %). All phases have same coincident trends for TiO_2, Al_2O_3, FeO, MgO and P_2O_5 on Harker diagrams (not shown). In the K_2O-diagram, the trondhjemites and granodiorites form three groups; trondhjemite with increasing trend (0.5-1.0 %), granodiorite with increasing trend (3-4.5 %) and trondhjemite-granodiorite group with intermediate values (1-3 %) at the same SiO_2 levels. The same grouping, though not so evident, is seen in the CaO, Na_2O, Sr, Ba and Rb diagrams. These features cannot be explained by differentiation, although differentiation is indeed seen in the groups, but could indicate mixing of a low-K trondhjemite melt with a high-K granodiorite melt. The low K_2O (0.5-1.0 %) and Rb (10-20 ppm) contents compared to high Sr (600-800 ppm) contents of the most mafic trondhjemites points to a low-K source rocks with amphibolitic to eclogitic residues, if a partial melting model is invoked. On the other hand the high La (19-36 ppm), Zr (190-230 ppm), Th (6-8 ppm) and Ta (1.0-1.2 ppm) contents favour an evolved source. The rather high

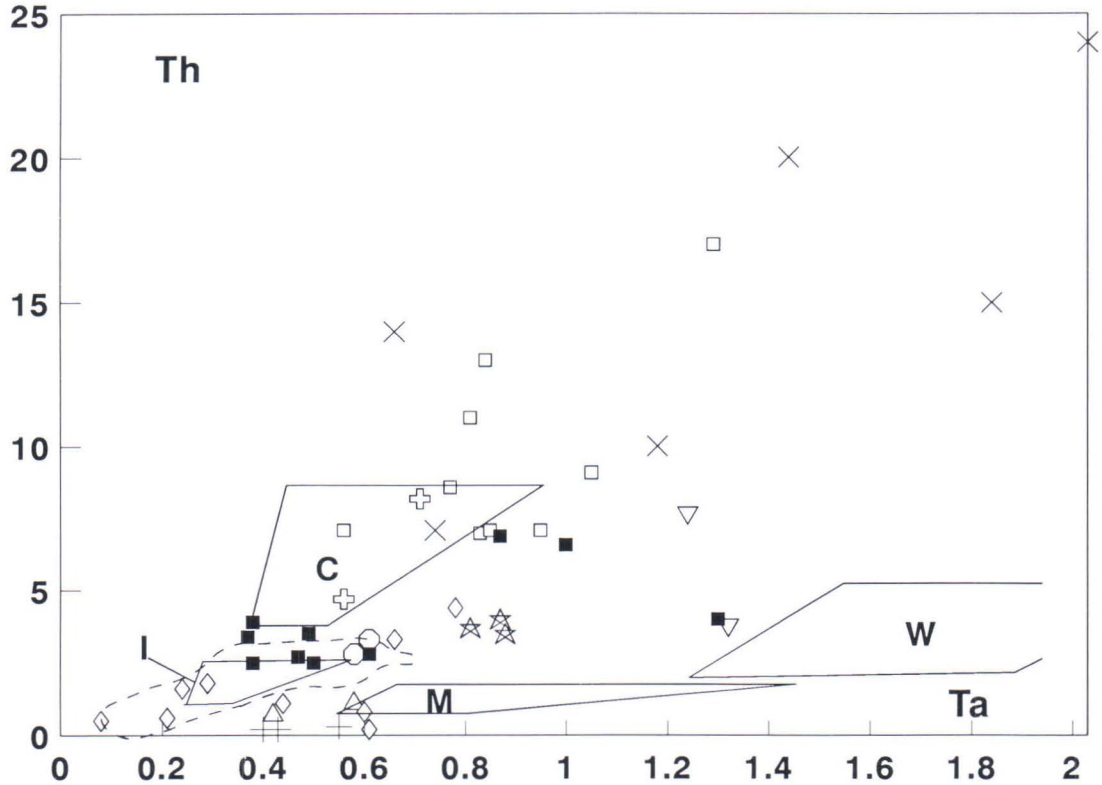

Fig. 26. Th-Ta diagram for granitoids, gneissic tonalites and related rocks. Outlined areas are calculated model melt compositions for 10 % melting with gabbroic residue and 30 % melting with amphibolitic residue (see Table 7 for MORB and IAT) of different basalts (starting compositions from Pearce 1982). I - Island Arc Basalts; C - Calc-alkaline Basalts; M - Mid-Ocean Ridge Basalts; W - Within Plate Basalts. Bulk distribution coefficients for Th 0.05 and for Ta 0.1 and 0.4. Dashed outlined area - Rastinpää-type gneissic tonalites and related rocks and Särkilampi-type gneissic tonalites and related rocks; filled squares - median compositions of syntectonic tonalites; open squares - median composition of syntectonic granodiorites; oblique crosses - median composition of syntectonic granites; open diamonds - tonalites and granodiorites in the Rautalampi area; crosses - pyroxene quartz diorites and tonalites; open circles - Kangasjärvi gneissic granodiorite; open triangles - Leväniemi gneiss; inverted open triangles - Kettuperä gneiss; open crosses - Venetpalo area; stars - Veteli granodiorite. Data from this study except for syntectonic granitoids (unpublished data available from the Geological Survey of Finland).

MgO (1.2-2.7 %) and low CaO (2.1-2.5 %) contents, as well as the high A/CNK-values (1.15-1.30), are not compatible with the experimental data for basalt melting. Other possible mechanisms for producing the low-K trondhjemites are low pressure liquid differentiation or Na-metasomatism. The internal structure of the pluton is not known and there is no published mineralogical evidence for these different processes, so that the origin of the variation in geochemistry remains debatable. The trondhjemite and granodiorite plot in the calc-alkaline fields in Figures 25 and 26, and both have very similar Th (6.9 and 7.0 ppm) and Ta (0.87 and 0.82 ppm) median contents. These features would indicate an origin cogenetic with high-K calc-alkaline rocks.

The Majajärvi pluton near Kankaanpää was briefly described by Nurmi et al. (1984). The extent of this mostly granodioritic body is not known, but it does have migmatitic contacts with the surrounding mica gneisses, mica schists and metavolcanics. Small amphibolite enclaves and mica-rich xenoliths are present. To the SE occurs a heterogeneous assemblage of gabbros, diorites, tonalites and amphibolites

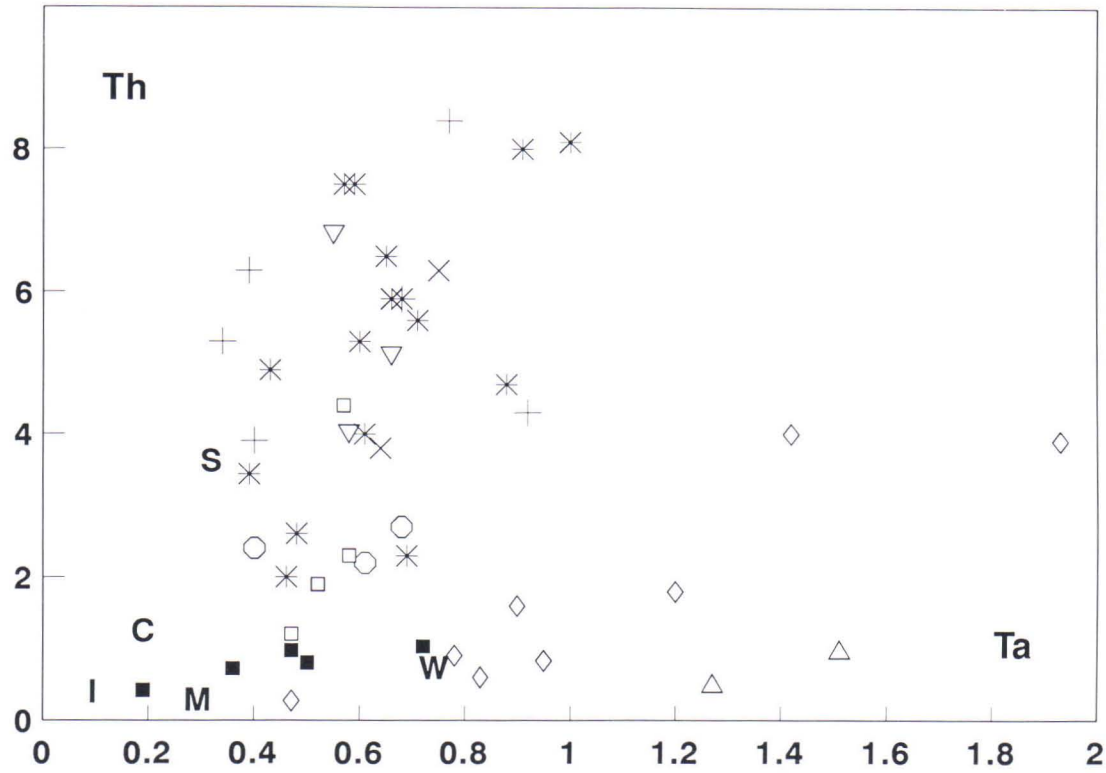

Fig. 27. Th-Ta diagram for mafic rocks. Average basalt compositions from Pearce (1982). I - Island Arc Basalts; C - Calc-alkaline Basalts; S - Shoshonitic Basalts; M - Mid-Ocean Ridge Basalts; W - Within Plate Basalts. Crosses - Käkövesi mafic volcanics: filled squares - Viitasaari gabbros; open squares - Mäntylä diorites; asterixes - Rautio diorites; open triangles - Varparanta gabbros and diorites; open diamonds - Jyväskylä diorites; oblique crosses - Värmälä diorite; inverted open triangles - Kankaanpää gabbros and diorites; open circles - Sydänmaa gabbro and diorite. Unpublished data available from the Geological Survey of Finland.

(ibid.). The gabbro-tonalite group has K_2O 0.5-2 % at SiO_2 values of 49-56 % and moderate La (12-20 ppm), Zr (80-140 ppm), Sr (500-350 ppm), Rb (10-90 ppm) and Ba (200-600 ppm) contents. These features, as well as Th-Ta relations (Fig. 27) indicate a mature calc-alkaline nature for these basaltic to basaltic andesitic rocks. The granodiorite is a rather heterogeneous intrusion and different trends with respect to TiO_2, Sr and Th on Harker-diagrams (not shown) indicate that it is not comagmatic with the gabbro-tonalite group. A partial melting origin is most probable and the heterogeneous nature is attributed to either a heterogeneous source or some combination of contamination and assimilation. The La (19-42 ppm), Sr (700-1000 ppm), Zr (200-240 ppm), Rb (60-90 ppm) and Ba (600-1400) contents at 60-67 % SiO_2 suggest an evolved source and according to the REE-curve an amphibolitic residue (Fig. 24). Possible source rocks include crustal calc-alkaline rocks and/or rather immature sediments, since they plot in or near calc-alkaline fields in Figures 25 and 26.

Two small granitoids intrusions occur in the Luhanka area (Sydänmaa and Hernemäki), which are briefly described in the unpublished field report by Nironen (1981) and in Kallio (1986). The concentric Sydänmaa intrusion is composed of an inner zone of diorite with enclaves of gabbro and plagioclase porphyry, and an outer granodiorite with enclaves of gabbro,

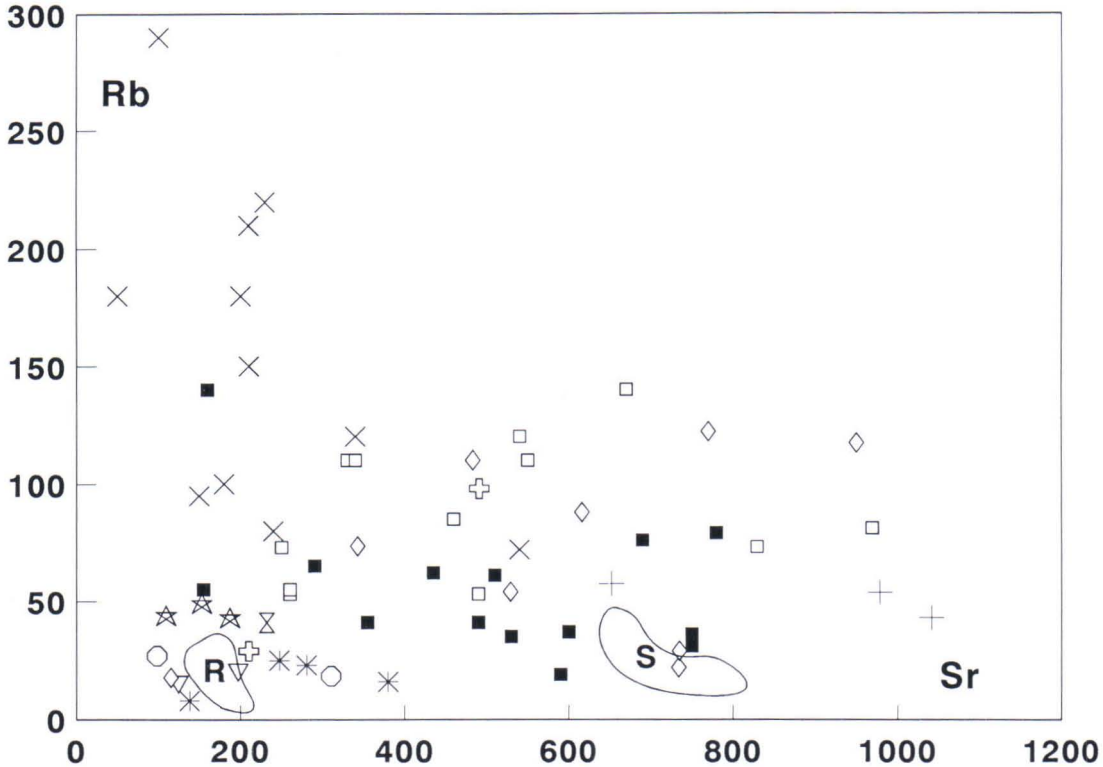

Fig. 28. Rb-Sr diagram for granitoids, gneissic tonalites and related rocks. Symbols are as in Figure 26 except for open hourglass - Saunakangas tonalite. The main fields shown are for Rastinpää-type (R) and Särkilampi-type (S) gneissic tonalites and related rocks. Data sources as in Figures 5 and 21.

plagioclase porphyry and mica-rich material. In the western part of the intrusion quartz diorite is slightly foliated. The diorite, gabbro and plagioclase porphyrite are geochemically similar and show enriched characteristics (K_2O/Na_2O=1, Rb 40-60 ppm, Ba 500-800 ppm, Sr 700-1000 ppm, P_2O_5 0.4-0.9 % and La 20-30 ppm). These features and the generally low SiO_2 (< 50%), high TiO_2 (1.9-2.2 %) and moderate Zr (130-150 ppm) are consistent with a contaminated WPB origin (see Fig. 27 for Ta-Th relation). This contamination is either due to the presence of a subduction component or assimilation during ascent. The outer granodiorite truncates the diorite and contains feldspar megacrysts of probable late-magmatic origin (Nironen 1981; Kallio 1986). The high K_2O (3.8-4.5 %), Ba (900-1300 ppm), La (42-87 ppm), Rb (100-200 ppm, med. 140 ppm), Sr (600-800 ppm, med. 670 ppm), Th (11-25 ppm, med. 17 ppm), Ta (1.0-1.6 ppm, med. 1.3 ppm) and Zr (240-360 ppm) at 62-68 % SiO_2 indicate an evolved crustal source (see Figs. 26 and 28). The quartz diorite is more mafic (SiO_2 54-62 %) and of high-K calc-alkaline nature, and it forms "differentiation trends" with granodiorite in Harker diagrams (not shown). The quartz diorite could represent either a magma batch cogenetic with the granodiorite or the source material type for the granodiorite (lower in the crust).

The unzoned feldspar megacrystic Hernemäki monzogranite is foliated and contains abundant large schist xenoliths. The SiO_2 con-

tents (67-72 %) of the Hernemäki pluton partly overlap with the Sydänmaa granodiorite, but the lower Al_2O_3 and Sr, and markedly greater FeO/MgO ratio in the Hernemäki monzogranite indicates that they are not comagmatic (data not shown). The high La (44-91 ppm), Rb (170-250 ppm), Th (11-17 ppm) and Ta (1.6-2.2 ppm) indicates an evolved crustal origin and the high Zr (360-500 ppm) possibly to a fluorine-bearing magma. The occurrence of garnet (Kallio 1986) and schist xenoliths in the monzogranite could indicate anatectic melting of the schists or strong assimilation. The rather homogenous nature of the monzogranite however favour melting somewhat lower in the crust. The low Sr-contents (180-220 ppm) reflect a Sr-poor source or else retention of plagioclase in the residue. The low A/CNK-values (0.98-1.05) points out to immature sediments with igneous characteristics, if the schist xenoliths are considered as restites.

The Palokka plutonic complex near Jyväskylä is part of the CFGC and apart from granitoids contains only small inclusions of supracrustal rocks (Nurmi et al. 1984). The age relations of the plutonic complex are as follows: diorite, quartz diorite, porphyritic monzogranite, equigranular monzogranite, microcline granite (syenogranite) and biotite granite. The diorite is rather heterogeneous geochemically and is probably partly altered and/or contaminated; this is seen especially in the great variation in Ba (200-3000 ppm) at SiO_2 47-58 %. Apart from one sample, the SiO_2 variation is from 52 to 58 % and FeO/MgO ratios are from 3.5 to 6, which thus classifies these rocks as tholeiitic basaltic andesites. The high La (20-30 ppm), Ba (normally 800-1000 ppm), K_2O (normally 1.5-2.0 %), Rb (50-80 ppm), Sr (400-700 ppm), moderate TiO_2 (0.8-1.6 %) and especially Th-Ta relations (Fig. 27) indicate a within-plate origin for these rocks. The quartz diorites (SiO_2 54-60 %) are geochemically similar to the diorites and coincide in Harker diagrams (not shown) and therefore same origin is proposed for them both. The tendency towards higher Th-contents in Fig. 27 can be attributed to either fractionation or contamination.

The porphyritic monzogranite (SiO_2 69-72 %), equigranular monzogranite (SiO_2 64-69 %), syenogranite (SiO_2 77-80 %) and biotite granite (SiO_2 69-76 %) are geochemically heterogeneous and form different groups in Harker diagrams (not shown). They differ from diorites and quartz diorites, and are not directly comagmatic with them. Although the equigranular monzogranite is heterogenous, it has major and trace element characteristics (normally La 40-50 ppm, Ba 1000-1600 ppm) that indicate a high-K calc-alkaline origin through melting or differentiation. The porphyritic monzogranite could represent a smaller degree melting of the same source. The microcline granite has variable contents of Ba (13-200 ppm), Rb (120-200 ppm), Sr(< 60 ppm), La (5-35 ppm), Th (4.4-29 ppm) and Ta (0.1-2.9 ppm) at SiO_2 77-80 % and K_2O 3.2-4.9 %. The biotite granite is more homogenous with Ba (600-1000 ppm), Rb (200-270 ppm), Sr (180-310 ppm), La (14-60 ppm), Th (12-32 ppm) and Ta (1.0-1.8 ppm) contents at SiO_2 69-73 % and K_2O 4.4-5.3 %. They both have crustal signatures but the genetic relation between the two phases is unclear.

The Mäntylä plutonic complex at Viitasaari is also part of the CFGC and is composed of metagabbro, gabbro pegmatoid, monzodiorite, quartz diorite, hypersthene granodiorite (Lahnanen pluton), hypersthene granodiorite (Hirvivuori pluton), hypersthene monzogranite (Saunamäki and Saarilampi plutons), fine-grained monzogranite, coarse-porphyritic monzogranite and equigranular monzogranite (Nurmi et al. 1984 and Nurmi 1984). The U-Pb zircon ages are 1885±1 Ma for the quartz diorite, 1882±2 Ma for a gabbro pegmatoid and 1886±6 Ma for hypersthene granite (Nironen and Front 1992). The metagabbro occurs as inclusions in the granitoids and is heterogeneous and locally banded, with amphibole- and pyroxene-rich layers alternating with plagioclase-rich layers. This heterogenous nature is

also seen in the geochemistry and hinders the interpretation of the origin. The low abundances of Zr (30-80 ppm) and rather high contents of K_2O (0.7-1.6 %), Ba (400-700 ppm), Rb (17-60 ppm) and La (10-22 ppm) at 44-52 % SiO_2 compared to the Ta-Th relation (Fig. 27) are difficult to interpret. The porphyritic monzodiorite truncates the meta-gabbro (Nurmi et al. 1984) and geochemically it resembles high-K calc-alkaline andesites, with TiO_2 (0.8-1.0 %), K_2O (2.0-2.6 %), Rb (50-75 ppm), Ba (750-850 ppm), Zr (120-160 ppm), Sr (700-750 ppm), La (20-27 ppm) and Th (1.0-2.3 ppm) at 57-62 SiO_2. The metagabbro and porphyritic monzodiorite are geochemically similar and are probably cogenetic if not comagmatic. According to the Ta-Th relation (Fig. 27) these rocks could indicate the mixing of within-plate magma with crustal derived rocks were the gabbros are mainly cumulate rocks.

The Mäntylä hypersthene-bearing quartz diorite contains abundant hornblende gneiss inclusions and according to Nurmi et al. 1984 (see also Nironen and Front 1992) it is older than the other hypersthene granitoids. The geochemical variation indicates differentiation within the intrusion and also alteration or assimilation/contamination, which is seen in the scatter of Rb and Th. The normal variations of selected elements over a range of SiO_2 from 56 % to 67 % are as follows: TiO_2 1.1-0.4 %, K_2O 0.6-2.1 %, Ba 400-900 ppm, Rb 18-60 ppm, Sr 500-350 ppm, La 13-24 ppm, Ta (0.3-0.6 ppm, due to possible contamination these are the maximum values), Th 0.4-2.0 ppm and Zr 90-200 ppm. It deviates from the typical synkinematic tonalites having higher FeO and FeO/MgO-ratio, and lower MgO and Sr contents.

The two hypersthene granodiorite plutons in the Mäntylä plutonic complex are geochemically similar, except for the different median contents of Th (Lahnanen 1.3 ppm and Hirvivuori 3.8 ppm). They differ from typical synkinematic granodiorites (points at 72.3% and at 72.7% SiO_2 in Fig. 25) in being strongly tholeiitic and also in having higher CaO. The monzogranites are, except for the coarse-porphyritic monzogranite and equigranular monzogranite, also tholeiitic in character and have slightly higher CaO and lower K_2O contents compared to other synkinematic granites. The internal geochemical variation of within and between different intrusions has not been studied. The median contents of Rb (55 ppm and 73 ppm) and Sr (260 ppm and 250 ppm) as well as the Th contents and high SiO_2 (see above) in granodiorites are consistent with an interpretation involving a small degree of melting with a gabbroic residue. The source rocks are probably evolved calc-alkaline rocks. Nironen and Front (1992) considered that the most evolved rocks including hypersthene granites, biotite granite and monzogranite are possibly comagmatic but that other lithologies represent different source materials. A generally deep source area with dry conditions was also postulated.

The areal extent of the Tienpää tonalite complex near Halsua in the NW-corner of the CFGC, is poorly known (Nurmi et al. 1984). The complex comprises hornblende tonalite, porphyritic tonalite, equigranular tonalite, tonalite porphyry and K-feldspar-bearing tonalite. The tonalites are geochemically rather similar with some gradation from a slightly more mafic hornblende tonalite to tonalite porphyry, as reflected in the median contents of major and some trace elements (see Appendix in Nurmi 1984). The tonalites are medium-K (except for the K-feldspar-bearing tonalites) calc-alkaline rocks with median SiO_2 values from 62 to 67 %. The A/CNK-values (normally 1.05-1.15) would indicate S-type affinities, but high Sr-contents (normally 400-600 ppm) and Na_2O-K_2O-relations are typical I-type characteristics. The abundances of Rb (normally 30-50 ppm), Ba (normally 400-700 ppm), Th (normally 1.4-4.5 ppm), Ta (normally 0.3-0.6 ppm), La (12-17 ppm) and Zr (normally 130-160 ppm) are also consistent with an I-type interpretation. The major element data and Th-Ta-relations (Fig. 26) indicate a calc-alkaline island arc origin either as intrusions cogenetic with calc-

alkaline andesitic-dacitic volcanics or as the melting products of calc-alkaline basalts. For the partial melting alternative the high Al- and Sr-nature of these rocks and the REE-curve (Fig. 24) would favour an amphibolitic residue.

The geochemistry of the Rautio pluton was studied in detail by Nurmi (1983). According to him the pluton can be divided geochemically into two plutonic suites, both of which include rocks ranging in composition from hornblendites to granodiorites. The hornblendites have been interpreted as cumulates and the diorites as continental margin type high-K andesites (see Fig. 27). The granodiorites originated from partial melting of igneous rocks (Nurmi 1983). The high contents of K_2O (3.3-3.8 %), Rb (110-140 ppm), Th (11-13 ppm) and high Th/U-ratio (normally 4.2) at 68-72 % SiO_2 in granodiorites (see Table 3 in Nurmi 1983) indicates an evolved crustal source. The REE data for two granodiorite samples shows a Eu-minimum (Fig. 24), probably due to plagioclase fractionation, since the high abundances of Sr (500-600 ppm) do not favour high amounts of plagioclase in the source residue.

The syntectonic Kalliokangas and Aukeaneva tonalite plutons of the Bothnia schist belt intruded turbiditic metagreywackes and metapelites during the third local deformation phase and they contain abundant micagneiss enclaves (Vaarma 1990). K-feldspar phenocrystic granodiorite occurs in the eastern margin of the Kalliokangas pluton. According to Vaarma (1990) these rocks are geochemically of calc-alkaline I-type and indicate a mature island arc origin. The lack of mafic and intermediate intrusive rocks favours the partial melting model. The most mafic samples in both intrusions (excluding sample 5 in Table 6, Vaarma 1990) have about comparable K_2O abundances (1.7 %), as well as Sr (360-380 ppm), La (16-15 ppm), Rb (58-46 ppm), and Th (3.4-3.0 ppm) at 67 % SiO_2. The most probable candidates for source rocks would be a calc-alkaline island arc setting. On the other hand the high median Ta abundance (1.28 ppm) compared to Th (4.0 ppm) suggests a source enriched in "within-plate component" (Fig. 26). The mica gneiss enclaves could indicate either melting of sedimentary source or assimilation of country rocks. The distinct I-type characteristics, as reflected in the low A/CNK-values of most mafic samples (<1.05) would require the presence of sedimentary rocks rich in volcanic material, if these mica gneiss enclaves are interpreted as restitic. The REE patterns (Fig. 24) indicate a plagioclase-bearing amphibolite residue (see also point at Rb 62 ppm and Sr 360 ppm in Fig. 28). The K-feldspar phenocrystic granodiorite has higher K_2O (2.5-3.1 %), La (31-39 ppm) and Th (6.2-7.6 ppm), somewhat lower Sr (240-280 ppm) and similar Rb (47-76 ppm) contents compared to the tonalite. Vaarma (1990) considered that these rocks were all comagmatic through differentiation.

The Kopsa tonalite porphyry clearly intrudes and truncates the surrounding metagreywackes, mica schists and metavolcanics (Nurmi et al. 1984). The tonalite has higher CaO and lower Sr compared to the normal fields for synkinematic tonalites. The median Th/U-ratio of 2.4 is rather low and this, compared with the Th-Ta-relation (Th 3.9 ppm and Ta 0.4 ppm, Fig. 26) and combined with the high CaO could indicate an origin cogenetic with island arc tholeiitic to calc-alkaline basalts either through differentiation or melting. The low abundances of Sr (point Rb 65 ppm and Sr 290 ppm in Fig. 28) favour the presence of plagioclase early in the crystallisation history or in the source residue.

There are no published descriptions available for the Äitsaari and Salosaari plutons at Ruokolahti and so their internal variations and relationships to surrounding rocks are not known. The median contents of the Äitsaari trondhjemite always plots in the main fields for synkinematic tonalites and therefore an origin through melting of island arc mafic rocks is proposed. The Salosaari tonalite has high K_2O (2.2-3.3 %) and plots in the granodiorite field on the K_2O-SiO_2-diagram (Fig. 21). The MgO

contents (3.4-3.9%) are also high at SiO_2 values of 61-65 % as also are La abundances (23-40 ppm). The high MgO values could indicate interaction with mafic rocks. The Salosaari tonalite clearly plots inside the modern calc-alkaline andesite field in Figure 25 (point at 65.2 % SiO_2). This and the high median Rb (79 ppm) and Sr (780 ppm) would indicate a more evolved source. The scatter in Th, Ta and U values indicate a heterogeneous nature and probable interaction between different magmas/assimilation.

IMPLICATIONS FOR THE EVOLUTION OF SVECOFENNIAN SCHIST BELTS AND GRANITOIDS

When analysing the crustal evolution of Svecofennian as a whole, and the petrogenesis of different rock types it is essential to have age constraints concerning emplacement, deformation and metamorphism for each intrusive complex and event, as well as the origin and evolution of associated supracrustal rocks and how they correlate with each other. Here, by concentrating on presently available geochemical and isotopical data, I will attempt to correlate them with the different Svecofennian schist belts and < 2.1 Ga evolution of cover sequences and magmatism in Archean craton and ultimately, summarize all these developments within a comprehensive plate tectonic model (p. 95).

Savo schist belt (SSB)

The SSB is a diverse and complex association of rocks occurring near the paleosuture and includes rocks of different age and origin. The SSB is here considered to include rocks from Vihanti to Virtasalmi occurring west of the proposed paleosuture but the possibility still exists that some rocks of the SSB are comparable with rocks east of it (see Ekdahl 1993). The rocks have been subdivided into different age groups, namely 1.93-1.91, 1.91-1.89, 1.89-1.88, 1.88-1.85, 1.85-1.81 and < 1.81 Ga according to available zircon U-Pb data. In some cases ages overlap and therefore other distinguishing criteria are also used. This classification is however partly subjective and should be considered as such. The possible occurrence of inherited zircons is a further complicating factor. The rocks with possible ages in excess of 1.93 Ga are considered in the 1.93-1.91 Ga group. The supracrustal associations of Pielavesi area are discussed here within the 1.93-1.91 Ga group and only briefly noted in conjunction with rocks of other age groups.

1.93-1.91 Ga group

Vihanti area

The basalt-rhyolite cycle of volcanics in the Vihanti area (Rauhamäki et al. 1980) cut by the Alpua gabbro 1901±12 Ma (Vaasjoki and Sakko 1988) is interpreted to belong to this age group. The Lampinsaari ore complex consist of mica gneiss with amphibolite intercalations followed by a felsic volcanic-sedimentary unit with abundant dolomites and calc-silicate rocks. The ore deposit consists principally of pyrite ore, disseminated copper ore, zinc ore and lead-zinc ore underlain by uranite-apatite mineralization (Rauhamäki et al. 1980). Migmatitic mica gneisses that are found both below and above these rocks are characterized

by abundant hornblende-bearing variants interpreted as the occurrence of weathering products of volcanics (ibid.).

Venetpalo area

The gneissic granodiorite of this study is comparable with oligoclase dome gneiss of Hautala (1968) and has geochemical characteristics comparable with those of the Rastinpää high-Y group gneissic tonalite at Rautalampi and the Kangasjärvi gneissic granodiorite. The Sm-Nd data favours an origin comparable with that proposed for the Kangasjärvi gneissic granodiorite (Lahtinen and Huhma in prep.). The supracrustal association, which includes altered rocks and sulphide mineralizations, is associated with 1.93-1.91 Ga group orthogneisses suggesting an age and origin comparable with that of 1.93-1.91 Ga group rocks in other parts of the SSB. The plagioclase porphyritic gneiss resembles synkinematic tonalites in composition but high Ca nevertheless indicates a rather primitive source. The proposed age is about 1.90-1.88 Ga.

Pyhäsalmi area

Volcanics in the Pyhäsalmi area can be divided into two main units, namely the Pyhäsalmi and Kuusaanjärvi formations (Kousa 1990). The Pyhäsalmi bimodal volcanics of island arc affinity are notable for hosting the VMSD-type Pyhäsalmi Zn-Cu deposit. They are low-K tholeiitic basalts and basaltic andesites, and low-K calc-alkaline rhyolites (Mäki 1986, Kousa 1990). A U-Pb age of 1921±2 Ma for zircon from rhyolite at Riitavuori (Kousa, Marttila and Vaasjoki in press.) is comparable with the age of 1930±15 Ma obtained for the Kettuperä gneiss (Helovuori 1979) which has been considered as basement to the volcanics (Helovuori 1979 and Marttila 1993). This rock has been interpreted as a subvolcanic sill or even volcanic unit in this study and it is geochemically comparable with, if not, cogenetic with felsic volcanics. A cogenetic origin was proposed also by Ward (1988a) and Kousa (1990), and this is verified by Sm-Nd data that show a positive ε_{Nd} -value about +3 for both the Kettuperä gneiss and the Riitavuori rhyolite, which thus precludes a remobilized Archean basement origin for the Kettuperä gneiss (Lahtinen and Huhma in prep.). The chemical composition of the Pyhäsalmi-type mafic volcanics show variable K_2O (Table 2 in Mäki 1986 and Table 1 in Marttila 1993) and it is not clear whether or not this is due to alteration or to a different origin. The main difference compared to the Rautalampi mafic volcanics is the lower TiO_2 and possible higher Ba at similar MgO values. The lack of more detailed trace element data from the Pyhäsalmi mafic volcanics hinders the interpretation but the low TiO_2 could indicate either a higher degree of mantle melting or a more depleted mantle compared to that in the Rautalampi area. The higher K and Ba level on the other hand could indicate a greater subduction component. This would favour a more depleted mantle component associated with higher proportion of subduction component compared to the Rautalampi area. The Kuusaanjärvi formation is included in 1.89-1.88 Ga group and will be discussed below (p. 89).

Kangasjärvi area

The Kangasjärvi gneissic granodiorite/tonalite is geochemical similar to the Rastinpää high-Y group gneissic tonalites and hence similar age and origin is proposed. It underlies volcanics, altered rocks and a VMSD-type Zn-Cu deposit but the nature of the contact is unknown (Rehtijärvi 1984). The footwall volcanics of this Zn-Cu deposit comprise intermediate calc-alkaline rocks followed by tholeiitic basalts and calc-alkaline felsic volcanics implying that volcanism was not bimodal in character (Rasilainen 1991). Although this seems to apply to the association as a whole, the uppermost volcanics are more or less bimodal and

could possibly be correlated with the bimodal volcanics at Pyhäsalmi.

Pielavesi area

The Pielavesi area has been extensively studied for decades, particularly because of the abundant mineralization in the district (see references in Ekdahl 1993). Ekdahl also divided the supracrustal associations in the area into different suites and presented new ideas on the evolution of the SSB as whole. The character of rock suites, especially their geochemistry, are considered here, while their implications for crustal evolution and plate tectonics are considered later in the text. Not all of the rock units and suites belong to the 1.93-1.91 Ga group but are still considered here in order to evaluate their relations to this group. They are also briefly considered later in their interpreted groups.

The Kirkkosaari, Joutsenniemi and Leväniemi domes in the Pielavesi area have been considered to represent depositional basement for the supracrustal associations divided to four different suites (Ekdahl 1993), namely the Salonsaari, Savijärvi, Säviä and Koivujoki Suites. The Salonsaari Suite (up to 1000 m) is the lowermost supracrustal unit recognized with mafic Salo Metavolcanite unit (10-50 m) at the base, followed by graphitic schist that passes into migmatitic mica gneisses. The upper part of the Salonsaari Suite is characterized by abundant graphite- and carbonate-bearing horizons. The Salo Metavolcanite consists of massive lava with sporadic pillow-like features. The upper part has features suggesting a pyroclastic or epiclastic origin. The Savijärvi Suite overlies migmatitic gneisses and quartz diorites of the Salonsaari Suite with a gradual increase in the abundance of carbonate, calc-silicate and graphitic lithologies. The Savijärvi Suite comprises felsic metavolcanics with narrow mafic units, carbonaceous and calc-silicate rocks, cherts, uranium- and phosphorus-bearing horizons, black schists and the Kalliokylä conglomerate, which possibly represents rapid syndeformational erosion and deposition prior to or initiating the Säviä Suite volcanism. The felsic volcanics locally show primary porphyritic and pyroclastic features but normally they are gradational into calc-silicate rocks.

The Säviä Suite, included within the main sequence of volcanic rocks occurring throughout the Pyhäsalmi-Pielavesi area, is characterized by mafic, intermediate and felsic metavolcanics and garnet-cordierite-anthophyllite gneisses (Ekdahl 1993). These are considered as hydrothermal alteration products of volcanics and host the Säviä Cu-Zn deposit. Uralite-porphyritic lavas with locally occurring pillow structures and metatuffs are the most important types of mafic volcanics. Intermediate metatuffs and metatuffites are the most volumetrically significant unit within the Säviä Suite. They show rapid variation in composition, usually manifest as distinct banding. Felsic pyroclastic layers normally alternate with more mafic units and do not form individual felsic volcanic units. The uppermost Koivujoki Suite consists of mica-hornblende gneisses, metaarkosites, mafic metavolcanics and the Kotajärvi Metalava and associated metagabbros. The mica-hornblende gneisses differ from the underlying migmatitic gneisses in being less intensely deformed and containing more biotite and hornblende. Mafic metavolcanic intercalations are typical in the upper part with, local graphite-bearing layers. Feldspathic graywackes rich in K-feldspar are relatively abundant locally. The Kotajärvi Metalava, or Kotajärvi Diorite of Salli (1983), varies from 50-100 m in thickness and contain also gabbroic variants.

Ekdahl (1993) summarized the evolution of Pielavesi area as follows. Low-K tholeiites in a submarine setting (Salo Metavolcanite) were deposited on a basement consisting of orthogneisses and paragneisses (> 1.93 Ga) followed by thick graywacke sequences (migmatitic mica gneisses of the Salonsaari Suite). This was gradually replaced by a shallow marine

association (Savijärvi Suite) and followed by extensive bimodal volcanism and Kuroko-type ore formation (Säviä Suite) preceded by local Kalliokylä conglomerate indicating depositional break prior the eruption of Säviä Suite. A later phase of flysch-type sedimentation (Koivujoki Suite) with the associated Kotajärvi Metalava occurs above Säviä Suite.

Orthogneisses in the basement domes contain abundant mafic inclusions and are granodioritic to tonalitic in composition and resemble the Rastinpää gneissic tonalite (Ekdahl 1993). Paragneisses were probably derived from hornblende- and mica gneisses and amphibolites but due to intense deformation the distinction between ortho- and paragneisses is sometimes difficult. The contact between the gneiss domes and the overlying Salo Metavolcanite appear to be concordant and sometimes gradational with alternating layers of gneiss and metavolcanite. According to Ekdahl this is due to either isoclinal folding or the presence of subvolcanic dykes. On the other hand, these relationships could be interpreted as apophyses intruding the supracrustal association if a sill origin is favoured (see Fig. 6c in Ekdahl 1993). The contact zone is intensely deformed and brecciated, reflecting the competence contrast between gneissic tonalite and supracrustal rocks.

The U-Pb age of 1925±5 Ma for the Kirkkosaari orthogneiss (Ekdahl 1993) places this rock within the 1.93-1.91 Ga group of gneissic tonalites. Orthogneiss samples from Sikoniemi (sample 4 in Appendix 1 in Ekdahl 1993) and Leväniemi (sample 6 ibid.) are geochemically comparable with the high-Y group Rastinpää gneissic tonalite and orthogneiss from Kylmälahti (sample 2) is comparable with low-Y group Rastinpää gneissic tonalites. The Sm-Nd data from the gneissic tonalites in Pielavesi area have positive ε_{Nd} values (+1.9 - +4.4), precluding their origin as remobilized Archean basement (Lahtinen and Huhma in prep.). An orthogneiss sample from Salonsaari (sample 1 in Appendix 1 in Ekdahl 1993) has different geochemical characteristics, with high Zr and Al_2O_3 and indicates a different origin; this unit should be studied in more detail. The compositions of two paragneisses are also included in the study by Ekdahl (samples 3 and 5 in Appendix 1, 1993) and they both show low K_2O/Na_2O ratios (0.5 and 0.1) combined with very low values of CIA index (46.0 and 49.5). This indicates compositions close to that of the igneous protoliths without any noticeable effects due to weathering. One mica gneiss (paragneiss) form Kirkkosaari has also been dated and gave a result 2234±69 Ma (Ekdahl 1993). The roundish type zircons and high error are consistent with sedimentary origin involving detrital mixing of Archean and early Proterozoic zircon populations as noted in other metasediments (Huhma et al. 1991). The Sm-Nd data is consistent with this interpretation (Lahtinen and Huhma in prep.). Unfortunately there are no published chemical analyses of this rock. The Saarela granodiorite, included also in the 'basement', has discordant U-Pb data with older zircons at about 2.1 Ga and younger zircons slightly older than 1.9 Ga (Ekdahl 1993). This could be interpreted as a 1.93 Ga age magmatic granodiorite/tonalite having assimilated sediments with mixed zircon populations. Ekdahl (1993) correlated the paleosomes with the Salonsaari Suite. The Laajamäki quartz diorite (1923±4 Ma), also in Pielavesi area intrudes a supracrustal formation (Vaasjoki and Sakko 1988) which, according to the Appended Map 1 in Ekdahl (1993), belongs to the Salonsaari Suite.

The supracrustal formations in Pielavesi area are briefly considered here, concentrating particularly on the geochemistry. Sample numbers used in this context are from Appendix 1 in Ekdahl (1993). He proposed in general a plate margin rather than intraplate origin for the volcanics in Salonsaari Suite. Metalava samples 7 and 8 are medium-high K tholeiites and are characterized by low SiO_2 (43-47%), high TiO_2 (1.4-1.5%), P_2O_5 (0.32-0.35%), Sr (364-736 ppm), Zr (79-104 ppm), Ba (283-170 ppm)

and Ce (45-42 ppm). These features differ from normal oceanic island arc tholeiites. The low Nb (4-5 ppm) combined with high Ti is problematic as combined with above shown characteristics indicate either the presence of subduction component or contamination. The third metalava sample 10 has low TiO_2 (0.59%) and Zr (37 ppm) possibly indicating a slightly different origin. All the samples of Salonsaari Suite show clear negative Nb-anomaly (data not shown) and metalava sample 10 and metatuffite 9 geochemically resemble the medium-K tholeiite from Toholampi.

Mafic tuffite sample 15 from the Savijärvi Suite is a medium-K tholeiitic basalt in composition with high TiO_2 (2.09%), P_2O_5 (0.31%), Y (42 ppm), Zr (220 ppm), Nb (15 ppm), Ba (212 ppm) and Ce (51 ppm). In particularly the Ba-Nb-Ce relation compared with the high Zr, Ti and Y is consistent with a WPB origin. Mafic tuffite 16 has low Zr (30 ppm) and Nb (2 ppm) at intermediate TiO_2 (0.93%) indicating island arc affinity. Intermediate tuffites 18 and 19 show great differences in their geochemistry, notably in K_2O (1.66% and 0.35%) and Zr (154 and 42 ppm). Felsic tuffites show also variable geochemistry. Ekdahl (1993) proposed general island arc affinities for the Savijärvi Suite and variations in composition were attributed to mixed volcano-sedimentary origin with possible phosphorous addition. This mixed origin hinders the interpretation of tuffites but the general WPB-affinity, also noted by Ekdahl, of sample 15 cannot be explained by such an origin and records an episode of WPB magmatism.

Ekdahl (1993) interpreted the Säviä Suite mafic and intermediate volcanism, with two exceptional samples, to indicate an overall plate margin setting. Mafic metavolcanic sample 23 and metatuffite sample 25 have Ba-Nb-Ce-P-Zr-Ti-Y relations compatible with E-MORB, marginal basin or WPB origins but an E-MORB origin can be excluded on geological grounds. Samples 26, 27, 28, 30, 31, 33 and 35 have geochemistry comparable with low-K tholeiitic basalts and basaltic andesites in the Rautalampi area and a similar origin is therefore proposed. Samples 32, 34, 36 and 37 are high-K tholeiitic basaltic andesites and their geochemical similarity indicates a cogenetic, possibly also comagmatic, origin. They are characterized by high Rb (52-72 ppm), Sr (707-839 ppm) and Ba (606-847 ppm). Compared to their enriched nature they have rather high Nb (8-10 ppm) and no distinct negative Nb-anomaly is observed. The mantle component has probably been of EM-type. Sample 24 is more mafic in composition but has a similar high-K tholeiitic nature and its origin is probably of similar to that of the above considered samples. Samples 39 and sample 41 are very high-K trachytes in composition. These samples have many geochemical features, such as rather high Rb, Sr, Ba and Nb, in common with the above high-K samples. The high-K tholeiitic basaltic andesites and the very high-K trachytes are probably not comagmatic although their similar geochemistry does suggest a cogenetic origin.

Sample 42 is geochemically comparable with the high-Y group felsic volcanics and gneissic tonalites in Rautalampi area and is also comparable with the high-Y orthogneisses in Pielavesi area. Samples 45-47 are calc-alkaline felsic volcanics with variable K_2O (1.9-4.6%) and high Rb, Sr (two samples) and Ba compared to sample 42. Sample 44 is also calc-alkaline with elevated K_2O (2.8%) but has low TiO_2 (0.17%) and Zr (87 ppm), possibly indicating different origin.

Ekdahl (1993) divided the Koivujoki Suite into tholeiitic more mafic group containing gabbros and to a more felsic group including the Kotajärvi Metalava, and other mafic metavolcanics exhibiting calc-alkaline character. Gabbros are, at least partly, cumulate rocks as evident from the high TiO_2 (3.63%) and P_2O_5 (1.97%) in sample 48. An interesting feature is the geochemical similarity between gabbro sample 50 and high-K basaltic andesite sample 32 from the Säviä Suite. The Kotajärvi Metalavas have alkaline affinities and are trachy-

andesites in composition (Ekdahl 1993). Their enriched nature is seen in high values of P_2O_5 (normally 0.4-0.8%), Sr (normally 900-1800 ppm), Zr (180-230 ppm) and Ba (1200-1600 ppm). Niobium values are low (3-9 ppm), except in sample 55 (26 ppm), and a pronounced negative Nb-anomaly either indicates the presence of subduction component or crystallization of a Ti-bearing phase. Although the crystallization history of these rocks, especially that of gabbros, is unknown and should be studied further, the high levels of P, Rb, Sr, Zr, Ba and Ce favour an EM-type mantle origin.

The contacts between the Salonsaari, Savijärvi and Säviä Suites are often gradational and have been observed from drilled sections. Samples considered above are distributed over the study area and their precise stratigraphical positions within the different suites is unknown. Samples 7 and 8 from the Salo Metavolcanite in the Salonsaari Suite have compositions that indicate either low degree melting of DM-type mantle or melting of an EM-type mantle accompanied by a subduction component. The possibility of crustal contamination cannot be excluded. Sample 15 from the Savijärvi Suite and samples 23 and 25 from the Säviä Suite have geochemical composition more compatible with a marginal basin or WPB origin, probably in a continental environment, than normal IAT. Notice that sample 15 from the Savijärvi Suite and sample 25 from the Säviä Suite are both from Kirkkosaari and are located within 10 m of one other. The other Säviä Suite samples can be divided into two geographical associations. The low-K tholeiitic basalts and basaltic andesites are of IAT-type and are comparable with similar mafic volcanics in Rautalampi area. These samples are located west of Kirkkosaari as also is felsic metavolcanic sample 42, which have been considered as cogenetic with the gneissic tonalites in Pielavesi area. High-K basalt and basaltic andesites, very high-K trachyandesite, very high-K trachyte and three of four calc-alkaline felsic volcanics are all from the Pangansalo-Haapajärvi association to the east of Kirkkosaari. These rock types are not found, according to data given by Ekdahl (1993), in the Säviä Suite west of Kirkkosaari, although the felsic volcanic 44 from Saarela is an exception to this. This sample also has high K_2O (2.8%) but otherwise differs from the calc-alkaline felsic volcanics found in Pangansalo-Haapajärvi association. If the samples from this association are representative of the whole area, it indicates different origin from the Säviä Suite proper. These geochemical differences, together with the occurrence of the Koivujoki Suite arkoses (rich in K-feldspar) nearby and the possible geochemical similarities with Koivujoki Suite gabbros indicate that Pangansalo-Haapajärvi association should either be included within Koivujoki Suite or considered as distinct subgroup within the Säviä Suite.

Based on the above considerations a slightly different scheme than that given by Ekdahl (1993) is proposed here. The orthogneisses of 'basement domes' are interpreted to represent tonalitic-granodioritic, possibly sill-like, intrusions and subvolcanic sills that are related to the low-K felsic volcanics in the bimodal Säviä Suite. Although, this is considered reasonable for most of the gneissic tonalites in Pielavesi area, orthogneiss sample 1 from Salonsaari differs from other tonalites and should be studied in more detail. The interpretation of the gneissic tonalites, at least most of them, as intrusive rocks complicates the interpretation of the stratigraphical sequence proposed by Ekdahl. On the other hand, if most of the rocks included into the Salonsaari and Savijärvi Suites occur below the Säviä Suite and have been intruded by 1.93 Ga tonalites, they have an inferred age >1.93 Ga, while the Säviä Suite would have an age comparable with the Kirkkosaari orthogneiss (1925±5 Ma). The U-Pb zircon age 1882±2 Ma from the Kotajärvi Diorite (Salli 1983), interpreted as metalava by Ekdahl (1993), possibly indicates a rather similar age for Koivujoki Suite. The Pangansalo-Haapajärvi association within the Säviä Suite

differ from the normal Säviä-type bimodal association and is correlated here in a broad sense with the Koivujoki Suite.

Rautalampi area

Zircon U-Pb ages from the Rastinpää gneissic tonalite 1922±12 Ma (Korsman et al.1984) and Toholampi gneissic tonalite 1914±4 Ma (Pyöreänsuonvuori in Vaasjoki and Sakko 1988) imply an age of about 1.92 Ga for cogenetic felsic volcanics. The age of Rastinpää gneissic tonalite is almost exactly the same as that of the in Kirkkosaari orthogneiss and Laajamäki quartz diorite in Pielavesi area, while the Toholampi gneissic tonalite has a slightly younger age. More data are needed however to verify the existence of different magmatic epochs. The mafic inclusions in gneissic tonalites and medium-K tholeiitic basalt associated with migmatic garnet-pyroxene gneiss could possibly correlate with Salonsaari Suite metavolcanics in Pielavesi area. Uranium- and phosphorus-bearing horizons are not at present found in Rautalampi area and thus the equivalents of Savijärvi Suite have been not yet identified from this region. The bimodal volcanism, VMSD-type Pukkiharju Zn-Cu prospect and associated altered rocks do however show lithological and geochemical similarities with the Säviä Suite bimodal volcanics.

Virtasalmi area

The Virtasalmi volcanic belt comprise predominant amphibolites with minor marble, calc-silicates or gneisses enveloped by voluminous fine-grained clastic metasediments and the Hällinmäki Cu deposit is genetically related to pretectonic alteration of the amphibolites (Lawrie 1992). The occurrence of both massive lava flows with local pillow lava and sills and/or dykes shows an igneous-volcanic derivation for the Virtasalmi amphibolites. Highly vesicular pillow lavas possibly suggests a relatively shallow water environment during eruption of the lavas. Lawrie (1992) claimed that all the amphibolites are WPB or E-MORB volcanics and are not related to supra-subduction zone magmatism, instead possibly indicating rifted-passive margin environment. These volcanics are intruded by syntectonic granitoids and are thus older than 1.89 Ga. Their geochemical composition and association differs from the mafic volcanics in Rautalampi area, but similarities exist with WPB-type volcanics in Pielavesi area. To the east of Virtasalmi the Viholanniemi metarhyolite has a zircon U-Pb age of 1906±4 Ma (Vaasjoki and Sakko 1988) while the Saunakangas trondhjemite-granodiorite has an age of 1903±10 (Huhma 1986). Therefore, the Virtasalmi amphibolites are probably older than 1.90 Ga but whether they are about 1.91 Ga, 1.92 Ga or older is still unresolved.

Discussion

The similarities in occurrence, geochemistry and age of the gneissic tonalites in Rautalampi, Kangasjärvi, Pielavesi, Pyhäsalmi and Venetpalo districts suggest that they form a coherent group with a common origin concentrated within the middle and northern part of the SSB. These rocks are associated with VMSD-type Zn-Cu deposits and mineralization in island arc tholeiitic to calc-alkaline basalts and andesites and low-K felsic volcanics (Huhtala 1979, Mäki 1986, Rasilainen 1991, Ekdahl 1993, this study). The geochemical similarity of these gneissic tonalites to the surrounding felsic volcanic rocks, their probable occurrence as sill-like intrusions and subvolcanic sills in supracrustal associations, lack of structural discordance and the fact that they have suffered the same deformation events as surrounding rocks, favours a cogenetic and possibly comagmatic origin for gneissic tonalites and felsic volcanics in this region. The geochemical data point to an origin through 10-15 % melting of low-K island arc tholeiite basalts for the Rastinpää, Toholampi, Kangasjärvi gneissic tonalites and most of Pielavesi orthogneisses

and Venetpalo gneissic granodiorite. The Kettuperä gneiss, interpreted as subvolcanic sill or volcanic rock, indicates a more enriched protolith. The residue from these rocks would have been gabbroic to slightly amphibolitic in nature and rich in plagioclase as reflected in the low Sr contents of these rocks. This would indicate pressures less than 10-15 kb (eg. Carrol and Wyllie 1989) and possibly a crustal thickness less than 30 km. The lead isotopic composition of the massive sulphide occurrences in the 1.93-1.91 Ga group forms a homogenous group, that was interpreted to show a large mantle derived lead contribution (Vaasjoki 1981). However, Vaasjoki and Sakko (1988) interpreted the same data to indicate that the lead was derived from the lower crust in Vihanti area. If the interpretation of a syngenetic origin for these ores is correct, the ore lead isotopic composition would equal the lead composition of the surrounding, mainly felsic volcanic host rocks (main carriers of Pb). Huhma (1986) calculated a tentative model based on the same data to show that about 1.5 % Archean crustal lead combined with mantle lead would produce the Pyhäsalmi type lead composition. There are slight but significant variations in ε_{Nd} values for the Rastinpää gneissic tonalite (+ 0.6 to +1.4, 1922 Ma) and for the Kettuperä granite gneiss (+2.7 to +3.2, 1930 Ma) and for the Riitavuori metarhyolite (+3.4, 1921 Ma)(Lahtinen and Huhma in prep.) and therefore it might be expected that differences in lead isotopic composition would occur, especially since the Sm-Nd data would require a greater amount of sediment input via subduction for the Rastinpää gneissic tonalite. In contrast, the Rautalampi lead isotopic composition has slightly lower 206/204 and 207/204 values (Vaasjoki 1981) than the Pyhäsalmi mine consistent with a slightly lower contribution of Archean crustal lead.

The Rastinpää T_{DM} ages of 2.2-2.3 Ga, when compared with the origin of the gneissic tonalite through melting of a low-K island arc tholeiite source, and the lead isotopic composition of the Pukkiharju Zn-Cu-mineralisation (a composition the same as for felsic rocks of the 1.93-1.91 Ga group in Rautalampi area has been assumed) gives an approximation for the age of the protolith as 2.0-2.1 Ga (Lahtinen and Huhma in prep.). The Kettuperä granite gneiss has T_{DM} ages of 1.98-2.02 Ga and on the same basis the approximate protolith age is 1.93-1.95 Ga (ibid.). This is very close to the intrusion age. As a whole the 1.93-1.91 Ga group appears to have been generated within a rather immature island arc with mainly mafic crust, ranging in age from 1.95-2.1 Ga. At about 1.92 Ga a tensional event occurred that produced bimodal magmatism and associated volcanogenic massive sulphide deposits in subsidence basins comparable to the setting of Kuroko deposits (Ohmoto and Takahashi 1983).

1.91-1.89 Ga group

There are three U-Pb zircon age determinations from the SSB belonging to this group; the Saunakangas trondhjemite-granodiorite 1903±10 (Huhma 1986), Alpua gabbro 1901±12 and Viholanniemi metarhyolite 1906±4 Ma (Vaasjoki and Sakko 1988). The Saunakangas trondhjemite-granodiorite migmatizes earlier metamorphosed paleosomes indicating that the Pieksämäki block was strongly metamorphosed before that date (Korsman et al. 1988). The ε_{Nd} value of +3.3 and T_{DM} age of 1.93 nevertheless indicate a very short crustal residence time (Huhma 1986). The age data for Saunakangas does on the other hand have a rather high analytical error which could indicate the presence of xenocrystic zircons and hence a slightly younger intrusive age. The present interpretation involving melting of evolved island arc rocks during a collisional event, leaving a gabbroic residue implies very rapid evolution from island arc rocks to accreted continental crust. There are no published geochemical data for the Alpua gabbro or Viholanniemi metarhyolite, although the Alpua gabbro is a syntectonic rock (Rauhamäki et al.

1980) and indicates that collision had commenced prior 1.90 Ga. The Viholanniemi metarhyolite is the host rock for Viholanniemi base metal mineralization, which has a lead isotopic composition different from the Vihanti-Pyhäsalmi type in that it is intermediate between it and the CFGC-type (Vaasjoki and Sakko 1988). The positive ε_{Nd} (+3.4) and T_{DM} age of 1936 Ma for the Viholanniemi metarhyolite also indicate a very short crustal residence time (Lahtinen and Huhma in prep.).

The age of the Särkilampi gneissic tonalite and related gneisses is problematic and the low zircon content probably prevents at least conventional zircon dating. An age older than 1.90 Ga is inferred but a younger age is also possible. The geochemical data indicate melting of tholeiitic to calc-alkaline island arc mafic rocks depleted in Th, U and possibly Rb and Ta (p. 67). Thus the source rocks have possibly suffered one melt or vapour phase segregation event. The proposed model involves melting of lower mafic crust depleted in Th and U in the same island arc system as the 1.93-1.91 Ga group, but at 1.91-1.90 Ga. The ages of other granitoids in Rautalampi are not known, but an age about 1.89 Ga is proposed. The within-plate signatures in some granitoids indicate a genetic link with the underplating stage discussed below (p. 110). The more calc-alkaline granitoids are either linked to underplating with melting of island arc rocks or record the latest stages of subduction during collision and melting of island arc rocks.

1.89-1.88 Ga group

Most zircon data fall into this group and thus record the major crust forming event in Svecofennides. The Silvola tonalite (1886±5 Ma, Nironen 1989a), Tuusmäki tonalite (1888±15 Ma, Korsman et al. 1984), Osikonmäki tonalite (1887±5 Ma, Vaasjoki and Kontoniemi 1991), Voinsalmi hypersthene granodiorite (1887±11 Ma, Patchett and Kouvo 1986), Vaaraslahti hypersthene granite (1884±5 Ma, Salli 1983), Haukimäki hypersthene granite (1887±16 Ma, Korsman et al. 1984) and Käpylä granite (1887±4 Ma, Vaasjoki and Sakko 1988) have identical ages which point out to major event at 1885 Ma. The Jusko quartz diorite (1892±15 Ma, Helovuori 1979) has been also included in this group. The hypersthene granitoids in Kiuruvesi area are cut by ophitic gabbros and diabases (1886±5 Ma, Marttila 1981). The Kotajärvi pyroxene diorite (1882±2 Ma, Salli 1983), norite (1880±3 Ma, Huhma 1986) from Laukunkangas Ni-Cu deposit and mafic plutonic rocks (1883±6 Ma, Gaál 1980) from Kotalahti Ni-Cu deposit show slightly younger ages.

The Kotajärvi diorite has been later named as Kotajärvi Metalava (Ekdahl 1993) and it and the Koivujoki Suite in general is included in this age group. The Kotajärvi Metalava has an enriched alkalic high-K tholeiitic affinity associated possibly with a subduction component. It was evidently derived from mantle of EM-type as indicated by rather low ε_{Nd} value of -0.3 (Lahtinen and Huhma in prep.). A small amount of contamination can not be excluded. The low Nb in these rocks could be due to the presence of a subduction component or else crystallization of a Ti-bearing phase associated with contamination. The Pangansalo-Haapajärvi association is also included in this age group. The high-K tholeiitic basaltic andesites and associated very high-K more felsic rocks indicate the occurrence of EM-type mantle derived melts, possibly related to a subduction component. However, the almost neglible negative Nb-anomaly in these rocks compared to their relatively high-K could also be interpreted as evidence against the presence of a subduction component.

Two cycles of volcanic activity occur in the Vihanti area, where the younger phase was terminated by the emplacement of hypabyssal plagioclase porphyries of age 1878±17 Ma (Vaasjoki and Sakko 1988). The Pihtipudas area contain granitoids and metavolcanic rocks with a pooled age for granitoids and felsic volcanics of 1883±20 Ma (Aho 1979). In the

Kuusaanjärvi formation, especially in its southern part, primary volcanic structures are extremely well preserved and ash-flow tuffs and possible ignimbrites are found locally (Kousa 1990). These features and volcanic conglomerate intercalations could indicate that volcanism was, at least partly, subareal. Geochemically these rocks differ from the Pyhäsalmi formation and show high-K calc-alkaline affinity with rocks variating from basaltic andesites to rhyolites but on the basis of the lithological similarities Kousa (1990) correlated them with the 1880 Ma Pihtipudas metavolcanics. These volcanic associations show somewhat younger ages of about 1882-1880 Ma compared to the bulk of the syntectonic granitoids, which have a modal age at 1886-1887 Ma. The ages of alkaline WPB-affinity dykes in the Rautalampi area are unknown but intrusive relationships with pegmatite indicate that they are the youngest rocks in Rautalampi area. They are included here within the 1880 Ma group but a younger age is also possible.

The emplacement of the Silvola and Kaartila plutons was consider synchronous with D_2 (Nironen 1989a) and therefore it is proposed that the undated Kaartila pluton is of the same age as Silvola. The Silvola tonalite was interpreted to represent melting of LREE enriched low-K island arc tholeiitic to calc-alkaline basalts (p. 70) possibly in the lower part (at the base) of tectonically thickened island arc crust. The Kaartila pluton indicates a smaller degree of melting of a rather similar source higher in the crust. The emplacement of the Varparanta stock is problematic, but it was probably synchronous with or later than D_2 (see Nironen 1989a for discussion and Gaál and Rauhamäki 1971). The interpreted origin is through melting of low-K tholeiitic to calc-alkaline island arc basalts and interaction with a mafic component at the base of thick crust (garnet in the residue, p. 71). The coeval gabbro and diorite are both geochemically of WPB type. The Laukunkangas Ni-Cu deposit (Grundström 1980) occurs near the Varparanta stock and contains strongly LREE-enriched norite with ε_{Nd} +0.2 (Huhma 1986), a result which could be attributed either to derivation from DM combined with crustal assimilation or from subcontinental lithospheric mantle (EM). The gabbros and diorites of Varparanta pluton are also characterized by high La contents and their close association points to a common origin and probably to shared geochemical characteristics. If we consider crustal contamination as the cause of the low ε_{Nd} value, it would indicate that the majority of the LREE result from contamination, because the ε_{Nd} values for Svecofennian granitoids and sediments are -1 - +3 and -1 - 0, respectively (Huhma 1986 and Huhma 1987). The low Th (< 1 ppm) and Rb (13-15 ppm) in the Varparanta gabbros excludes contamination as a cause for high La and the same argument is applied to the Laukunkangas norite. Therefore, the Laukunkangas norite and Varparanta gabbro inclusions are interpreted to be of the same age and to have been derived from subcontinental lithospheric mantle (EM). The Kotalahti Ni-Cu deposit occur in spatial association with Archean rocks (Gaál 1980), but is a similar deposit type and age, consistent with an origin similar to that of the Laukunkangas deposit.

The hypersthene quartz diorites and tonalites in the Rautalampi area are considered to have been cogenetic with within-plate magmatism, either due to melting or crystallization combined with possible mixing and/or assimilation with crustal derived melts. The Voinsalmi hypersthene granodiorite and Vaaraslahti hypersthene granite both have ε_{Nd} -0.6 and T_{DM} ages of about 2.2-2.3 Ga, and this data was interpreted as representing a mixture of depleted mantle and an Archean component (Patchett and Kouvo 1986). On the other hand, it could indicate the occurrence of 2.0-2.1 Ga age crust (discussed p. 87) and/or the effect of subcontinental lithospheric mantle with heterogenous isotopic compositions. The emplacement of these rocks has been interpreted as syn- or post-D_2 (Korsman et al. 1984). The Käpylä biotite gran-

ite, which is not foliated and has a locally developed rapakivi-type texture (Vaasjoki and Sakko 1988) is also included within this group. The Tuusmäki and Osikonmäki tonalites in Rantasalmi-Sulkava area were emplaced coevally with the other tonalites, but this preceded D_2 deformation in Rantasalmi-Sulkava area (Tuusmäki tonalite) and stresses the difference in tectonic history between Rantasalmi-Sulkava area and areas to the N and NW (Korsman et al. 1984, 1988).

1.88-1.85 Ga group

Two syntectonic intrusive rocks in Vihanti area have ages of 1874±13 and 1860±13 Ma (Vaasjoki and Sakko 1988) and are clearly younger than other syntectonic granitoids and also younger than the posttectonic Käpylä granite in the same area. The Kumiseva gabbro with a U-Pb zircon age of 1879±5 Ma, gabbro from Tyypekinlampi of age 1875±0.3 Ma and a gabbro pegmatoid dated at 1874±0.6 Ma also record a younger mafic magmatic event (Ekdahl 1993). The Lammasaho granodiorite 1853±12 Ma sets a minimum age constraint for the end of the tectonic activity in Lampaanjärvi block (Vaasjoki and Sakko 1988). The late tectonic Hiltula granodiorite (1850±7 Ma, Vaasjoki and Kontoniemi 1991) in the Rantasalmi area also belong to this age group.

1.85-1.79 Ga group

Southern Finland is characterized by a (1.84-1.82 Ga) metamorphic belt with voluminous late-kinematic granites (Korsman et al. 1984, 1988, Huhma 1986, Suominen 1991, Ehlers et al. 1993). The Rantasalmi-Sulkava area fall within this region, and was partly thrust over older metamorphic complexes during D_1-D_2 deformation, before the onset of this metamorphism (Korsman et al. 1988). However, at least D_2 in this area is clearly younger than the D_2 event recognized in areas N and NW of Rantasalmi (ibid.). The metamorphism has been attributed primarily to tectonically thickened crust, although the high geothermal gradient in granulite facies Sulkava area would indicate an additional external heat source (Korsman et al. 1984). The present metamorphic block structure was caused mainly by D_3 deformation (Korsman et al. 1984,1988) and this was cut by the Pirilä-type granitoids (Kilpeläinen 1988) dated at 1815±7 Ma (Vaasjoki and Sakko 1988). The occurrence of post-kinematic granitoids parallel to this metamorphic belt suggests some genetic relationship between these and the metamorphism. The ages and ε_{Nd} values of the Åva (1797±4 Ma and +0.2, Patchett and Kouvo 1986), Parkkila (1794±5 Ma and +0.3, ibid.) and Luonteri (1802±22 Ma, Hirvensalo in Korsman et al. 1984) post-kinematic granitoids are all close to each other. The Luonteri and Parkkila granitoids within the SSB are characterized by significant enrichment in LREE (about 1000x chondritic) and high TiO_2 (1.2-2.5 %), P_2O_3 (0.5-1.5 %), Ba (2000-6000 ppm), Sr (1000-4000 ppm) and Zr (400-1100 ppm) (Nurmi 1984, Nurmi and Haapala 1986 and Nurmi, unpublished data). The available F-data from Luonteri (0.4-0.9 %) are high and, with other data, reveal an alkalic signature to these rocks and a proposed origin by differentiation with alkalic basaltic magmas in an extensional environment (Pitkänen 1985). Similar rocks have been found near Renko, associated with microcline granites and even greater amounts of this type of magmatism in the area are postulated (Lahtinen in prep.). The Sm-Nd data for Parkkila could be interpreted as due to crustal assimilation, but as discussed before that would require that almost all of the LREE result from contamination. The mafic rocks (about 53 % SiO_2) share the same characteristics and although some amount of crustal melting and assimilation is very probable, it would more likely depress these extreme geochemical characteristics. Therefore, a subcontinental lithospheric mantle (EM) origin is favoured for the basic magmas. The Puruvesi granite (1797±19 Ma) has ε_{Nd} -6.9 and indicates a

strong Archean component (Huhma 1986), but according to its age relation it could be related to this same event by crustal melting.

Bothnian Schist Belt (BSB)

Age data are available from the Rautio batholith (1883±10 Ma, Huhma 1986), Veteli granodiorite (1913 Ma Lonka pers. comm., ref. in Ekdahl 1993) and Kalliokangas tonalite (1923±10 Ma, Vaarma 1990) intruding the BSB. The Ylivieska gabbro (1883 Ma, Patchett and Kouvo 1986) and Rauhala base metal deposit are also included in the BSB. The galena lead isotopic composition of the Rauhala deposit conforms to the CFGC group and the similarity of lead isotopic data between the deposit and adjacent quartz diorite (about 1880 Ma) indicates a common source (Vaasjoki 1989). In the southern part of the BSB there are age determinations from plagioclase porphyrite (1886±3 Ma, Mäkitie and Lahti 1991), tonalite (1882±9 Ma, ibid) and quartz monzonite (1871±1 Ma, Mäkitie 1990).

The Rautio batholith was probably intruded at a comparatively late orogenic stage and the associated diorites were interpreted as continental margin-type high-K andesites, and granodiorites as crustal melts (Nurmi 1983). The very positive ε_{Nd} +3.0 and +3.2 for granodiorites (Huhma 1986) indicate very short crustal residence times and very rapid development of mature island arc/continental margin type crust from depleted mantle. The Ylivieska gabbro is also characterized by a positive ε_{Nd} of + 2.6 (Patchett and Kouvo 1986). The Kopsa tonalite clearly truncates surrounding schists (Nurmi et al. 1984) and has probably a cogenetic origin with tholeiitic to calc-alkaline island arc basalts, either due to differentiation or melting. In and near the Evijärvi area are the Tienpää tonalite, belonging to the CFGC, and the Veteli granodiorite and Evijärvi tonalite, which intrude the BSB (see Fig. 1). The Tienpää tonalite has geochemical features indicating a close genetic link with calc-alkaline rocks. The Veteli granodiorite intruded during D_2 (Heikkilä granodiorite in Vaarma 1990) and has been considered to represent small degree melting of evolved island arc rocks with a gabbroic residue. The Kalliokangas and Aukeanneva tonalite stocks intruded during D_3 (ibid.) and a melting origin involving calc-alkaline island arc rocks with a plagioclase-bearing amphibolite residue is proposed. The Ta-Th relation indicates a source enriched in a within-plate component. The older age of the Kalliokangas tonalite is problematic because it was interpreted by Vaarma (1990) to have intruded later than the Veteli granodiorite (see above). The age is probably nevertheless a maximum age considering the heterogenous nature of zircon population (Vaasjoki 1994, pers. comm.). These features point out to an age of less than 1.91 Ga. The positive ε_{Nd} value of +3.9 with a T_{DM} of 1913 Ma for the Veteli granodiorite (Lahtinen and Huhma in prep.) indicate very rapid growth from DM source to evolved rocks. The small analytical error indicates a concordant age but the possibility of small component of inherited zircons exist and hence a slightly younger age is also possible.

The mafic volcanics in the Evijärvi area occur as intercalations within turbiditic metagreywackes-pelites and have been divided to different associations with WPB-E-MORB, N-MORB-low-K IAT and transitional IAT to CAB geochemical features (Vaarma 1990). The slightly depleted LREE nature and high ε_{Nd} values (+2.9 - +4.5) of two basalt samples support a depleted mantle origin (Huhma 1986) and is in accordance with the interpretation by Vaarma (1990). This could indicate a slightly different source for Kalliokangas and Aukeanneva (partly WPB) and Veteli (more calc-alkaline in nature). The main currently known characteris-

tics of the BSB are the occurrence of volcanics with both marginal basin (back-arc ?) affinities and island arc affinities (Vaarma 1990). He proposed a general stratigraphic scheme with rocks of island arc affinity being the youngest volcanics but there is some field evidence to suggest the opposite as well. Although some inconsistencies exist concerning the isotope data, an age ≥ 1.91 Ga is proposed for these volcanics. The plutons studied so far would indicate an increase in maturity and crustal thickness when progressing eastwards into the CFGC. The Sm-Nd data for the Rautio batholith and Veteli granodiorite support very rapid development of mature island arc/continental margin type crust from depleted mantle.

Tampere-Hämeenlinna area

The geochemistry of sedimentogenic, volcanogenic and plutonic rocks in the Tampere-Hämeenlinna area have studied by Lahtinen (in prep.) and only a brief summary will be given here. The well-preserved Tampere Schist Belt (TSB) has been actively examined since the time of Sederholm (1897) and includes numerous subsequent studies (see Kähkönen 1989 and Nironen 1989a for references). The volcanics are dominated by calc-alkaline medium-K and high-K intermediate rocks that vary from tholeiitic low-K to medium-K rocks to shoshonites with an overall similarities to mature island arc or continental margin rocks (Kähkönen 1989). The age of these volcanics have been bracketed between 1904-1889 Ma (Kähkönen et al. 1989). The Haveri Ti-rich basalts may indicate initial island arc stage (Mäkelä 1980), or they represent E-MORB or marginal basin basalts (Kähkönen 1989). The age and stratigraphical position of the Haveri Formation is still open, although the 1.99 Ga whole-rock Pb-Pb age implies an older age (Vaasjoki and Huhma 1987). The age data for the Hämeenkyrö granodiorite (1885±2 Ma, Nironen 1989a), Värmälä granodiorite (1878±3 Ma, ibid.) and Varissaari gabbro (1885±5, Patchett and Kouvo 1986) show slightly younger and geologically reasonable ages for intrusives in the TSB. The Nokia batholith contains a heterogeneous zircon population, indicating some inheritance (Nironen 1989a). The age of plutonic clasts in the metaconglomerates, high in the stratigraphy, are 1890-1884 Ma (Nironen 1989a), while an Archean granitoid cobble has been dated from the Ahvenlammi conglomerate (Kähkönen and Huhma 1993), low in the stratigraphy. This confirms the existence of Archean provenance, at least within the lowermost TSB sediments, as found in the detrital zircons (Huhma et al. 1991 and Claesson et al. 1993).

The geochemistry of the Hämeenkyrö, Värmälä, Pukala and Majajärvi plutons is characteristic of granites formed in mature island arc/continental margin environments. The ε_{Nd} values of Hämeenkyrö granodiorite (+ 0.1, Patchett and Kouvo 1986), Varissaari gabbro (+ 1.3, ibid.) and dacite at Ylöjärvi (- 0.7, Huhma 1987) could either indicate a substantial subducted Archean sediment component or the existence of older crust and/or EM. The available Nd-Sm data (ε_{Nd} -1 - 0, Huhma 1987) and detrital zircon data (Huhma et al. 1991, Claesson et al. 1993) from the TSB indicate a minor Archean and predominantly Proterozoic (1907-2039 Ma) source for sediments. This would therefore require unrealistic amounts of sediment subduction in a two component model involving depleted mantle combined with a subduction component at 1.9 Ga. The preferred model therefore involves the presence of 2.0-2.1 aged crust and a combination of three components (mantle, subduction component and pre-existing crust of 2.0-2.1 Ga age) in different ratios in different rocks of the TSB. The geochemistry of volcanics indicates that the mantle component was not pure DM type and

a more or less EM-type mantle under the TSB is proposed instead (Lahtinen in prep.). The Hämeenkyrö pluton intruded before the end of D_1 while the Värmälä pluton was syntectonic with respect to D_1 and both acted as rigid bodies during the late stage of D_1 deformation (Nironen 1989a). The beginning of the collisional event can thus be constrained to between 1885-1890 Ma.

There are no age data for the two small granitoid intrusions in Luhanka area, which represents the probable eastward continuation of the TSB. An age of 1892±3 Ma (Kallio 1986) has been obtained from feldspar porphyritic granodiorite south of these intrusions. The Sydänmaa intrusion includes quartz diorite of high-K calc-alkaline aspect, granodiorite of crustal origin and mafic rocks of WPB origin associated with a subduction or assimilation component. The Hernemäki monzogranite probably originated probably at a higher crustal level and also has an evolved crustal origin (p. 77).

The Mica gneiss-migmatite Belt (MB) between the TSB and HSB is characterized by abundant metasedimentary rocks that in the northern part are comparable with sediments in the TSB (Lahtinen in prep.). Volcanics are normally lacking in the northern part, but MORB-affinity volcanics are found from the southern part. These and the spatially associated mafic/ultramafic plutonic bodies probably mark a suture in the southern MB. Granitoid compositions record sediment assimilation and can be divided into high-K calc-alkaline types and to those having a within-plate component. Alkaline affinity WPB dykes are the youngest rocks in the MB (ibid.).

The Hämeenlinna Schist Belt (HSB) is characterized by island arc type volcanics followed by mafic volcanics with gradually increasing within-plate affinity (Lahtinen in prep., see also Hakkarainen 1990). The granitoids normally show calc-alkaline I-type characteristics with greater abundance of medium-K variants near the main Häme volcanic belt. The Aulanko granodiorite (1886±14 Ma, Patchett and Kouvo 1986) in the HSB has been assigned an origin by melting of calc-alkaline basaltic to basaltic andesitic island arc rocks (p. 72). The ε_{Nd} value of +1.9 (Patchett and Kouvo 1986) as well as geochemistry indicates rather rapid recycling of juvenile material and different evolution history and less mature crust under the HSB compared to the TSB. This is also seen in geochemical differences between island arc-type volcanics in the HSB and TSB (Lahtinen in prep.).

Central Finland Granitoid Complex (CFGC)

The CFGC is a poorly defined area of mainly plutonic rocks separated by minor schist belts. The plutons range compositionally from gabbro to the more voluminous granites (Front and Nurmi 1987). Mingling of mafic and felsic magmas is common and indicates that magma mixing was probably locally important (Lahtinen in prep.). The geochemistry of evolved granitoids (> 60 % SiO_2) in the CFGC reveals higher FeO/MgO and K_2O/Na_2O ratios than the granitoids in schist belts, but otherwise they have typical I-type characteristics (Front and Nurmi 1987). The same features were evident in granitoids to the W and N of Rautalampi and from Mäntylä hypersthene granitoid complex (included in the study of Front and Nurmi 1987). The hypersthene granodiorites of the Mäntylä complex (1886±6 Ma, Nironen and Front 1992) have been considered to represent melting of evolved calc-alkaline rocks (p. 78). The gabbros (1882±2 Ma, ibid.) and related rocks are difficult to interpret, but could have originated by mixing of within-plate magma with crustally derived rocks/melts. The granites of the Palokka plutonic complex indicate an evolved igneous source and the associated diorites and quartz diorites are of within-plate origin (p. 77). The Lehesjärvi microcline granite (see

below) is the youngest phase (Nurmi et al. 1984).

The ages and ε_{Nd} data from other granitoids in CFGC are as follows; Koppelojärvi granite (1879±14 Ma and -0.4, Patchett and Kouvo 1986), Keuruu granodiorite (1883±14 Ma and -0.2, Huhma 1986), Kiukaanniemi granodiorite (1883±17 Ma and +1.2, ibid.), Hankavesi granite (1886±10 Ma and + 0.7, ibid.) and Lehesvuori granite (1889±15 Ma and -0.9, ibid). There are also age data from the Mustajärvi supracrustal formation within the CFGC, namely felsic volcanics of ages 1907±13 Ma and rhyolite dated at 1872±12 Ma, with associated granite and plagioclase porphyrite having ages of 1893±6 Ma and 1883±15 Ma respectively (Vaasjoki and Lahti 1991). The available age data do not show any clear age zonation for granitoids, but indicates the existence of two volcanic cycles (ibid.). The Sm-Nd data of granitoids is not so homogenous as age data, but do not show any evidence for Archean crust (Huhma 1986). The Sm-Nd data and galena Pb isotopic data (Vaasjoki 1981) have been interpreted as representing a mixture of Archean and juvenile components (Huhma 1986), but as discussed above, this now seems unrealistic and the occurrence of ≥ 2.0 Ga aged crust is proposed instead (Lahtinen and Huhma in prep.). Lahtinen (in prep.) have studied the granitoids occurring to the N of the TSB. Some gneissic granitoids are probably cogenetic with the volcanics and an age of about 1.90-1.89 Ga is proposed. The high-K calc-alkaline intermediate granitoids are about 1885 Ma in age. The high-K granites show some tholeiitic features and have an interpreted age of about 1880 Ma. An interesting feature is the abundant occurrence of very high-K granitoids of tholeiitic affinity forming a continuous belt north of the TSB, and coinciding with a prominent regional magnetic anomaly (Lahtinen and Korhonen in prep.). Their origin is attributed to melting of WPB affinity rocks in the base of the crust. The Koppelojärvi granite belongs to this group but also shows effects of crustal contamination. The U-Pb zircon age of 1879±14 Ma (Patchett and Kouvo 1986) has high analytical error and could indicate small amount of inherited zircon. This feature and the distinct post-tectonic nature, including a contact aureole (Sjöblom 1990) indicate a 1875-1870 Ma age for these within-plate affinity granitoids and are consistent with the presence of subcontinental lithospheric mantle beneath continental crust in the CFGC.

Other areas

The Käkövesi batholite indicates a mature arc environment with a genetic link to calc-alkaline/shoshonitic rocks, but the lack of age data for emplacement, deformation and metamorphism inhibits further interpretation and correlation. The same holds for Äitsaari tonalite with a proposed island arc source and for the Salosaari tonalite, which had a still more evolved source. The Mäntsälä pluton is interpreted to contain two different melts which have also mixed with each other, including a granodiorite melt with a crustal source and trondhjemite with a rather primitive island arc source. Age and isotopic data are needed to verify these interpretations and their genetic significance. The southernmost volcanic belts (KMB in Fig. 1) and plutons are considered only briefly here. The age data (see Suominen 1991 for more age data) and ε_{Nd} data from the Orijärvi granodiorite (1891±13 Ma and -0.7, Huhma 1986), Mörskar granite (1881±9 Ma and +0.2, Patchett and Kouvo 1986), Skåldo gabbro (1885±7 Ma and +0.1, ibid.) and Svartgrund pluton (1891±11 Ma and +0.4) indicate the existence of older crust in these areas and the low ε_{Nd} values of mafic rocks indicate either strong crustal contamination or the presence of heterogenous mantle. The galena lead isotopic

compositions from this belt (Vaasjoki 1981) can also be interpreted as representing crustal lead evolution. North of this belt is the Kalanti district, which is characterized by ε_{Nd} values of +3.1 - +3.2 in diorites/gabbros (normally 1872-1892 Ma) and lower ε_{Nd} values of +1.7 - +2.4 in trondhjemites (1892 Ma) with inherited zircons (Patchett and Kouvo 1986). This would indicate different crust compared to that to the south and this area could be probably correlated with the HSB. The Hyvinkää gabbro (1880±5 Ma) also has an intermediate ε_{Nd} value of +2.6 (ibid.).

The Forssa group volcanics to the south of the HSB are intermediate to felsic in composition with calc-alkaline island arc-affinities. Similar rocks have also been found from the HSB where they underlie tholeiitic Häme group volcanics (Hakkarainen 1990). The subaquatic mafic/intermediate to felsic volcanics in the Enklinge region also have an orogenic origin and are underlain by marbles, greywackes and volcaniclastic sediments (Ehlers and Lindroos 1990). The mafic volcanics in the Nagu-Korpo area interlayered with metasedimentary gneisses and show clear WPB-affinity, interpreted as indicating an initial rifting stage (Ehlers et al. 1986). The Orijärvi area is characterized by bimodal volcanics but andesites are also locally present at Kisko (Hakkarainen and Väisänen 1993). In the Orijärvi domain volcanics are succeeded by limestones, iron formations, pelites and volcanic conglomerates. The mafic volcanics have an LILE-enriched MORB signature consistent with an initial rifting stage (ibid.). The Aijala-Orijärvi area also hosted the now exhausted Cu-Zn-Pb deposits (Latvalahti 1979). These volcanic rift basins with shallow-water type lithologies have been reactivated repeatedly as zone of inherent weakness and were also the locus for later 1.84-1.83 Ga granite intrusion (mainly sheets) along subhorizontal ductile shears during transpressional deformation (Ehlers et al. 1993). The occurrence of low-Ti metabasalts (also pillow-lavas) and meta-andesites in the Pellinge area (Laitala 1984) with ε_{Nd} values from +0.9 to +2.4 at 1887 ±14 Ma (Patchett and Kouvo 1986) indicates the occurrence of either island arc or low-Ti continental WPB type magmatism at this time.

DISCUSSION AND SYNTHESIS: A PLATE TECTONIC MODEL

The Svecofennian has been thought to resemble convergent plate margins and different plate tectonic models (see Gaál and Gorbatschev 1987, Gaál 1990, Park 1985,1991, Ekdahl 1993 for references) have been applied, commencing with pioneering work of Hietanen (1975). The sequential accretion of two (Gaál and Gorbatschev 1987) or more island arcs (Park et al. 1984, Park 1985) to the N or NE has been postulated, although the timing of the latter model is not supported by age data from volcanic rocks (Vaasjoki and Sakko 1988). Gaál (1990) revised the model of Gaál and Gorbatschev (1987), based on the work by Ward (1987), to account for an earlier subduction stage to the west. A terrane collage model has been also applied (Park 1991). Ekdahl (1993) revised the model of Gaál (1986) and proposed prolonged subduction with three successive events in a N-NE direction towards the continental margin with back arc spreading. A very important consequence of the proposed multiple episodes is that magmatic, structural and metamorphic overprinting at 1.89-1.88 Ga destroyed older features and thus complicates the correlation of older events.

The Fennoscandian Shield has been divided into Archean and Svecofennian domains where the latter represents Paleoproterozoic crustal growth during the Svecofennian orogeny (Gaál and Gorbatschev 1987), formerly named as Svecokarelidic orogeny (Simonen 1980b). The

Karelian formations deposited or situated on the Archean Karelian Province have been traditionally divided into Sariolian, Jatulian and Kalevian units, with their formal lithostratigraphic significance still being a matter of debate (e.g. Väyrynen 1939, Simonen 1980b and Laajoki 1986). Because no consensus about the definitions of these units has been reached, they are used here in a broad context. Karhu (1993) used the carbon isotopes to classify the Karelian formations into four distinct stages. Stage I includes sedimentary carbonates which were deposited before the ^{13}C enrichment event (2.4-2.6 Ga) whereas stage II is poorly constrained. Stage III is well characterized by ^{13}C enrichment in sedimentary carbonates deposited on the Archean craton at about 2.2 to 2.1 Ga and correlates with Jatulian. Stage IV carbon isotopes show a drop in the ^{13}C values of sedimentary carbonates during the time interval 2.11-2.06 Ga. This drop starts from the uppermost Jatulian dolomite formations and continues throughout the lowermost Ludian formations. According to Karhu the boundary between Jatulian and Ludian, adopted from Sokolov (1980), coincides with an abrupt facies change from quartzite-dolomite (upper Jatulian) to graphitic shales and impure carbonates seen for example in the Kiihtelysvaara area (Pekkarinen 1979).

The massive Kalevian graywackes e.g. in Höytiäinen Province of southeastern Finland form autochthonous units (Ward 1988b) and occur above or are coeval with the Ludian group. On the contrary, Gaál (1990) considers these greywackes to represent young fore-basin fillings postdating 1.97 Ga ophiolites and older turbidites (upper Kalevian see below). These autochthonous graywackes in Höytiäinen Province are named as lower Kalevian to differentiate them of the allochtonous thrust sheets and nappe complexes of the Savo Province (Koistinen 1981 and Ward 1987, 1988b) which are considered as upper Kalevian (Kontinen and Sorjonen-Ward 1991). These monotonous upper Kalevian mica schists occur above the Outokumpu Assemblage (Koistinen 1981), whose age is constrained by the Horsmanaho gabbro dated by U-Pb zircon at 1972±12 Ma (Huhma 1986). A similar stratigraphic sequence is found in the Kainuu Schist Belt, in which the Jormua Ophiolite Complex has U-Pb zircon ages from metagabbro and metatrondhjemite of 1960±12 Ma and 1954±11 Ma respectively (Kontinen 1987). There is a problem with the boundary between Svecofennian supracrustals and Kalevian sediments in that Gaál and Gorbatschev (1987) consider Kalevian rocks as constituents of the Svecofennian orogeny. On the other hand Ekdahl (1993) proposes correlation of his Savijärvi Suite with the upper Jatulian. The Svecofennian has been divided to lower, middle and upper Svecofennian (Simonen 1980b) and this is in a broad sense accepted here, although the main focus is on regional variations rather than correlations.

2.5-2.1 Ga stage

This stage is only briefly discussed here (see e.g. Gaál and Gorbatschev 1987) and represents continental stage (supercontinent ?), without any clear evidence for wide spread felsic magmatism as is indirectly seen in the detrital zircon data (Huhma et al. 1991, Claesson et al. 1993, and Sorjonen-Ward et al. 1994). Layered mafic intrusions of about 2.5 Ga from the Kola Peninsula (Bayanova 1993) and of about 2.44 Ga from Kemi-Koillismaa Belt in northeastern Finland (Alapieti 1982) record the most voluminous magmatism in the early Paleoproterozoic time. The greenstone belts of northern Finland are probably early Paleoproterozoic but a late Archean age has also been proposed (see e.g. discussion in Gaál 1990). Mafic sills and dykes, also known karjalites because of their distinctive mineralogy and chemistry

(Väyrynen 1938, Vuollo 1994), are dated at 2.2 Ga and are widespread throughout the eastern and northern parts of the Fennoscandian Shield.

2.1-2.06 Ga stage

This stage is interpreted to represent the main rifting of continent (Fig. 29). The 2.1 Ga stage is exemplified by the widespread occurrence of tholeiitic dyke swarms over the whole Archean part of Fennoscandian Shield area (Vuollo 1994). This stage is also exemplified by intracratonic Höytiäinen Province comprising sporadic tholeiitic volcanism associated with chemogenic and pelitic deposition followed by coarse clastics (Ward 1987). The tholeiitic volcanism has been divided to older and younger metabasites (Pekkarinen and Lukkarinen 1991). The older metabasites in the Kiihtelysvaara-Tohmajärvi district have a U-Pb zircon age of 2100-2120 Ma (ibid., Huhma 1986) and the younger metabasites can possibly be correlated with 2062 Ma volcanism at Kuopio and Siilinjärvi (Pekkarinen and Lukkarinen 1991). The older metabasites at Kiihtelysvaara have continental WPB-affinity associated with crustal contamination (ibid.) but the contemporaneous Tohmajärvi complex show less evidence for crustal contamination (Nykänen 1992), possibly indicating the occurrence of rifting centre in small basin underlain by extended crust in this area. The Oravaara basalt, dated at 2.10 Ga by the zircon U-Pb method (Huhma 1986) from Tohmajärvi has a LREE depleted character with ε_{Nd} +2.6 (Huhma 1986), indicating depleted mantle origin and thus differs from the Kiihtelyvaara basalts, which have an LREE enriched nature (Pekkarinen and Lukkarinen 1991).

The Northern Ostrobothnia Schist Belt shows some lithological similarities with the Höytiäinen Province but the lack of age data from the associated volcanics precludes reliable correlation at present. On the other hand the age of felsic volcanic clast (2093±35 Ma) from the Koiteli Conglomerate Formation stratigraphically associated with Kiiminki Volcanic Formation indicate an age ≤ 2.1 Ga for this volcanic unit and overlaying greywackes (Honkamo 1988, 1989), while simultaneously indicating the occurrence of felsic magmatism at 2.1 Ga. Honkamo (1989) proposed an evolved oceanic basalt origin for the Vepsä volcanics but a WPB-origin affected by a small amount of contamination is also possible (see Fig. 11 in Honkamo 1989). The high analytical detection limit for Rb, Ba, Th and Ta hampers the interpretation but an overall MORB-like affinity was proposed for the volcanics occurring above the Vepsä Formation (ibid.). Although there are problems with the analytical data, the sporadic high K and possibly also Rb and Ba values in these volcanics do not favour a MORB origin in purely oceanic environment, and is instead consistent with crustal contamination or the presence of subduction component. This possibility was also considered by Honkamo (1989). The model of a basin developing in a cratonic or epicratonic rift environment with deposition of a thick sequence of greywackes and conglomerates, at least partly on Archean basement, was favoured by Honkamo (1989).

The felsic volcanics from the Koivusaari Formation in Siilinjärvi have a U-Pb zircon age of 2062±2 Ma (reference to Lukkarinen in prep. in Pekkarinen and Lukkarinen 1991). The Koivusaari Formation can be correlated with the mafic volcanic Vaivanen Formation at Kuopio (Lukkarinen 1990) which is underlain by arkosites with conglomerates, quartzites, dolomites with skarn and black schist interbeds and overlain by turbiditic greywackes (ibid. and Aumo 1983). The mafic volcanics show normally MORB or WPB affinity but one alkalic member with high levels of LREE (alkaline WPB ?)

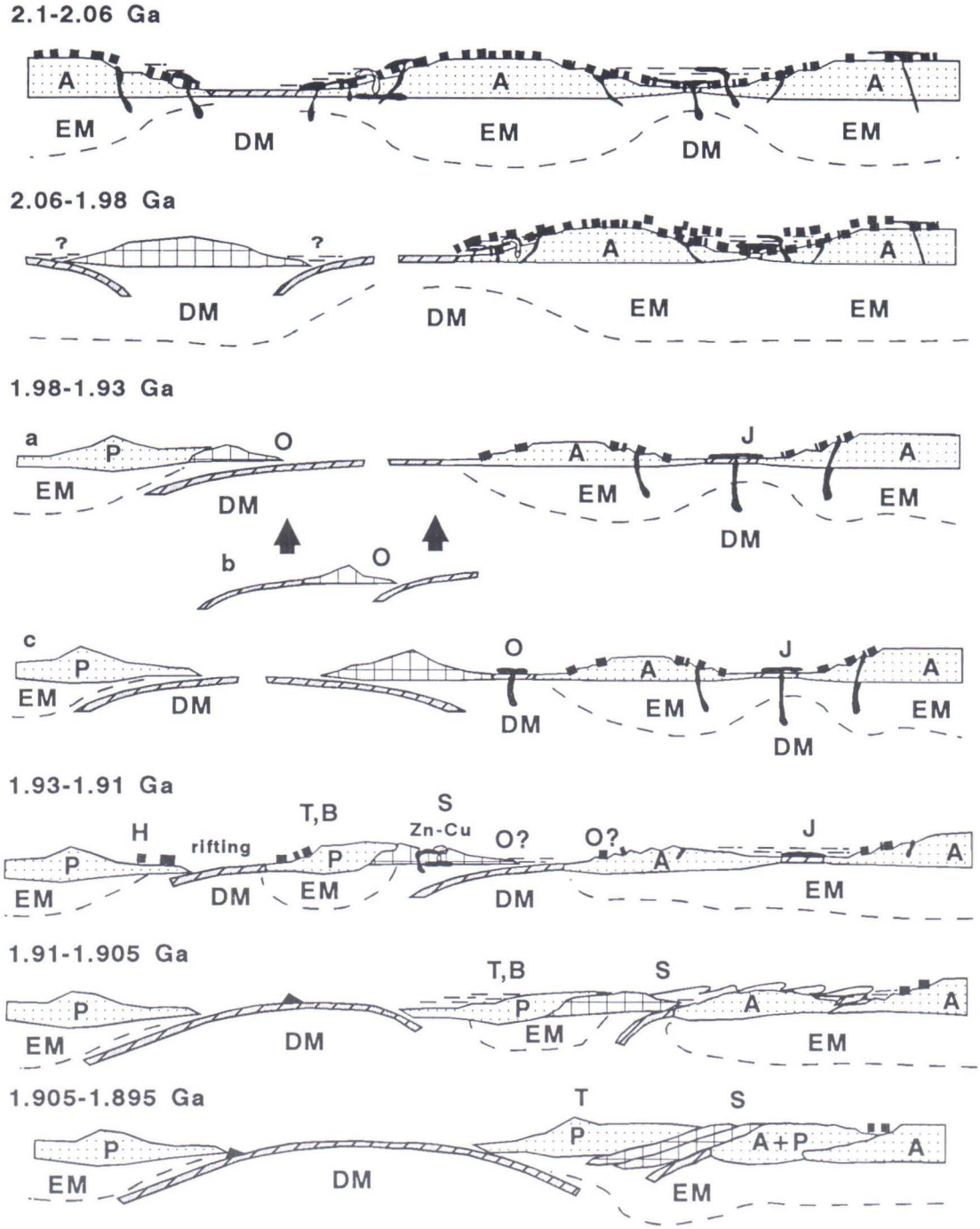
2.1-2.06 Ga
A
A
A
EM
DM
EM
DM
EM
2.06-1.98 Ga
?
?
A
A
DM
DM
EM
EM
1.98-1.93 Ga
a
P
O
A
J
A
EM
DM
EM
EM
DM
b
O
c
P
O
J
A
A
EM
DM
DM
EM
EM
DM
1.93-1.91 Ga
H
T,B
S
Zn-Cu
O?
O?
J
rifting
P
P
A
A
EM
DM
EM
DM
EM
1.91-1.905 Ga
T,B
S
P
P
A
A
EM
DM
EM
EM
1.905-1.895 Ga
T
S
P
P
A+P
A
EM
DM
EM

Fig. 29.

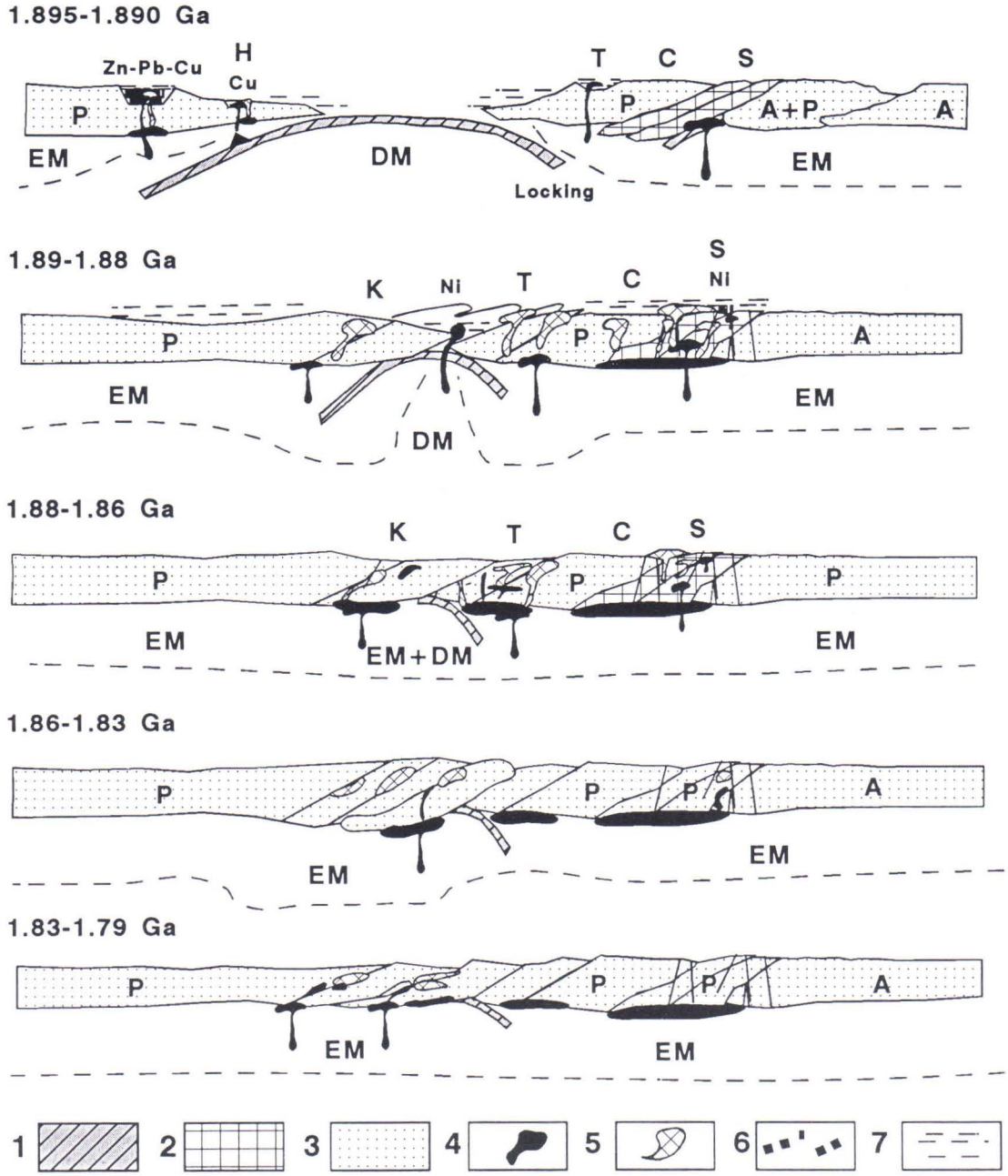

Fig. 29. Sequence of schematic profiles depicting 2.1-1.79 Ga plate tectonic evolution of the Svecofennian and Karelian domains. Not to scale. Only selected magmatic episodes are shown and subduction related magmatism is not shown.1 - Oceanic crust, 2 - primitive island arc crust, 3 - continental crust, 4 - mafic magmatism, 5 - felsic magmatism, 6 - cratonic/passive margin/rift sedimentation, 7 - turbidites. DM - depleted mantle; EM - enriched mantle; A - Archean; P - Proterozoic; O - Outokumpu; J - Jormua; S - SSB; T- TSB; H- HSB; B-BSB; C - CFGC; K - KMB.

occurs in the Koivusaari Formation (Lukkarinen 1990). The bimodal nature of volcanism favours a melting origin for the felsic volcanics, possibly from Archean basement, rather than fractionation. Lukkarinen (1990) proposed that the volcanic rocks in Kuopio and Siilinjärvi area erupted in subaqueous cratonic rift or rifted continental margin environment. Layered mafic intrusions in the Otanmäki area are dated at 2060 Ma (Talvitie and Paarma 1980) and also belong to this stage.

In this model (Fig. 29) the cratonic environment changed to continental margin environment at about 2.10-2.06 Ga ago. Voluminous mafic magmatism at 2.1 Ga affected large areas of the Archean craton as exemplified by the abundant occurrence of Fe-tholeiitic dykes (Vuollo 1994). Coeval and cogenetic continental floodbasalts have probably been also abundant although they are now uncommon at present erosion level. Minor felsic magmatism occurred also at this stage. Several rift basins (Ward 1987) formed at the same stage along the evolving continental margin. The upwelling of more depleted-like mantle produced MORB affinity volcanism in marginal basins either on extensively thinned continental crust or in small ocean basins. One of these basins developed into a wider ocean and marks the continental break-up line. Extension along the continental margin continued until at least 2.06 Ga with both mafic and felsic magmatism. The mafic magmatism was partly alkaline in nature. The lower Kalevian greywackes, at least partly, belong to this stage as proposed by Ward (1987) for the Höytiäinen province. The Höytiäinen province and similar lower Kalevian basins seem to have a depositional history dating back to > 2.1 Ga with sedimentation possibly continuing until 1.91-1.89 Ga and thus geochemical data on metasediments and especially ion microprobe single zircon age determinations are needed to resolve questions of timing, provenance and correlation.

2.06-1.98 Ga stage

This stage is the most hypothetical one in the present model since there are no dated igneous rocks from the Svecofennian or Karelian domains in Finland belonging to this age group. Detrital zircon ages from Finnish Svecofennian and upper Kalevian metasediments show that voluminous plutonic magmatism took place in this time interval either somewhere within the Fennoscandian Shield or within the provenance area supplying these sediments (Claesson et al. 1993). The results of this study and Lahtinen and Huhma (in prep.) are used to advocate the widespread occurrence of primitive and mature crust of 2.1-1.95 Ga underlain by both DM and EM respectively in different parts of the Svecofennian throughout the main 1.93-1.88 Ga magmatic events. The tectonic, magmatic and accretionary history of the 2.06-1.98 Ga stage is unknown and can have had many stages of subduction, collisions and rifting probably spatially unrelated to their present position near to the Archean craton. According to Ward (1988b) the Höytiäinen province was not deformed compressively before the initiation of Savo province (upper Kalevian) deposition. This would exclude the possibility of collision tectonics during this time period and indicates that the Archean craton margin records passive margin sedimentation during this period without any evidence as yet for magmatic or collisional activity.

As noted before igneous rocks belonging to this age group have not been found from the Svecofennian although the occurrence of basement about 2.1-2.0 Ga age in many areas seem to be inevitable, based on the geochemical and isotopical data (this study, Lahtinen and Huhma in prep.). On the other hand, Ekdahl (1993) correlates the Salo Metavolcanite, Salonsaari Suite and Savijärvi Suite in the Pielavesi area

with upper Jatulian, sensu Ekdahl, (2080-1970 Ma) deposits of the Karelian domain, based on similarities between stratigraphic sequences. The "marine Jatulian association", especially the U-P horizon, is considered by Ekdahl (1993) as the key horizon and he correlates the U-P horizons in the Savijärvi Suite and the Lampinsaari U-P horizon with Jatulian U-P horizons, such as U-P enriched layers in the Petonen Formation that were deposited prior to the extrusion of the Vaivanen Formation volcanics (possibly 2.06 Ga, Lukkarinen 1990). Ekdahl (1993) considers the U-P horizons in both Svecofennian and Karelian domains to represent a widespread shelf association in different basins, thus constituting a useful stratigraphic marker horizon. Marine phosphate assemblages characterized by volcanic sediments and bedded sulphides are also found associated with active tectonism near plate boundaries (Riggs 1986). Äikäs (1989) summarized the occurrences of phosphorites and phosphatic rocks in Finland and indicated the occurrence of different types and ages of deposition, thus militating against time-stratigraphic correlations between deposits. The close association of many U-P horizons in Svecofennian domain with massive sulphide deposits is also noteworthy (e.g. Ekdahl 1993). This is seen in Vihanti area where Lampinsaari phosphatic rocks are found to interfinger with and overlap with massive sulphide deposition (Rauhamäki et al. 1980). These phosphatic rocks have been considered as metamorphic derivatives of phosphatic felsic tuffs (Rehtijärvi et al. 1979). The carbon isotopes from sedimentary carbonates in the Savijärvi Suite also show different values when compared with Jatulian and Ludian sedimentary carbonates indicating different age and origin, providing that the secular change in carbon isotope compositions proposed by Karhu (1993) are valid. These data, combined with the 1.92-1.93 Ga age data for the Vihanti-Pyhäsalmi type massive sulphide deposits could therefore imply a similar age for U-P horizons associated with these deposits.

The Salonsaari Suite, including the Salo Metavolcanite is thus interpreted here as older than 1.92-1.93 Ga, although its exact age remains unknown. The occurrence of mica gneiss with conventional U-Pb zircon age 2234±69 Ma lower in the proposed stratigraphy (Ekdahl 1993) indicates that sediments probably had a mixed Archean and Paleoproterozoic source. There is therefore need for ion microprobe analyses from single zircons to reveal the maximum age of deposition but a tentative age of about 1.98-1.93 Ga is proposed. The evolution of rocks in Svecofennian and Karelian domains are considered to have had separate histories during this 2.06-1.98 Ga stage, with the Karelian domain recording passive margin evolution and the basement of Svecofennian rocks forming concurrently in a possibly exotic position.

1.98-1.93 Ga stage

This stage is exemplified by the Jormua ophiolite (1960±12 Ma, Kontinen 1987), Outokumpu ophiolite (Koistinen 1981, Vuollo 1994) with an age of 1972±18 Ma (Huhma 1986) and the 1965±10 Ma tholeiitic dykes described by Vuollo et al. 1992. An age comparable with the T2 dykes, of 1967±24 Ma, has also been obtained from albite diabase in the Nilsiä area (Paavola 1984). The T2 dykes occur within and near to the edge of the extended part of the Archean craton, intruding with Paleoproterozoic quartzites in Koli area. The Jormua ophiolite complex mafic rocks indicate a MORB origin in an incipient ocean, possibly at a passive margin or in an intracontinental rift environment (Kontinen 1987). The Outokumpu ophiolite hosts massive sulphide deposits and has geochemical affinities with supra-subduc-

tion zone (SSZ) ophiolites (Vuollo 1994, see also Park 1984, 1988). The presence of alkaline rocks of shoshonite and shonkinite affinity in the Outokumpu assemblage also suggest an intra-oceanic island arc origin for this assemblage (Park 1992). The T2 dykes have been interpreted as having similarities with island arc tholeiites and inferred connection with subduction (Vuollo et al. 1992). The geochemical data given by Vuollo et al. (1992) for T2 dykes show similar patterns to the 2.1 Ga T1 dykes on MORB-normalized diagrams (Pearce 1982) (data not shown), although at lower level, indicating a difference in degree of partial melting, as concluded by Vuollo et al. (1992). The similarities with the T1 dyke patterns, high level of Rb and inclined trend from Ba to Yb, together with the lack of a negative Nb anomaly is nevertheless more consistent with a WPB origin than subduction related origin. Also, the lack of subduction related magmas, especially those of more intermediate and silicic compositions also renders a subduction related origin less likely and instead a continental WPB origin is proposed for the T2 dykes .

Many ophiolites show island arc geochemical characteristics and have been interpreted to have formed by sea-floor spreading above a subduction zone (e.g. Pearce et al. 1984b, Elthon 1991 and Cawood and Suhr 1992). Forearc ophiolites are also considered to form the basement to some intraoceanic arcs and these are thus by definition older than the initiation of subduction (Charvet and Ogawa 1994). Pearce et al. (1984b) indeed suggested that most SSZ ophiolites form in a pre-arc setting during the initial stages of subduction and later became basement to fore-arc basins. Cawood and Suhr (1992) proposed a model for Appalachian SSZ ophiolites where ophiolite generation occurs close to an irregularly shaped continental margin and is synchronous with arc-continent collision, with emplacement closely following its generation. On the other hand ophiolitic emplacement can occur before continental collision as exemplified by the Spontang ophiolites, which were emplaced about c. 20 Ma before collision (Searle et al. 1987). Ophiolite emplacement is considered to be related to the subduction of a thin and rifted continental margin below relatively thin oceanic lithosphere (Dewey et al. 1988, Coward 1994). The SSZ Coast Range ophiolite was also accreted about 18-24 Ma after its generation (McLaughlin et al. 1988). Extremely oblique subduction can also produce a SSZ type ophiolite beneath a spreading ridge, as in the forearc basin of the Andaman Sea (Hamilton 1979). The Mineoka ophiolite has been interpreted as a forearc ophiolite with initial emplacement occurring during the early phases of arc-continent collision (Pickering and Taira 1994).

The Outokumpu and Jormua ophiolites currently occur within and beneath upper Kalevian turbiditic mica schists and at least the Outokumpu assemblage is allochtonous, while having acted in part as the depositional substrate for the mica schists (e.g. Koistinen 1981, Ward 1988b). The bulk of the allochthonous sequence is upward facing in Savo province and the apparent tectonic translation direction has been towards NE and NNE (Ward 1987,1988b). Although the total amount of tectonic translation is unknown the primary location of the ophiolite during generation was evidently some considerable distance southwest of its present position. If we consider the proposed SSZ nature we get at least 100 km displacement from the craton edge but this is probably a minimum estimate as the attenuated craton margin has probably been shortened during collision and later events (Kohonen et al. 1991). Detrital zircon ages from metagreywackes constrains maximum depositional ages to about 1.94 Ga and 1.92 Ga (Claesson et al. 1993) for upper Kalevian sedimentation above Jormua and Outokumpu ophiolites respectively. If the ophiolites were indeed the depositional substrate for upper Kalevian sediments these data indicate a time lag about 20-50 Ma between the generation of ophiolites and the onset of deposition of

upper Kalevian sediments. Accordingly this is the minimum time difference between ophiolitic generation and collision, and represents a somewhat longer time interval than that proposed or documented for typical younger SSZ ophiolites. On the other hand if the ophiolites were tectonically disrupted slices emplaced within upper Kalevian rocks the above considered arguments are not necessarily valid. The relationships between upper Kalevian sediments and the ophiolitic rocks as presently understood nevertheless support a primary depositional interpretation.

Three possible models for ophiolite generation during this stage are shown in Figure 29. Models a and b are rather similar, in which the Outokumpu ophiolite has been generated in a supra-subduction zone environment either near a pre-existing island arc far from its present position, associated with the initiation of a new subduction episode (a), or instigating the formation of an intraoceanic island arc near to the Archean craton (b). These models can be further divided into two alternatives, depending whether the Jormua ophiolite is correlated with the Outokumpu ophiolite or with T2 dyke generation. The T2 dykes record the occurrence of rifting and mafic magmatism in the Archean craton at this time and the possibility exists that rifting has culminated in marginal basin development (Jormua) with formation of a small ocean (Kontinen 1987). Although the age determinations show overlapping ages the slightly older age for the T2 dyke (1965 Ma) compared to Jormua (1960 Ma) fits this model. The Sm-Nd isochron age of 1980±27 Ma of gabbro intrusion from the Suisaarian formation in the Onega region (Pukhtel et al. 1992) indicate the occurrence of coeval and widespread magmatism further east within the craton. The slightly older age of Outokumpu ophiolite (1972 Ma) compared to Jormua and T2 dykes could indicate that its initiation near to the Archean craton (model b) was caused by undefined mantle process (mantle plume ?), also manifest by further affecting coeval mafic magmatism in the eastern part of the Archean craton. Other possibilities include the occurrence of thin oceanic lithosphere near to Archean craton or marginal basin associated with the generation of Outokumpu ophiolite. In these alternatives the Jormua ophiolite has been thrust into its to present position during collision.

Alternative c could also be referred to as a multi-rift model (Tamaki 1988) including all the 1.96-1.97 Ga magmatism. This model has similarities with previously considered models (e.g. Park 1988, Ekdahl 1993), but the main difference is the multi-rift nature. The multi-rift type includes complicated spreading systems possibly associated with oblique rifting (Tamaki 1988, Charvet and Ogawa 1994). In this model the Outokumpu ophiolite is a back-arc basin affected by subduction processes and later by island arc volcanism during termination of subduction. The Jormua ophiolite and the T2 dykes are instead related to rifting with basin development in the Archean craton. Although an elegant solution for explaining contemporaneous mafic magmatism and ophiolite generation this model has some major problems in that the island arc should have been accreted to the craton margin before the onset of subduction and extension that produced the multi-rift basins. According to Ward (1988b) there is no evidence for tectonic deformation in the Höytiäinen province prior the deposition and deformation of the upper Kalevian of the Savo province. The maximum depositional age of about 1.92 Ga for upper Kalevian sediments in the Outokumpu area also places a maximum age constraint upon collision. If the island arc system and the adjacent back-arc basin was not related with magmatism in the Archean craton, then the model becomes similar to model b. Although model c can not be totally discarded models a and b are favoured.

There are no comparable age data from the Svecofennian domain but the possibility exists that the Salonsaari Suite in the Pielavesi area (Ekdahl 1993) belongs to this stage. The WPB -affinity rocks in Savijärvi suite are also tenta-

tively related to this age group and possibly indicate a rifting event. The volcanics of the Haveri Formation in the TSB and the volcanics in the BSB could also be included within either this stage or the 1.93-1.915 Ga stage. The evolution of the Svecofennian domain is rather undefined during this stage but a back-arc spreading initiating about 1.96-1.92 Ga ago is proposed. This separated (suture, p. 93) the area including the SSB, BSB, CFGC and TSB from the area including the HSB, KMB (Fig. 1) and possibly also areas to the south e.g. crystalline basement of Estonia (see also Lahtinen in prep.).

1.93-1.91 Ga stage

The model for this stage is based on models a or b from the preceding 1.98-1.93 Ga stage. The Outokumpu ophiolite forms either the fore-arc basement for the island arc (model a) or is accreted into the accretionary prism (model b) of the same island arc in the west. An alternative model is the accretion of the Outokumpu ophiolite to the leading edge of the craton about 10-20 Ma before the collision. According to Koistinen (1981) the ophiolite assemblage and mica schists were affected by tectonic juxtaposition during pre-D_1 possibly indicating that the emplacement occurred either during pre-collision time or during initiation of collision. All three possible settings for the Jormua ophiolite are feasible but a passive margin marginal basin setting is preferred. The position of the upper Kalevian sediments are somewhat problematic. They were locally deposited directly on Archean basement (Gaál et al. 1975, Ward 1988b) and have been described as Svecofennian post-arc flysch (Park 1985), molasse from the Lapland granulite belt (Barbey et al. 1984), pericontinental turbidites including Kalevian as whole (Laajoki 1986). Kontinen and Sorjonen-Ward (1991) interpreted based on geochemical and petrographical data that upper Kalevian rocks both in the Kainuu Schist Belt and Savo province are similar and show effective mixing of detritus. They accumulated within a single large-scale depositional system associated with continental rivers supplying material to large trailing margin deltas followed by downslope sedimentation. The low CIA-values indicate immature sediment origin that is atypical for trailing edge sediments but Kontinen and Sorjonen-Ward (1991) proposed either rapid erosion or short distance to rapidly rising orogenic domain to explain this problem. They favoured the model of Barbey et al. (1984) and considered the upper Kalevian sediments to represent foredeep sediments related to uplift in Lapland and Kola Peninsula. However, although SHRIMP zircon data from a Lapland granulite sample reveal a similar age spectrum to that of Kalevian sediments, it is debatable whether the granulites were actually uplifted at 1.92 (Sorjonen-Ward et al. 1994). It is therefore possible that granulite protoliths and Kalevian share the same provenance, but less certain that the granulites were the source for the Kalevian.

The upper Kalevian is probably not a single entity and possibly includes rocks with different histories. In this model the onset of voluminous upper Kalevian sedimentation started either about 1.93-1.92 Ga ago due to collision in some remote hypothetical location or at about 1.91 Ga ago connected with collision from the west. One possibility for the hypothetical source area is the Osnitsk-Mikashevichi Belt, which has ages between 2.02 and 1.97 Ga (Shcherbak 1991) but the source area may not exist any more, at least not in its original position - it may have been totally destroyed, modified or tectonically transported to a far distant location.

The apparent absence of sediments connected with rifting at about 1.96 Ga in the Kainuu Schist Belt is problematic. Two possibilities

exists if a parautochtonous setting is correct. Either the marginal-uplift was not associated with severe erosion or syn-rift sediments accumulated only near basin margins and have either not been observed or they have been thrusted and eroded during later events. The upper Kalevian sediments above the Outokumpu ophiolite have a maximum depositional age of 1.92-1.93 Ga. An intra-oceanic island arc origin with accretion at this time, either to the accretionary prism of the island arc in the west, or alternatively to the leading edge of the craton, explains the apparent absence of sedimentation during 1.97-1.92 Ga (Fig. 29). The same holds for a fore-arc basement origin where the basement (ophiolite) in the accretionary prism or fore-arc region was later transferred to a position where it formed the substrate for sedimentation during initial collision at about 1.915-1.905 Ga. This could in principle be due to tectonic erosion caused by subduction of a seamount according to the Coulomb wedge model (von Huene and Lallemand 1990), accompanied by subsidence. Another possibility is the lack of an accretionary prism during Mariana-type subduction (Charvet and Ogawa 1994) before ocean closure at about 1.92-1.91 Ga, followed by a dramatic increase in supply of sediment to the trench during development of the accretionary prism.

Subduction took place to the west under the island arc with transitional crust. The mafic island arc basement was a collage of 1.94-2.0 Ga aged volcanics underlain by DM type mantle, at least under the SSB rocks (Fig. 29). At about 1.93 Ga ago typical island arc basalts and andesites were erupted in the SSB, as exemplified by mafic inclusions in gneissic tonalites in the Rautalampi area, andesites at Kangasjärvi and island arc affinity rocks in the Savijärvi Suite at Pielavesi. Shallow water conditions prevailed during this time with a lack of voluminous sedimentation during the initiation of basin formation. The U-P horizons in the SSB belong to this stage and are likely to be spatially related to the initiation of hydrothermal activity during subsidence. This was followed by rapid subsidence about 1.92 Ga ago in an extensional environment associated with bimodal magmatism. The felsic volcanics, hypabyssal sills and tonalitic sills have a cogenetic origin due to the melting of mafic island arc volcanics. This bimodal volcanism was accompanied by hydrothermal alteration and the formation of VMSD. Subsidence was not uniform in space and time, producing crustal blocks at different elevations. Some phosphatic deposition also coincided with sulphide mineralization, as seen in the Vihanti area. This scenario resembles the environments of formation of deep marine (2500 m) Kuroko deposits that are hosted by bimodal volcanics underlain by andesites and shallow water deposits (≤ 500 m), recording rapid subsidence (1-2 Ma) (Tanimura et al. 1983, Gruber and Merrill 1983). The aborted rift model proposed for the Kuroko deposits (Cathless et al. 1983) is also considered valid for 1.92 Ga VMSD in the SSB. The more Pb-deficient nature compared to the Kuroko deposits is due to the more immature nature of the SSB island arc, as reflected in low K values in both mafic and felsic rocks. The rifted and confined nature of basins associated with VMSD and related rocks are probably the reason for the preservation of these blocks during later events.

Magmatic activity also took place in the W, affecting the present basement for the CFGC. The area was probably underlain by more evolved crust derived from 2.0-2.1 Ga aged crust with the development of EM type subcontinental lithospheric mantle (Lahtinen and Huhma in prep.). The evolution of the Tampere-Hämeenlinna area has been considered in detail by the author in another paper (Lahtinen in prep.) and is only briefly discussed here. The TSB and the northern part of the Mica gneiss-migmatite Belt (MB) south of it is considered to represent an incipient rift basin during this time period. The occurrence of blastoclastic subarkose associated with diopside gneiss and limestone at Kangasala (Matisto 1976) is ten-

tatively correlated with this stage. The southern terrain including the HSB and KMB (Fig. 1) contain some quartzites, as at Tiirismaa and according to Neuvonen (1954) the lowermost units in the HSB contain limestones and quartzites. The metasediments from the lowermost part of the HSB show geochemical evidence for strong weathering and a long recycling history (Lahtinen in prep.). Although no age data are available, at least the lowermost sediments in the HSB are considered to belong to this stage. It is proposed that at about 1.920-1.915 Ga southward dipping subduction under the southern terrain commenced, although there are no age data to confirm this. It may be that magmatic activity started somewhat later, at about 1.91-1.90 Ga, but the occurrence of a U-Pb zircon age of 1918±10 Ma from metavolcanic in the Tapa zone of northern Estonia (Petersell in prep.) could be included to this stage.

1.910-1.905 Ga stage

This stage is characterized by the collision of the western arc with the Archean craton and partly overlaps with the 1.93-1.91 Ga stage (Fig. 29). The initiation of collision probably commenced slightly earlier and continued until 1.90-1.89 Ga, as discussed later. Arc-trench systems are often curved and collision is often oblique or involves strike slip movements (Coward 1994). Oblique collision starting in the N, associated with rotation and later collision in the S is the proposed model for the collision between western arc and Archean continent. The collision possibly started as thin-skinned thrust sheets with associated collisional foredeep sediments. Early collision-related sediments can be subducted below the overthrusting plate during arc-continent collision and thus the earliest preserved sediments can postdate the onset of collision (Coward 1994). The earlier foredeep sediments can also be thrusted later and form a component in later foredeep sedimentation due to migrating large-scale ramping of overthrust faults (Molnar and Lyon-Caen 1988). Flexural extension of the upper parts of continental crust often form deep foredeep basins (Bradley and Kidd 1991) preserved under later thrusts during foreland propagation. The upper Kalevian rocks and possibly also some metasediments occurring in the SSB are thus interpreted, at least partly, to represent accretion prism sediments thrusted on the foreland and also as foredeep sediments preserved under the propagating major thrusts. The late advance of the thrust belt has driven the foreland subsidence in front of it (ibid.) and thus the youngest foredeep deposits should occur further to the east. The amount of crustal shortening is unknown but shortening factors up to 50-80% are possible during mountain belt generation (Le Pichon et al. 1982, Coward 1994). This would indicate a 400-1000 km wide zone from the unstretched part of craton to the hinterland part of island arc during initial collision. These are probably minimum values as later 'mega shear' tectonics also lead to crustal shortening.

During this stage or slightly earlier a hinterland basin formed in the BSB and TSB infilled with debris eroded from the mountain belts. The position of the TSB is either due to rotation and/or oblique rifting during this time or earlier (1.91-1.98 Ga). Oblique convergence can cause initial splitting and rifting oblique to the arc trend (e.g. Sibuet et al. 1987, Jolivet et al 1989). The age of Haveri Formation volcanics and the MORB affinity volcanics in the BSB occupy in key positions in defining the exact timing. The Haveri Formation volcanics show evidence of an EM type mantle and crustal environment (Lahtinen in prep., Lahtinen and Huhma in prep.) and thus differs from the volcanics with MORB affinity in the BSB (Huhma 1986 and Vaarma 1990). However, they could be correlated with volcanics with WPB affinity in the BSB (Vaarma 1990), indicating the ini-

tial stage of rifting. Volcanics with MORB affinity have also been found in the Vammala area (Peltonen submitted) in the MB, and also from other areas from the southern part of the MB (Lahtinen in prep.). These are not pure MORB rocks but nevertheless relate to the initiation of ocean basin development in Figure 29 and at present, probably mark a suture (ibid.). The collision started to affect the hinterland basin in the BSB with adjacent thrusting about 1.91 Ga. The collision probably started prior to 1.91 Ga according to the intrusion of the Veteli granodiorite (1913 Ma) during D_2 (Vaarma 1990) although the possibility exists that a small amount of inherited zircons is present in the Veteli granodiorite. An origin for the Veteli granodiorite by melting of evolved island arc rocks with DM affinity, and associated calc-alkaline intermediate magmatism reflect the latest phases of island arc magmatism at 1.91 Ga. The Saunakangas tonalite (1903±10 Ma, Huhma 1986) in the Pieksämäki area migmatizes paleosomes associated with D_1-D_2 deformation (Korsman et al. 1988) and thus indicates that collision in the southern part of the SSB started prior 1.90 Ga. The position of the Viholaniemi metavolcanic with age 1906±4 Ma (Vaasjoki and Sakko 1988) remains uncertain - it could be either correlated with the youngest volcanic phase in the SSB island arc or with earliest phase of TSB volcanism.

The hinterland basin in the TSB, possibly associated with oblique rifting, was not affected by collision due to this area being an extensional environment. The stratigraphy of the main greywacke association above the lower volcanics in the TSB is rather well constrained (e.g. Kähkönen and Leveinen in prep.). The age of deposition of the main metagreywacke association in the TSB, from about 1.91 Ga to 1904 Ma, is also well constrained by the detrital zircon data (Huhma et al. 1991, Claesson et al. 1993) and age data from the overlying occurring metavolcanics (Kähkönen et al. 1989). The detrital zircons in sample A1 contain Paleoproterozoic 1.91-2.01 Ga (70%) and late Archean 2.71-2.75 Ga (23%) population (Huhma et al. 1991). Sample A57, which is also from the TSB has a very similar Paleoproterozoic 1.92-2.04 Ga (63%) population but the Archean zircons (37 %) are normally older 2.92-3.44 Ga with only one zircon having an age of 2.77 Ga (Claesson et al. 1993). These samples show differences in the source area during erosion, as also reflected in the change in geochemistry of basement related metasediments in the TSB and MB (Lahtinen in prep.). The recycled sediment nature with quartz depletion (ibid.) and the occurrence of Archean detritus in these basement related sediments show that the source was composed of a mixture of earlier sedimentary cover and volcanic and plutonic rocks formed during the 2.0-1.91 Ga evolution of the island arc. The occurrence of an Archean cobble in the Ahvenlammi conglomerate in the lower part of the main metagreywacke association (Kähkönen and Huhma 1993) indicates either proximity to Archean source rocks or multiple transportation and reworking.

Although no age data is available, continuing subduction to the south is postulated, with some island arc affinity volcanics from the HSB and KMB (Fig. 1) possibly belonging to this stage. The same applies to some rocks of the crystalline basement in Estonia which have similar Sm-Nd data to those of Svecofennian rocks (Puura and Huhma 1993).

1.905-1.895 Ga stage

During this main collision stage between the western island arc and the Archean continent the tectonic style changed to thick-skinned deformation that also involved Archean rocks. In Figure 29 the thickening of arc crust is also indicated, as crustal scale thrust sheets possibly

affected lithospheric mantle as well (see e.g. Coward 1994). It is not clear if collision continued or ceased due to onset of subduction in the south (see below) but the syntectonic Alpua gabbro (1901±12 Ma, Vaasjoki and Sakko 1988) and the migmatizing Saunakangas tonalite in Pieksämäki area (1903±10 Ma, Huhma 1986, Korsman et al. 1988) belong to this stage. The difficulty of correlating the Viholanniemi metarhyolite was considered above but the similar and positive ε_{Nd} values of +3.3 and +3.4 for Saunakangas tonalite and Viholanniemi metarhyolite indicate very short crustal residence time and derivation from newly formed island arc rocks (Lahtinen and Huhma in prep.). The heterogenous Onkivesi granodiorite, considered to be the youngest intrusion in the Iisalmi region, has an U-Pb zircon age 1908±16 Ma (due to heterogeneity ?) and possibly record the presence of an older Archean component (Paavola 1988). The age of foliated quartz diorite from the boundary zone of Archean craton and the western arc at Maaninka is 1902 Ma (ibid) and is tentatively correlated with the previously considered syncollisional intrusions.

The TSB volcanics, excluding the Haveri Formation volcanics, have ages ranging 1904 Ma to 1889 Ma, which well correlate with the inferred stratigraphic sequence (Kähkönen et al. 1989). The TSB volcanics show mature continental island arc (CIA) or active continental margin affinities (Kähkönen 1989, Lahtinen in prep.) and the Sm-Nd data indicate the occurrence of EM type subcontinental lithospheric mantle beneath crust about 2.0 Ga in age (Lahtinen and Huhma in prep.). The main collision stage in the NE was accompanied by reversal of arc polarity in the oblique rift basin and the onset of subduction under the TSB at about 1905 Ma ago (Lahtinen in prep.). This subduction event was a short-lived one (15-20 Ma) and ended about 1.89 Ga ago, as discussed below. The volcanism was accompanied by deposition of sediments having either a direct arc origin or recording mixing with older sediments (ibid.). The Hirsilä Schist Belt just north of the TSB is included in this stage. The CFGC contains some small supracrustal formations and one rhyolite sample from the Mustajärvi area gives an age 1907±13 Ma (Vaasjoki and Lahti 1991) which is also tentatively correlated with the TSB volcanism at this stage. Subduction to the south probably continued during this stage although no comparable age data from volcanics are available. One gneissose granodiorite from Algersö has a U-Pb zircon age 1899±8 Ma (Suominen 1991).

1.895-1.890 Ga stage

There are no age data from the SSB for this stage and it is thus unclear whether the SSB underwent compression or not during this time. One possibility is that subduction roll back (Dewey 1980) in the TSB area led to termination of the collision but this is highly speculative. The 1895±15 Ma age from the Lapinlahti gabbro in the Archean Iisalmi block (Paavola 1988) could indicate the onset of magmatic underplating at 1.89 Ga (see below), which also affected the margin of the Archean craton. Lahtinen (in prep.) proposes a tentative model of ridge subduction under the HSB being responsible for the earliest deformation phase in some volcanics noted by Jokela (1991). This probably commenced about 1.895 Ga ago and records the initial stages of collision. Subduction of the oceanic ridge was associated with rifting in the fore-arc region producing the main tholeiitic association (Häme group) preceded by hydrothermal activity in rift basin sediments (Lahtinen in prep., see also Hakkarainen 1990). As the rifting proceeded mafic rocks showing no subduction component and having continental WPB affinity were formed (Lahtinen in prep.). The upper age limit of the

main volcanic formation in the HSB is probably constrained by the U-Pb zircon age 1888±11 Ma (Vaasjoki 1994, pers. comm.) obtained from felsic volcanics that according to Neuvonen (1954) occur above the main volcanic formation. Marginal basins on the active continental margin further south were formed showing intermediate MORB-WPB affinities (Ehlers et al. 1986 and Hakkarainen and Väisänen 1993) during this stage or earlier. The VMSD and associated bimodal volcanism in the Orijärvi area are also included in to this stage. The age of the volcanism in the Orijärvi area is uncertain but is probably bracketed by the intrusive Orijärvi granodiorite (1891±13 Ma, Huhma 1986) and by detrital zircon data (Claesson et al. 1993) that indicate 1.93 Ga as maximum deposition age for the Orijärvi metagreywackes. The tholeiitic and komatiitic metavolcanics in the Rantasalmi area (Kousa 1985) occur stratigraphically above iron formations and intermediate and felsic volcanics (Makkonen and Ekdahl 1988) and are tentatively correlated with these marginal basins.

According to Lahtinen (in prep.) a tensional stage in the TSB at about 1.89 Ga produced the Veittijärvi-type conglomerates with plutonic clasts having similar age. The sediments of this stage reflect a direct arc provenance with large variations in chemical composition, favouring the existence of isolated basins with local sediment sources (ibid.). The Upper Volcanic Unit at Ylöjärvi (Kähkönen 1989), which overlies these sediments is correlated with this stage and the U-Pb zircon age of 1889 Ma (Kähkönen et al. 1989) indicates its approximate age. The locking of the subduction slab caused by downwarping, possibly due to onset of collision and subduction of an oceanic ridge beneath the HSB, is tentatively proposed as the reason for the cessation of subduction under the TSB about 1.89-1.88 Ga. This changes the position of the TSB and CFGC from overriding plate to passive underthrusting plate subducting and colliding to the south. This is a one possible model for stabilizing the proposed northeastward-dipping subduction slab (BABEL Working Group, 1990) into the continental mass and maintaining the major conductivity anomalies seen in the same areas by Korja et al. (1993).

1.89-1.88 Ga stage

This is the main collision stage between the TSB and CFGC in the N with the area to the S (including HSB and KMB in Fig. 1). The collision also affected the SSB and adjacent Archean craton and its cover. The thick-skinned deformation style related with oblique collision associated with rotation changed to strike-slip and shear tectonics modifying and partly coinciding with previous structures. The thrust and fold belts in many mountain belts are crossed by such features (see discussion and references in Coward 1994) and similar overprinting of the earlier low-angle structures, affecting also Archean basement, by shear systems in North Karelia is proposed by Koistinen (1981), Ward (1987), Ward and Kohonen (1989) and Kohonen et al. (1991). Due to crustal and lithospheric thickening the Raahe-Ladoga Zone (RLZ) changed from a zone of compressive deformation to an extensional environment despite the collision in progress to the south. This model is similar to that proposed by England and Houseman (1989) for the Himalayan collision of India and Asia.

The RLZ represents a major shear system initiated at about 1.89 Ga, and which was active for a long period of time. Other shear zones affecting the Fennoscandian Shield have also been described (e.g. Berthelsen and Marker 1986, Ward et al. 1989, Kohonen et al. 1991 and Kärki et al. 1993). Ductile shearing changed to brittle deformation during uplift as seen for example in the brittle Kinturi fault that cuts the Kumiseva gabbro, which has a U-Pb

zircon age of 1879±5 Ma (Ekdahl 1993). Although the RLZ partly follows the proposed suture zone, it cuts across it in many places and affects also both the CFGC and the area underlain by Archean basement. During this stage compressional, tensional and strike-slip tectonics operated, producing very complex tectonic histories differing from place to place.

In this model the 1.89-1.88 Ga rocks in the SSB were produced by an intense underplating event subsequent to the accretion of the pre-1.90 Ga island arc (SSB) and associated continental crust (GFGC) to the Archean craton to form tectonically thickened mixed crust. The magmatic underplating commenced at about 1.89 Ga and the remnants of this magmatism (layers of mafic rocks and possibly also dense restitic products of intracrustal melting) are seen as high velocity zones in the lower part of the very thick crust in this area (e.g. Luosto 1991). The basic magmatism has been mainly within-plate type (p. 89), originating from subcontinental lithospheric mantle (EM type) and records the change from DM originated melts to EM originated melts, as a result of collision. This WPB magmatism affected the crust, generating mainly tonalitic rocks from island arc material (of 2.0-1.91 Ga age) and granodiorites and granites from more evolved crustal rocks by melting at the crust-mantle boundary. The tonalites showing more primitive origin are confined to the SSB and granitoids from more mature crust characterize the CFGC. The tonalite dikes in the Archean Iisalmi block show a mixed age population with younger zircons of 1924-1880 Ma age (Paavola 1988), possibly recording the effects of magmatic underplating in the marginal part of the Archean craton. The crust has been exceptionally thick beneath parts of the SSB according to inferred presence of residual garnet (p. 71) in source to the Varparanta stock. Melting also occurred higher in the crust, as for example with the Kaartila pluton. Some of the granodiorites and granites having within-plate affinities were produced by melting of within-plate underplate. The presence of hypersthene could indicate melting of dry rocks or flooding by CO_2 accompanied by within-plate magmatism. This model is very similar to the rifting model of Korsman et al. (1984) and can also account for the block structure and high-T metamorphism linked, at least partly, with the hypersthene granitoids and high heat flow due to magmatic underplating. The Vaaraslahti hypersthene granite (1884±5 Ma, Salli 1983) has a metamorphic contact aureole overprinting regional granulite facies metamorphism (Hölttä 1988) but the age of the proposed older metamorphic event is not precisely constrained. The Ni-bearing intrusions in the RLZ belong to this stage and are genetically related to the within-plate magmatism. The difference between Rantasalmi-Sulkava district and areas further to the north and northwest in the SSB is seen in a tectono-metamorphic discordance recorded by the deformed (D_2) and metamorphosed conglomerate at Haukivuori, which contains clasts of 1885±6 Ma age (Korsman et al. 1988), indicating the occurrence of a younger collisional event (see discussion below).

This magmatic underplating affected also partly the CFGC as discussed above but the origin of younger 1.88 Ga volcanism in the northern part of the SSB and CFGC is open. They could be either volcanic rocks extruded during the extensional event or connected with the terminal phase of subduction further south under the TSB at 1.89-1.88 Ga ago if not the result of both processes. The Settijärvi Conglomerate, correlated with conglomerates at Haukivuori (see above), with plutonic clasts dated at 1888±7 Ma (Vaasjoki and Sakko 1988) underlies such younger volcanics (Isohanni et al. 1980) and indicates rapid erosion. One possible model is the occurrence of small strike slip basins associated with the RLZ tectonics. Small strike-slip basins subside very rapidly with subsidence rates exceeding sediment supply rates (Woodcock and Schubert 1994) and thus their preservation potential is probably good. The same interpretation is given to for

other conglomerates in the SSB that occur high in the stratigraphy.

The collision between the northern continental island arc/active continental margin (TSB), which was the underthrusting plate, and the overriding southern active continental margin with its associated marginal basins (e.g. at Nagu-Korpo and Orijärvi within the KMB, Fig. 1), started to influence the southern part of Finland at about 1.89 Ga ago or slightly earlier. This is seen in the early 1.89 Ga sill-like granitoids that were intruded along S_1 axial-plane schistosities (Ehlers et al. 1993). The Hämeenkyrö (1885 Ma) and Värmälä plutons (1878 Ma) in the TSB are syntectonic with respect to D_1 (Nironen 1989a) and therefore possibly indicate slightly younger ages for the onset of collision related thrusting in the TSB compared to further south. The suture zone (p. 93) is considered to have approximately coincided with the belt associated containing volcanics of MORB affinity and sedimentary rocks characterized by low S/Se ratios (Lahtinen and Lestinen in prep.). The proposed suture, at least partly, corresponds to the Vammala Ni-belt. The intrusion of the Vammala ultramafic cumulate-textured bodies coincided with the peak of regional metamorphism and deformation (Peltonen submitted) and have ages about 1.885 Ga or somewhat younger. Peltonen (submitted) interpreted these intrusions as representing middle crustal expressions of arc magmatism but the data can instead be interpreted to indicate assimilated DM derived melts, as has been proposed for similar intrusion further east in the same belt (Lahtinen in prep.). The suture in the southern part of the SSB between the Kiuruvesi and Haukivesi complex and the Rantasalmi-Sulkava area, as proposed by Korsman et al. (1988) could be the eastern continuation of this zone. The presence of such suture is also favoured by the geochemical and isotopic data that indicate that the TSB differs from the HSB in having a thicker and more mature crust (Lahtinen in prep.). The detrital zircon data from the TSB and upper Kalevian sediments contain only 8-14% of zircons of the 2.0-2.1 Ga age, in contrast with a sample from Orijärvi that has 58% of zircons in this age group. This indicates a different source area in the KMB, at least at Orijärvi, compared e.g. to the TSB thus favouring the occurrence of ocean between them.

A shear zone between the TSB and the MB to the south was interpreted by Nironen (1989b) as a south dipping thrust, possibly related to southward subduction to the south. In this model (Fig. 29) it is interpreted as a reactivated trenchward dipping backstop or backthrust that subsequently took part in the northward vergent thrusting. The continent-continent collision was probably associated with crustal-scale thick-skinned thrust sheets and lithosphere thickening (e.g. Coward 1994). This collision was accompanied by local domains of extension or transpression associated with the intrusion of calc-alkaline I-type plutonic rocks possible. The Hämeenkyrö and Värmälä intrusions show evidence for passive emplacement (Nironen 1989a) and tensional fractures, pull-apart basins or en echelon "P-shear" tensional bridges (e.g. Tikoff and Teyssier 1992) are possible mechanisms. During crustal thickening the lower crust becomes hotter and enables spreading to take place and if the initial Moho temperatures exceed 700°C extension may follow immediately after release of compression (Sonder et al. 1987). This is one possibility for the thinning of the thickened lithosphere predicted in the model (Fig. 29) but an alternative origin due to thinning of the lithospheric mantle as an inevitable response to orogenic lithospheric thickening is favoured (England and Houseman 1989, Turner et al. 1992). This model implies the rise of asthenospheric-lithospheric thermal boundary higher, thus increasing the overall thermal budget of the orogen. Lithosphere thinning may produce increase in horizontal buoyancy forces and the convergent deformation will terminate or be partitioned in regions of lower potential energy and/or strength (Turner et al. 1992). This can

either keep the crust in dynamic equilibrium or induce extensional collapse of the orogen (ibid.). The thick crust north of the TSB (Luosto 1991) and the absence of voluminous magmatic activity younger than 1.86 Ga in this area could indicate that no pronounced crustal thinning occurred during 1.88-1.86 Ga. This area could on the other hand mark the northern edge of crustal thinning that started about 1.88-1.87 Ga ago and continued throughout the 1.85-1.79 Ga stage. The above considered model could explain the high T-low P metamorphism, which requires very high heat flow (Korsman et al. 1984, Hölttä 1986, van Duin 1992, Hölttä et al. 1993), the abundance of granitoids associated with mafic rocks in the MB and the almost coeval occurrence of compressional and neutral to tensional tectonics. The mafic underplating connected with this lithospheric thinning probably started at about 1.885-1.880 Ga ago.

The metamorphism and deformation in the southern area can be divided into older (1.89-1.88 Ga) and younger (1.87-1.81 Ga, probably 1.85-1.83 Ga) episodes (Korsman et al. 1988, Ehlers et al. 1993 and Hölttä et al. 1993). The age of the older metamorphic peak is probably about 1.88 Ga, and is bracketed by the deformed tonalitic gneisses at 1.89 Ga cut by discordant dykes at 1.87 Ga (Ehlers et al. 1993). Hence, the proposed model implies the initiation of collision at about 1.890-1.895 Ga, affecting first the rocks in the southern part, followed by progressive development of northwards propagating thrust sheets. The associated foredeep sediments should occur north of the TSB, but their absence can be attributed to later erosion. Possible candidates nevertheless include schist remnants in the CFGC and the well preserved metasediments in the northern part of the BSB and the SSB e.g. Koivujoki Suite in Pielavesi (Ekdahl 1993). One metagreywacke sample from the Bothnian basin in Sweden shows maximum deposition age about 1.88 Ga (Claesson et al. 1993) and is thus another potential candidate.

1.88-1.86 Ga stage

The northern part of the SSB was intruded by some gabbros and granitoids (Vaasjoki and Sakko 1988, Ekdahl 1993) that belong to this stage and record waning phase of prolonged magmatic underplating in the region surrounding the RLZ. The titanite age of 1.86 Ga from the Mustikkamäki pyroxene quartz diorite show that cooling to 400-450 °C took place at this time in Pielavesi (Hölttä 1988). According to Vaasjoki and Sakko (1988) the magmatic and tectonic activity ceased in these areas about 1.86-1.85 Ga ago.

Magmatic underplating was connected with lithospheric thinning beneath the southern collision zone, as discussed above. High-K calc-alkaline magmas associated with continental WPB type mafic magmas, possibly affected further by a subduction component, were the main granitoid types in the TSB and areas to the north of it (southern CFGC) during the change from pure collision regime to a neutral or tensional stage during lithosphere thinning (Lahtinen in prep.). The upwelling of the thermal boundary and the high amount of mafic magmatic underplate and mafic intrusions higher in the crust faciliated melting of pre-existing evolved rocks, possibly in the middle crust, generating high-K granitoids of tholeiitic affinity that have an interpreted age about 1.88 Ga. These and the very high-K granitoids of tholeiitic affinity north of the TSB (southern CFGC) coincide with a regional magnetic anomaly about 200 km long and 20-30 km wide that also has a continuation to the northwest (Korhonen 1992). These rocks normally contain magnetite but are not solely responsible for the regional anomaly and a deeper component is therefore also needed (Lahtinen and Korho-

nen in prep.). They are interpreted as the products of melting of WPB affinity underplate at the base of crust, in some cases associated with small amount of crustal contamination (Lahtinen in prep.). One example of such an assimilated variant is the Koppelojärvi granite that has clear post-tectonic nature with contact aureole (Sjöblom 1990). The U-Pb zircon age of 1879±14 Ma (Patchett and Kouvo 1986) has high analytical error and given the presence of a crustal component could indicate a small contribution from inherited zircons. Considering its unequivocal post-tectonic nature an age about 1875-1870 Ma has been proposed. Therefore, the regional magnetic anomaly associated with abundant very high-K granites of anorogenic/post-orogenic affinity is considered to represent a large scale linear feature at the margin of the zone of thinned lithosphere, with a large amount of subcontinental lithosphere (EM) derived mafic magmas at the base and probably also at higher levels of the crust. The 1.87-1.86 Ga granitoids containing a substantial Archean component in the Karelian domain (Huhma 1986) are also tentatively correlated with this stage. They possibly occur in area crossed by the eastern continuation of this belt and the RLZ magmatic underplating zone.

Granitoid magmatism and volcanism also took place further north in the CFGC during this stage as exemplified by the rhyolite in the Mustajärvi area dated at 1872±12 Ma (Vaasjoki and Lahti 1991). WP affinity magmatism also occurred in the southern part of the TSB and in the MB during this stage. These include high-Ti granitoids, although calc-alkaline I-type granitoids, possibly representing melting of an igneous source, are also found (Lahtinen in prep.). These granitoids intruded a thick sedimentary sequence and consequently often show evidence of assimilation, while the overall reducing environment prevented magnetite crystallization (Lahtinen and Korhonen in prep.). Alkaline WPB affinity dykes are associated with some younger granites, such as in the Kylmäkoski area (Lahtinen in prep.).

The geochemistry of rocks south of the HSB have not been extensively studied and the evolution of this area in this respect remains open. The occurrence of gabbros and diorites dated at 1.88-1.87 Ga with ε_{Nd} values of +2 - +3 in the Kalanti district and Hyvinkää-Soukkio area (Patchett and Kouvo 1986, Huhma 1986) are interpreted as recording the existence of young lithospheric mantle formed from DM under previously rather thin and immature crust during 1.91-1.89 Ga. The presence of about 1.88 Ga aged granites in the Åland islands (Suominen 1991) are included to this stage. Pyroxene granodiorites occur as large plutons in the Turku district and the Kakskerta pyroxene granodiorite contains a bimodal zircon population with at 1.88 Ga and 1.84 Ga possibly indicating primary crystallization at 1.88 Ga, followed by new zircon growth during later high-grade metamorphism (ibid.). Pyroxene granodiorite at Houtskär and a garnet tonalite dyke at Brändö have similar ages at 1862 Ma (ibid.). These features indicate magmatic activity between 1.88-1.86 Ga associated with the proposed lithospheric thinning. The titanite age of 1864±14 Ma from granodiorite with a 1.90 Ga zircon age at Föglö indicates (ibid.) cooling down through 450-500 °C (Mattinson 1982) at 1.86 Ga. This was possibly related to uplift and erosion and indicates that not all southern areas were heated over 500 °C during younger 1.85-1.82 Ga event at the present crustal level.

1.86-1.79 Ga stage

This stage has been subdivided into 1.86-1.83 and 1.83-1.79 Ga stages in Figure 29. The younger collisional event was first documented by Korsman et al. (1988) from the Rantasalmi-Sulkava area where rocks affected by younger metamorphic event were partly thrust over old-

er metamorphic complexes. Ehlers et al. (1993) concluded that the late Svecofennian granite-migmatite (LSGM) zone is a distinct tectonic unit of crustal thickening showing strong deformation coupled with high-grade metamorphism and intrusion of 1.84-1.83 Ga old granite sheets during transpressional deformation. They further implied that the zone is located in a former extensional basin acting as inherited zone of weakness. According to Hölttä et al. (1993) older structures are deformed by the main regional folding F3 that is younger than 1.87 Ga and indicates, at least partly, a new or prolonged episode of thrusting. The vergence of F3 is towards NW and the fold limbs are often strongly sheared dividing the region to folded and sheared domains. The deformation continued in D3 shear zones possibly associated with 1.83 Ga magmatism in extensional shear zones (ibid., see also Ehlers et al. 1993). Metamorphic grade reached locally granulite facies with later decompression reactions in pelitic rocks that are possibly associated with the intrusion of granites (Hölttä et al. 1993). The coeval mafic magmatism (diorite dyke) with potassium granites in Turku indicate mantle activity (ibid.). A quartz monzonite of alkaline WPB affinity, possibly associated with microcline granite at Renko (Lahtinen in prep.), is dated at 1813±2 Ma (pers. comm., M. Vaasjoki 1994) and thus has a younger age than the normal 1.83-1.84 Ga microcline granites.

The model presented in Figure 29 tries to integrate the evidence presented above. The 1.88-1.86 Ga lithospheric thinning stage was associated with uplift and erosion, as for example indicated by cooling ages such as the 1.86 Ga titanite age referred to above. About 1.86-1.85 Ga ago a further stage of intra-continental thrusting commenced, either due to a new collision further away or continued collision after termination of lithospheric thinning. New crustal scale thrust sheets were formed that coincided in part with older structures. The continuous high heat flow from the mantle-crust boundary associated with high Moho temperatures, in combination with crustal thickening and highly radiogenic heat production within granitoids promoted melting of the underthrust migmatitic mica gneisses to create potassium granites. The mafic magmatism at 1.82 - 1.81 Ga (see above) possibly relates to further heat addition by mantle underplating. The granites were possibly emplaced in sub-horizontal contemporaneous mid-crustal shear zones, thrusts (Ehlers et al. 1993) and/or in extensional, reversed and strike-slip faults (Hölttä et al. 1993). Hence, the generation of these granites and their emplacement are directly linked with thrusting and either coeval or later extensional shear. One potential explanation for this is the sucking of magmas due to these deep-seated shear zones (e.g. Pitcher 1993). These granites have been classified as S-type granites (Nurmi and Haapala 1986). Lahtinen (in prep.) studied the microcline granites in the Hämeenlinna area and proposed a S-type minimum melt origin associated with variable amount of restitic component to explain the geochemical variation. Differences in the source components were also noted. The occurrence of restitic bands as noticed in garnet-rich 'layers' and proposed source heterogeneity indicates that these rocks are not totally homogenizated at the accretion level. The rather high viscosities of these felsic magmas probably limit exchange between different pulses and homogenization would not be entirely achieved, even at the meter scale (Deniel et al. 1987).

The migmatites south of the HSB record granulite facies metamorphic conditions and show gains in Cl, F and Rb and loss of C, S, B, Cu, As, Te and Bi, indicating loss of fluid phase (Lahtinen in prep.). The S-type granites in the same area show similar depletion patterns and differ from the S-type granites in the MB metamorphosed in upper amphibolite conditions. This indicates origin by melting of previously metamorphosed high-grade sedimentary rocks deeper in the crust because the migmatites at the present erosion level still contain both melt and restite components and the proportion of

granites implies an unrealistically too high degree of melting at these compositions. The proposed two-stage evolution allows generation of a large volume of these S-type granites. The earlier metamorphism has generated large amounts of mainly subsolidus leucosomes (Hölttä et al. 1993) but the critical melt fraction has been exceeded only locally. Further burial of these migmatites associated with increase in temperature and an external fluid phase have melted the earlier leucosomes, accompanied by further melting of restite. There are also indications of the role of a more igneous-like source. Due to an increase in the total melt fraction and pumping by tectonic processes it is possible to separate high viscosity melts and emplace them at higher levels. The proposed external fluid phase could be derived from dehydration of more weakly metamorphosed underthrust sediments (Le Fort 1981) and/or fluids expelled from the alkaline WPB-type melts. The migmatites considered above show high F values which could indicate a genetic link with fluids derived from mantle derived alkaline WPB melts (Lahtinen in prep.). This is a tentative model and should be tested. More data are also needed from other areas in the LSGM to establish whether these characteristics are widespread.

The possible existence of 1.83 Ga alkaline WPB magmatism, possibly related to the later extensional phase, should also be established. Post-kinematic magmatism (1.80 Ga) occurring parallel to the northern contact of the LSGM in the SW-NE direction represents an alkali-basaltic magmatism in an extensional environment (p. 90) and so the occurrence of continuous mafic underplating from 1.83 Ga to 1.79 Ga is possible. The zircon age (of 1813 Ma) from quartz monzonite at Renko (see above) indicate that this type of magmatism took also place somewhat earlier than previously documented. The 1.80 Ga titanite age for a granitoid at Kökär with 1.88 Ga zircons (Suominen 1991) and 1.81 Ga titanite in a 1.89 Ga granitoid and similar zircon and titanite ages (1812 Ma and 1815 Ma respectively) from a gabbro pegmatoid in Kalanti area (Patchett and Kouvo 1986) show that cooling and possibly also erosion started about 1.81 Ga ago. The granulites in the Turku area had attained a depth corresponding to 2 kb by 1.72 Ga (Hölttä 1986). The 1.83-1.84 Ga granites are deformed by shear zones (D4), that were possibly extensional, and which exhibit mylonitic textures (Hölttä et al. 1993). Korja and Heikkinen (1993, see also Korja et al. 1993) proposed, based on deep seismic reflection images, the existence of SE dipping extensional listric shear zones correlated with post-collisional collapse of the Svecokarelian orogeny. The listric normal faults related to the Moho detachment zone are younger and cross cut the upper detachment zone of older listric faults (ibid.). They also advocated intrusion of the rapakivi granites (1.6-1.5 Ga) along the younger listric faults during subsequent extensional environment. The change in thickness of high velocity layer from values greater than 16 km to values less than 12 km (Korja et al. 1993) and especially the 12 km contour has a marked SW-NE trend which is the same direction as seen in the LSGM. The abrupt transition between granulite and amphibolite facies metamorphic blocks in the LSGM (Korsman et al. 1984, 1988, Hölttä 1986) was also attributed by Korja and Heikkinen (1993) to extensional deformation. On the other hand the occurrence of two different episodes of metamorphism at 1.885 Ga and 1.83 Ga ago would doubtless produce a complex pattern of both thrusted and strike-slip boundaries that possibly also affected rocks influenced by the younger metamorphic event. Hölttä et al. (1993) describe a shear zone filled with potassium granite in the Turku area that possibly represents a metamorphic block boundary. Later extensional faults, possibly of more than one generation, can either coincide or cross cut older structures resulting in a complex geometry.

Granulite facies rocks are also widespread in Estonia, especially in the southern part (Puura et al. 1983) and they are at present considered

to be cogenetic with Svecofennides (Puura and Huhma 1993). The granulite facies rocks in Estonia were metamorphosed at slightly higher pressures than those in southern Finland (Höltä and Klein 1991). Further south, in the early Proterozoic West Lithuanian granulite terrain, pressures up to 8 kb have been determined (Skridlaite 1993). The occurrence of weakly gneissose potassic granodiorite in southern Estonia dated at 1833±7 Ma by the U-Pb zircon method and the youngest reset zircon ages of about 1.85 Ga obtained from aluminiferous gneisses (Petersell in prep.) indicate the existence of this younger metamorphic event in Estonia as well. The presence of two fine-grained gneisses (metavolcanics) with ages of 1827±7 Ma and 1828±8 Ma (Petersell in prep.) are in this context problematic. They were metamorphosed in the amphibolite facies (ibid.) so that the resetting of zircon U-Pb systematics seems unlikely but can not be totally excluded. If the ages are true crystallization ages, two possible interpretations can be made. Either they are true volcanics or they are annealed mylonites derived from granitoids deep in the crust. The 1816±3 Ma U-Pb zircon age from undeformed granitoid in northern Latvia may indicate the lack of orogenic events in this area after 1.8 Ga (Mansfeld et al. 1993).

The above considered features could be interpreted as recording the onset of extensional faulting about 1.81-1.80 Ga ago or possibly slightly earlier (1.83-1.82 Ga). The large-scale extensional faults in many cases dip in the same direction as the thrusts and may link with them at depth (e.g. Coward 1994) and so the extensional stage faults possibly partly follow the older structures. Although this event is correlated with 1.82 - 1.80 Ga alkali WPB magmatism, extension continued apparently until 1.5 Ga above a zone of major crustal thinning located along the Gulf of Finland (Korja et al. 1993). However, it is uncertain whether this represents a single prolonged episode or several discrete events coinciding with and reworking older structures.

Some closing remarks

There are many crucial unresolved problems, such as whether the Jormua ophiolite is a parautochthonous rather than allochtonous unit, the nature and evolution of the CFGC basement and the pre-1.89 Ga evolution of the southern part of Svecofennian domains. More age data, including single zircon age determinations, are also needed to precisely date the proposed evolutionary stages. Although this model is simplistic and may even be inaccurate in some details, it serves to draw attention to the complex nature of Svecofennian evolution. The model advocates three different collisional stages (orogenies) during Svecofennian evolution at 1.91-1.90 Ga, 1.89-1.88 Ga and 1.86-1.84 Ga where the 1.86-1.84 Ga stage may be a direct continuation of the 1.89-1.88 Ga stage. The granitoids show successive geochemical evolutionary trends during the two earlier stages. A small volume of granitoids were associated with arc magmatism at 1.93-1.91 Ga and 1.90-1.89 Ga respectively followed by more abundant synkinematic granitoids during initial collision at 1.90 Ga and 1.890-1.885 Ga respectively and the latest, voluminous phase is during post-collision stage at 1.885-1.880 and 1.875-1.860 Ga respectively. The important feature is that the post-collisional stages associated with mafic underplating have been very important in crustal reworking and in producing a vast amount of granitoids and mafic plutons from the mafic underplate accreted to the lower crust. The crustal scale thrusting has thickened the crust (see also Korja et al. 1993) and the later mafic underplating further increased crustal thickness and stabilized the crust, allowing the present thickness to be maintained. The 1.83-1.79 Ga and later exten-

sional processes have thinned the crust south of the TSB to achieve the present crustal profiles (see Korja et al. 1993).

The evolution of the Fennoscandian Shield as a whole is beyond scope of this study but two comments are made here. In the Norvijaur area north of Skellefte is a 1926 Ma aged polyphase deformed granitoid beneath an unconformity (Skiöld et al. 1993). The analytical data record a large error, possibly due to inherited zircon cores and geochemically the granitoid resemble the 1.89 Ga Jörn granitoids in Skellefte area (ibid.). A very interesting feature is the occurrence of a 1954±6 Ma granitoid in the Knaften area south of Skellefte (Wasström 1993). These granitoids show some subvolcanic features and contain abundant K-feldspar, being locally granitic. They differ from the 1.93-1.91 gneissic tonalites of this study but could possibly be correlated with the evolved crust predicted in the model for 1.98-1.93 Ga stage (Fig. 29).

CONCLUSIONS

- The 1.93-1.91 Ga gneissic tonalites and associated supracrustal rocks in the Rautalampi area formed within an immature 1.93-1.91 Ga island arc. The tonalites represent low degree melting of low-K tholeiitic island arc basalts with gabbroic to amphibolitic residue remaining at the base of the primitive island arc crust. Tonalites are cogenetic and probably comagmatic with felsic volcanics and they were preceded by mafic and intermediate island arc volcanism. Aborted rifting at about 1.92 Ga produced bimodal volcanism that is associated with VMSD type Zn-Cu mineralization and altered rocks. The garnet±cordierite±orthoamphibole/orthopyroxene rocks and gneisses at Rautalampi can be divided into two groups. That which occurs below the ore zone is characterized by a lack of sulphides and clearly had both felsic and mafic protoliths, and is considered to represent altered rocks of lower semi-conformable alteration zone. The other group is characterized by the occurrence of sulphides and had mixed mafic-felsic to sedimentary protoliths and represent the lateral continuation of the ore zone.

- Similar features characterize the central and northern part of the SSB and an analogous origin for gneissic tonalites, volcanics, altered rocks and Zn-Cu mineralization in the Kangasjärvi, Pielavesi, Pyhäsalmi and Venetpalo areas is proposed.

- The 1.89-1.88 Ga granitoids in the SSB and adjacent areas in the CFGC are attributed to with magmatic underplating of EM type melts. The granitoids show diverse sources. Tonalites in the SSB represent melting of rather primitive island arc crust that was locally thick enough to stabilize garnet in the residue. Associated granitoids in the CFGC record a more evolved source. Pyroxene granitoids are genetically related to WPB magmas either due to differentiation or melting associated with crustal contamination.

- A plate tectonic model is proposed for the 2.1-1.79 Ga evolution of Svecofennian and Karelian domains. Three different collisional stages (orogenies) at 1.91-1.90 Ga, 1.89-1.88 Ga and 1.86-1.84 Ga are discriminated and involved thin-skinned and crustal scale thrusting episodes. Lithospheric thickening during these events was followed by upwelling of the thermal boundary leading to magmatic underplating and high heat flow. The granitoids can be divided to early arc related, synkinematic collision-related and post-collision granitoids that have different ages in different areas. Post-collisional magmatism has been a major new crust forming process. The crustal thickening is due to both crustal scale thrusting and magmatic underplating that also eventually stabilized the crust at its present thickness.

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In this paper I have attempted to summarize the 2.1-1.79 Ga evolution of Svecofennian and Karelian domains within a comprehensive plate tectonic model. Of course, much of this research has been collaborative in nature and many of the ideas that I present have evolved through complex interaction with others; I especially acknowledge discussions with Hannu Huhma, Jarmo Kohonen, Kalevi Korsman, Yrjö Kähkönen and Peter Sorjonen-Ward. Discussions and the free exchange of ideas during numerous field trips in the SSB and within Kalevian rocks with Elias Ekdahl, Asko Kontinen, Jukka Kousa, Heikki Lukkarinen, Timo Mäki, Jorma Paavola, Matti Pajunen, Pentti Pietikäinen, Heikki Puustjärvi and Matti Vaasjoki are also greatly acknowledged.

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REFERENCES

Aho, L., 1979. Petrogenetic and geochronological studies of metavolcanic rocks and associated granitoids in the Pihtipudas area, Central Finland. Geol. Surv. Finland, Bull. 300, 22 p.

Äikäs, K., 1977. Suonenjoen-Rautalammin alueen geologia kallioperäkartoituksen ja geofysikaalisten matalalentokarttojen avulla tutkittuna. Unpubl. M. Sc. thesis. Helsinki: Univ. Helsinki, 54 p.

Äikäs, O., 1989. Phosphate resources in Early Proterozoic supracrustal rocks, Finland, with reference to the Baltic Shield. In: Notholt, A.J.G., Sheldon, R.P. & Davidson, D.F. (eds.) Phosphate Deposits of the World: 2, Phosphate rock resources. Cambridge: Cambridge University Press, 429-436.

Alapieti, T., 1982. The Koillismaa layered igneous complex, Finland - its structure, mineralogy and geochemistry, with emphasis on the distribution of chromium. Geol. Surv. Finland, Bull. 319, 116 p.

Aumo, R., 1983. The Proterozoic metasedimentary and metavolcanic mantle of the Kuopio domes. In: Laajoki, K. & Paakkola, J. (eds.) Exogenic processes and related metallogeny in the Svecokarelian geosynclinal complex. Geol. Surv. Finland, Guide 11, 86-99.

BABEL Working Group, 1990. Evidence for Early Proterozoic plate tectonics from seismic reflection profiles in the Baltic Shield. Nature 348, 34-38.

Bachinski, D.J., 1976. Metamorphism of cupriferous iron sulphide deposits, Notre Dame Bay, Newfoundland. Econ. Geol. 71, 443-452.

Barbey, P., Couvert, J., Moreau, B. & Capdevila, R., 1984. Petrogenesis and evolution of an Early Proterozoic collisional orogenic belt: The granulite belt of Lapland and the Belomorides (Fennoscandia). Bull. Geol. Soc. Finland 56, 161-168.

Barker, F., 1979. Trondhjemite: definition, environment and hypotheses of origin. In: Barker, F. (ed.) Trondhjemites, Dacites and Related Rocks. Amsterdam: Elsevier, 1-12.

Bayanova, T.B., 1993. Geochronology of the platinum-bearing basic-ultrabasic layered intrusions of the Kola region, Baltic Shield. Symposium on the Svecofennian Domain and Annual Meeting of IGCP-275, Abstr. Dept. Geol. Miner. Åbo: Åbo Akademi, 10-11.

Beeson, R., 1988. Identification of cordierite-anthophyllite rock types associated with sulphide deposits of copper, lead and zinc. Trans. Inst. Min. Metall. (sect. B: Appl. Earth Sci.) 97, 108-115.

Berthelsen, A. & Marker, M., 1986. Tectonics of the Kola collision suture and adjacent Archean and Early Proterozoic terrains in the northeastern region of the Baltic Shield. Tectonophysics 126, 31-55.

Bloomfield, A.L. & Arculus, R.J., 1989. Magma mixing in the San Francisco Volcanic Field, AZ. Contrib. Mineral. Petrol. 102, 429-453.

Boynton, W.V., 1984. Cosmochemistry of the Rare Earth Elements: meteorite studies. In: Henderson, P. (ed.) Rare Earth Element Geochemistry (Developments in Geochemistry 2). Amsterdam: Elsevier, 63-114.

Bradley, D.C. & Kidd, W.S.F., 1991. Flexural extension of the upper continental crust in collisional foredeeps. Geol. Soc. Am. Bull. 103, 1416-1438.

Brown, G.C., 1982. Calc-alkaline intrusive rocks: their diversity, evolution, and relation to volcanic arcs. In: Thorpe, R.S. (ed.) Andesites, Orogenic Andesites and Related Rocks. Chichester: Wiley, 437-461.

Brown, G.C., Thorpe, R.S. and Webb, P.C., 1984. The geochemical characteristics of granitoids in contrasting arcs and comments on magma sources. J. Geol. Soc. Lond. 141, 411-426.

Carlson, R.W., 1991. Physical and chemical evidence on the cause and source characteristics of flood basalt volcanism. Austr. J. Earth Sci. 38, 525-544.

Carrol, M.R. & Wyllie, P.J., 1989. Experimental Phase Relations in the System Tonalite-Peridotite-H_2O at 15 kb; Implications for Assimilation and Differentiation Processes near the Crust-Mantle Boundary. J. Petrol. 30, 1351-1382.

Castro, A., Moreno-Ventas, I. & de La Rosa, J.D., 1991. H-type (hybrid) granitoids: a proposed revision of the granite-type classification and nomenclature. Earth Sci. Rev. 31, 237-253.

Cathles, L.M., Gruber, A.L. Lenagh, T.C. & Dudas, F.Ö., 1983. Kuroko-type massive sulfide deposits of Japan: Products of an aborted island arc rift. Econ. Geol. Mon. 5, 96-111.

Cawood, P.A. & Suhr, G., 1992. Generation and obduction of ophiolites: Constraints from the Bay of Islands Complex, western Newfoundland. Tectonics 11 (4), 884-897.

Chappel, B.W. & White, A.J.R., 1974. Two contrasting granite types. Pac. Geol. 8, 173-174.

Charvet, J. & Ogawa, Y., 1994. Arc-trench tectonics. In: Hancock, P.L. (ed.) Continental Deformation. Oxford: Pergamon Press, 180-199.

Claesson, S., Huhma, H., Kinny, P.D. & Williams, I.S., 1993. Svecofennian detrital zircon ages - implications for the Precambrian evolution of the Baltic Shield. Precambrian Res. 64, 109-130.

Coward, M., 1994. Continental collision. In: Han-

cock, P.L. (ed.) Continental Deformation. Oxford: Pergamon Press, 264-288.

Deniel, C., Vidal, P., Fernandez, A., Le Fort, P. & Peucat, J-J., 1987. Isotopic study of the Manaslu granite (Himalaya, Nepal): inferences on the age and source of Himalayan leucogranites. Contrib. Mineral. Petrol. 96, 78-92.

De Paolo, D.J., 1981. A neodymium and strontium isotopic study of the Mesozoic calc-alkaline granitic batholiths of the Sierra Nevada and Peninsular Ranges, California. J. Geophys. Res. 86, 10470-10488.

Dewey, J.F., 1980. Episodicity, sequence and style at convergent plate boundaries. Geol. Ass. Canada, Spec. Pap. 20, 553-573.

Dewey, J.F., Shackleton, R.M., Chang C.F. & Sun, Y., 1988. The tectonic evolution of the Tibetian plateau. Phil. Trans. R. Soc. Lond., Ser. A327, 379-413.

Drummond, M.S., Ragland, P.C. & Wesolowski, D., 1986. An Example of trondhjemite genesis by means of alkali metasomatism: Rockford granite, Alabama Appalachians. Contrib. Mineral. Petrol. 93, 98-113.

van Duin, J.A., 1992. The Turku granulite area, SW Finland: a fluid-absent Svecofennian granulite occurrence. Academisch Proefschrift. Amsterdam: Vrije Universiteit, 234 p.

Eberz, G.W. & Nicholls, I.A., 1990. Chemical modification of enclave magma by post-emplacement crystal fractionation, diffusion and metasomatism. Contrib. Mineral. Petrol. 104, 47-55.

Ehlers, C. & Lindroos, A., 1990. Early Proterozoic Svecofennian volcanism and associated plutonism in Enklinge, SW Finland. Precambrian Res. 47, 307-318.

Ehlers, C., Lindroos, A. & Jaanus-Järkkälä, M., 1986. Stratigraphy and geochemistry in the Proterozoic mafic volcanic rocks of the Nagu-Korpo area, SW Finland. Precambrian Res. 32, 297-315.

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

Ekdahl, E., 1993. Early Proterozoic Karelian and Svecofennian formations and the Evolution of the Raahe-Ladoga Ore Zone, based on the Pielavesi area, central Finland. Geol. Surv. Finland, Bull. 373, 137 p.

Elthon, D., 1991. Geochemical evidence for formation of the Bay of Islands ophiolite above a subduction zone. Nature 354, 140-143.

England, P.C. & Houseman, G.A., 1989. Extension during continental convergence with application to the Tibetian Plateau. J. Geophys. Res. 94, 17561-17579.

Eskola, P., 1914. On the petrology of the Orijärvi region in southwestern Finland. Bull. Comm. géol. Finlande 40, 277 p.

Eskola, P., 1932. On the origin of granitic magmas. Mitt. Mineral. Petrogr. 42, 455-481.

Eskola, P., 1960. Granitentstehung bei Orogenese und Epiorogenese. Geol. Rundsch. 50, 105-123.

Eskola, P., 1963. The Precambrian of Finland. In: Rankama, K. (ed.) The Precambrian 1. New York: Wiley, 145-263.

Fleming, P.D. & Fawcewtt, J.J., 1976. Upper stability of chlorite-quartz in the system MgO-FeO-Al_2O_3-SiO_2-H_2O at 2 kbar water pressure. Am. Mineral. 61, 1175-1193.

Franklin, J.M., Lydon, J.W. & Sangster, D.F., 1981. Volcanic-associated massive sulfide deposits. Econ. Geol., 75th Anniv. Vol., 485-627.

Friesen, R.G., Pierce, G.A. & Weeks, R.M., 1982. Geology of the Geco base metal deposit. In: Hutchinson, R.W., Spence, C.D. & Franklin, J.M. (eds.) Precambrian sulphide deposits. Geol. Ass. Canada, Spec. Pap. 25, 343-363.

Front, K., 1981. Hämeenkyrön batoliitin litogeokemia, kivilajit ja suhde ympäröiviin liuskeisiin. Unpubl. M. Sci. thesis. Helsinki: Univ. Helsinki, 59 p.

Front, K. & Nurmi, P.A., 1987. Characteristics and geological setting of synkinematic Svecokarelian granitoids in southern Finland. Precambrian Res. 35, 207-224.

Furman, T. & Spera, F.J., 1985. Co-mingling of acid and basic magma with implications for the origin of mafic I-type xenoliths: Field and petrochemical relations of an unusual dike complex at Eagle Lake, Sequoia National Park, California, USA. J. Volcanol. Geotherm. Res. 24, 151-178.

Gaál, G., 1972. Tectonic control of some Ni-Cu deposits in Finland. 24th Int. Geol. Congr., Montreal 4, 215-224.

Gaál, G., 1980. Geological setting and intrusion tectonics of the Kotalahti nickel-copper deposit, Finland. Bull. Geol. Soc. Finland 52, 101-128.

Gaál, G., 1986. 2200 million years of crustal evolution. The Baltic Shield. Bull. Geol. Soc. Finland 58, 149-168.

Gaál, G., 1990. Tectonic styles of Early Proterozoic Ore Deposition in the Fennoscandian shield. Precambrian Res. 46, 83-114.

Gaál, G. & Rauhamäki, E., 1971. Petrological and structural analysis of the Haukivesi area between Varkaus and Savonlinna, Finland. Bull. Geol. Soc. Finland 43, 265-337.

Gaál, G. & Gorbatschev, R., 1987. An Outline of the Precambrian evolution of the Baltic Shield. Precambrian Res. 35, 15-52.

Gaál, G., Koistinen, T. & Mattila, T., 1975. Tectonics and stratigraphy in the vicinity of Outokumpu, North Karelia, Finland. Geol. Surv. Finland, Bull. 271, 67 p.

Gill, J., 1981. Orogenic Andesites and Plate tectonics. Berlin: Springer-Verlag, 390 p.

Grant, J.A., 1968. Partial melting of common rocks as a possible source of cordierite-anthophyllite assemblages. Am. J. Sci. 266, 908-931.

Green, T.H. & Ringwood, A.E., 1968. Genesis of the calc-alkaline igneous rock suite. Contrib. Mineral. Petrol. 18, 105-162.

Gruber, A.L. & Merrill, S., 1983. Paleobathymetric significance of the foraminera from the Hokuroku district. Econ. Geol. Mon. 5, 55-70.

Grundström, L., 1980. The Laukunkangas nickel-copper occurrence in southeastern Finland. Bull. Geol. Soc. Finland 52, 23-53.

Goetz, P.A. & Froese, E., 1982. The Sherrit Gordon massive sulphide deposit. In: Hutchinson, R.W., Spence, C.D. & Franklin, J.M. (eds.) Precambrian sulphide deposits. Geol. Ass. Canada, Spec. Pap. 25, 557-569.

Hakkarainen, G., 1990. Litostratigraphic and geochemical relations in the Hämeenlinna-Somero volcanic belt. In: Kähkönen, Y. (ed.) IGCP Project 217 Proterozoic Geochemistry. National Working Group of Finland. Symposium, Proterozoic Geochemistry. Helsinki '90. Abstr., 12-15.

Hakkarainen, G. & Väisänen, M., 1993. Stratigraphy and geochemistry of Orijärvi area, SW Finland. Symposium on the Svecofennian Domain and Annual Meeting of IGCP-275, Abstr. Dept. Geol. Miner. Åbo: Åbo Akademi, p. 27.

Hamilton, W.B., 1979. Tectonics of the Indonesian region. U.S. Geol. Surv., Prof. Pap. 1078, 345 p.

Hanski, E.J., 1983. Alkuaineiden jakautuminen mineraalien ja silikaattisulan kesken: Jakautumiskertoimet. Univ. Oulu, Arkeisten alueiden malmiprojekti Rep. 15, 179 p.

Hautala, P.O., 1968. Venetpalon alueen petrologia ja rakenne. Unpubl. M.Sc. thesis. Oulu: Univ. Oulu, 161 p.

Hawkesworth, C.J., Hergt, J.M., McDermott, F & Ellam, R.M., 1991. Destructive margin magmatism and the contributions from the mantle wedge and subducted crust. Austr. J. Earth Sci. 38, 577-594.

Hawkesworth, C.J. & Ellam, R.M., 1989. Chemical fluxes and wedge replenishment rates along Recent destructive plate margins. Geology 17, 46-49.

Heinonen, R., 1987. Ruokolahden Käköveden batoliitin petrologia ja geokemia. Unpubl. M.Sci. thesis. Helsinki: Univ. Helsinki. 51 p.

Helovuori, O., 1979. Geology of the Pyhäsalmi ore deposit, Finland. Econ. Geol. 74, 1084-1101.

Helz, R.T., 1973. Phase relations of basalts in their melting at P_{H2O} = 5 kb as a function of oxygen fugacity, Part. I. Mafic phases. J. Petrol. 14, 249-302.

Helz, R.T., 1976. Phase relations of basalts in their melting at P_{H2O}= 5 kb. Part. II. Melt compositions. J. Petrol. 17, 139-193.

Hietanen, A., 1975. Generation of potassium-poor magmas in the northern Sierra Nevada and the Svecofennian in Finland. J. Res. U.S. Geol. Surv. 3, 631-645.

Hildreth, W., 1981. Gradients in Silicic Magma Chambers: Implications for Lithospheric Magmatism. J. Geophys. Res. 86, 10153-10192.

Hoffer, E. & Grant, J.A., 1980. Experimental investigation of the formation of cordierite-orthopyroxene paragenesis in pelitic rocks. Contrib. Mineral. Petrol. 73, 15-22.

Holden, P., Halliday, A.N., Stephens, W.E. & Henney, P.J., 1991. Chemical and isotopic evidence for major mass transfer between mafic enclaves and felsic magma. Chem. Geol. 92, 135-152.

Holloway, J.R. & Burnham, C.W., 1972. Melting relations of basalt with equilibrium water pressure less than total pressure. J. Petrol. 13, 1-29.

Hölttä, P., 1986. Observations on the metamorphic reactions and PT conditions in the Turku granulite area. In: Korsman, K. (ed.) Development of deformation, metamorphism and metamorphic blocks in eastern and southern Finland. Geol. Surv. Finland, Bull. 339, 43-58.

Hölttä, P., 1988. Metamorphic zones and the evolution of granulite grade metamorphism in the early Proterozoic Pielavesi area, central Finland. Geol. Surv. Finland, Bull. 344, 50 p.

Hölttä, P. & Klein, V., 1991. PT-development of granulite facies rocks in southern Estonia. In: Autio, S. (ed.) Current Research 1989-1990. Geol. Surv. Finland, Spec. Pap. 12, 37-47.

Hölttä, P., Rastas, J. & Väisänen, M., 1993. The tectono-metamorphic evolution of the crust in the Turku district, S-W Finland. GGT/SVEKA Working Meeting in Kuopio 25-26.11.1993, Abstr., p. 5.

Honkamo, M., 1988. Haukiputaan ja Kiimingin kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheets 2533 and 3511, 46 p.

Honkamo, M., 1989. Pohjois-Pohjanmaan vulkaniitit. Summary: Volcanic rocks in northern Ostrobothnia. Geol. Surv. Finland, Rep. Invest. 89, 46 p.

Huhma, H., 1986. Sm-Nd, U-Pb and Pb-Pb isotopic evidence for the origin of the Early Proterozoic Svecokarelian crust in Finland. Geol. Surv. Finland, Bull. 337, 48 p.

Huhma, H., 1987. Provenance of Early Proterozoic

and Archaean metasediments in Finland: a Sm-Nd isotopic study. Precambrian Res. 35, 127-143.

Huhma, H., Claesson, S., Kinney, P.D. & Williams, I.S., 1991. The growth of Early Proterozoic crust: new evidence from Svecofennian detrital zircons. Terra Nova 3, 175-179.

Huhtala, T., 1979. The Geology and Zinc-Copper Deposits of the Pyhäsalmi-Pielavesi District, Finland. Econ. Geol. 74, 1069-1083.

von Huene, R. & Lallemand, S., 1990. Tectonic erosion along the Japan and Peru convergent margins. Bull. Geol. Soc. Am. 102, 702-720.

Humpris, S.E., 1984. The mobility of the rare earth elements in the crust. In: Henderson, P. (ed.) Rare Earth Element Geochemistry (Developments in Geochemistry 2). Amsterdam: Elsevier, 317-342.

Johnston, A.D. & Wyllie, P.J., 1988. Interaction of granite and basic magmas: experimental observations on contamination processes at 10 kbar with H_2O. Contrib. Mineral. Petrol. 98, 352-362.

Jokela, J., 1991. Valkeakosken alueen kallioperän integroitu rakennetulkinta. Unpubl. M. Sci. thesis. Turku: Univ. Turku, 88 p.

Jolivet, L., Huchon, P. & Rangin, C., 1989. Tectonic setting of western Pacific marginal basins. Tectonophysics 160, 23-48.

Kähkönen, Y., 1989. Geochemistry and petrology of the metavolcanic rocks of the early Proterozoic Tampere Schist Belt, southern Finland. Geol. Surv. Finland, Bull. 345, 104 p.

Kähkönen, Y. & Huhma, H., 1993. An Archaean cobble in a Svecofennian conglomerate near Tampere, southern Finland. In: Autio, S. (ed.) Current Research 1991-1992. Geol. Surv. Finland, Spec. Pap. 18, 31-36.

Kähkönen, Y., Huhma, H. & Aro, K., 1989. U-Pb Zircon ages and Rb-Sr Whole-rock Isotope Studies of Early Proterozoic Volcanic and Plutonic Rocks near Tampere, Southern Finland, Precambrian Res. 45, 27-43.

Kahma, A., 1973. The main metallogenic features of Finland. Geol. Surv. Finland, Bull. 265, 29 p.

Kallio, J., 1986. Joutsan kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 3122, 56 p.

Kamineni, D.C., 1979. Metasedimentary cordierite-gedrite rocks of Archaean age near Yellowknife, Canada. Precambrian Res. 9, 289-301.

Kano, H., 1961. Petrology of the metasomatic cordierite rocks from the Northern Tanzawa Mountainland, Central Japan - Studies in the Mg-Fe metasomatism in Japan, Part 1. J. Miner. Coll., Akita Univ., Ser. A 1, 1-26.

Karhu, J., 1993. Paleoproterozoic evolution of the carbon isotope ratios of sedimentary carbonates in the Fennoscandian Shield. Geol. Surv. Finland., Bull. 371, 87 p.

Kärki, A., Laajoki, K. & Luukas, J., 1993. Major Paleoproterozoic shear zones of the central Fennoscandian Shield. Precambrian Res. 64, 207-223.

Keleman, P.B., 1990. Reaction Between Ultramafic Rock and Fractionating Basaltic Magma I. Phase Relations, the Origin of Calc-alkaline Magma Series, and the Formation of Discordant Dunite. J. Petrology 31, 51-97.

Kilpeläinen, T., 1988. Evolution of deformation and metamorphism as a function of time in the Rantasalmi-Sulkava area, southeastern Finland. In: Korsman, K. (ed.) Tectono-metamorphic evolution of the Raahe-Ladoga zone. Geol. Surv. Finland, Bull 343, 77-87.

Knuckey, M.J., Comba, C.D.A. & Riverin, G., 1982. Structure, metal zoning and alteration at the Millenbach deposit, Noranda, Canada. In: Hutchinson, R.W., Spence, C.D. & Franklin, J.M. (eds.) Precambrian Sulphide Deposits. Geol. Ass. Canada, Spec. Pap. 25, 255-295.

Kohonen, J., Luukkonen, E. & Sorjonen-Ward, P., 1991. Nunnanlahti and Holinmäki shear zones in North Karelia: Evidence for major Early Proterozoic ductile deformation of Archean basement and further discussion of regional kinematic evolution. In: Autio, S. (ed.) Current Research 1989-1990. Geol. Surv. Finland, Spec. Pap. 12, 11-16.

Koistinen, T.J., 1981. Structural evolution of an Early Proterozoic strata-bound Cu-Co-Zn deposit, Outokumpu, Finland. Trans. R. Soc. Earth Sci. 72, 115-158.

Kontinen, A., 1987. An Early Proterozoic ophiolite - the Jormua mafic-ultramafic complex, northeastern Finland. Precambrian Res. 35, 313-341.

Kontinen, A. & Sorjonen-Ward, P., 1991. Geochemistry of metagraywackes and metapelites from the Paleoproterozoic Nuasjärvi Group, Kainuu Schist Belt and the Savo Province, north Karelia: Implications for provenance, lithostratigraphic correlation and depositional setting. In: Autio, S. (ed.) Current Research 1989-1990. Geol. Surv. Finland, Spec. Pap. 12, 21-22.

Korhonen, J., 1992. Kallioperän magneettiset ominaisuudet. Suomen Kartastovihko 123-126 (English summary). National Board of Survey and Geographical Society of Finland, 26-29.

Korja, A., Korja, T., Luosto, U. & Heikkinen, P., 1993. Seismic and geoelectric evidence for collisional and extensional events in the Fennoscandian Shield - implications for Precambrian crustal evolution. Tectonophysics 219, 129-152.

Korja, A. & Heikkinen, P.J., 1993. Extensional structures of southwestern Finland based on the Babel reflection results. GGT/SVEKA Working

Meeting in Kuopio 25-26.11.1993, Abstr., 4.

Korsman, K., Hölttä, P., Hautala, T. & Wasenius, P., 1984. Metamorphism as indicator of evolution and structure of the crust in Eastern Finland. Geol. Surv. Finland, Bull. 328, 40 p.

Korsman, K., Niemelä, R. & Wasenius, P., 1988. Multistage evolution of the Proterozoic crust in the Savo schist belt, eastern Finland. In: Korsman, K. (ed.) Tectono-metamorphic evolution of the Raahe-Ladoga zone. Geol. Surv. Finland, Bull 343, 89-96.

Kousa, J., 1985. Rantasalmen tholeiittisista ja komatiittisista vulkaniiteista. Summary: The tholeiitic and komatiitic metavolcanics in Rantasalmi, southeastern Finland. Geologi 37 (2), 17-22.

Kousa, J., 1990. Paleoproterozoic metavolcanic rocks in the borderzone of Savo and Pohjanmaa, Central Finland. In: Kähkönen, Y. (ed.) IGCP Project 217 Proterozoic Geochemistry. National Working Group of Finland. Symposium, Proterozoic Geochemistry. Helsinki '90. Abstr., 29-30.

Kouvo, O. & Kulp, J.L., 1961. Isotopic composition of Finnish galenas. Ann. N.Y. Acad. Sci. 91, 476-491.

Koyaguchi, T., 1986. Textural and compositional evidence for magma mixing and its mechanism, Abu volcano group, Southwestern Japan. Contrib. Mineral. Petrol. 93, 33-45.

Kuusola, J., 1982. Savonlinnan eteläpuolella sijaitsevien Kaartilan ja Silvolan granitoidimassiivien petrologia, geokemia ja tektoninen asema. Unpubl. M. Sci. thesis. Helsinki: Univ. Helsinki, 78 p.

Laajoki, K., 1986. The Precambrian supracrustal rocks of Finland and their tectono-exogenic evolution. Precambrian Res. 33, 67-85.

Lagerblad, B. & Gorbatschev, R., 1985. Hydrothermal alteration as a control of regional geochemistry and ore-formation in the central Baltic Shield. Geol. Runsch. 74, 33-49.

Lahtinen, R., 1988. Rautalammin Pukkiharjun sulfidiesiintymän ja sen sivukivien geokemia ja petrologia. Unpubl. Ph. Lic. thesis. Helsinki: Univ. Helsinki, 167 p.

Lahtinen, R., 1989. Application of the zoning of ore-associated elements to exploration for volcanogenic massive sulphide deposits: a case study of the Pukkiharju Zn-Cu prospect, central Finland. J. Geochem. Expl. 32, 395-398.

Lahtinen, R. & Johanson, B., 1987. Barium feldspar from Pukkiharju base-metal mineralization, Central Finland. Bull. Geol. Soc. Finland 59, 77-80.

Laitala, M., 1984. Pellingin ja Porvoon kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheets 3012 and 3021, 53 p.

Latvalahti, U., 1979. Cu-Zn-Pb ores in the Aijala-Orijärvi area, southwest Finland. Econ. Geol. 74, 1035-1059.

Lawrie, K.C., 1992. Geochemical characterisation of a polyphase deformed, altered, and high grade metamorphosed volcanic terrane: implications for the tectonic setting of the Svecofennides, south-central Finland. Precambrian Res. 59, 171-205.

Leake, B.E., 1990. Granite magmas: their sources, initiation and consequences of emplacement. J. Geol. Soc. Lond. 147, 579-589.

LeBas, M.J., LeMaitre, R.W., Streckeisen, A. & Zanettin, B., 1986. A chemical classification of volcanic rocks based on total alkali-silica diagram. J. Petrol. 27, 745 - 750.

Le Fort, P., 1981. Manaslu leucogranite: a collision signature of the Himalaya. A model for its genesis and emplacement. J. Geophys. Res. 86, 10545-10568.

Le Pichon, X., Angelier, J. & Sibuet, J.C:, 1982. Plate boundaries and extensional tectonics. Tectonophysics 81, 239-256.

Lukkarinen, H., 1990. Petrography, geochemistry and tectonic setting of the Paleoproterozoic metavolcanic rocks of the Kuopio and Siilinjärvi areas, Mid-Finland. In: Kähkönen, Y. (ed.) IGCP Project 217 Proterozoic Geochemistry. National Working Group of Finland. Symposium, Proterozoic Geochemistry. Helsinki '90. Abstr., 39-40.

Luosto, U., 1991. Crustal structures of Eastern Fennoscandia. Tectonophysics 189, 19-27.

Luosto, U., Lanne, E., Korhonen, H., Guterch, A., Grad, M., Materzok, R. & Perchuc, E., 1984. Deep structure of the Earth's crust on the SVEKA profile in central Finland. Ann. Geophys. 2, 559-570.

Mäkelä, K., 1980. Geochemistry and origin of Haveri and Kiipu, Proterozoic strata-bound volcanogenic gold-copper and zinc mineralizations from southwestern Finland. Geol. Surv. Finland, Bull. 310, 79 p.

Mäki, T., 1986. Lithogeochemistry of the Pyhäsalmi zinc-copper-pyrite deposit, Finland. Prospecting in areas of glaciated terrain 1986. IMM, Kuopio 1-2.9 1986, 69-82.

Mäkitie, H., 1990. Progressiivinen metamorfoosi ja deformaatio Seinäjoen-Ilmajoen alueella. Unpubl. Ph. Lic. thesis. Helsinki: Univ. Helsinki, 80 p.

Mäkitie, H. & Lahti, S.I., 1991. Seinäjoen kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 2222, 60 p.

Makkonen, H.,1981. Säviän liuskejakson granaattikordieriitti-antofylliittikivet. Unpubl. M.Sci. thesis. Oulu: Univ. Oulu, 84 p.

Makkonen, H. & Ekdahl, E., 1988. Petrology and

structure of the Early Proterozoic Pirilä gold deposit in southeastern Finland. Bull. Geol. Soc. Finland 60, 55-66.

Mc Laughlin, R.J., Blake, M.C. Jr., Griscom, A., Blome, C.D. & Murchey, B., 1988. Tectonics of formation, translation, and dispersal of the Coast Range ophiolite of California. Tectonics 7, 1033-1056.

Mansfeld, J., Sundblad, K., Motuza, G., Vetrennikov, V. & Claesson, S., 1993. Tentative correlations between the Precambrian crust in the Baltic states and the Svecofennian domain in Sweden. Symposium on the Svecofennian Domain and Annual Meeting of IGCP-275, Abstr. Dept. Geol. Miner. Åbo: Åbo Akademi, p. 40.

Marttila, E., 1976a. Evolution of the Precambrian volcanic complex in the Kiuruvesi area, Finland. Geol. Surv. Finland, Bull. 283, 109 p.

Marttila, E., 1976b. Suonenjoen-Rautalammen alueen geologiset tutkimukset kesällä 1976. Outokumpu Oy Malminetsintä, Unpubl. Rep., 16 p.

Marttila, E., 1981. Kiuruveden kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 3323, 48 p.

Marttila, E., 1993. Pyhäjärven kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 3321, 64 p.

Matisto, A., 1976. Kangasalan kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 2141, 27 p.

Mattinson, J.M., 1982. U-Pb "blocking temperatures" and Pb loss characteristics in young zircon, sphene and apatite. The Geol. Soc. Am., 95th Ann. Meeting, Abstr. with Programs 14, 558.

Miyashiro, A., 1974. Volcanic rock series in island arcs and active continental margins. Am. J. Sci. 274, 321 - 355.

Molnar, P. & Lyon-Caen, H., 1988. Some simple physical aspects of the support, structure, and evolution of mountain belts. Spec. Pap. geol. Soc. Am. 218, 179-217.

Moore, J.M. & Waters, D.J., 1990. Geochemistry and origin of cordierite-orthoamphibole/orthopyroxene-phlogopite rocks from Namaqualand, South Africa. Chem. Geol. 85, 77-100.

Neuvonen, K.J., 1954. Stratigraphy of the schists of the Tammela-Kalvola area, southwestern Finland. Bull. Comm. géol. Finlande 166, 85 p.

Nironen, M., 1981. Kenttäraportti Luhangan Sydänmaan granitoidi-intruusioiden tutkimuksista. Univ. Helsinki, Rep. 5 3122/MN/81, 4 p.

Nironen, M., 1982. Kenttäraportti Mäntsälän granodioriittimassivin tutkimuksista. Univ. Helsinki, Rep. 13 2044/MN/82, 3 p.

Nironen, M., 1989a. Emplacement and structural setting of granitoids in the Early Proterozoic Tampere and Savo Schist Belts, Finland - implications for contrasting crustal evolution. Geol. Surv. Finland, Bull. 346, 83 p.

Nironen, M., 1989b. The Tampere Schist Belt: Structural style within an Early Proterozoic volcanic arc system in southern Finland. Precambrian Res. 43, 23-40.

Nironen, M. & Front, K., 1992. The 1.88 Ga old Mäntylä complex central Finland: emplacement and deformation of mafic to felsic plutonic rocks and associated Mo mineralization. Bull. Geol. Soc. Finland 64, 75-90.

Nesbitt, H.W. & Young, G.M., 1982. Early Proterozoic climates and plate motions inferred from major element chemistry of lutites. Nature 299, 715-717.

Nurmi, P.A., 1983. Trace element variations in the mid-Proterozoic Rautio batholith, Finland: petrogenetic implications. In: Augistithis, S.S. (ed.) The Significance of Trace Elements in Solving Petrogenetic Problems & Controversies. Athens: Theophrastus Publ., 353-376.

Nurmi, P.A., 1984. Applications of lithogeochemistry in the search for Proterozoic porphyry-type molybdenum, copper and gold deposits, southern Finland. Geol. Surv. Finland, Bull. 329, 40 p.

Nurmi, P.A. & Haapala, I., 1986. The Proterozoic granitoids of Finland: Granite types, metallogeny and relation to crustal evolution. Bull. Geol. Soc. Finland 58, 203-233.

Nurmi, P.A., Front, K., Lampio, E. & Nironen, M., 1984. Etelä-Suomen svekokarjalaiset porfyyrityyppiset molybdeeni- ja kupariesiintymät, niiden granitoidi-isäntäkivet ja litogeokemiallinen etsintä. Summary: Svecokarelian porphyry-type molybdenum and copper occurrences in southern Finland: their granitoid host rocks and lithogeochemical exploration. Geol. Surv. Finland, Rep. Invest. 67, 88 p.

Nykänen, V., 1992. Tohmajärven emäksisen kompleksin tektoninen synty-ympäristö ja differentiaatio (English summary). M. Sci. thesis. Univ. Oulu, Res Terrae, ser. B 17, 74.

Ohmoto, H. & Takahashi, T., 1983. Submarine calderas and Kuroko genesis. Econ. Geol. Mon. 5, 39-54.

Pääjärvi, A., 1991. Geological Survey of Finland. Geological map of Finland 1:100 000. Pre-quaternary rocks. Sheet 3224 Karttula.

Paavola, J., 1984. Nilsiän kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 3334, 57p.

Paavola, J., 1988. Lapinlahden kartta-alueen kalliopera (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 3332, 60p.

Park, A.F., 1984. Nature, affinities and significance of metavolcanic rocks in the Outokumpu assemblage of eastern Finland. Bull. Geol. Soc. Finland 56, 25-52.

Park, A.F., 1985. Accretion tectonism in the Proterozoic Svecokarelides of the Baltic Shield. Geology 13, 725-729.

Park, A.F., 1988. Nature of the Early Proterozoic Outokumpu Assemblage, Eastern Finland. Precambrian Res. 38, 131-146.

Park, A.F., 1991. Continental growth by accretion: A tectonostratigraphic analysis of the evolution of the western and central Baltic Shield, 2.50 to 1.75 Ga. Geol. Soc. America, Bull. 103, 522-537.

Park, A.F., 1992. Metagabbros, metasyenites and related rocks in the Outokumpu assemblage, east Finland. Bull. Geol. Soc. Finland 64, 91-111.

Park, A.F., Bowes, D.R., Halden, N.M. & Koistinen, T.J., 1984. Tectonic evolution at an early Proterozoic continental margin: the Svecokarelides of eastern Finland. J. Geodynamics 1, 359-386.

Parry, S. & Hutchinson, R.W., 1981. Origin of complex alteration assemblage, Four Corners Cu-Zn prospect, Quebec, Canada. Econ. Geol. 76, 1186-1201.

Patchett, P.J. & Kouvo, O., 1986. Origin of continental crust of 1.9-1.7 Ga age: Nd isotopes and U-Pb ages in the Svecokarelian Terrain of South Finland. Contrib. Mineral. Petrol. 92, 1-12.

Pearce, J.A., 1982. Trace element characteristics of lavas from destructive plate margins. In: Thorpe, R.S. (ed.), Andesites. Orogenic Andesites and Related Rocks. Chichester: Wiley, 525 - 548.

Pearce, J.A., 1983. Role of sub-continental lithosphere in magma genesis at active continental margins. In: Hawkesworth, C.J. & Norry, M.J. (eds.) Continental Basalts and Mantle Xenoliths. Cheshire: Shiva, 230 - 249.

Pearce, J.A., Harris, N.B.W. & Tindle, A.G., 1984a. Trace element discrimination diagrams for the tectonic interpretation of granitic rocks. J. Petrol. 25, 956-983.

Pearce, J.A., Lippard, S.J. & Roberts, S., 1984b. Characteristics and tectonic significance of suprasubduction zone ophiolites. In: Kokelaar, B.P. & Howells, M.F. (eds.) Marginal Basins Geology. Geol. Soc. London, Spec. Publ. 16, 77-94.

Pekkarinen, L.J., 1979. The Karelian formations and their depositional basement in the Kiihtelysvaara-Värtsilä area, east Finland. Geol. Surv. Finland, Bull. 301, 141 p.

Pekkarinen, L.J. & Lukkarinen, H., 1991. Paleoproterozoic volcanism in the Kiihtelysvaara-Tohmajärvi district, eastern Finland. Geol. Surv. Finland, Bull. 357, 30 p.

Peltonen, P., submitted. Petrogenesis of ultramafic rocks in the Vammala Nickel Belt: Implications for magmatism and crustal evolution of the early Proterozoic Svecofennian arc terrane. Lithos.

Perfit, M.R., Gust, D.A., Bence, A.E., Arculus, R.J. & Taylor, S.R., 1980. Chemical characteristics of island arc basalts: implications for mantle sources. Chem. Geol. 30, 227-256.

Pickering, K.T. & Taira, A., 1994. Tectonosedimentation: with examples from the Tertiary-Recent of Southeast Japan. In: Hancock, P.L. (ed.) Continental Deformation. Oxford: Pergamon Press, 320-354.

Pitcher, W.S., 1987. Granites and yet more granites forty years on. Geol. Rundsch. 76, 51-79.

Pitcher, W.S., 1993. The Nature and Origin of Granite. Glasgow: Blackie Academic&Professional, 321 p.

Pitkänen, P., 1985. Anttolan Luonterin postorogeenisen intruusion petrologia ja geokemia. Unpubl. M. Sci. thesis. Helsinki: Univ. Helsinki, 81 p.

Pukhtel, I.S., Zhuravlev, D.Z., Ashikmina, N.A., Kulikov, V.S. & Kulikova, V.V., 1992. Sm-Nd age of the Suisarian suite on the Baltic Shield. Doklady Akademii Nauk 326, 706-711. (in Russian).

Puura, V., Vaher, R., Klein, V., Koppelmaa, H., Niin, M., Vanamb, V. & Kirs, J., 1983. The Crystalline Basement of Estonian Territory. Nauka, Moscow, 208 p. (in Russian).

Puura, V. & Huhma, H., 1993. Paleoproterozoic age of the East Baltic granulitic crust. Precambrian Res. 64, 289-294.

Puustjärvi, H., 1981. Aijala-Orijärven alueen kordieriittia ja antofylliittia sisältävistä kivistä. Unpubl. M.Sc. thesis. Turku: Univ. Turku, 100 p.

Rasilainen, K., 1991. Geochemistry and wall rock alteration in at the Kangasjärvi massive sulphide deposit, Central Finland. In: Autio, S. (ed.) Current Research 1989-1990. Geol. Surv. Finland, Spec. Pap. 12, 107-110.

Rauhamäki, E., Mäkelä, T. & Isomäki, O-P., 1980. Geology of the Vihanti mine. In: Häkli, T.A. (ed.) Precambrian ores of Finland. Guide to excursions 078 A+C, Part 2 (Finland). 26th Int. Geol. Congr. Paris 1980. Geol. Surv. Finland, 14-24.

Rehtijärvi, P., 1984. Distributions of phosphorous, sulphur and sulphur isotopes in a strata-bound base metal deposit, Kangasjärvi, Finland. Geol. Surv. Finland, Rep. Invest. 65, 16 p.

Rehtijärvi, P., Äikäs, O. & Mäkelä, M., 1979. A middle Precambrian Uranium- and Apatite-Bear-

ing Horizon Associated with the Vihanti Zinc Ore Deposit, Western Finland. Econ. Geol. 74, 1102-1117.

Reid, J.B., Evans, O.C. & Fates, D.G., 1983. Magma mixing in granitic rocks of the central Sierra Nevada, California. Earth Planet. Sci. Lett. 66, 243-261.

Reinhardt, J., 1987. Cordierite-anthophyllite rocks from north-west Queensland, Australia: metamorphosed magnesian pelites. J. Metam. Geol. 5, 451-472.

Rice, A., 1981. Convective fractionation: a mechanism to provide cryptic zoning (macrosegregation) layering, crescumulates, banded tuffs and explosive volcanism in igneous processes. J. Geophys. Res. 86, 405-417.

Riggs, S.R., 1986. Proterozoic and Cambrian phosphorites - specialist studies: phosphogenesis and its relationship to exploration for Proterozoic and Cambrian phosphorites. In: Cook, P.J. & Shergold, J.H. (eds.) Phosphate Deposits of the World: 1, Proterozoic and Cambrian Phosphorites. Cambridge: Cambridge University Press, 352-368.

Roedder, E., 1979. Silicate liquid immiscibility in magmas. In: Yoder, H.S., Jr (ed.) The evolution of igneous rocks: fiftieth anniversity perspectives. Princeton, NJ, Princeton University Press, 483-520.

Rogers, G.R. & Hawkesworth, C.J., 1989. A geochemical traverse across the north Chilean Andes: Evidence for crust generation from the mantle wedge. Earth and Planet. Sci. Lett. 91, 271-285.

Rosenberg, R.J., Kaistila, M. & Zilliacus, R., 1982. Instrumental neutron activation analysis of solid geochemical samples. J. Radioanal. Chem. 71, 419 - 428.

Rutter, M.J. & Wyllie, P.J., 1989. Experimental study of interaction between hydrous granite melt and amphibolite. Geol. Mag. 126, 633-646.

Saksela, M., 1936. Über die geologischen Kartierung und die Einteilung der Granite im finnischen Grundgebirge. Bull. Comm. géol. Finland 115, 275-292.

Sakuyama, M., 1983. Petrology of arc volcanic rocks and their origin by mantle diapirs. J. Volcanol. Geotherm. Res. 18, 297-320.

Salli, I., 1983. Pielaveden kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 3314, 29p.

Savolahti, A., 1966. On rocks containing garnet, hypersthene, cordierite and gedrite in the Kiuruvesi region, Finland. Part 1: Juurikkajärvi. Bull. Comm. géol. Finlande 222, 343-386.

Schreurs, J. & Westra, L, 1985. Cordierite-orthopyroxene rocks: the granulite facies equivalents of the Orijärvi cordierite-anthophyllite rocks in West Uusimaa, southwest Finland. Lithos 18, 215-228.

Searle, M.P., Windley, B.F., Coward, M.P., Cooper, D.J.W., Rex, A.J., Rex, D.C., Tindong, L., Xuchang, X., Jan, M.Q., Thakur, V.C. & Kumar, S., 1987. The closing of Tethys and the tectonics of the Himalaya. Bull. Geol. Soc. Am. 98, 678-701.

Sederholm, J.J., 1897. Über eine archäische Sedimentformation im südwestlichen Finland und ihre Bedeutung für die Erklärung der Enstehungsweise des Grundebirges. Bull. Comm. géol. Finlande 6, 254 p.

Sederholm, J.J., 1932. On the geology of Fennoscandia with special reference to the pre-Cambrian. Bull. Comm. géol. Finland 98, 30 p.

Sharma, M., Basu, A.R., Cole, R.B. & DeCelles, P.G., 1991. Basalt-Rhyolite Volcanism by MORB-Continental Crust Interaction: Nd, Sr-Isotopic and Geochemical Evidence from Southern San Joaquin Basin, California. Contib. Min. Petrol. 109, 159-172.

Shaw, D.M., 1970. Trace element fractionation during anatexis. Geochim. Cosmochim. Acta 34, 237-243.

Shcherbak, N.P., 1991. The Precambrian subdivisions of the Ukrainian Shield and their comparison with USSR and international scales. Geol. Zh. 4, 3-9. (in Russian, with an English summary).

Sibuet, J.C., Letouzey, J., Barbier, F., Charvet, J., Foucher, J.P., Hilde, T.W.C., Kimura, M. Chiao, L.Y., Marsset, B., Muller, C. & Stephan, J.F., 1987. Back arc extension in the Okinawa Trough. J. Geophys. Res. 92, 14041-14063.

Simonen, A., 1948. On the Petrology of the Aulanko Area in Southwestern Finland. Bull. Comm. géol. Finlande 143, 66 p.

Simonen, A., 1960. Plutonic rocks of the Svecofennides in Finland. Bull. Comm. géol. Finland 189, 101 p.

Simonen, A., 1980a. Prequaternary rocks of Finland. 1:1 000 000. Espoo: Geol. Surv. Finland.

Simonen, A., 1980b. The Precambrian of Finland. Geol. Surv. Finland, Bull. 304, 58 p.

Sjöblom, B., 1990. Mäntän kartta-alueen kallioperä (English summary). Geol. Surv. Finland, Geological map of Finland 1:100 000. Explanation to the map of rocks, sheet 2231, 64 p.

Skiöld, T., Öhlander, B., Markkula, H., Widenfalk, L. & Claesson L-Å., 1993. Chronology of Proterozoic orogenic processes at the Archean continental margin in northern Sweden. Precambrian Res. 64, 225-238.

Skridlaite, G., 1993. The prograde metapelitic granulites of western Lithuania. Symposium on the Svecofennian Domain and Annual Meeting of

IGCP-275, Abstr. Dept. Geol. Miner. Åbo: Åbo Akademi, p. 53.

Smith, M.S., Dymek, R.F. & Schneiderman, J.S., 1992. Implications of trace element geochemistry for the origin of cordierite-ortoamphibole rocks from Orijärvi, SW Finland. J. Geol. 100, 545-559.

Sokolov, V.A., 1980. Jatulian formations of the Karelian ASSR. In: Silvennoinen, A. (ed.), Jatulian Geology in the Eastern Part of the Baltic Shield. Proceedings of a Finnish-Soviet Symposium held in Finland, 21th-26th August, 1979. Rovaniemi. The Committee for Scientific and Technical Co-operation between Finland and Soviet Union, 163-220.

Sonder, L.J., England, P.C., Wernicke, B.P. & Christiansen, R.L., 1987. A physical model for Cenozoic extension of western North America. In: Coward, M.P., Dewey, J.F. & Hancock, P.L. (eds.) Continental Extensional Tectonics. Spec. Publ. geol. Soc. Lond. 28, 187-202.

Sorjonen-Ward, P., Claoué-Long, J. & Huhma, H., 1994. Shrimp isotope studies of granulite zircons and their relevance to Early Proterozoic tectonics in Northern Fennoscandia. In: Lanphere, M.A., Dalrymple, G.B. & Turrin, B.D. (eds.) Abstracts of the eight International Conference on Geochronology, Cosmochronology, and Isotope Geology. U.S. Geol. Surv., Circ. 1107, 299.

Sparks, R.S.J. & Marshall, L.A., 1986. Thermal and mechanical constraints on mixing between mafic and silicic magmas. J. Volcan. Geothermal Res. 29, 99 - 124.

Sparks, R.S.J., Huppert, H.E. & Turner, J.S., 1984. The fluid dynamics of evolving magma chambers. Phil. Trans. R. Soc. Lond. A310, 511-534.

Spulber, S.D. & Rutherford, M.J., 1983. The Origin of Rhyolite and Plagiogranite in Oceanic crust: An Experimental Study. J. Petrology 24, 1-25.

Stern, C.R. & Wyllie, P.J., 1978. Phase compositions through crystallization intervals in basalt-andesite-H_2O at 30 kbar with implications for subduction zone magmas. Am. Mineral. 63, 641-663.

Stern, C.R., Huang, W.-L., & Wyllie, P.J., 1975. Basalt-andesite-rhyolite-H_2O: crystallization intervals with excess H_2O and H_2O-undersaturated liquidus surfaces to 35 kbars, with implications for magma genesis. Earth Planet. Sci. Lett. 28, 189-196.

Suominen, V., 1991. The chronostratigraphy of southwestern Finland, with special reference to Postjotnian and Subjotnian diabases. Geol. Surv. Finland, Bull. 356, 100 p.

Talvitie, J., 1975. Fractures, dynamic model and Ni-Cu mineralized basic intrusives in Central Finland. Geol. Surv. Finland, Rep. Invest. 10, 3-12.

Talvitie, J. & Paarma, H., 1980. On Precambrian basic magmatism and the Ti-Fe ore formation in central and northern Finland. In: Siivola, J. (ed.) Metallogeny of the Baltic Shield. Proceedings of the symposium held in Helsinki, Finland, June 12-21, 1978. Geol. Surv. Finland, Bull. 307, 98-106.

Tamaki, K., 1988. Geological structure of the Japan Sea and its tectonic implications. Bull. Geol. Surv. Japan 39, 269-365.

Tanimura, S., Date, J:, Takahashi, T. & Ohmoto, H., 1983. Stratigraphy and structure of the Hokuroku district. Econ. Geol., Mon. 5, 24-38.

Tikoff, B. & Teyssier, C., 1992. Crustal-scale, en echelon "P-shear" tensional bridges: A possible solution to the batholithic room problem. Geology 20, 927-930.

Treloar, P.J., Koistinen, T.J. & Bowes, D.R., 1981. Metamorphic development of cordierite-amphibole rocks and mica schists in the vicinity of the Outokumpu ore deposit, Finland. Trans. R. Soc. Edinburgh, Earth Sci. 72, 201-215.

Tuominen, H.V. & Mikkola, T., 1950. Metamorphic Mg-Fe enrichment in the Orijärvi region as related to folding. Bull. Comm. géol. Finlande 150, 67-92.

Turner, S., Sandiford, M. & Foden, J., 1992. Some geodynamic and compositional constraints on "postorogenic" magmatism. Geology 20, 931-934.

Vaarma, M., 1990. Pohjanmaan liuskevyöhykkeen geologia Evijärven alueella. Unpubl. Ph. Lic. thesis. Helsinki: Univ. Helsinki, 101 p.

Vaasjoki, M., 1977. Rapakivi granites and other postorogenic rocks in Finland: their age and the lead isotopic composition of certain associated galena mineralizations. Geol Surv. Finland, Bull. 294, 71 p.

Vaasjoki, M., 1981. The lead isotopic composition of some Finnish galenas. Geol. Surv. Finland, Bull. 316, 30 p.

Vaasjoki, M., 1989. Isotopic studies on the Rauhala base metal deposit and its environment in western Finland. In: Autio, S. (ed.) Current Research 1988. Geol. Surv. Finland, Spec. Pap. 10, 37-38.

Vaasjoki, M. & Huhma, H., 1987. Lead isotopic results from metabasalts of the Haveri formation, southern Finland: an indication of early Proterozoic mantle derivation. Terra Cognita 7, 159.

Vaasjoki, M. & Sakko, M., 1988. The evolution of the Raahe-Ladoga zone in Finland: isotopic constraints. In: Korsman, K. (ed.) Tectono-metamorphic evolution of the Raahe-Ladoga zone. Geol. Surv. Finland, Bull. 343, 7-32.

Vaasjoki, M. & Lahti, S.I., 1991. Zircon U-Pb age determinations from the Mustajärvi area, western Finland. In: Autio, S. (ed.) Current Research 1989-1990. Geol. Surv. Finland, Spec. Pap. 12, 49-52.

Vaasjoki, M. & Kontoniemi, O., 1991. Isotopic

studies from the Proterozoic Osikonmäki gold prospect at Rantasalmi, southeastern Finland. In: Autio, S. (ed.) Current Research 1989-1990. Geol. Surv. Finland, Spec. Pap. 12, 53-57.

Vaasjoki, M., Rämö, T. & Sakko, M., 1991. New U-Pb ages from the Wiborg rapakivi area: constraints on the temporal evolution of the rapakivi-anorthosite-diabase dyke association of southeastern Finland. Precambrian Res. 51, 227-243.

Vallance, T.G., 1967. Mafic rock alteration and isochemical development of some cordierite-anthophyllite rocks. J. Petrol. 8, 84-96.

Väyrynen, H., 1938. Notes on the geology and tectonics of Karelia and Onega region in the summer of 1937. Bull. Comm. géol. Finlande 123, 66-80.

Väyrynen, H., 1939. On the geology and tectonics of the Outokumpu ore field and region. Bull. Comm. géol. Finlande 124, 1-91.

Vernon, R.H., 1983. Restite, xenoliths and microgranitoid enclaves in granites. J. Proc. R. Soc. New South Wales 116, 77-103.

Vuollo, J., Piirainen, T. & Huhma, H., 1992. Two Early proterozoic tholeiitic diabase dyke swarms in the Koli-Kaltimo area, eastern Finland - their geological significance. Geol. Surv. Finland, Bull. 363, 32 p.

Vuollo, J., 1994. Paleoproterozoic basic igneous events in eastern Fennoscandian Shield between 2.45 and 1.97 Ga. Ph.D. thesis. Acta Universitaes Ouluensis, ser. A 250, 47 p.

Wahl, W.A., 1936. The granites of the Finnish part of the Svecofennian Archaean mountain chain. Bull. Comm. géol. Finlande 115. 489-505.

Walford, P.C. & Franklin, J.M., 1982. The Anderson Lake mine, Snow Lake, Manitoba. In: Hutchinson, R.W., Spence, C.D. & Franklin, J.M. (eds.) Precambrian sulphide deposits. Geol. Assoc. Canada, Spec. Pap. 25, 481-523.

Ward, P., 1987. Early Proterozoic deposition and deformation at the Karelian craton margin in southeastern Finland. Precambrian Res. 35, 71-93.

Ward, P., 1988a. Mesoscopic scale structural analysis. In: Gaál, G. (ed.) Exploration target selection by integration of geodata using statistical and image processing techniques: an example from central Finland, Part I (text). Geol. Surv. Finland, Rep. Invest. 80, 50 - 53.

Ward, P., 1988b. Early Proterozoic Kalevian lithofacies and their interpretation in the Hammaslahti-Rääkkylä area, eastern Finland. In: Laajoki, K. & Paakkola, J. (eds.) Sedimentology of the Precambrian formations in eastern and northern Finland, 1986. Geol. Surv. Finland, Spec. Pap. 5, 29-48.

Ward, P. & Kohonen, J., 1989. Structural provinces and style in the Proterozoic of North Karelia: preliminary correlations and discussion. In: Autio, S. (ed.) Current Research 1988. Geol. Surv. Finland, Spec. Pap. 10, 23-29.

Ward, P., Härkönen, I., Nurmi, P.A. & Pankka, H.S., 1989. Structural studies in the Lapland greenstone belt, Northern Finland and their application to gold mineralization. In: Autio, S. (ed.) Current Research 1988. Geol. Surv. Finland, Spec. Pap. 10, 71-77.

Wasström, A., 1993. The Knaften granitoids of Västerbotten County, northern Sweden. In: Lundqvist, T. (ed.), Radiometric dating results. Geol. Surv. Sweden, Res. Pap. C 823, 60-64.

Watson, E.B. & Harrison, T.M., 1983. Zircon saturation in felsic liquids: experimental data and applications to trace element geochemistry. Contrib. Mineral. Petrol. 70, 407-419.

Welin, E., Vaasjoki, M. & Suominen, V., 1983. Age differences between Rb-Sr whole rock and U-Pb zircon ages of syn- and postorogenic Svecokarelian granitoids in Sottunga, SW Finland. Lithos 16, 297-305.

Whalen, J.B., Currie, K.L. & Chappell, B.W., 1987. A-type granites: geochemical characteristics, discrimination and petrogenesis. Contrib. Mineral. Petrol. 95, 407-419.

White A.J.R. & Chappel, B.W., 1977. Ultrametamorphism and granitoid genesis. Tectonophysics 43, 7-22.

Winchester, J.A. & Floyd, P.A., 1976. Geochemical magma type discrimination; application to altered and metamorphosed basic igneous rocks. Earth Planet. Sci. Lett. 28, 459-469.

Woodcock, N.H. & Schubert, C., 1994. Continental Strike-Slip Tectonics. In: Hancock, P.L. (ed.) Continental Deformation. Oxford: Pergamon Press, 251-263.

Wyllie, P.J., 1984. Constrains imposed by experimental petrology on possible and impossible magma sources and products. Phil. Trans. R. Soc. Lond. A310, 439-456.

Yoder, H.S., 1969. Calc-alkalic andesites: experimental data bearing on the origin of their assumed characteristics. In: Birney, A.R. (ed.) Proceedings of the Andesite Conference. Oreg. Dep. Geol. Miner., Ind. Bull. 65, 77-89.

Zartman, R.E., Futa, K. & Peng, Z.C., 1991. A comparison of Sr-Nd-Pb isotopes in young and old continental lithospheric mantle: Patagonia and eastern China. Austr. J. Earth Sci. 38, 545-557.

Zindler, A. & Hart, S.R., 1986. Chemical geodynamics. Ann. Rev. Earth Planet. Sci. 14, 493-571.

Zorpi, M.J., Coulon, C. & Orsini, J.B., 1991. Hybridization between felsic and mafic magmas in calc-alkaline granitoids - a case study in northern Sardinia, Italy. Chem. Geol. 92, 45-86.

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