POSIVA 2002-04

Structure and geological evolution of the bedrock of southern Satakunta, SW Finland

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February 2002

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TILAUSNUMEROT: 9632/01/LIW

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Posiva-raportti: 2002-04

STRUCTURE AND GEOLOGICAL EVOLUTION OF THE BEDROCK OF SOUTHERN SATAKUNTA, SW FINLAND

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Hyv. Posiva-raporttina
12.3.2002

Posiva-raportti – Posiva Report

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Töölönkatu 4, FIN-00100 HELSINKI, FINLAND
Puh. (09) 2280 30 – Int. Tel. +358 9 2280 30

Raportin tunnus – Report code
POSIVA 2002-04

Julkaisuaika – Date
February 2002

Tekijä(t) – Author(s)

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Toimeksiantaja(t) – Commissioned by

Posiva Oy

Nimeke – Title

STRUCTURE AND GEOLOGICAL EVOLUTION OF THE BEDROCK AT SOUTHERN SATAKUNTA, SW FINLAND

Tiivistelmä – Abstract

The southern Satakunta area lies on the west coast of Finland, mainly covering the mainland (with main towns Pori and Rauma), but also including the coastal archipelago and part of the Bothnian Sea. Near the centre of the area lies the island of Olkiluoto, on which Finland's site for a deep repository for spent nuclear fuel is located. The purpose of the present report is to compile and interpret all available geological and geophysical data relevant to understanding the regional geological setting of the Olkiluoto site. The area described is covered by four 1:100 000 scale geological map sheets, published by the Geological Survey of Finland, which, together with low-altitude aeromagnetic maps, provide the basis for a new 1:250 000 geological map compilation. This shows that the bedrock of southern Satakunta can be subdivided into three main zones: a pelitic migmatite belt in the southwest, a central, NW-SE trending area of sandstone, and a psammitic migmatite belt in the northeast. The migmatite belts formed during the Svecofennian orogeny, 1900-1800 Ma ago (Palaeoproterozoic). The sandstone area is the remnant of an alluvial basin, preserved now in a NW-SE trending graben, bounded on both sides by normal fault zones. The sandstones are thought to be at least 1400-1300 Ma old (Mesoproterozoic), and they are cut by Postjotnian olivine diabase dykes, 1270-1250 Ma in age.

The Svecofennian migmatite belts show a complex history of formation, with various phases of anatexis/metamorphism, deformation and intrusion. In the pelitic migmatite belt, in which the Olkiluoto site is situated, four phases of ductile deformation (D_1-D_4) and two phases of regional highT/lowP metamorphism and migmatite formation can be recognised, together with synorogenic (tonalite, granodiotite) and late orogenic (potassium granite) intrusions. Subsequently, this very heterogeneous complex was intruded by anorogenic rapakivi granites, with ages 1580-1550 Ma. One pluton, the Eurajoki stock, approaches to within 5 km of the Olkiluoto site. The results of gravimetric surveys have indicated that the margin of the Eurajoki stock slopes westward underneath the site, but to depths in excess of 3000 m.

Plate tectonic reconstructions of the Precambrian of Finland, partly based on the results of major deep seismic sounding experiments, such as the international GGT/SVEKA project (along a NE-SW transect through the Satakunta area), indicate the pelitic and psammitic migmatite belts in Satakunta represent parts of the early Proterozoic Southern Finland and Central Finland continental arcs, respectively. Collision of these arc complexes took place 1890 - 1880 Ma ago, when the rocks were deformed and metamorphosed for the first time. The highT/lowP metamorphism was caused by mafic underplating, which led to a strong increase in temperature, and recrystallisation and partial remelting of the rocks in the upper crust. The collision of the arc complexes is characterised by an intense magmatic activity, which appears as synorogenic granitoids. In the next stage, 1860 - 1810 Ma ago, mafic underplating caused a second high-temperature metamorphic event and partial melting of the sedimentary rocks in southern Finland, producing the late-orogenic potassium. granites, dated at 1840 - 1830 Ma.

The Subjotnian rapakivi granites associated with mafic rocks, the Jotnian Satakunta sandstone formation and the Postjotnian diabase dykes and sills represent the cratonisation stages of the Svecofennides. Rapakivi granites and related mafic rocks were generated in an anorogenic extensional regime by partial melting of the upper mantle and lower crust. The Jotnian Satakunta sandstone is a fluvial sediment formation deposited in a deltaic environment. The development of the graben or rift valley, where the sandstone was deposited, may have begun already during the Subjotnian, ca. 1650 Ma ago. The olivine diabase dykes represent the feeding channels of Postjotnian flood basalts. However, no such volcanic rocks have preserved in the area. The intrusion of the diabase dykes caused the sinking of the graben after the sedimentation of the sandstone. Sedimentary rocks younger than the Precambrian have not been preserved in the study area.

Avainsanat - Keywords
Migmatites, plutonic rocks, rapakivi granites, sandstone, olivine diabase, deformation, metamorphism, geophysics, crustal structure, Precambrian, plate tectonics, Svecofennian orogeny, Satakunta, SW Finland

ISBN	ISSN
ISBN 951-652-111-8	ISSN 1239-3096
Sivumäärä – Number of pages	Kieli – Language
125	English

Posiva-raportti – Posiva Report

Posiva Oy
Töölönkatu 4, FIN-00100 HELSINKI, FINLAND
Puh. (09) 2280 30 – Int. Tel. +358 9 2280 30

Raportin tunnus – Report code
POSIVA 2002-04

Julkaisuaika – Date
Helmikuu 2002

Tekijä(t) – Author(s)

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Toimeksiantaja(t) – Commissioned by

Posiva Oy

Nimeke – Title

ETELÄISEN SATAKUNNAN KALLIOPERÄN RAKENNE JA KEHITYSHISTORIA

Tiivistelmä – Abstract

Eteläisen Satakunnan vanhimmat kivilajit ovat savi- ja hiekkasyntyisiä kiillegneissejä, jotka suureksi osaksi esiintyvät seoskivilajeina eli migmatiitteina. Niitä leikkaavat syväkivet ovat granodioriitteja ja tonaliitteja sekä graniitteja ja pegmatiitteja. Alueen eteläosa koostuu Laitilan rapakivimassiiviin kuuluvasta rapakivigraniitista, johon ns. satelliittiesiintymänä liittyy Eurajoen rapakivistokki. Satakunnan hiekkakivimuodostuma on kerrostunut kallioperän lohkoliikuntojen muodostamaan hautavajoamaan. Kaikkia edellä mainittuja kivilajeja leikkaavat vaaka- ja pystyasentoiset oliviinidiabaasijuonet. Alueen koillisosassa sijaitseva Sääksjärvi on alkuperältään meteoriittikraatteri, joka on syntynyt meteoriitin törmäyksessä noin 560 miljoonaa vuotta sitten.

Pintasyntyiset kivilajit kerrostuivat merellisissä saarikaariympäristöissä. Svekofennisen vuorijonomuodostuksen päävaiheessa Etelä- ja Keski-Suomen kaarikompleksit törmäsivät 1890 - 1880 miljoonaa vuotta sitten. Törmäyksessä eteläinen saarikaari (ml. hiekkakivimuodostuman lounaispuoliset pintakivet) työntyi pohjoista saarikaarta (ml. hiekkakivimuodostuman koillispuolisen alueen pintakivet) vasten ja sen yli ja viimeksi mainitun sisältänyt laatta painui edellisen alle. Törmäyksessä kivilajit joutuivat syvälle maankuoreen ja deformoituivat sekä metamorfoituivat ensimmäisen kerran 800 - 670°C:en lämpötilassa ja 5 - 6 kb:n paineessa. Törmäystä luonnehti voimakas magmaattinen toiminta, mikä ilmenee ns. synorogenisina (vuorijonomuodostuksen päävaiheen aikaisina) syväkivinä, jotka tutkimusalueella ovat koostumukseltaan granodioriitteja, tonaliitteja ja trondhjemiitteja. Kallioperän seuraavassa kehitysvaiheessa vaipasta peräisin olevan kivisulan asettuminen kuoren ja vaipan rajapinnalle aiheutti korkean lämpötilan (700 - 800 °C) metamorfoosin ja sedimenttikivien lähes täydellistä sulamista Etelä-Suomessa. Vaiheelle ovat ominaisia 1840 - 1830 miljoonan vuoden ikäiset ns. myöhäisorogeniset (vuorijonomuodostuksen myöhäisvaiheen aikaiset) kaligraniitit.

Svekofennisen kallioperän kratonisoitumisvaihetta tutkimusalueella edustavat 1580 - 1550 miljoonan vuoden ikäiset rapakivigraniitit, Satakunnan hiekkakivimuodostuma sekä 1270 - 1250 miljoonaan vuoden ikäiset oliviinidiabaasijuonet. Rapakivigraniitit ovat peräisin kivisulista, jotka muodostuivat kuoren alaosan sulaessa osittain, ja ne ovat asettuneet paikalleen kuoren ollessa vetojännityksen luonnehtimassa tilassa. Satakunnan hiekkakivimuodostuman yläosat kerrostuivat noin 1400 - 1300 miljoonaa vuotta sitten, mutta muodostuman kehittyminen on voinut alkaa jo rapakivien paikalleen asettumisen aikoihin. Oliviinidiabaasijuonet ovat alkuperältään mantereellisten laakiobasalttien tulokanavia ja edustavat alueen viimeistä proterotsooista kehitysvaihetta. Selkämerellä hiekkakiven päällä olevia kambri- ja ordovikikautisia sedimenttikiviä ei esiinny tutkimusalueella.

Avainsanat - Keywords
Migmatiitit, syväkivet, rapakivi, hiekkakivi, oliviinidiabaasi, deformaatio, metamorfoosi, geofysiikka, kuoren rakenne, prekambrinen, laattatektoniikka, svekofenninen orogenia, Satakunta

ISBN
ISBN 951-652-111-8

ISSN
ISSN 1239-3096

Sivumäärä – Number of pages
125

Kieli – Language
Englanti

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PREFACE

This study was carried out under contract to Posiva Oy (order 9632/01/LIW). On behalf of Posiva Oy the work has been supervised by Liisa Wikström, whereas Seppo Paulamäki at the Geological Survey of Finland was the contact person on behalf of the authors.

The report has been prepared by the experts who have participated in the studies of the Olkiluoto site. Seppo Paulamäki is responsible for Sections 1, 2, 3, 6 and 7, Markku Paananen for Sections 4 and 5 and Seppo Elo for Section 4.7. Within the Geological Survey the report has been reviewed by research professor Kalevi Korsman, senior scientist Veli Suominen, geologist Matti Pajunen, geologist Pekka Pihlaja, and geologist Arja Lehtonen. The manuscript profited from their critical comments and suggestions.

The authors wish to thank Pekka Anttila, Fortum Engineering and Prof. Alan Geoffrey Milnes of GEA Consulting, Sweden, for their valuable comments. A.G. Milnes is also thanked for corrections to the English of the text.

1 INTRODUCTION

In Finland, two companies utilise nuclear energy to generate electric power – Teollisuuden Voima Oy (TVO) and Fortum Power and Heat Oy (formerly Imatran Voima Oy). The companies are preparing for the final disposal of the spent nuclear fuel waste deep in the bedrock. In 1996, they established a joint company, Posiva, to run the programme of site investigations and other research and development for spent fuel disposal. Posiva will ultimately construct and operate the future disposal facility. On the basis of site investigations at several study sites since 1987, Posiva submitted an application to the Government in May 1999 for a Decision in Principle to build a final disposal facility for spent fuel in Olkiluoto, Eurajoki. A positive decision was made at the end of 2000 by the Government, and in May 2001 the Finnish Parliament ratified the Decision in Principle. The construction of the facility should start after 2010, and the operation of the final disposal facility will start in 2020.

This report is carried out under contract to Posiva, and it summarises the essential existing geological and geophysical data on the southern Satakunta area, in which the Olkiluoto site lies. The study area comprises the 1:100 000 map-sheets 1132, 1134, 1141 and 1143 with main towns Pori and Rauma (Fig. 1-1). The purpose of the present report is to compile and interpret all available geological and geophysical data relevant to understanding the regional geological setting of the Olkiluoto site. The report will serve as a basis for future regional geological studies, which will supplement the present knowledge of the bedrock at Olkiluoto and its geological evolution. The literature concerning the study area has been collected from the online databases of the Geological Survey of Finland (GTK). Additionally, it has been possible for the authors to benefit from the expertise of the research scientists of the GTK who have been working in the Satakunta area.

Chapter 2 describes the rock types of the study area, starting with a short introduction to the Precambrian geology of Finland and southern Satakunta. The 1:100 000 geological map-sheets 1132 Rauma (Suominen & Torssonen 1993), 1134 Kokemäki (Hämäläinen 1994), 1141 Luvia (Pihlaja & Kujala 1994) and 1143 Pori (Pihlaja 1994), published by the Geological Survey of Finland, provide, together with low-altitude aeromagnetic maps, the basis for a new 1:250 000 geological map compilation presented in Fig. 2.1-3 and App. 1. In Chapter 3 the deformation and metamorphism is elucidated with the help of studies made in the study area and elsewhere in southwestern Finland within comparable bedrock areas.

The geophysical characterisation in Chapter 4 is based on the examination of the aeromagnetic map and on the various existing interpretations, mainly concerning the properties and occurrence of the rapakivi granites, diabases and sandstone. The petrophysical data has been compiled from the petrophysical register of the GTK. In addition the report summarises the results of a gravity survey carried out in the Olkiluoto area during the summer 2000 by Elo (2001). The aims of the survey were to study the contacts of the Eurajoki rapakivi stock and its possible subsurface extension into the Olkiluoto island, to estimate the thickness of the diabase sills on the northern side of the Eurajoki rapakivi stock, and to obtain information on the nature of the granites and gneisses in Olkiluoto.

In Chapter 5 the deep crustal structure of the area has been examined on the basis of the deep seismic BABEL profiles and the GGT-SVEKA transect, which combine the geophysical data (seismic, electric, gravimetric, thermic, magnetic and petrophysical) with the geological data (lithology, structural geology, metamorphism, isotopes and geochemistry).

In Chapter 6 the geological evolution of the bedrock of southern Satakunta is examined in the context of the plate tectonic evolution of Precambrian bedrock in southern and southwestern Finland. The summary of the geological evolution of the study area is presented in Chapter 7.

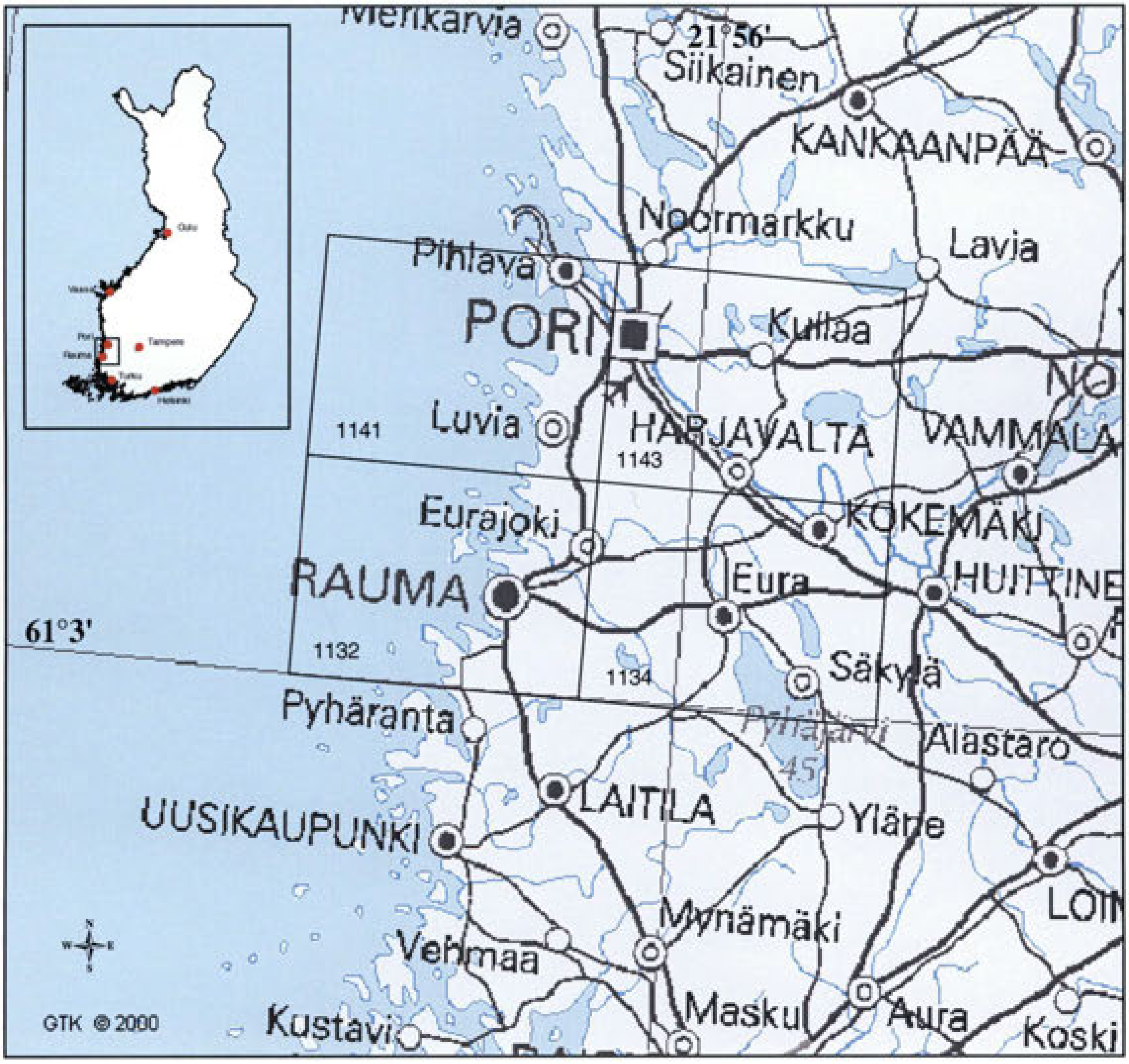


Figure 1-1. The location of the study area (1: 100 000 map-sheets 1132 Rauma, 1134 Kokemäki, 1141 Luvia and 1143 Pori). Base map© National Land Survey, permission no. 41/MYY/01, map scale 1:1000 000.

2 PETROGRAPHY

2.1 Introduction

2.1.1 Precambrian bedrock of Finland

The crystalline bedrock of Finland (Fig. 2.1-1) is a part of the extensive Precambrian Fennoscandian Shield. The oldest part of the Finnish bedrock is formed by 3100 - 2500 Ma old Archaean basement complex in northeastern Finland and consists mainly of tonalitic to granodioritic gneisses and migmatites (Fig. 2.1-2 and no. 37 in Fig. 2.1-1) (Gaál & Gorbatschev 1987). Within the complex narrow Archaean greenstone belts occur, composed of metavolcanics and metasediments (Fig. 2.1-1: nos. 34 and 35), and intruded by tonalitic to granitic magmas (Fig. 2.1-1: no. 33) (Luukkonen 1992).

About 2440 Ma ago layered gabbro intrusions (Fig 2.1-1: no. 31) were emplaced in northern and northeastern Finland (Alapieti 1982) together with corresponding mafic dyke swarms (Vuollo 1994). The Archaean craton is discordantly overlain by 2500 - 2000 Ma old Sumian-Sariolan, Kainuan, Jatulian and Kalevian metasediments and metavolcanics (Fig. 2.1-2 and nos. 20, 27 - 29, 30 and 32 in Fig. 2.1-1), which are cut by 1970 - 2200 Ma diabase dykes (Laajoki 1991). About 1960 Ma old ophiolite complexes in eastern Finland (Fig. 2.1-2 and no. 22 in Fig. 2.1-1) represent an ancient ocean floor.

Most of the Finnish bedrock is composed of Palaeoproterozoic metamorphic rocks (Fig. 2.1-1: nos. 19 and 21) and igneous rocks (Fig. 2.1-1: nos. 13-18) of the Svecofennian Domain. The oldest Svecofennian rocks are 1930 - 1920 Ma old gneissic tonalites and volcanites of the Pyhäsalmi primitive island arc (Fig. 2.1-2). The main parts of the Svecofennian are composed of the Central Finland continental arc and the Southern Finland sedimentary-volcanic complex (Korsman et al. 1999) (Fig. 2.1-2). The former consists of the Central Finland granitoid complex, psammitic migmatite belt and the Tampere schist belt between the two. The Southern Finland sedimentary-volcanic complex consists of metasediments and metavolcanites of the pelitic migmatite belt. The pelitic and psammitic migmatite belts can be distinguished on the basis of predominant granitic and trondhjemitic to granodioritic leucosomes, respectively. The long history of volcanism, sedimentation and igneous activity of the Svecofennian Domain culminated in the Svecofennian orogeny 1900 – 1800 Ma ago (Koistinen 1996). Later the crust was intruded by anorogenic rapakivi granites, 1650 – 1540 Ma in age (Fig. 2.1-1: no. 8) with associated mafic dyke swarms (Fig. 2.1-1: no. 10). The youngest basement rocks are the 1400 – 1300 Ma old Jotnian sandstones (Fig. 2.1-1: no. 7) cut by Postjotnian (1270 – 1250 Ma) olivine diabase dykes and sills (Fig. 2.1-1: no. 6), and the 1100 and 1000 Ma dykes of Salla and Laanila in northern Finland (Fig. 2.1-1: no. 5).

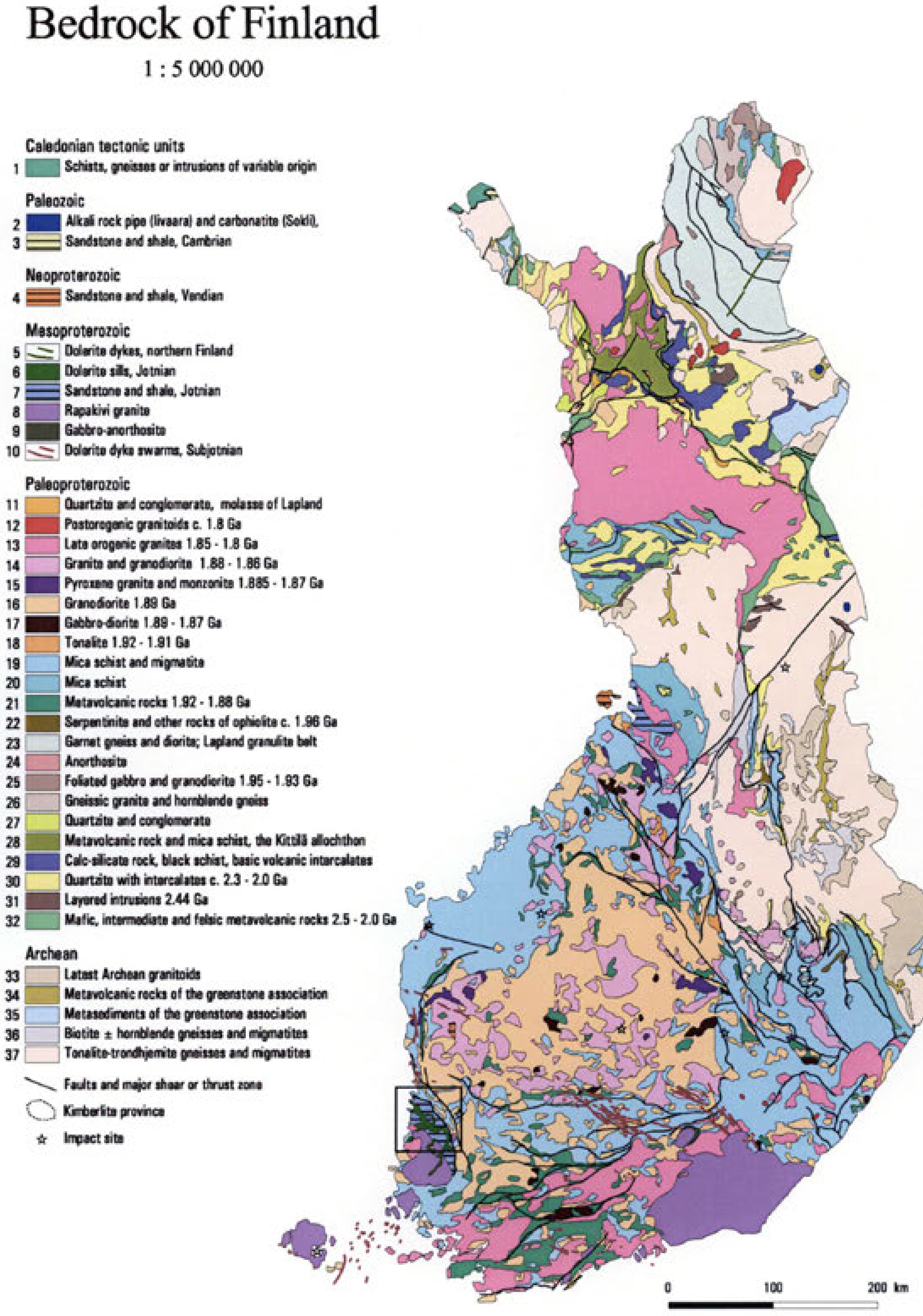


Figure 2.1-1. *Bedrock of Finland 1:5000 000. Simplified from the Bedrock Map of Finland 1:1000 000 (Korsman et al. 1997). Location of the southern Satakunta area is marked with a square.*

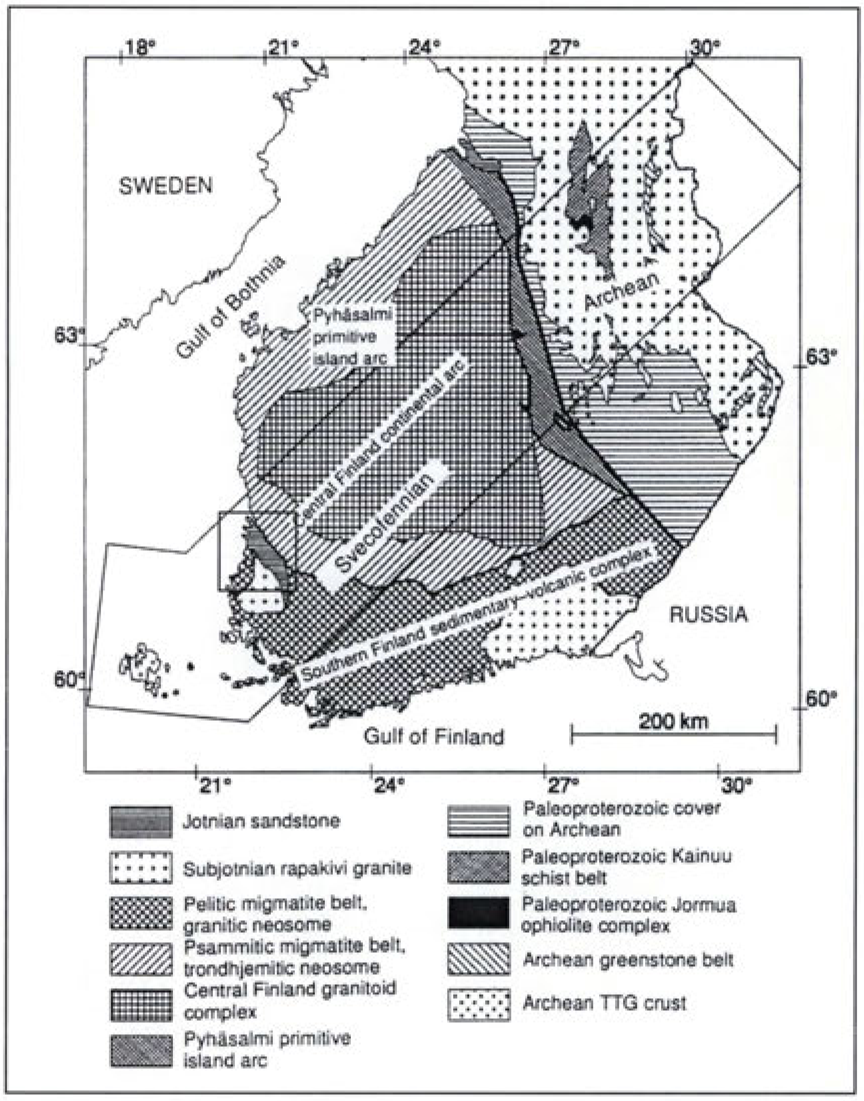


Figure 2.1-2. *Main geotectonic units of the Svecofennides (Korsman et al. 1999). The location of the southern Satakunta area is marked with a square, and the GGT/SVEKA transect is outlined by a polygon. TTG crust = tonalite-trondhjemite-granodiorite crust.*

2.1.2 Main geological features of southern Satakunta

The bedrock map of the southern Satakunta area (Fig. 2.1-3) has been compiled and simplified from four sheets of the geological map of Finland in the scale of 1: 100 000: 1132 Rauma (Suominen & Torssonen 1993), 1134 Kokemäki (Hämäläinen 1994), 1141 Luvia (Pihlaja & Kujala 1994) and 1143 Pori (Pihlaja 1994). Appendix 1 shows the geological map with a topographic relief. The continuity of the rock types at depth has been interpreted in three cross-sections (App. 2 - 4).

The oldest rocks in southern Satakunta area are supracrustal rocks, which consist mainly of migmatite mica gneisses, metasedimentary in origin, being deformed and metamorphosed during the Svecofennian orogeny 1900 – 1800 Ma ago, and containing cordierite, sillimanite and garnet porphyroblasts (Suominen et al. 1997, Veräjämäki 1998). In the northeastern part of the area they belong to the psammitic migmatite belt, whereas the southwestern part of the area belongs to the pelitic migmatite belt (Fig. 2.1-2 and Fig. 2.1-3). Amphibolites, uralite porphyrites and hornblende gneisses, which are derived from mafic and intermediate volcanics, occur as rare narrow zones.

The mica gneisses are intruded by 1890 - 1880 Ma old plutonic rocks, which consist of tonalites and granodiorites, occurring conformably with the structures of the mica gneisses (Pietikäinen 1994, Suominen et al. 1997, Veräjämäki 1998). Coarse-grained granites and pegmatites occur as large bodies and are present in older rocks, migmatizing them or occurring as cross-cutting veins.

The northern part of the anorogenic Laitila rapakivi batholith, about 1580 Ma in age, is located in the area (Vorma 1976, Vaasjoki 1996). The Eurajoki rapakivi stock is a satellite massif to the Laitila batholith, and can be divided into two types, hornblende-bearing Tarkki granite and younger, light-coloured Väkkärä granite (Haapala 1977). Both are somewhat younger than the Laitila batholith.

The Jotnian Satakunta Sandstone, at least 1400 – 1300 Ma in age, is a fluvial sediment formation deposited in a deltaic environment and has been preserved in a NW-SE trending graben structure (Kohonen et al. 1993). The sedimentary material of the sandstone derives from the Svecofennian supracrustal and plutonic rocks. Based on one borehole and geophysical interpretation, the thickness of the sandstone is at least 600 m, and probably as much as 1800 m thick (Elo 1976). The sandstone is cut by Postjotnian olivine diabase dykes, 1270 – 1250 Ma in age (Suominen 1991). Aeromagnetic map shows that these, in turn, are cut by north-south trending magnetic anomalies (*see* App. 5 and 6), which have been interpreted as being a swarm of younger diabase dykes (Veräjämäki 1998). Lake Sääksjärvi in the northeastern part of the map area is an impact crater of early Cambrian age.

One of the main structural features of the region is the NW-SE trending Kynsikangas shear zone (Fig. 2.1-3, App. 1), which is about 70 km long and hundreds of metres wide (Pietikäinen 1994, Veräjämäki 1998, Pajunen et al. 2001).

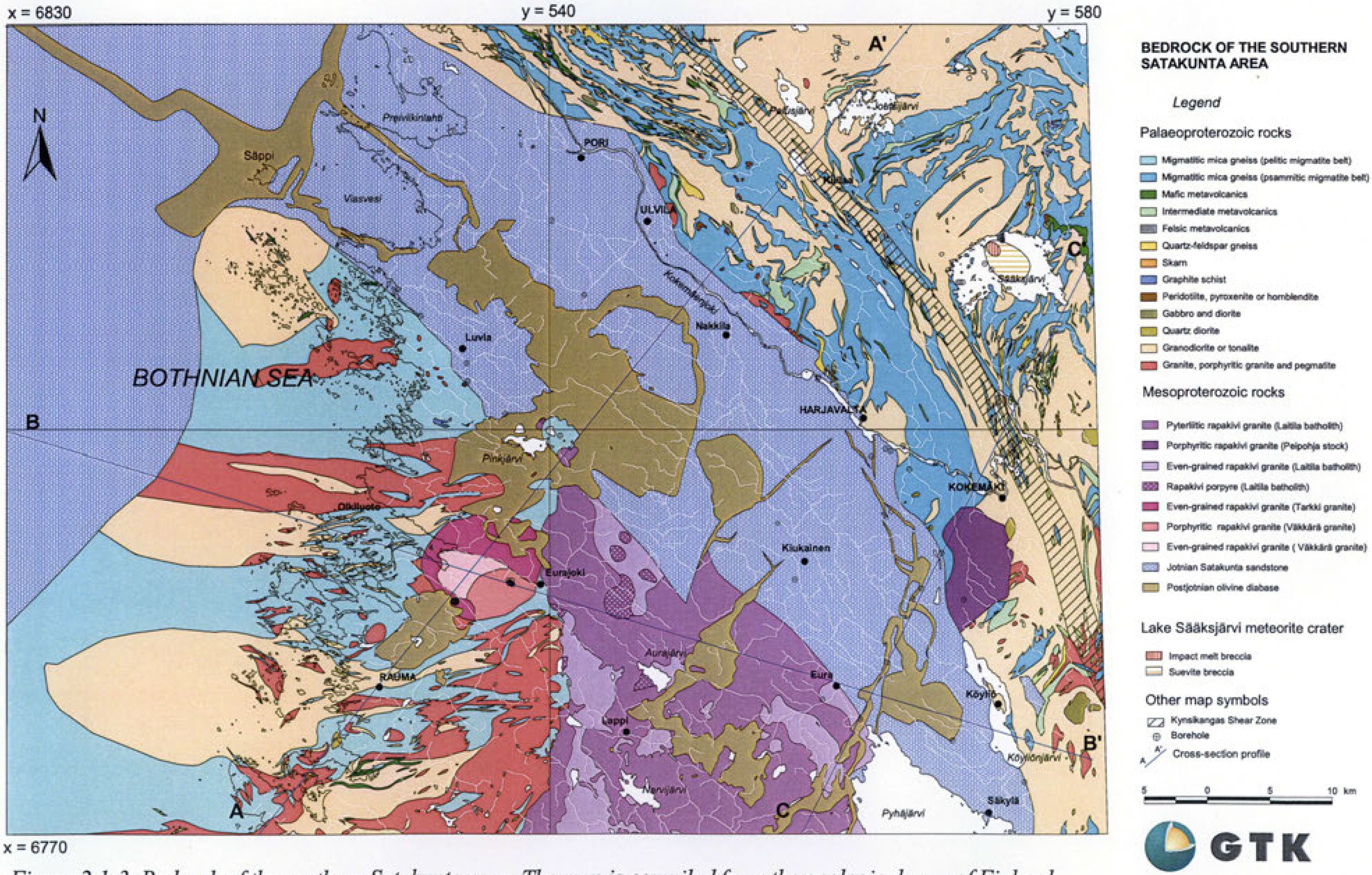


Figure 2.1-3. Bedrock of the southern Satakunta area. The map is compiled from the geological map of Finland 1: 100 000, map sheets 1132 Rauma (Suominen & Torssonen 1993), 1134 Kokemäki (Hämäläinen 1994), 1141 Luvia (Pihlaja & Kujala 1994) and 1143 Pori (Pihlaja 1994).

Base map (c) National Land Survey, permission no. 41/MYY/01

2.2 Rock types of the southern Satakunta area

2.2.1 Pelitic migmatite belt

Supracrustal rocks

The mica gneisses of the pelitic migmatite belt (Fig. 2.1-3, App. 1) are metasedimentary, which is demonstrated by the quartz-feldspar-rich and biotite-rich layers within the mica gneisses. The former represent psammitic layers, whereas the latter were originally pelitic layers. All the mica gneisses are variably migmatized (Fig. 2.2-1). The older component of the migmatite, or palaeosome, is mainly composed of quartz, plagioclase, biotite and often also potassium feldspar. It contains macroscopic garnet and cordierite porphyroblasts, as well as sillimanite porphyroblasts found only as microscopic inclusions in the biotite and cordierite grains.

The detailed petrographic and geochemical investigations of the borehole samples of the Olkiluoto site in the 1132 Rauma map-sheet area have shown three main palaeosome types: mica gneisses, quartzitic gneisses and amphibolitic palaeosomes (Gehör et al. 1996). The mica gneisses and quartzitic gneisses have been derived from the same sedimentation sequence, the former representing the metamorphism of an originally pelitic material and the latter arenitic material. Between these end members, there is a continuous series, which can be seen in both the mineral and the chemical composition of the palaeosome. In moving from the pelitic to more quartzitic compositions the percentage of quartz increases and that of biotite decreases.

Mica gneiss, which is the most common palaeosome both in the core samples and the outcrops, typically contains 20 – 40% biotite, 20 – 45% quartz, 10 – 30% plagioclase and less than 20% potassium feldspar. The quartzitic palaeosomes are represented by quartz-feldspar gneisses and quartzite gneisses, which contain more than 40% quartz, 20 – 50% plagioclase, 15% biotite and less than 10% potassium feldspar. In outcrops they occur as interlayers in the mica gneisses. The quartzitic palaeosomes in core samples also contain amphibole-bearing types (skarn quartzites), which are most likely derived from originally calcareous sediments.

The younger component of the migmatite, or neosome, is granitic in composition, and occurs as coarse-grained and pegmatitic veins parallel to the foliation and banding. The principal minerals of the neosome are quartz, potassium feldspar and plagioclase. Occasionally it contains the same porphyroblasts as the paleosome.

In the southwestern part of the 1141 Luvia map-sheet area the mica gneisses contain less neosome than the mica gneisses southwest of the Satakunta sandstone in general. Also mica schists occur among the mica gneisses. Andalusite or its restite occurs as a porphyroblastic mineral.

Mafic metavolcanics, hornblende gneisses and amphibolites, are present at the southwestern part of the 1132 Rauma map-sheet area (Fig. 2.1-3, App. 1). The hornblende gneisses usually occur as 1 to 15 cm wide bands in the granodiorites and the quartz diorites but also wider zones, up to 300 m wide, are met with. They grade into amphibole-bearing quartz diorites without clear contacts, and the contact against the mica gneiss can be gradual within a few metres (Suominen et al. 1997). Amphibolites occur as inclusions or as narrow, discontinuous zones, which are up to 150 m long and 10 m wide. Mica gneisses contain occasional mafic streaks and fine- or coarse-grained breccias and inclusions. The uralite porphyritic and amphibolitic streaks occur parallel to the foliation and are presumably intercalations of volcanic origin (Suominen et al. 1997).

Plutonic rocks

Most of the plutonic rocks among the pelitic migmatites southwest of the Satakunta sandstone are granodiorites, tonalites, granites and pegmatites. Mafic rocks, gabbros and diorites, usually occur only as small xenoliths.

The largest intrusions of the area are composed of trondhjemites, tonalites and granodiorites (Suominen et al. 1997). The contacts between the different rock types are gradual. Consequently, they have been combined in the geological map, presented in Figure 2.1-3 and Appendix 1, where they are labelled as "granodiorite or tonalite". The most felsic of the above rock types, trondhjemites, are medium-grained and massive or weakly foliated. They are cut and brecciated by pegmatite and in places by aplite. In the zone seaward the town of Rauma the plutonic rocks are dark-coloured tonalites or quartz diorites. The tonalites and granodiorites are fine- to medium-grained and practically even-grained (Fig. 2.2-2), although the grain size of potassium feldspar and plagioclase is usually greater then the other minerals. Biotite, and often hornblende, occur as mafic minerals.

This group of plutonic rocks includes also gneissose granodiorites, which are fine- or medium-grained, almost always hornblende-bearing, and include a few centimetres wide light- and dark-coloured streaks (Suominen et al. 1997). They sometimes resemble coarse-grained mica gneisses but are lacking in the porphyroblasts typical of mica gneisses. Consequently, they have been interpreted as plutonic rocks. However, according to the geochemical investigation of the core samples, the gneissose tonalites, at least on the island of Olkiluoto, most likely originated from partial melting and recrystallisation of the mica gneisses (Gehör et al. 1996).

Granites, porphyritic granites and pegmatites (Fig. 2.1-3, App. 1) form a heterogeneous group of rocks, containing numerous inclusions and granitised restites of the mica gneiss. They contain garnet and sometimes also cordierite from the assimilated mica gneisses. The chemical composition of the granites resembles the so-called S-type granites, which derive from partial melting of the sedimentary rocks (Suominen et al. 1997).

Figure 2.2-1. *Intensely migmatized mica gneiss (veined gneiss). Map-sheet 1132 08B; x = 6789 250, y = 1524 980. Photo by Pasi Virtanen/GTK.*

Figure 2.2-2. *Granodiorite with mafic inclusions. Map-sheet 1132 04D; x = 6778 200, y = 1516 360. Photo by Pia Fagerström/GTK.*

2.2.2 Psammitic migmatite belt

Supracrustal rocks

In the 1143 Pori and northeastern part of the 1141 Luvia map-sheet areas (Fig. 2.1-3, App. 1), the supracrustal rocks are mainly psammitic, in places turbiditic, variably migmatized mica gneisses, where the neosome is predominantly trondhjemitic (Pihlaja 2000) (Fig. 2.2-3). The migmatites in places contain narrow calcareous and amphibolitic interlayers (Mancini et al. 1996 a). The mineral assemblage of the Al-rich paleosome of the migmatitic mica gneisses is sillimanite-garnet-cordierite-quartz±potassium feldspar. The mica gneisses occasionally contain arkose gneiss and quartz-feldspar gneiss intercalations and in places narrow graphite and mica schist sections and graphite-bearing interbeds.

Mafic and intermediate metavolcanics occur in association with the mica gneisses, for example, northeast of Nakkila and east of Lake Sääksjärvi (Fig. 2.1-3, App. 1). Occasional pyroclastic primary structures suggest that at least some of them are volcanic ash or ejecta in origin (Pihlaja 2000). Around Lake Sääksjärvi the metavolcanic rocks (amphibolites) have the mineral assemblage hornblende-diopside-plagioclase-quartz±sphene (Mancini et al. 1996a).

In the 1134 Kokemäki map-sheet area, the main minerals of the mica gneisses are plagioclase, quartz and biotite, in places also potassium feldspar and garnet (Veräjämäki 1998). The mica gneisses are variably migmatized veined gneisses, the neosome being granite or tonalite in composition. The amount of the granitic neosome increases towards the southern part of the map-sheet area. No primary structures have been found in the mica gneiss paleosomes, although the banding of the mica gneiss most likely represent the original layering (Veräjämäki 1998). Also the quartz-feldspar gneiss interlayers indicate variations in the composition of the original sediment. Tuffitic interlayers occur widely, and occasionally narrow sulphide-bearing interlayers. Calcareous concretions are in places common in the mica gneisses but actual calcareous beds are not present. In places, the mica gneisses contain ultramafic and mafic xenoliths.

In the Kynsikangas shear zone, orto- and ultramylonites, and augen-gneisses occur (Pietikäinen 1994) (Fig. 2.2-4). The mylonitic foliation is subparallel to the foliation of the veined gneisses, and paleosomes of the migmatites occur as megacrysts. The contact of the shear zone to northeast is clear, but indistinct to southwest.

Metavolcanic rocks in the Kokemäki map-sheet area include metatuffs, lapilli tuffs, meta-agglomerates, hornblende gneisses, amphibolites, and possibly quartz-feldspar gneisses (Veräjämäki 1998). In the geological map (Fig. 2.1-3, App. 1) all but quartz-feldspar gneisses are combined into basic and intermediate volcanics. The metatuffs, lapilli tuffs and meta-agglomerates are originally volcanic ash, bombs and ejecta, and they occur as less than 500 m wide and 10 km long sections. The metatuffs and lapilli tuffs are fine-grained, the lapillituffs containing plagioclase and hornblende fragments, 2 - 3 mm in size. They are usually andesites in their chemical composition.

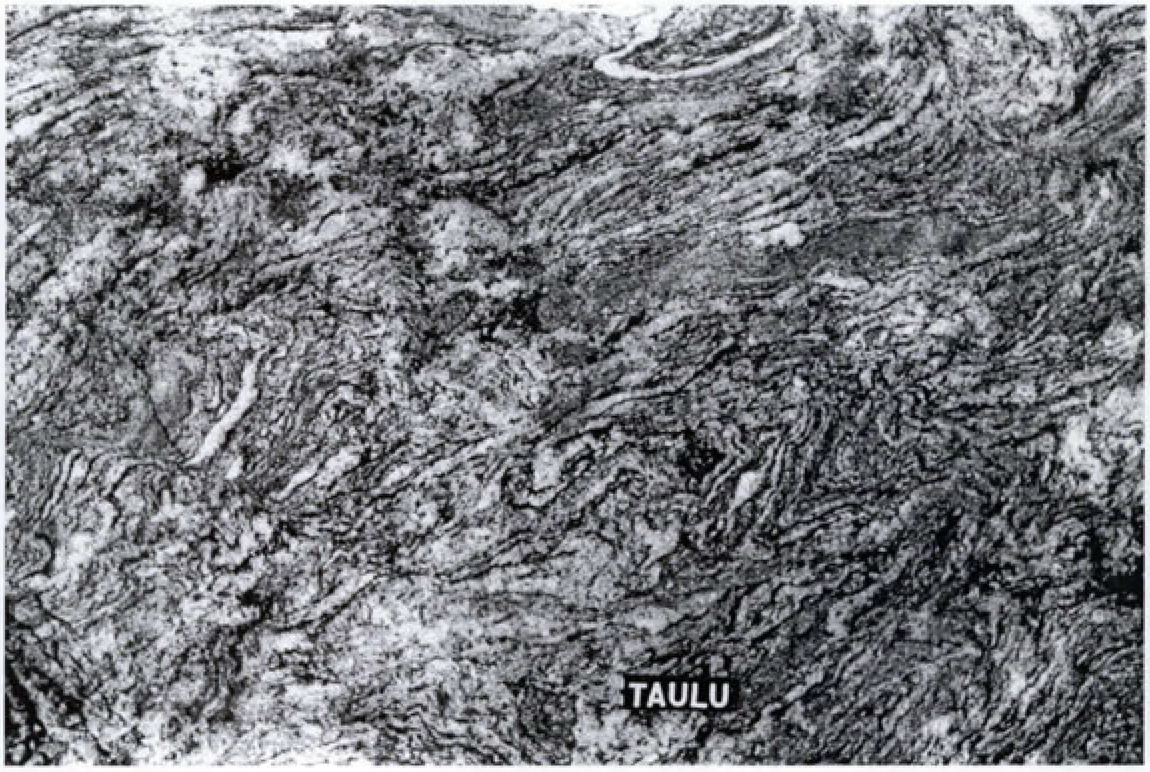


Figure 2.2-3. *Migmatitic mica gneiss (veined gneiss) Hyvelä, Pori. Map-sheet 1143 03; x = 6827 830, y = 1543 150. Photo by Hannu Kujala/GTK.*

Figure 4.1-4. *Cataclastic, garnet-bearing mica gneiss from the Kynsikangas shear zone. Häyhtiönmaa, Kokemäki. Map-sheet 1143 08; x = 6811 950, y = 1566 770. Photo by Hannu Kujala/GTK.*

Layers of meta-agglomerates less than three metres thick and containing volcanic bombs and ejecta occur in association with the metatuffs and lapillituffs. The bombs and ejecta are strongly deformed and usually intermediate in composition. The matrix between the them is composed of fine-grained tuff.

Hornblende gneisses and amphibolites occur as narrow sequences and interlayers in other supracrustal rocks, as well as inclusions in the plutonic rocks. Even though the strong deformation has destroyed the original structures, they are presumably volcanic rocks, with some sedimentary component, in origin (Veräjämäki 1998). The main minerals of the amphibolites and hornblende gneisses are plagioclase and hornblende, the hornblende gneisses additionally containing quartz and biotite.

Quartz-feldspar gneisses, east of lake Köyliönjärvi, are fine-grained, in places banded rocks, quartz, plagioclase, biotite and potassium feldspar being the main minerals. They contain interlayers with hornblende phenocrysts, suggesting a volcanic origin.

Plutonic rocks

The plutonic rocks of the Kokemäki map-sheet area are mainly granodiorites, tonalites and granites with some small associated peridotite, gabbro, diorite and quartz diorite intrusions (Veräjämäki 1998) (Fig. 2.1-3, App. 1). Gabbros and diorites usually occur as small intrusions parallel to structures of the bedrock, and as numerous fragments in other rocks. They are medium-grained, hornblende, plagioclase and biotite being the main minerals, in places also garnet, apatite, augite and cummingtonite.

The granodiorites and tonalites form both large and small intrusions, which have a large number inclusions of the country rocks. They also form migmatites with older supracrustal rocks. The granodiorites and tonalites are medium-grained, foliated, often porphyritic, and in places sheared and mylonitic. The main minerals are plagioclase, quartz, potassium feldspar, biotite and hornblende. Potassium feldspar is usually secondary and its amount seems to be dependent on the vicinity of the granite (Veräjämäki 1998).

Granites and coarse-grained pegmatitic granites occur as large gneissose bodies containing abundant inclusions of the surrounding rocks. They also form veins and wider dykes, which intrude the older supracrustal and plutonic rocks to form migmatites. The granites of the Kokemäki map-sheet area are typically migmatizing the rocks, which have earlier been migmatized by tonalites and granodiorites (Veräjämäki 1998).

In the Pori map sheet area and the northern part of the Luvia map-sheet area, northeast of the Satakunta sandstone formation, the plutonic rocks are predominantly tonalites and quartz diorites (Pietikäinen 1994, Pihlaja 2000). In the less deformed and altered tonalites the main mafic mineral is hornblende. However, when the deformation and alteration increases, biotite becomes more abundant and the mineral composition approaches that of the trondhjemites. Mafic plutonic rocks, amphibolites and mica gneisses occur as inclusions in the tonalites. Gneissose tonalites (Fig. 2.2-5), and especially the migmatized tonalites are heterogeneous, containing restites of the assimilated mica gneiss.

Granites are scarce in the Pori and Luvia map-sheet areas. They are primarily microcline-bearing and pegmatitic granites occurring as small intrusions in the migmatites and also in the tonalites (Pihlaja 2000). The largest pegmatite granite intrusions are located in the northeastern margin of the Satakunta sandstone and in the archipelago of Luvia (Fig. 2.1-3, App. 1). Pegmatite occurs also as small patches, veins and dykes in the mica gneisses, tonalites and quartz diorites.

Several small, often differentiated, mafic and ultramafic intrusions occur north and east of Pori (Fig. 2.1-3, App. 1). They vary from peridotites, pyroxenites and hornblendites to gabbros and diorites in their mineral composition (Mäkinen 1987, Mancini et al. 1996a, 1996b). The Hyvelä intrusion, north of Pori, is concentric in composition, the more mafic rocks (norites) being in the centre of the intrusion (Mäki 1982). The intrusion hosts a small, uneconomic Ni-Cu mineralisation. The rocks of the Rottapäkki intrusion, east of Hyvelä, are lherzolites composed of olivine, orthopyroxene and clinopyroxene (Mäkinen 1987). The unexposed Sääksjärvi intrusion, northeast of Harjavalta, is mainly metaperidotitic and composed of olivine, amphibole and orthopyroxene (Mancini et al. 1996a, 1996b).

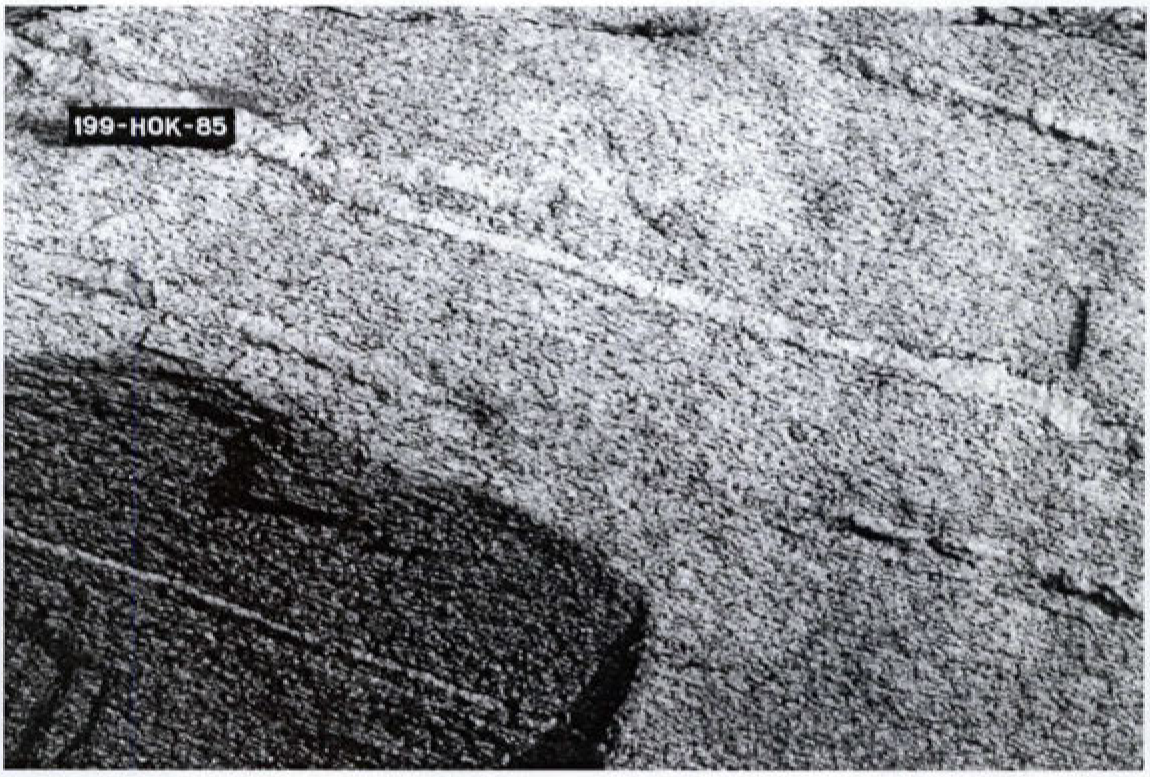


Figure 2.2-5. *Gneissose tonalite. Lutajärvi, Lavia. Map-sheet 1143 12; x = 6827 340, y = 1573 650. Photo by Hannu Kujala/GTK.*

2.2.3 Rapakivi granites

The rapakivi granites of the study area belong to the Laitila rapakivi batholith and the Eurajoki stock, emplaced ca. 1580 Ma ago (Vaasjoki 1977). According to gravimetric survey the average thickness of the Laitila batholith is 5 km, the maximum thickness being about 20 km (Laurén 1970). The present erosion level is apparently close to the roof of the batholith (Vorma 1989). The contacts of the rapakivi granites against the country rocks are sharp and cut the structures of the older bedrock (Veräjämäki 1998). In the contact zone, the rapakivi granites contain fragments of country rocks.

The Laitila rapakivi batholith is composed of several different types, which differ from each other in texture and mineral composition (Vorma 1976). The Eurajoki rapakivi stock, which is somewhat younger than the Laitila batholith, is a satellite massif to the Laitila batholith, and can be divided into two types, hornblende-bearing Tarkki granite and even-grained or porphyritic Väkkärä granite (Haapala 1977). On the basis of geophysical investigations the Kokemäki or Peipohja rapakivi stock, located northeast of the Satakunta sandstone, is connected with the Laitila batholith under the sandstone

(Laurén 1970, Elo 1982). The Söörmarkku felsic dyke in the northeastern part of the Luvia map-sheet area is coeval with the rapakivi granites (Pihlaja 2000).

In the Kokemäki map-sheet area, the Laitila rapakivi batholith can be divided into four types, which most likely represent different intrusion phases: pyterlitic rapakivi granite, porphyritic rapakivi granite, even-grained rapakivi granite and rapakivi granite porphyre (Veräjämäki 1998).

The pyterlitic rapakivi granite (Fig. 2.1-3, App. 1) is the most common rapakivi variant within the Laitila batholith (Vorma 1976, Veräjämäki 1998). It is characterised by potassium feldspar ovoids, 2 - 4 cm in diameter (Fig. 2.2-6). The potassium feldspar phenocrysts are usually lacking in the plagioclase mantle, typical of the wiborgitic rapakivi granite of the Wiborg batholith in southeastern Finland. However, potassium feldspar ovoids with plagioclase mantle are present in some of the darker pyterlite varieties, in which the pyterlite in places grades into wiborgite (Veräjämäki 1998). The main minerals of the pyterlite are potassium feldspar, quartz, plagioclase, biotite, and in the darker pyterlite types also hornblende. Accessory minerals include fluorite, apatite, zircon and anatase. The contacts of the pyterlite with the other rapakivi types are either gradual or sharp.

The Peipohja rapakivi stock (Fig 2.1-3, App. 1) consists of porphyritic rapakivi granite. It contains euhedral or subhedral, 2 - 6 cm long, potassium feldspar phenocrysts in a medium-grained matrix, composing of potassium feldspar, quartz, biotite, plagioclase and hornblende (Veräjämäki 1998). Ovoids, which are only weakly developed, are lacking in the plagioclase mantles.

The even-grained rapakivi granite is composed of medium- to coarse-grained granite with small potassium feldspar ovoids and small-grained granite, with or without the phenocrysts (Fig. 2.2-7). Usually it grades into pyterlitic rapakivi granite but also sharp contacts are known. The even-grained rapakivi usually occurs at the margins of the Laitila batholith (see Fig. 2.1-3, App. 1).

The rapakivi porphyre (Fig. 2.1-3, App. 1) has a fine-grained matrix with 1 to 4 cm long euhedral or ovoidal potassium feldspar phenocrysts. The ovoids occasionally have the plagioclase mantle. The contact between the rapakivi porphyries and the even-grained rapakivi granites are either gradual or sharp. According to Vorma (1976), the rapakivi porphyries were originally located close to the roof of the batholith.

Gabbro-anorthosites associated with the rapakivi granites do not occur within the study area. The nearest such rocks, belonging to the Kolinummi anorthosite/leucogabbro body, are located 0.5 km southeast of the Laitila batholith.

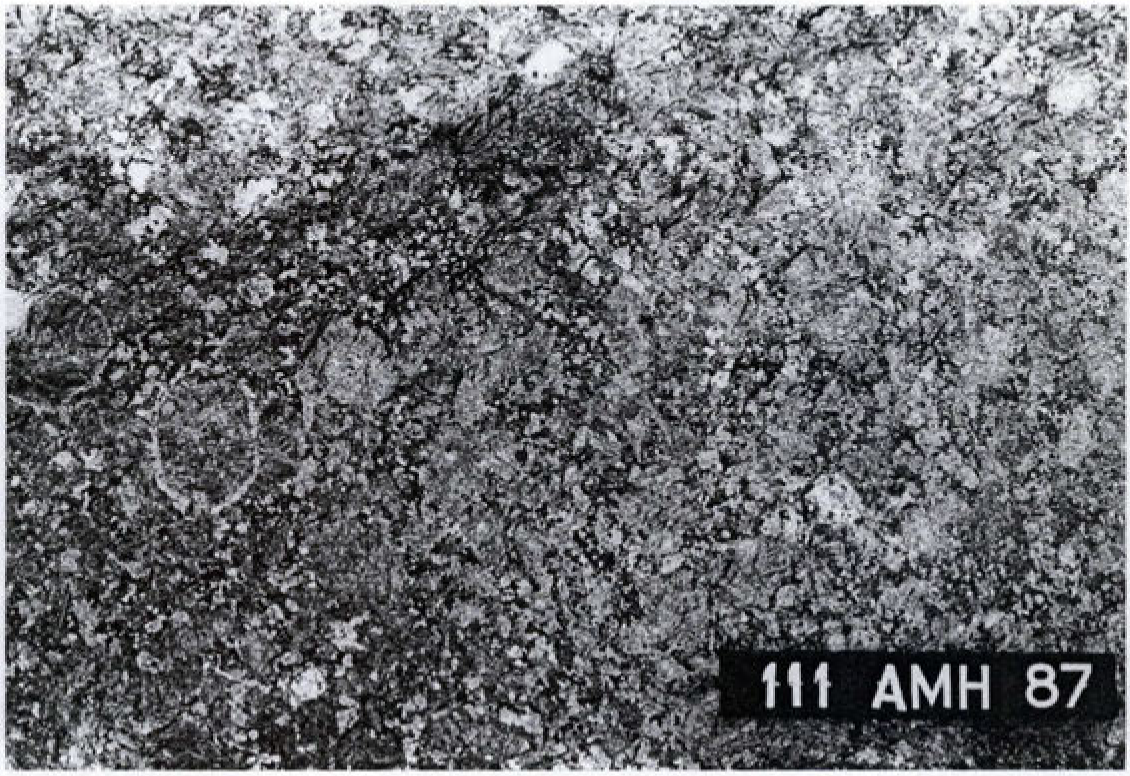


Figure2.2-6. *Pyterlitic rapakivi granite. Ylinen-Juva, Eurajoki. Map-sheet 1134 02 A; x = 6784 780, y = 1544 660. Photo by Arja Hämäläinen/GTK.*

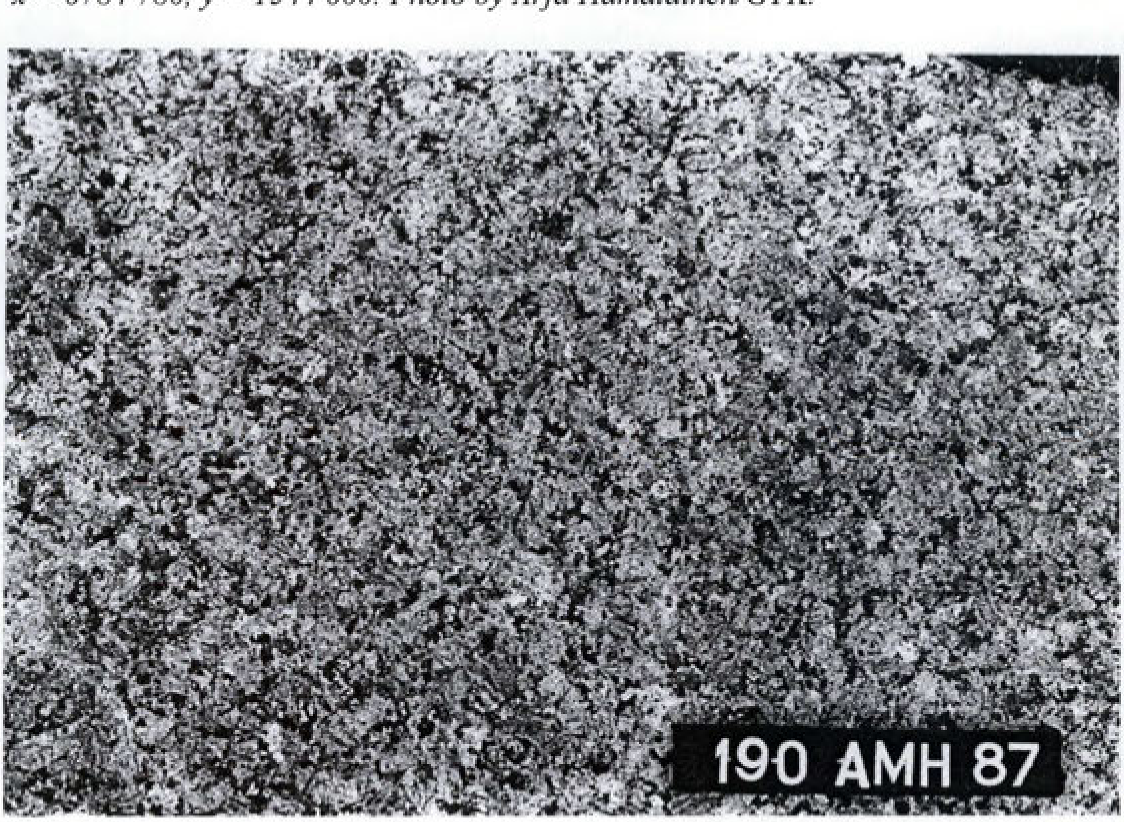


Figure 2.2-7. *Even-grained rapakivi granite. Lutta, Eurajoki. Map-sheet 1134 02 A; x = 6784 700, y = 1541 580. Photo by Arja Hämäläinen/GTK.*

The Eurajoki rapakivi stock is composed of two main rapakivi granite types: older hornblende-biotite granite (Tarkki granite) and younger, topaz-bearing microcline-albite granite (Väkkärä granite) (Haapala 1977, Suominen et al. 1997). The contact between the marginal Tarkki granite and the surrounding Svecofennian migmatites is sharp and almost vertical. The contact between the Tarkki granite and the central Väkkärä granite is also sharp, the Väkkärä granite dipping gently outwards (Haapala 1977, Elo 2001) (*see* Chapter 4.3).

The Tarkki granite (Fig. 2.2-8) is homogenous, medium- and even-grained rock, the main minerals being potassium feldspar, quartz, plagioclase, biotite and hornblende (Haapala 1977). It contains potassium feldspar ovoids, 3 - 6 cm in diameter, mantled with plagioclase. The ovoids are sparsely distributed, the distance between individual ovoids being often several metres. The granite is cut by quartz porphyre dykes, some of which are topaz-bearing. Both the mineral and the chemical composition of the dykes resemble those of the Väkkärä granite, indicating a close genetic connection between the two (Haapala 1977).

The Väkkärä granite (Fig. 2.2-9) is composed of several types, which differ from each other in texture. The contacts between the different types are in places sharp, in other places gradual (Haapala 1977). The porphyritic and coarse-grained granites are topaz-bearing, the phenocrysts being potassium feldspar and quartz. Accessory minerals include, in addition to topaz, fluorine-rich siderophyllite, monazite, ilmenite, Nb- and Ta-rich cassiterite, columbite and bastnaesite. Miarolitic cavities are common in the porphyritic granites, indicating the presence of a separate fluid phase during late stages of crystallisation of topaz-bearing granite, and greisen-type Sn-Be-W-Zn mineralisation is closely associated with it (Haapala 1997). The topaz-bearing Väkkärä granite resembles typical tin granites in chemical composition. It is enriched in fluorine, lithium, gallium, rubidium, tin and niobium, and depleted in strontium, barium and zircon. The even-grained Väkkärä granite is mainly composed of potassium feldspar, quartz and biotite, accessory minerals being zircon, ilmenite, anatase, monatzite, as well as topaz and cassiterite as secondary minerals. The contact type of Väkkärä granite against the Tarkki granite has potassium feldspar, plagioclase and quartz phenocrysts, 1 - 10 mm in diameter, in a fine-grained matrix. The contact type contains fragments of Tarkki granite and sends apophyses into it (Haapala 1977).

E-W-trending greisen veins and associated quartz veins occur in the Tarkki granite. The width of the veins varies from a few centimetres up to two metres, and the drillings have shown that at least some of the veins and vein swarms extend over 100 m in depth and over 300 m in a horizontal direction (Haapala 1977). In the Väkkärä granite the veins are more randomly oriented, and also lensoid, rounded and irregular greisen bodies occur. The main minerals of the greisen veins are quartz, micas and iron-rich chlorite. Often they also contain abundant topaz, fluorine, garnet, beryl, genthelvite an bertnandite (Haapala 1977). The most common ore minerals include sphalerite, cassiterite, chalcopyrite, wolframite, gahnite, molybdenite, rutile, secondary iron oxide, pyrite, pyrrhotite, arsenopyrite and galena. The greisen veins were caused by hot, hydrous fluids, migrating in interstices and fractures of the rapakivi granites (Haapala 1977). The greisen veins both in the Tarkki and Väkkärä granite were formed by fluids emanated from the Väkkärä granite.

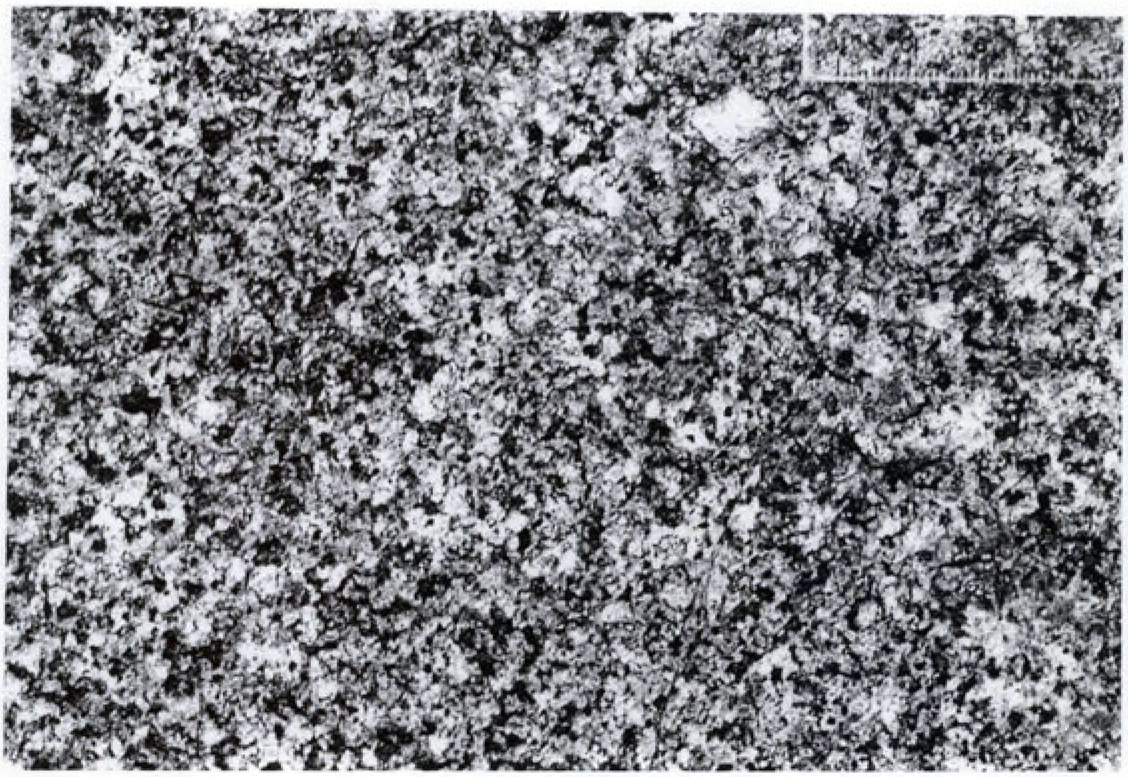


Figure 2.2-8. *Tarkki rapakivi granite. Iisakki-Tarkki, Eurajoki. Map-sheet 1132 11B; x = 6786 860, y = 1533 500. Photo by Pia Fagerstöm/GTK.*

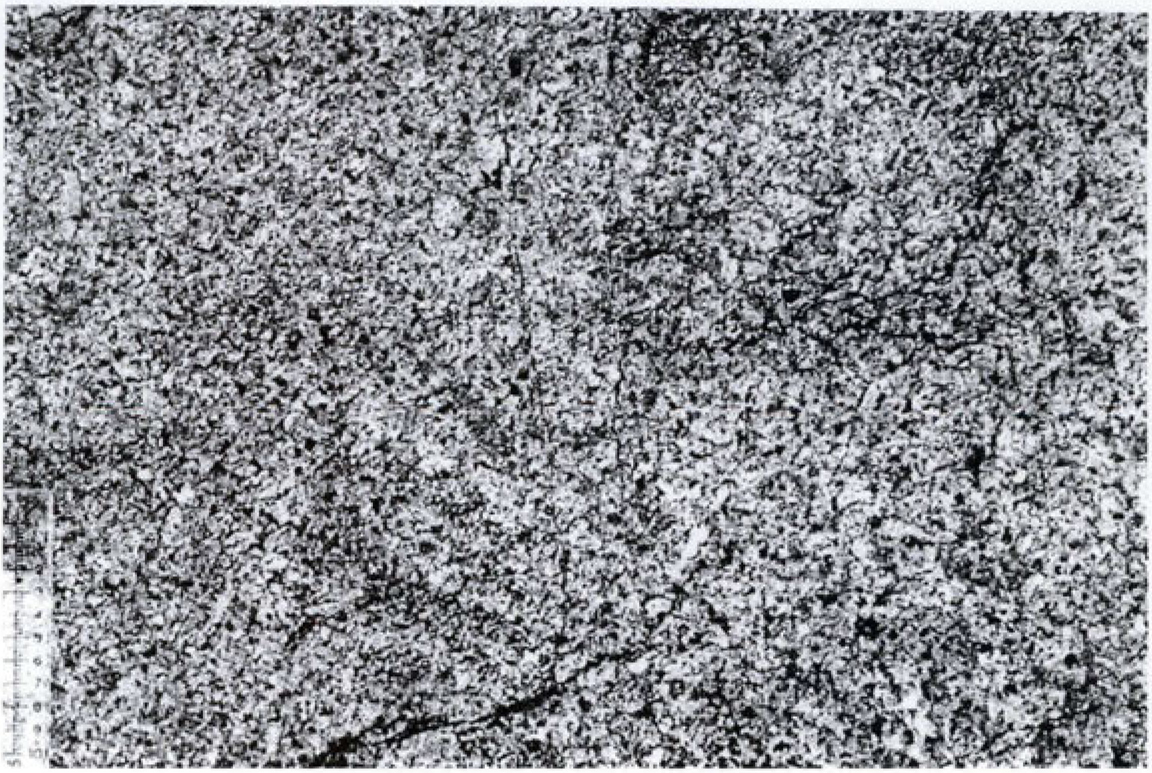


Figure 2.2-9. *Väkkärä rapakivi granite. Racetrack, Eurajoki. Map-sheet 1132 11D; x = 6787 100, y = 1535 500. Photo by Pia Fagerström/GTK.*

2.2.4 Satakunta sandstone

The Jotnian Satakunta sandstone (Fig. 2.1-3, App. 1) has been preserved in a NW-SE-trending graben structure formed by block movements. The layering of the sandstone is predominantly subhorizontal, but in the northeastern contact the sandstone layers are dipping about 35° to the southwest, towards the centre of the basin. No direct evidence of the thickness of the sandstone is available but, in a borehole located south of Pori, it is at least 600 m, and according to gravimetric studies, the maximum thickness could be about 1800 m (Elo 1982). In the southeastern part of the sandstone, between Lake Pyhäjärvi and Lake Köyliönjärvi, the thickness is probably less than 200 m (Elo et al. 1993). The sandstone is very poorly exposed. Only about 30 outcrops are known, and the only contact observations are from the contacts between the sandstone and the Postjotnian olivine diabases cutting it. The Satakunta sandstone is connected to the Bothnian Sea sedimentary basin, which is almost completely composed of Jotnian sandstone (Axberg 1980, Winterhalter et al. 1981). Offshore, northwest of the town of Pori, the sandstone is interpreted to be underlain by marine continuation of the Reposaari (Pori) rapakivi granite (Korja et al. 2001).

In the Satakunta sandstone the sedimentary materials range from massive gravels and pebbly sands over moderately sorted, medium to coarse grained sands to mudstones, and rare glauconite-rich layers. Among the slightly rounded clasts quartz dominates over microcline and plagioclase, rock fragments and micas being the most common accessories (Simonen & Kouvo 1955, Marttila 1969). The sandstone matrix is fine-grained and consists of quartz, clay minerals and chlorite. Basal conglomerates with rounded quartz pebbles have been observed in drill core samples along the southwestern contact of the sandstone in Luvia (Kohonen et al. 1993).

In the chemical composition of the sandstone, SiO_2 varies from 76 to 84% and Al_2O_3 from 6 to 12%, the K_2O content being typically over 3.5% (Marttila 1969). On the basis of detrital heavy minerals (e.g., garnet, ilmenite, zircon, monazite), the provenance of the sandstone was the Svecofennian rocks (Marttila 1969, Kohonen et al. 1993). According to paleocurrent analysis, the main transport direction was toward northwest or north, i.e., from the direction of the rapakivi batholith (Kohonen et al. 1993). However, during the sedimentation of the sandstone the rapakivi granite apparently was unexposed, because the sandstone is lacking in the minerals typical of the granite. This is further supported by the U-Pb study of detrital zircons (Vaasjoki & Sakko 1987), which excludes rapakivi as major source. Both the paleocurrent pattern and the grain size distribution support the hypothesis that the proximal part of the basin was located in the southeast (Kohonen et al. 1993).

The sedimentological features of the outcrops indicate a deposition in a continental fluviatile environment (Marttila 1969, Kohonen et al. 1993). These lithofacies are characterised by structureless fine-gravel conglomerates, trough-cross-bedded, planar-cross-bedded, laminated, and massive-structureless sandstones, as well as horizontally laminated mud- and siltstones (Fig. 2.2-10). The red colour of the sandstone is interpreted by Simonen and Kouvo (1955) to indicate a terrestrial, oxidising environment. However, according to Kohonen et al. (1993) the colouring may be partly epigenetic in origin.

The Satakunta sandstone hosts the world's oldest known micrometeorites (Deutsch et al. 1998). Despite their high terrestrial residence age (ca. 1400 Ma), the spherules are excellently preserved, show only very minor signs of mechanical abrasion, and match in their chemical properties unaltered stony spherules.

Figure 2.2-10. *Cross-bedded Jotnian sandstone. Lammainen, Harjavalta. Map-sheet 1143 04; x = 6803 950, y = 559 630. Photo by Arja Hämäläinen/GTK.*

2.2.5 Sub- and Postjotnian diabase dykes and sills

Diabase dykes probably of Subjotnian age (ca. 1650 Ma) occur at the Toukari area of Pori and at Harjakangas in Noormarkku (Pihlaja 1987). The two Toukari dykes are N-S-trending and are dipping steeply to the west. The main minerals of the dykes are plagioclase and uralitised augite. The diabase has amygdule-like spherules consisting of chlorite, quartz and carbonate. The Harjakangas dyke is 20 to 70 m wide, 2 km long and N-S-trending. The main minerals of the coarse-grained diabase are plagioclase, augite, olivine and hyperstene. On the basis of geochemical properties the Subjotnian diabases resemble tholeitic, continental intraplate basalts (Pihlaja 1987).

The Postjotnian Satakunta olivine diabases, dated at 1270 - 1250 Ma (Suominen 1991, Vaasjoki 1996b) cut all the previous rock types of the study area (Fig. 2.2-11). In the rapakivi area, olivine diabase occurs as long, narrow dykes, whereas within the sandstone are they form large horizontal sills (Hämäläinen 1985) (Fig. 2.1-3, App. 1). In the Kokemäki map-sheet area, the thickness of the sills is estimated to be 40 - 80 m (Kurimo et al. 1992 a), in the Rauma map-sheet area even 200 - 300 m (Suominen et al.

1997, Elo 2001). The diabases are tholeitic in chemical composition, and resemble continental flood basalts.

The main minerals of the olivine diabases are plagioclase, augite, olivine, opaques (ilmenite, magnetite) and rarely biotite (Hämäläinen 1987, Suominen et al. 1997, Veräjämäki 1998). Usually they also contain small amounts of quartz and potassium feldspar, which form micrographic intergrowths (Kahma 1951). The diabases are subophitic in texture (Fig. 2.2-12), indicating a hypabyssal origin. The diabase sill on the island of Säppi, offshore Pori (Fig. 2.1-3), is composed of several different varieties, which are all coarse-grained and have sharp contacts with each other (Inkinen 1963).

The contacts of the olivine diabases with the country rocks are usually sharp but occasionally the diabase magma has partially melted the country rock, making the contact diffuse (Veräjämäki 1998). The diabases in places have palingenic veins, which have resulted from the partial remelting of the country rocks (sandstone, rapakivi granites and migmatites). They are from a few millimetres up to 10 cm wide and proceed 2 to 3 m, at the most, into the diabase from the contact. The veins are mainly composed of quartz and alkali feldspars. In addition they contain some biotite, chlorite and especially near the diabase contact also hornblende.

In the subhorizontal (15°E) contact of the Sorkka diabase, located between Eurajoki and Rauma (see Fig. 2.1-3), against the surrounding migmatitic gneisses, the medium-grained diabase becomes fine-grained or aphanitic towards the contact (Kahma 1951). Also at the contact between the Sorkka diabase and the Tarkki rapakivi granite, the diabase is rather aphanitic, but becomes medium-grained a few metres away from the contact. Large numbers of apophyses proceed from the diabase dykes into the rapakivi granite. In Nakkila, the fine-grained apophyses from the diabase into the sandstone are 10 to 20 cm wide and on an average 30 m long, and they have caused the partial remelting of the sandstone (Laitakari et al. 1986).

The Kokemäki low-altitude aeromagnetic map (App. 5 and 6) suggests that the olivine diabase dykes are cut by younger diabase dykes trending almost north-south (Veräjämäki 1998). These dykes are not exposed at the study area but according to observations in the 1133 Yläne map-sheet area (Vorma & Niemelä 1994) they are 1 to 5 m wide and, contrary to the olivine dykes, often weakly porphyritic. The age of the dykes is still unknown but their different mode of occurrence suggests them to be considerably younger than the olivine diabases (Amantov et al. 1996).

Figure 2.2-11 *Contact between the sandstone (on the left) and the diabase. Nakkila, map-sheet 1143 04B; x = 6805 190, y = 1550 760. Photo by Paavo Vuorela/GTK.*

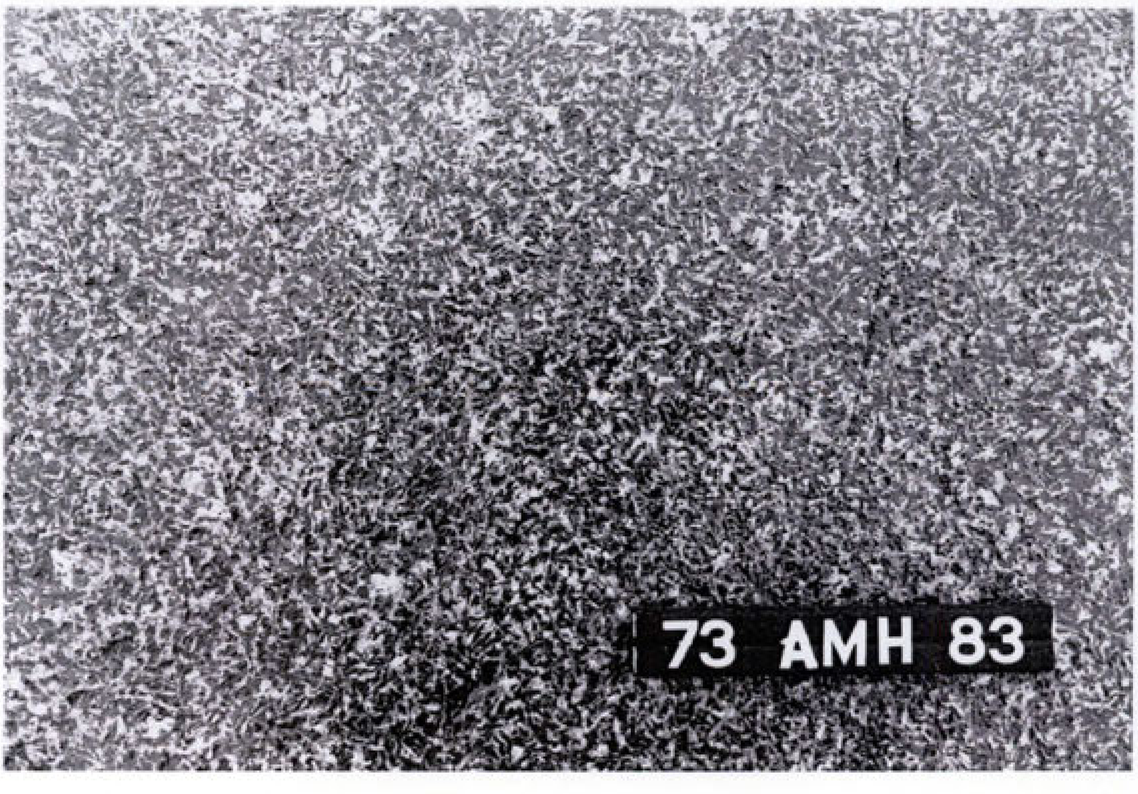


Figure 2.2-12. *Coarse-grained olivine diabase. Mäntyluoto, Pori. Map-sheet 1142 07; x = 6832 020, y = 524 710. Photo by Arja Hämäläinen/GTK.*

Narrow diabase dykes with an average width of 20 cm (Fig. 2.2-13) occur in the 1132 Rauma map-sheet area (Suominen et al. 1997). The diabases are composed of lamellar plagioclase in a very fine-grained matrix of biotite and other mafic minerals, which are hard to identify. They also contains quartz-, carbonate- and opaque-filled amygdules, which probably originated as gas cavities or vesicles, indicating that the dykes have intruded at shallow depth. The age of the dykes is unclear, since their chemical composition differs both from the Subjotnian and Postjotnian diabases (*cf.* Suominen et al. 1997, p. 33).

2.2.6 Lake Sääksjärvi meteorite crater

Lake Sääksjärvi, northeast of the Satakunta sandstone, hosts about 4.5 km wide impact crater of early Cambrian age (ca. 560 Ma). Below water level it contains unexposed suevite breccia and impact melt breccia (Papunen 1973, Müller et al. 1990, Elo et al. 1992, Pihlaja & Kujala 2000). The deep drilling in the centre of the gravity low penetrated 180 m of suevites and breccias before reaching the deformed Svecofennian mica gneiss (Pihlaja & Kujala 2000). No coherent impact melt layer was found in drilling.

The suevite breccia (Fig. 2.2-14) contains mineral and rock fragments of the Svecofennian rocks, as well as fluidal glass fragments and vesicles filled with clay minerals and zeolites (Pihlaja & Kujala 2000). The impact melt rocks show characteristic shock effects for quartz, feldspars and biotite, including planar deformation features, diapletic quartz glass, ballen-quartz, checkerboard plagioclase and kink bands in biotite (Papunen 1973, Müller et al. 1990, Elo et al. 1992, Pihlaja & Kujala 2000). Both impacts melts and suevites are strongly enriched in Ir, Pt, Pd, Ni and Cr. Of the major elements, Na and Ca are depleted, and Mg and especially K enriched in impact rocks compared with target rocks, as a consequence of replacement of structurally damaged shocked plagioclase with a secondary potassium feldspar and clay minerals (Pihlaja & Kujala 2000).

Figure 2.2-13. *About 10 cm wide diabase in migmatite. Pask-Aikko, Rauma. Map-sheet 1132 09A; x = 6790 200, y = 1522 200. Photo by Pia Fagerström/GTK.*

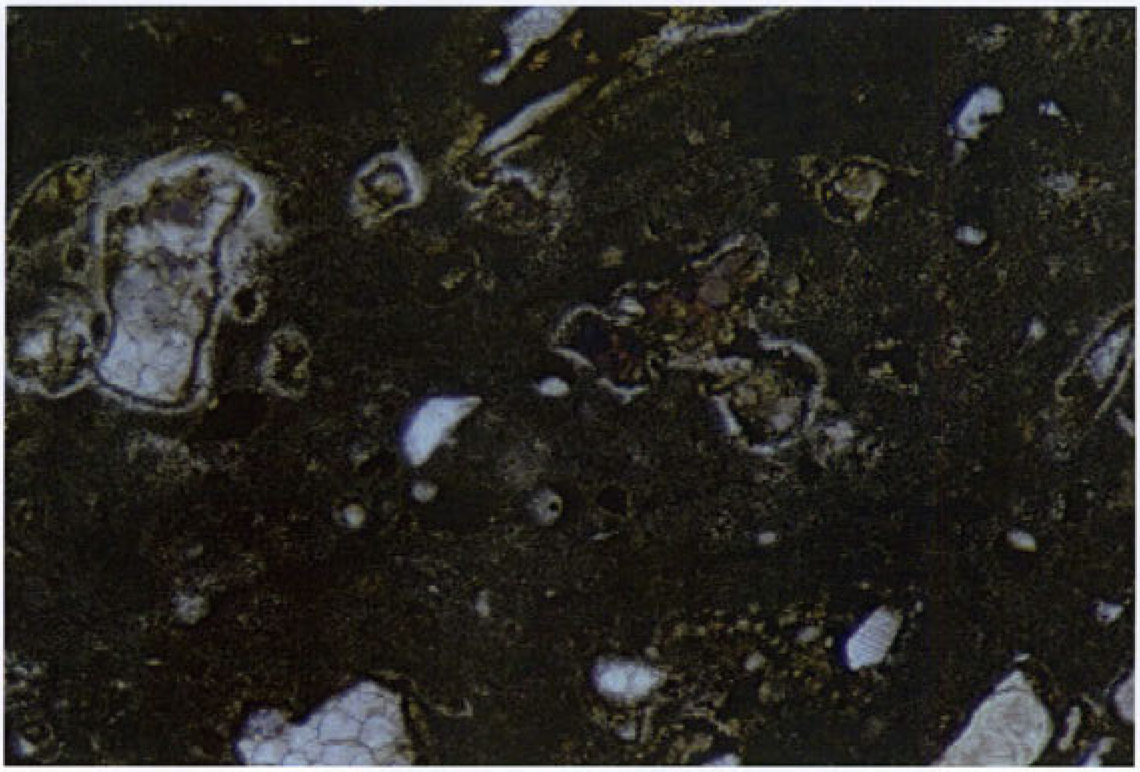


Figure 2.2-14. *Suevite. Sääksjärvi, Kokemäki. Borehole K/1143/91/R1, map-sheet 1143 11A; x = 6812 300, y = 1574 120. Microscope photo by Hannu Kujala/GTK.*

2.3 Ore potential of the bedrock in southern Satakunta

No ore mineralisations of economic importance have been found within the Rauma map-sheet area (Suominen et al. 1997). Eurajoki rapakivi granite stock hosts small, uneconomic greisen-type Sn-Be-W-Zn mineralisations (Haapala 1977a, Haapala 1977b). In the Tarkki granite of the Eurajoki stock the ore minerals, including cassiterite, sphalerite, chalcopyrite and galena, occur in the greisen and quartz veins. The distribution of tin in the greisen veins is uneven, and can locally be as much as 20 weight-% (Haapala 1989).

North and northeast of the Satakunta sandstone there are small mafic and ultramafic intrusions, which belong to the Vammala Ni-belt (see Papunen & Gorbunov 1985, Peltonen 1995). The estimated ore potential within the Hyvelä gabro intrusion, north of Pori, is about 800 000 t of ore, assaying 0.52% Ni and 0.26% Cu (Stenberg & Häkli 1985). According to the ore database of the Geological Survey of Finland, uneconomic Ni-Cu mineralisations are also present within the Harjunpää, Torisevanahde, Levanpelto, Routtuperko, Piilijoki, and Sääksjärvi intrusions (Fig. 2.3-1). The Ni-Cu content is also reflected in the basal till north and northeast of the Satakunta sandstone (Fig. 2.3-2).

According to the ore database, no ore mineralisations exist in the Kokemäki map-sheet area. Only a few boulders containing Ni and Cu are known. Within the Kakkuri gabbro, there are small amounts of Ti-Fe oxides (Veräjämäki 1998).

Peltonen (1993) has evaluated the ore potential of the Finnish bedrock. Except for the area north and northeast of the Satakunta sandstone, the ore mineral resource potential of the study area is considered to be less than average.

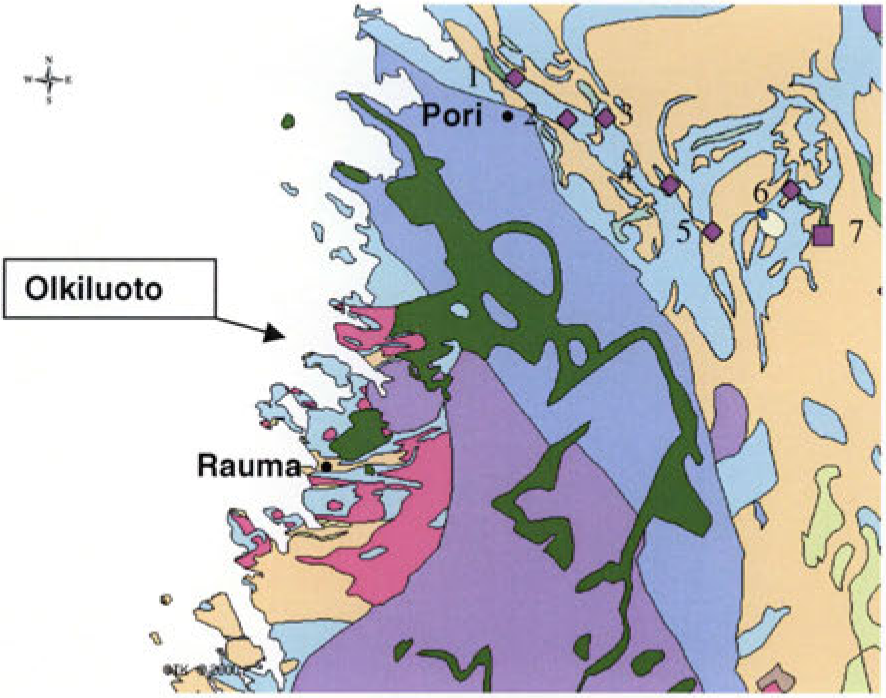


Figure 2.3-1. *Ni-Cu mineralizations. 1) Hyvelä, 2) Harjunpää, 3) Torisevanahde, 4) Levanpelto, 5) Routtuperko, 6) Sääksjärvi, 7) Piilijoki. Diamond = drilled Ni-Cu, square = Cu-Ni mineralisation (not drilled). Base map the bedrock map of Finland 1:1000 000. Map scale 1 cm = 8.75 km. Source: www.gsf.fi/explor.*

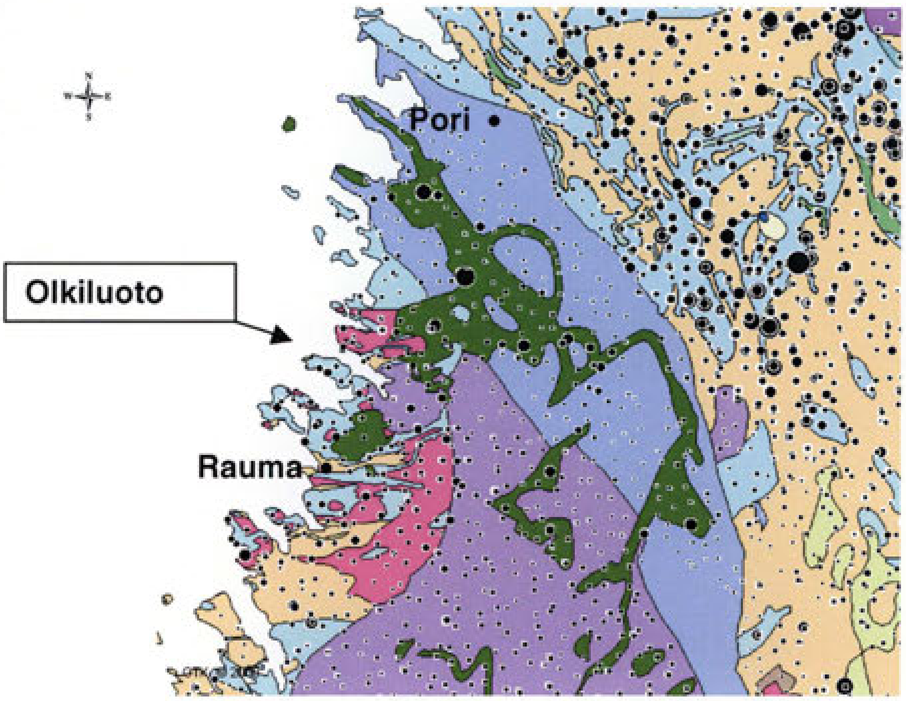


Figure 2.3-2. *Ni and Cu in basal till. The size of the symbols is in relation with the total concentration of the element. Base map the bedrock map of Finland 1:1000 000. Map scale 1 cm = 8.75 km. Source: www.gsf.fi/explor.*

3 DEFORMATION AND METAMORPHISM

3.1 Deformation

The detailed structural geological studies in the southern Satakunta area are rather scarce. Consequently, the deformation history of the study area must be elucidated with the help of studies made elsewhere in southwestern Finland within comparable bedrock areas (Fig. 3.1-1).

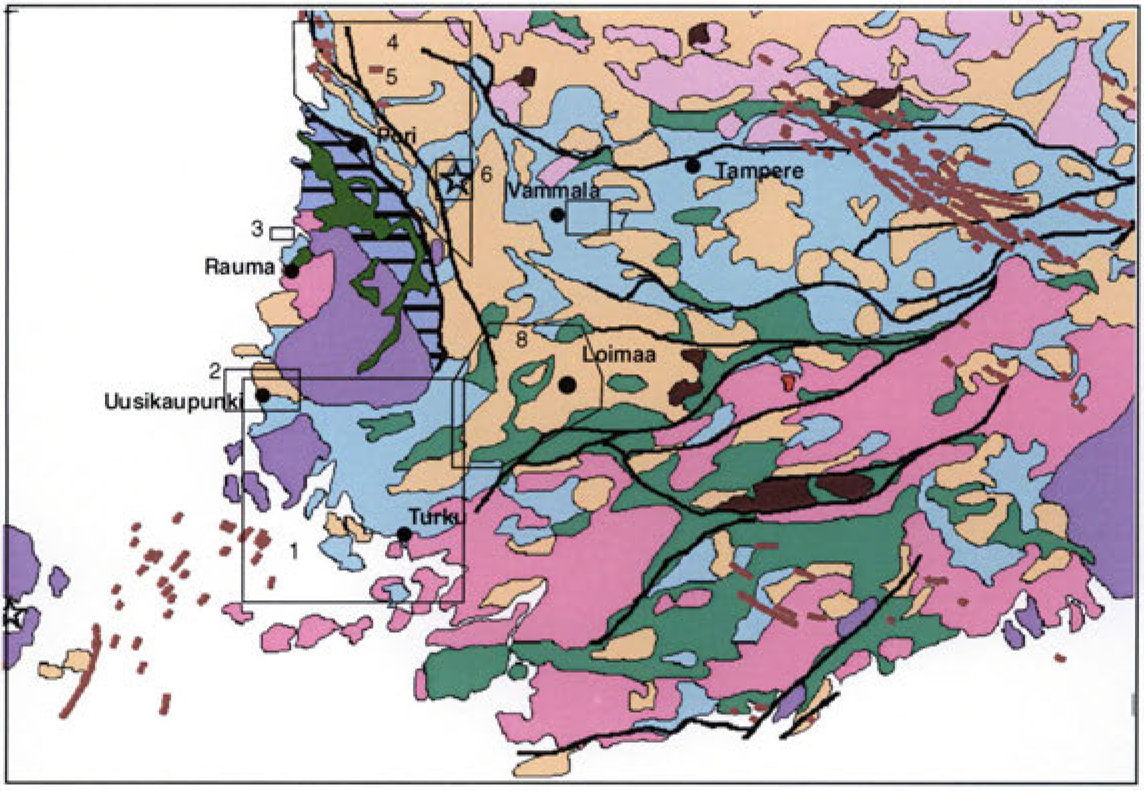


Figure 3.1-1. *Location of the structural geological study areas discussed in Section 3.1. 1. Turku area: Väisänen et al. 1994, Väisänen & Hölttä 1999; 2. Uusikaupunki: Selonen & Ehlers 1998; 3. Olkiluoto: Paulamäki & Koistinen 1991, Blomqvist et al. 1992; 4. Pori-Pomarkku: Pietikäinen 1994; 5. Pori: Pajunen 1999, Pajunen et al. 2001; 6. Sääksjärvi: Mancini et al. 1996a; 7. Vammala Stormi: Kilpeläinen & Rastas 1992, Kilpeläinen et al. 1994, Kilpeläinen 1998; 8. Loimaa: Nironen 1999. For the legend of the map see Fig. 2.1-1.*

3.1.1 Pelitic migmatite belt

Structural geological studies in southwestern Finland

<u>Turku area</u>

The earliest tectonic structure in the Turku area is the biotite foliation S_1 parallel to the bedding, identified only in the hinge zones of F_2 folds and as inclusion trails in porphyroblasts (Väisänen et al. 1994, Väisänen & Hölttä 1999). No F_1 folds have been identified. The most dominant foliation is usually penetrative S_2 with tectonic/metamorphic segregation. F_2 folding is recumbent and varies from isoclinal to tight, the fold axes plunging moderately to east or southeast. The recumbent attitude of the folds suggests a layer-parallel shearing and over- or underthrusting during D_2, although no major thrust zones have been identified (Väisänen & Hölttä 1999). The tonalite-trondhjemite intrusions were emplaced before or during D_2 and were deformed during D_2 and later during subsequent deformation phase D_3.

The regional, open to tight F_3 folding of the deformation phase D_3 is coaxial with F_2 and deformed the earlier, subhorizontal D_1/D_2 structures to upright or asymmetrically inclined structures (Fig. 3.1-2). The F_3 fold axis is usually horizontal or moderately plunging. Especially near the rapakivi granites, the folds are overturned to recumbent. F_3 fold limbs are often strongly sheared. Up to a few hundreds of metres wide shear zones are usually subparallel to the general D_3 trend and are subvertical or dip southeast. Late orogenic potassium granites dated 1840 - 1830 Ma were emplaced during late D_3. They occur both as large bodies and as leucosomes in metasedimentary rocks. The granite dykes are mainly parallel to the axial plane of the F_3 fold, but some dykes also cut the F_3 folds. The late-D_3 shear zones are more local than other D_3 shear zones, being a few metres to 10 m wide (Väisänen & Hölttä 1999). They have caused the alteration of the late-orogenic granites into protomylonites, mylonites and ultramylonites.

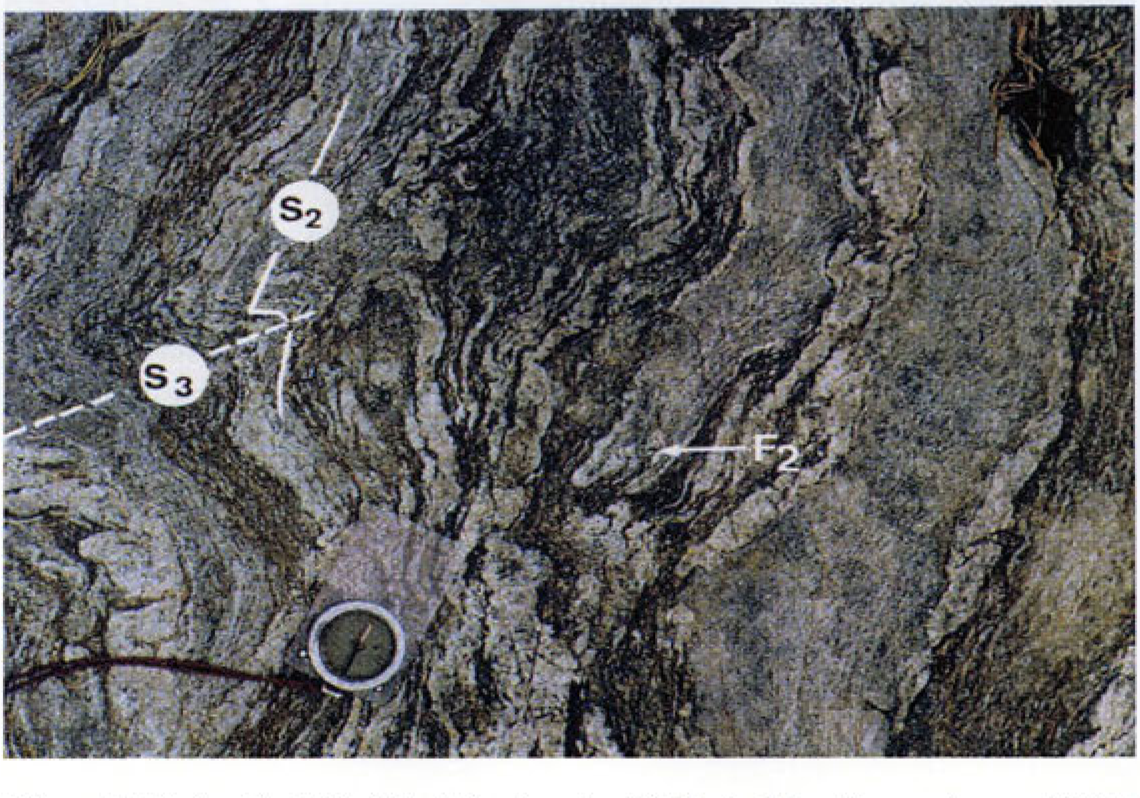


Figure 3.1-2. *Isoclinal F_2 fold deforming the lithological banding and open F3 fold deforming the D_2 neosome (Väisänen & Hölttä 1999). Vellua, map-sheet 1042 12; x = 6737 400, y = 1531 800.*

D_4 structures are a few tens of metres wide local shear zones, cross-cutting the previous structures and also the late-orogenic potassium granites. The general trend of the D_4 structures is from N-S to NNE-SSW. Both ductile and brittle structures occur, usually both in the same shear zone. Väisänen et al. (1994) tentatively associate the late-D_3 and D_4 shear zones with the older listric shear zones detached at the middle to lower crustal boundary, which are suggested by Korja and Heikkinen (1995) on the basis of deep seismic experiments. The timing of the D_4 shear zones is uncertain. They may be associated with the post-orogenic granite-intrusions dated at 1770 - 1815 Ma (*cf.* Branigan 1987, Ehlers et al. 1993). Väisänen & Hölttä (1999) suggest that they may have been re-activated 200 Ma later in association with the emplacement of the rapakivi granites, as evidenced by the sheared Kolinummi anorthosite.

According to Väisänen & Hölttä (1999) D_2 and D_3 deformations represent thickening and shortening of the upper crust, while D_4 may indicate transition to extensional regime and collapsing of the Svecofennian orogen.

Uusikaupunki area

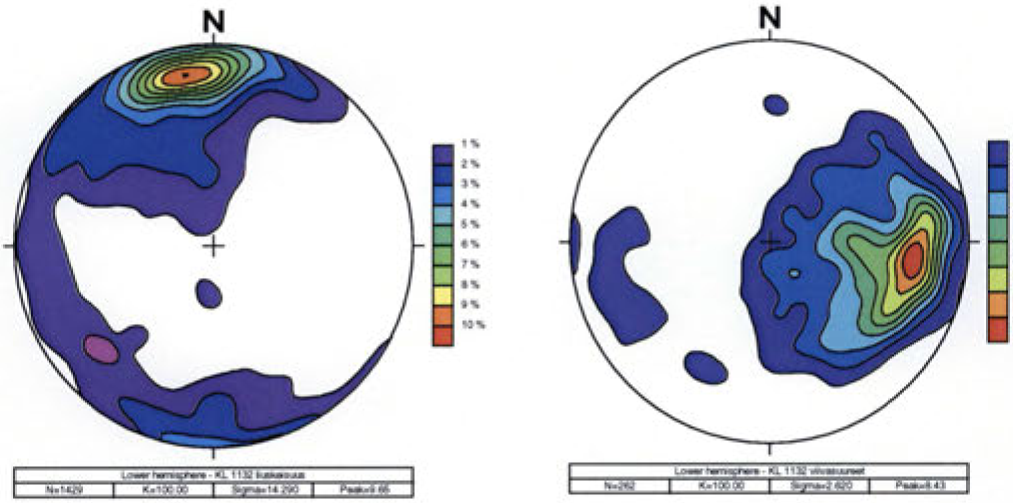
The deformation phases defined by Selonen & Ehlers (1998) at the Uusikaupunki area (see Fig. 3.1-1) are comparable with afore-mentioned deformation phases at the Turku area. Three phases of deformation are identified in the supracrustal rocks. The earliest D_1 deformation produced a foliation defined by biotite. No F_1 folds have been identified. The D_2 deformation is characterised by recumbent or reclined F_2 folds with NE-SW-trending subhorizontal to gently plunging (0 - 30°) fold axis and axial planes. A new penetrative foliation S_2 developed in the F_2 fold hinges. The D_3 deformation is associated with regional scale F_3 folds with E-W-trending axial planes. The axial planes are subvertical in the northern and central parts of the area, whereas in the southeast they are overturned towards the north. The fold limbs are often sheared. The Uusikaupunki trondhjemite is cut by post-D_3 dextral and sinistral shear zones.

Selonen & Ehlers (1998), referring to Lindroos & Ehlers (1994) and Selonen et al. (1996), suggest that during D_2 the supracrustal rocks were thrust towards the northwest in the Uusikaupunki area. Selonen & Ehlers (1998) interpret the foliation of the Uusikaupunki trondhjemite as S_2 and suggest that trondhjemite intruded as gently dipping sheets before or during the D_2 deformation. The D_3 deformation represents north-south compression accompanied by east-west shearing.

Structural geological studies in the southern Satakunta area

Rauma map sheet area

In the Rauma map-sheet area the dominant trend of foliation, fold axes and lineations is east-west (Fig. 3.1-3). The area exhibit an extensive folding with subhorizontal fold axis (Fig. 3.1-3B), re-folded by younger folding (Suominen et al. 1997). These deformation phases are apparently comparable with the deformation phases D_2 and D_3 at the Turku and Uusikaupunki areas. The main folding phase (D_2) is characterised by intense formation and subsequent folding of neosome.



A) **B)**

Figure 3.1-3. *Tectonic observations of the 1132 Rauma map-sheet area: A) foliation, N = 1429, B) fold axis and lineation, N = 262. Schmidt's equal area, lower hemisphere projection.*

Olkiluoto site

In the Olkiluoto area, within the Rauma map-sheet, the oldest recognisable structural feature is a slightly segregated, penetrative foliation S_1 (sub)parallel to the bedding S_0 (Paulamäki & Koistinen 1991). The next deformation phase (D_2) can be sub-divided into four deformation episodes, which are all characterised by intense deformation and continuous production of neosome. During the main stage of this deformation event (D_{2B}), the original bedding of the pelitic sediments was replaced by foliation and metamorphic segregation banding (S_{2B}) striking E-W, and which is associated with abundant production of granitic neosome. The earlier structural features ($S_0 + S_1 + S_{2A}$) have been preserved only within the boudinaged fragments of more competent layers, which occur inside the palaeosome-neosome matrix. Tonalite and granodiorite were intruded before or during D_{2B}. In the deformation phase D_{2C} the neosome veins formed earlier were folded isoclinally and new concordat veins were introduced. The deformation partitioned into domains producing semiconcordant and coalescent shears

(S_{2C}). The deformation phase D_{2D} represents the waning stage of semiconcordant deformation and neosome formation. It is characterised by large, rotated blocks bounded by shears (S_{2D}).

In the deformation phase D_3, the migmatite was folded by the F_3 folding (Fig. 3.1-4), producing NE-trending tight or chevron folds with some granite veins and weak shearing parallel to the axial planes, dipping SE. The folds of the deformation phase D_4 trend ca. N-S and have vertical axial planes with associated parallel weak shearing. No neosome development took place in this phase. The latest identified plastic structures are the F_5 folds of the deformation phase D_5. They are mostly just small flexures, the fold axes of which dip SE. Granulation of minerals along narrow zones occurs parallel to the axial plane, indicating that D_5 would appear to represent a transition from plastic to more brittle deformation.

The deformation phases D_1 to D_3 are comparable with the deformation phases in the Turku and Uusikaupunki areas. However, the deformation phases D_4 and D_5 have not been identified at the Uusikaupunki area. The D_4 deformation of the Turku area characterised by shear zones also include plastic features and, thus, may be comparable with the Olkiluoto D_4 structures. On the basis of the direction of the fold axis and the folding style, the D_4 deformation may correspond the D_4 deformation described by Nironen (1999) in the Loimaa area (*see* p. 46). North-south-trending fold structures are also described by Hobgood (1984) and Schreurs & Westra (1986) in southern Finland.

Figure 3.1-4. *Tight, overturned folding of the D_3 deformation in the migmatitic mica gneiss. Roadcutting 5 km east of Olkiluoto, Eurajoki. Map-sheet 1132 09C, x = 6790 070, y = 1528 920. Photo by Seppo Paulamäki/GTK.*

Brittle deformation

The brittle deformation has been examined in detail only at the Olkiluoto site, in Eurajoki, in connection with the nuclear waste disposal studies. After the plastic deformation phases, the bedrock of the area began to respond to stress in a more brittle fashion. Later deformation features include dextral and sinistral, brittle minor faults striking N-S and NW-SE, with a displacement of 1 to 3 cm (Paulamäki & Koistinen 1991).

The surface fractures form a very distinct system, where one main fracture direction is parallel to the foliation and migmatitic banding S_{2B-D} striking approx. ENE-WSW, the second is perpendicular to it and the third intersects these at an oblique angle (NE-SW) (Front et al. 1998). At least part of the third fracture set is also parallel to the foliation, since in the southeastern part of the area the strike of the foliation rotates from E-W to NE-SW. First two fracture strikes coincide with the direction of the maximum and minimum horizontal stress, respectively. In the drill core samples, fractures dipping gently (0 - 40°) to the SSE (160 - 165°) dominate.

Fracture minerals

Blomqvist et al. (1992) have studied the fracture infillings from a fracture zone at 613 - 618 m in borehole OL-KR1 at Olkiluoto. Based on mineral association, a total of 21 fracture infillings were detected. They were classified into five groups according to decreasing temperature and age. Two oldest groups are hydrothermal (T <300°C). The first one is considered to be associated with the emplacement of the Laitila rapakivi batholith. The hydrothermal fluids from the batholith caused alteration of the mica gneiss, including silicification and formation of albite and muscovite. The fracture infillings are characterised by muscovite-greisen fractures, silicified microbreccias, albite veins and quartz veins. The younger hydrothermal group is characterised by clay minerals, especially illite, crystallised on chlorite shear planes and fractures. Other typical fracture infillings include pyrrhotite, baryte, laumontite-leonhardite, analsime, adular and fluorite. On the basis of three illite samples dated at 1031 Ma, 1353 Ma and 1365 Ma, the group is presumably associated with hydrothermal activity before or during the Postjotnian magmatism, represented by olivine diabase dykes (Blomqvist et al. 1992). This hydrothermal activity was concentrated along fracture zones.

Pyrite veins, calcite-chamosite breccias and infillings of the third group cut the two previous groups. The fourth group is characterised by clay minerals with formation temperatures between 150°C and 40°C. Plagioclase is altered to kaolinite. The youngest group consists of monomineralic calcite infillings, or breccias with anatase on the walls of minor cavities. The dated calcites indicate crystallisation ages of less than 300 000 years (Blomqvist et al. 1992).

3.1.2 Psammitic migmatite belt

<u>*Pori map sheet area*</u>

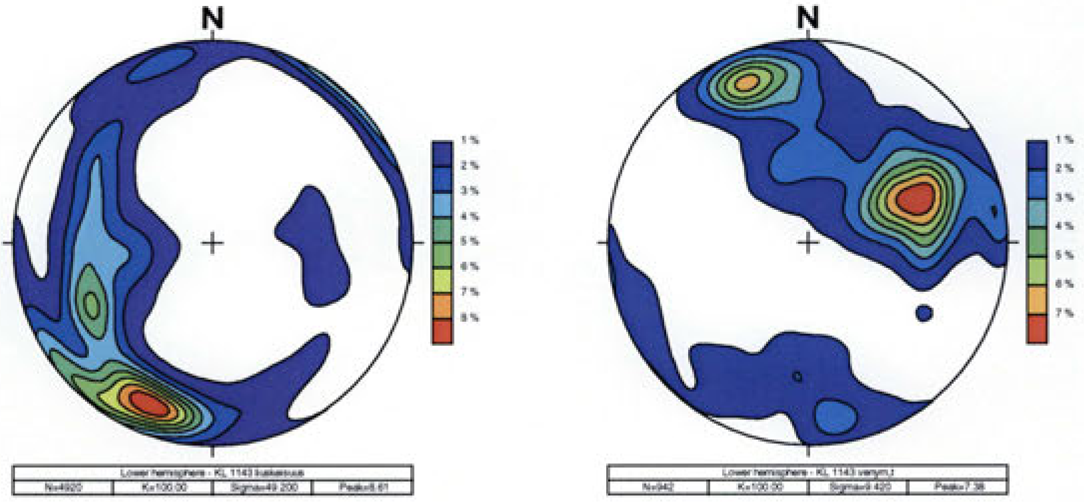
Ductile deformation

The tectonic history of the area north and northeast of the Satakunta sandstone formation have been studied by Pietikäinen (1994), Mancini et al. (1996a) and Pajunen et al. (2001). The Kynsikangas shear zone (see Fig. 2.1-3, App. 1) divides the area into several blocks, of which the southeastern part of the Pomarkku block, located northeast of the shear zone, and the Pori block between the sandstone and the shear zone are within the study area. The structural geological interpretation of the area by Pajunen et al. (op. cit.) is presented in Fig. 3.1-7.

Undisputed primary structures have not been observed and the signs of the earliest deformation (D_1) are weak, because later deformation phases have strongly overprinted or destroyed the D_1 structures. D_1 deformation is only visible in the supracrustal rock as a nearly bedding-parallel foliation S_1 (Pajunen et al. 2001).

Folds of the deformation phase D_2 (Fig. 3.1-6) vary from isoclinal to semi-open, and they are associated with a strong axial plane foliation S_2, which is the earliest foliation observed in the tonalites (Pietikäinen 1994, Mancini et al. 1996a, Pajunen et al. 2001). The foliation is mainly trending WNW-ESE and dipping to NNE (Fig. 3.1-5. and 3.1-7). Progressive metamorphism and partial melting of the sedimentary rocks, producing the trondhjemitic migmatites, have a close temporal relationship with the D_2 deformation (Pajunen et al. 2001, see also Koistinen et al. 2000). In the mica gneisses, the secondary banding occurs as biotite-rich bands, a few millimetres wide. In the tonalites the banding is caused by variation of the amount of mafic minerals, and the bands can be several metres wide (Pietikäinen 1994). The foliation S_2 is accompanied by shearing along the D_2 fold limbs (Fig. 3.1-7). According to tectonic analysis by Pajunen et al. (2001), shearing of the migmatites was strongest at the later stages of the D_2 deformation. The contact of the Lake Sääksjärvi ultramafic intrusion is parallel to the S_2 foliation of the country rock, which suggests that the intrusion was emplaced before or during the D_2 deformation (Mancini et al. 1996a).

D_3 deformation occur both as small-scale folds deforming the D_2 structures, and regional folds, for example, on the margin of the Pomarkku block. In places migmatitic leucosome occurs parallel to the axial plane (Mancini et al. 1996a). The magnetic pattern visible in the low-altitude aeromagnetic map (App. 5 and 6) on the margin between the Pomarkku block and the Kynsikangas shear zone is due to D_3 folding (Pietikäinen 1994).



A) B)

Figure 3.1-5. *The tectonic observations of the map-sheet 1143 Pori : A) foliation, N = 4920, B) lineation , N = 942. Schmidt's equal area, lower hemisphere projection.*

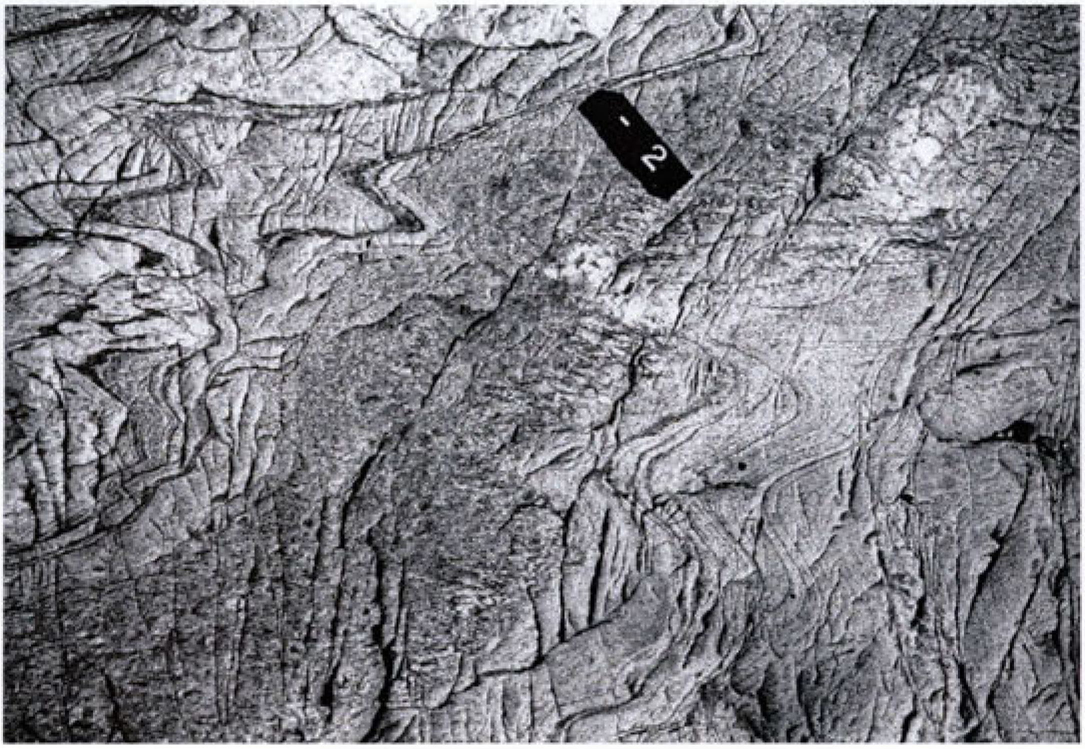


Figure 3.1-6. *Banded and folded (F_2) felsic metavolcanite. Harjakangas, Noormarkku. Map-sheet 1143 03. Photo by Hannu Kujala/GTK.*

The folds of the D_3 deformation phase vary from open to tight depending on the area (Pajunen et al. 2001). In the Pomarkku block (sub-area 2a in Fig. 3.1-7), the folding is usually open and the fold axis is subhorizontal. S_3 foliation is weak but the deformation phase is associated by a strong, almost horizontal S_2/S_3 intersection lineation L_2/L_3, trending E-W. Overturned folds towards the south occur locally indicating overthrusting along the D_3 shear zones towards the south (Pajunen et al. 2001). The Pori block (sub-area 1 in Fig. 3.1-7) is characterised by tight folding. The stretching lineation L_3 is mostly steep, dipping to NNW and ENE. Recrystallised, blastomylonitic shear zones, characterised by granitisation and, for example, biotitisation of amphibolites occur on the block (Pajunen et al. 2001).

The D_2-D_3 shear deformations are ductile and the rock types are notably recrystallised. Consequently, identification of the early fault rocks - mylonite gneisses and blastomylonites - from the normal gneisses is often difficult (Pajunen 1999). In places, the early shearing is accompanied by metasomatism, for example, growth of the potassium feldspar porphyroblasts.

The Kynsikangas shear zone (sub-area 7b in Fig. 3.1-7) represents the late-D_3 deformation, which deforms the earlier D_3 shear zones (Pietikäinen 1994, Pajunen 1999, Pajunen et al. 2001). Lensoid rock fragments with earlier structures, surrounded by shear zones, are typical. Compared to the earlier shear deformations, the Kynsikangas shear zone is more linear, has a strong lineation and more brittle structures. It is associated with proto-, orto- and ultramylonitic fault rocks, in which the recrystallisation is more pronounced than in the shear zones of the main phase of the D_3 deformation (Pajunen et al. 2001). The foliation of the mica gneisses near the Kynsikangas shear zone is subparallel to the mylonitic foliation within the zone. Strong horizontal stretching lineation indicates horizontal movements (Pajunen et al. 2001). According to Pietikäinen (1994) the movement along the shear zone has been sinistral, and the Pori block has moved upward compared with the Pomarkku block. The shear zone is cut by undeformed Subjotnian diabase dykes.

The N-S-trending, dextral folding and shearing of the deformation phase D_4 is only locally visible (Mancini et al. 1996a, Pajunen et al. 2001). The open, gentle folding deforms the early structures causing ovoid interference structures (F_X in Fig. 3.1-7). Compared with the earlier deformation phases this deformation is more brittle and represents a transition from the ductile deformation towards the brittle one.

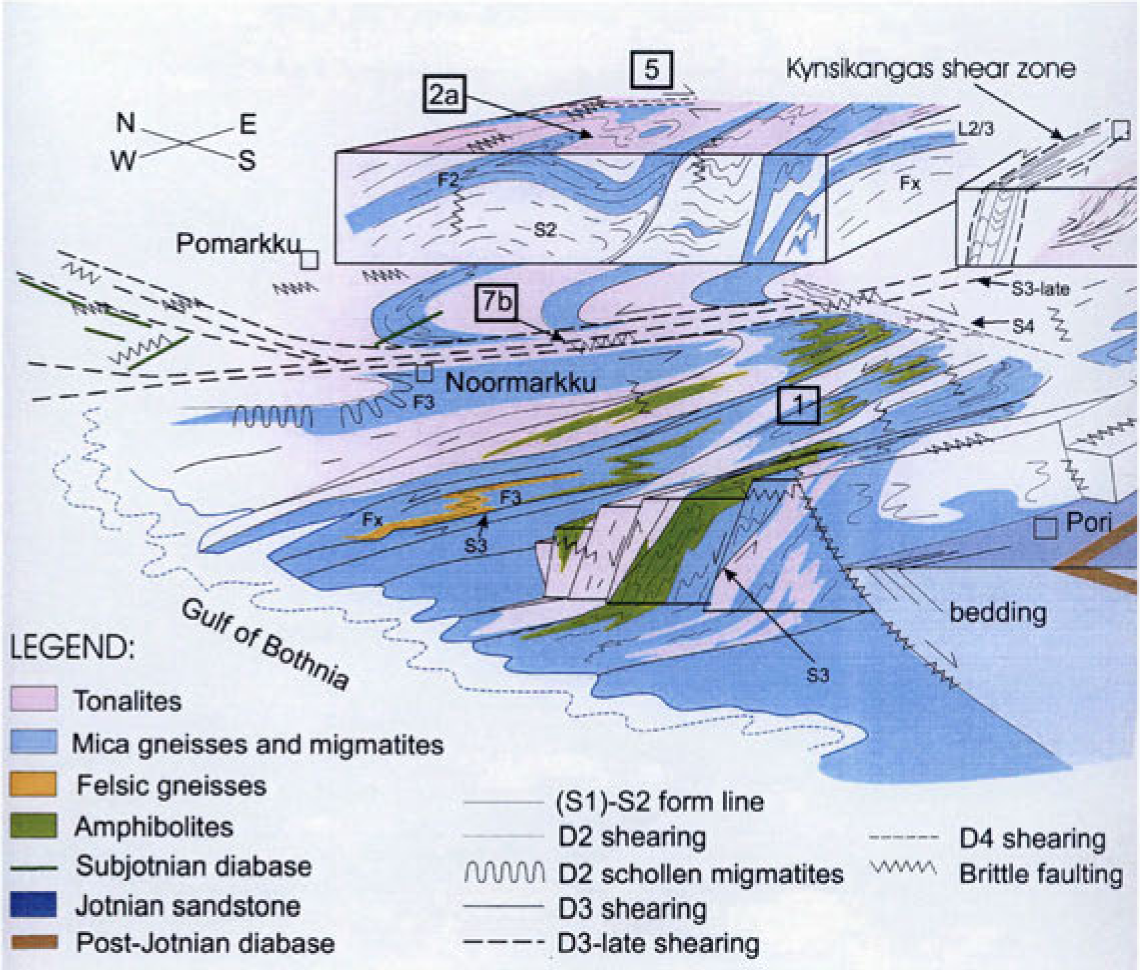


Fig. 3.1-7. Structural geological interpretation of the Pori area north and northeast of the Jotnian Satakunta sandstone (Pajunen et al. 2001).

Brittle deformation

The aeromagnetic data in the area northeast of the Satakunta sandstone shows weak magnetic lineaments cutting the geological units (Pajunen et al. 2001). The main trends of the lineaments, interpreted as surface traces of faults and fracture zones, are 10 - 20°, 50 - 60°, ca. 80°, 120 - 130° and 140 - 150° (Fig. 3.1-8).

Within the Pomarkku block (sub-areas 2a-2e in Fig. 3.1-7) and in northwestern part of the Pori block (sub-areas 1a-1e in Fig. 3.1-7) there are linear structures trending ca. 80°, which are locally associated with brittle, altered zones. The structures are probably related to Subjotnian tectonic processes, because the Subjotnian diabase dykes have intruded in this direction (Pajunen et al. 2001). The lineaments trending 140 - 150° are concentrated in the Kynsikangas shear zone (sub-areas 7a and 7b), and are locally closely connected with the Subjotnian diabase dyke swarms. The linear structures trending 120 - 130° are expressed as jumpy changes in the intensity of the magnetic anomaly, indicating vertical block movements (Pajunen et al. 2001). Some of these can be seen also within the Satakunta sandstone formation and the Postjotnian diabases are

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The aeromagnetic data in the area northeast of the Satakunta sandstone shows weak magnetic lineaments cutting the geological units (Pajunen et al. 2001). The main trends of the lineaments, interpreted as surface traces of faults and fracture zones, are 10 - 20°, 50 - 60°, ca. 80°, 120 - 130° and 140 - 150° (Fig. 3.1-8).

Within the Pomarkku block (sub-areas 2a-2e in Fig. 3.1-8) and in northwestern part of the Pori block (sub-areas 1a-1e in Fig. 3.1-8) there are linear structures trending ca. 80°, which are locally associated with brittle, altered zones. The structures are probably related to Subjotnian tectonic processes, because the Subjotnian diabase dykes have intruded in this direction (Pajunen et al. 2001). The lineaments trending 140 - 150° are concentrated in the Kynsikangas shear zone (sub-areas 7a and 7b in Fig. 3.1-8), and are locally closely connected with the Subjotnian diabase dyke swarms. The linear structures trending 120 - 130° are expressed as jumpy changes in the intensity of the magnetic anomaly, indicating vertical block movements (Pajunen et al. 2001). Some of these can be seen also within the Satakunta sandstone formation and the Postjotnian diabases are probably controlled by these structures. Magnetic lineaments trending 40 - 45° and in places cutting the Postjotnian diabase dykes occur in the southeastern part of the area. Some of the late brittle, prehnite-filled fault fractures are in this direction (Pajunen et al. 2001). The above magnetic lineaments are cut by N-S trending linear magnetic structures, which correspond with the topographic lineaments, such as linear valleys, rivers and lakes. Several semi-brittle to brittle faults occur in this direction.

The semi-brittle to brittle structures observed at the area are microscopically weakly recrystallised mylonite-series rocks and microbreccias (Pajunen et al. 2001). Pseudotachylitic structures are found near the sandstone margin. The brittle movements have often occurred in the old zones of weakness inherited from the earlier ductile deformation, and locally the same zone has moved several times (Pajunen et al. 2001). In the Kynsikangas shear zone, there are often weakly sheared fractures filled with carbonate, mica, epidote and clay minerals, which are cutting the mylonites of the shear zone, but not the diabase dykes. The fractures have been interpreted to indicate crustal fracturing during the emplacement of the rapakivi granites (Pietikäinen 1994).

The main fracture sets in the Pori region show a clear connection to tectonic structures (Pajunen et al. 2001). In many sub-areas within the region, one fracture set is parallel to the main foliation.

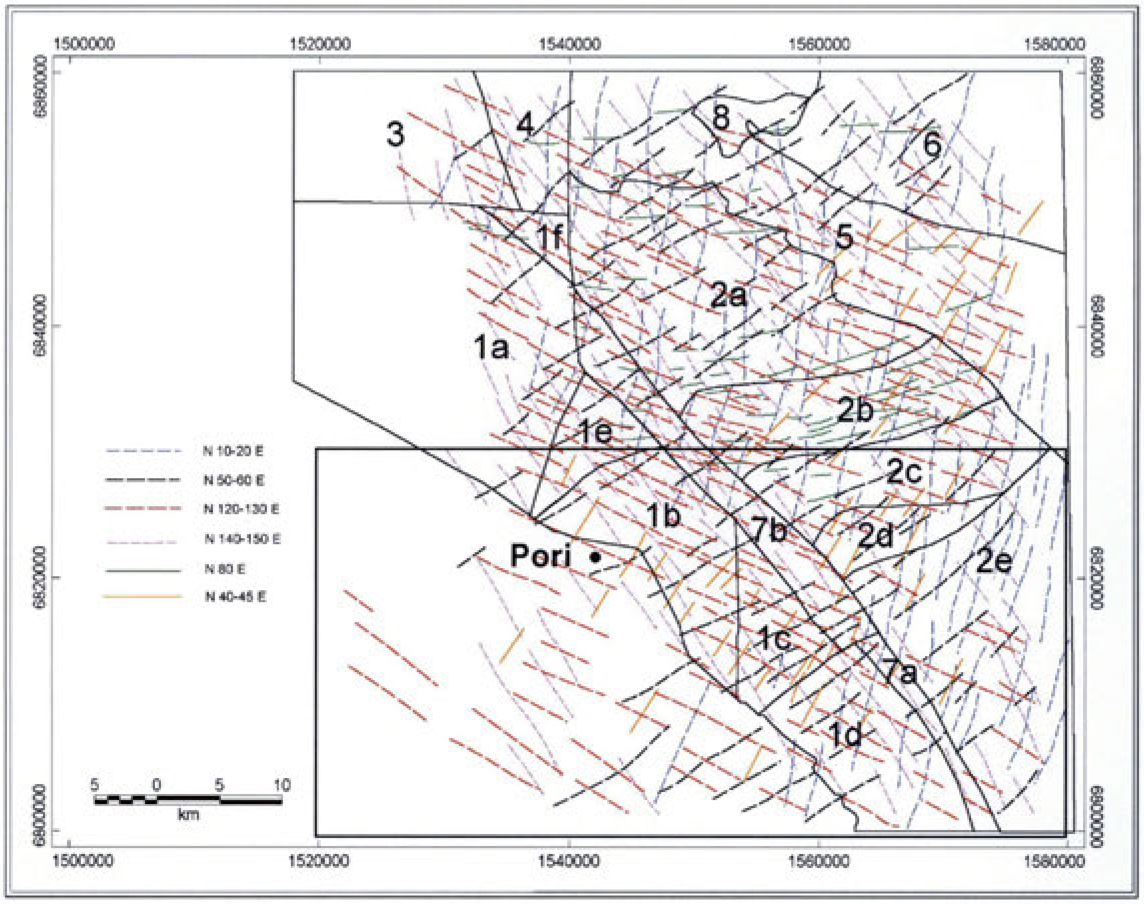
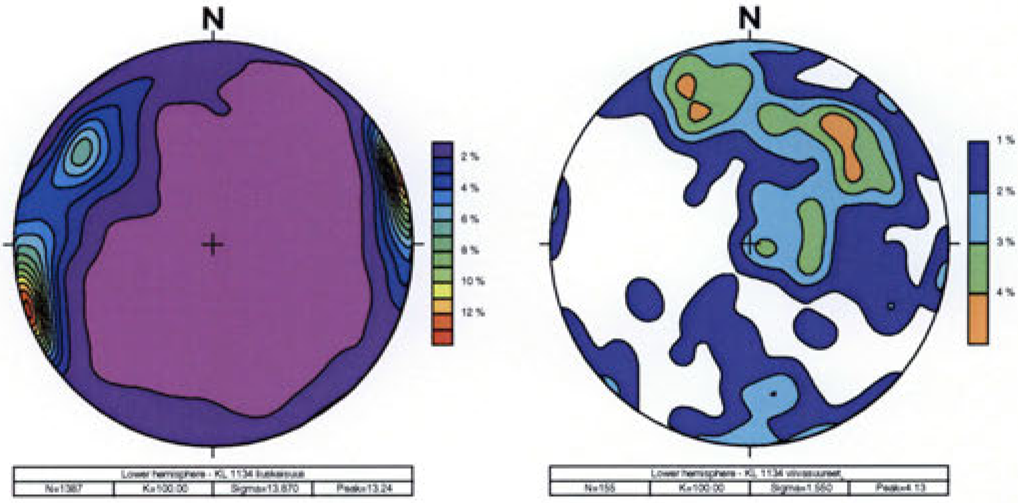


Figure 3.1-8. Main orientations of the lineaments in the Pori area, interpreted from the aeromagnetic maps (Pajunen et al. 2001). The sub-areas 1b-1c, 2c-2d and 7a-7b are within the northern part of the study area, marked with a rectangle.

Kokemäki map sheet area

In the Kokemäki map-sheet area, tectonic observations have been made in association with the general geological mapping but the structures have not been divided into separate deformation phases. Consequently, comparison with the above deformation phases is not possible. Foliation, most likely S_2, is mainly striking NNW-SSE and dipping steeply to ENE or WSW (Fig. 3.1-9). The measured fold axes plunge gently towards NE or NNW (Fig. 3.1-9) (Veräjämäki 1998). At the map scale the most prominent fold structures are the Kiettare anticline and the Ruotanajärvi antiform to the southeast of Lake Köyliönjärvi. Northeast of Lake Köyliönjärvi, there are N-S-trending folds, which can be seen also in the low-altitude aeromagnetic maps (App. 5 and 6).



A) B)

Figure 3.1-9. *Tectonic observations of the 1134 Kokemäki map-sheet area: A) foliation, N = 1387, B) fold axis and lineation , N = 155. Schmidt's equal area, lower hemisphere projection.*

The deformation associated with the Kynsikangas shear zone is represented by subhorizontal, NNW-plunging folding (Veräjämäki 1998). The sinistral movement along the shear zone is probably indicated by NE-striking foliation in the rocks east of Lake Köyliönjärvi and the strikes turning from SW to NE west of Kokemäki. The NNW-striking foliation east of Kokemäki is associated with the shear zone itself.

<u>*Vammala Stormi area*</u>

The deformation of the psammitic belt in the Pori region can be compared with the Vammala Stormi area (Fig. 3.1-1), which, according to Kilpeläinen (1998), represents the psammitic migmatite belt at its most typical. The rocks of the Stormi area are comparable with the Pori area, and are mainly composed of migmatized, arenaceous metasedimentary rocks and the granodiorites and quartz diorites cutting them.

The oldest tectonic and metamorphic feature is the penetrative biotite foliation S_1 parallel to the layering (Kilpeläinen et al. 1994, Kilpeläinen 1998). However, the intense deformation and metamorphism have usually destroyed the original layering. The S_1 foliation is comparable with the S_2 foliation of the Pori area. During the D_2 deformation the earlier horizontal structures were folded into vertical position by a horizontal folding with vertical axial plane and east-west plunge. The contacts of the granodiorites and quartz diorites and their gneissose orientation are always parallel to the S_2 foliation of the metasedimentary rocks. Consequently, the plutonic rocks were intruded either during or before D_2 deformation (Kilpeläinen 1998). On the basis of the granitoids, the age of the D_2 deformation is about 1885 - 1880 Ma.

The above structures are cut by two sets of folds of deformation phase D_3. The usually sinistral, asymmetric folds with NW-SE-striking axial plane are dominant. Most of them are open and the fold axis is vertical or plunges to ENE. The other D_3 folding is similar but the axial plane strikes 60°. The axial planes of D_3 folding commonly show weak crenulation. In the zones of intense D_3 folding, the older trondhjemitic veins of the migmatites are cut by granitic veins parallel to the fold axis. The age of the D_3 deformation is determined by the 1870±4 Ma old Lavia porphyritic granite, which has intruded contemporaneously with D_3 (Kilpeläinen 1998). The 1840 - 1830 Ma old granitic migmatization mainly concentrated in zones of intense D_3 at the southern part of the Vammala area is also found to be coeval with D_3. However, this does not have to mean that the deformation was long-lasting, since the D_3 structures are not necessarily contemporary everywhere (Kilpeläinen 1998).

Like in the Pori area the deformation phase D_4 is a weak event and only randomly visible. The folds are N-S-trending and together with D_3 folds form dome-and-basin interference structures (Kilpeläinen & Rastas 1992).

Loimaa area

Nironen (1999) has studied the deformation of the Loimaa area, located southeast of the Satakunta area (Fig. 3.1-1). The area is situated near the boundary between the psammitic and the pelitic migmatite belts but due to the high degree of metamorphism the boundary is unclear. The dominant foliation is S_2 of the deformation phase D_2, subparallel to the primary lithological layering. S_1 foliation of the earlier deformation D_1 is identified only as biotite inclusion trails within the potassium feldspar grains. The penetrative foliation of the tonalites and granodiorites is parallel to the gneissic banding and it is deformed by D_3 folding. Thus, the tonalites and granodiorites were intruded before the D_2 deformation or more likely during the early stages of the deformation at 1890 - 1880 Ma (Nironen 1999).

In the northern part of the area, the lithological layering is folded by open to isoclinal F_3 folds with a fold axis plunging steeply to east or west. The leucosome of the migmatitic gneisses is subparallel to the S_2 foliation and also to the F_3 axial plane, indicating that the migmatization occurred during D_3. The age of the deformation phase is determined by the Pöytyä granodiorite, which was emplaced during D_3 and dated at 1870 Ma.

The ductile folding of the deformation phase D_4 varies from open to tight, the axial plane trending north-south. The deformation probably happened sometime in the period 1850 - 1800 Ma (Nironen 1999).

Nironen (1999) has compared the deformation of the Loimaa area with the deformation at the Pori and Vammala Stormi areas (Fig. 3.1-10). The predominant foliation and compositional banding at the Pori area is similar to the penetrative foliation S_2 of the tonalites and granodiorites of the Loimaa area. Hence, it seems that D_2 structure at both areas are comparable. According to Nironen (1999), the late-D_3 structures of the Pori area can be correlated with D_4 structures of the Loimaa area. This correlation is, however, unclear because in the Pori area the late-D_3 is characterised by shearing, whereas in Loimaa F_4 folding was probably caused by crustal shortening. The D_2 and

D_3 structures of the Vammala area have developed during N-S compression, which is compatible with the D_3 deformation of the Loimaa area. The NNW-SSE and N-S trending folding interpreted by Kilpeläinen (1998) as D_3 at the Vammala area are interpreted by Nironen (1999) as D_4 structures of the Loimaa area.

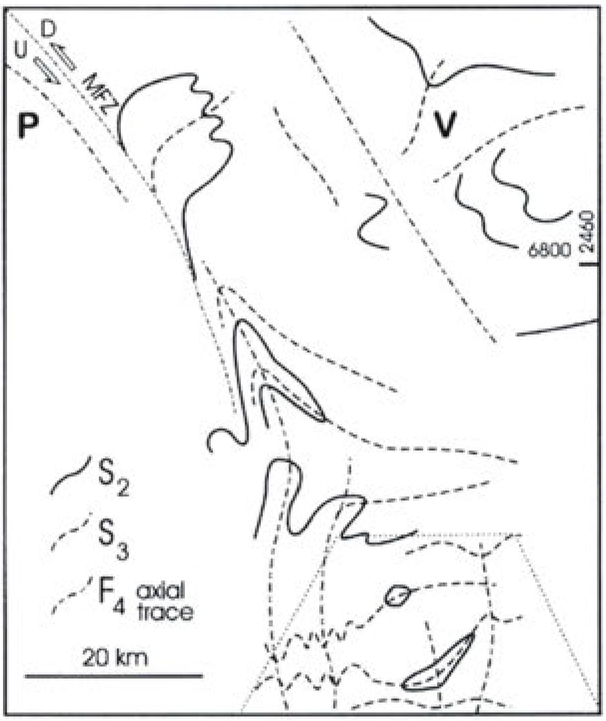


Figure 3.1-10*. Correlation of the structures at the Loimaa, Pori (P) and Vammala (V) areas (Nironen 1999). The relative sense of movement in the Kynsikangas shear zone (MFZ) is shown by arrows and block movement directions (U = up, D = down). The Loimaa area is shown by a dotted line.*

3.1.3 Post-Svecofennian rocks

The anorogenic rapakivi granites are undeformed. The gently curving lineaments within the Laitila rapakivi granite batholith are probably associated with the internal development of the rapakivi magma (Veräjämäki 1998). In addition to large-scale circular structures, the rapakivi granites are characterised by two sets of vertical, perpendicular fractures and a horizontal fracturing.

The greisen veins cutting the rapakivi granite represent ancient minor faults. The veins are mostly subvertical and strike E-W. In the vein swarm between Hakkila and Lapijoki in Eurajoki the movement direction has been dextral strike-slip, i.e., the northern side of the vein has moved to the east in relation to the southern side (Haapala 1977). In the vein swarm to the north of the Eurajoki centre, the sinistral strike-slip movement is dominant.

The Satakunta sandstone is at least partly bounded by NW-SE-trending, probably vertical faults. The layering within the sandstone is usually horizontal but in the northeastern margin the sandstone layers are tilted 35° towards the centre of the formation, indicating block movements after the deposition of the sandstone (Fig. 3.1-11).

Both in the rapakivi and sandstone areas there are NW-SE- and NE-SW-striking fractures, which are filled with olivine diabases (Veräjämäki 1998). These are cut by fractures striking almost N-S.

Figure 3.1-11. *Gently dipping sandstone layers. Lammaistenkoski hydro power plant, Harjavalta. Map-sheet 1143 04C; x = 6803 780, y = 1559 790. Photo by Marjatta Virkkunen/GTK.*

3.2 Metamorphism

The Svecofennian Domain is characterised by a high temperature/low pressure metamorphism. Typical metamorphic minerals, porphyroblasts, in pelitic metasediments are andalusite, staurolite, garnet, cordierite, sillimanite and potassium feldspar. The temperature of the metamorphism is estimated at 650 - 700°C and the pressure 4 - 5 kb, corresponding the upper amphibolite facies. The peak temperature of 700 - 800°C was attained in the migmatite areas under pressures of 4 - 6 kb.

On the basis of metamorphism the Svecofennian Domain can be divided into two belts (Fig. 2.1-2). The northern belt is characterised by tonalitic migmatites, in which the paleosome consists of psammitic metasediments and the neosome is tonalite or granodiorite. The southern belt is characterised by potassium granitic migmatites, produced by the partial melting of the pelitic metasediments (Korja et al. 1994, Korsman et al. 1999). The difference in leucosome composition depends mainly on the primary compositions of the sedimentary protoliths (Korsman et al. 1999). In the Al-rich metasedimentary rocks of the pelitic migmatite belt, the breakdown reactions of muscovite and biotite led first to formation of garnet-cordierite gneisses and later to formation of potassium-rich melts. In the psammitic migmatite belt, with alkalis and calcium in excess over aluminium, no breakdown of biotite and muscovite occurred. Potassium was retained in the biotite and therefore the melts were less potassic than those in the pelitic migmatite belt.

In the psammitic migmatite belt the peak temperature of 800 - 670°C and the pressure of 5 - 6 kb were attained at about 1885 Ma during a thermal pulse caused by the intrusion of the granitoids dated at 1889 - 1880 Ma, and the metamorphic evolution ceased soon after (Fig. 3.2-1). The metamorphism took place during or after the D_2 deformation but before the D_3 deformation (Korsman et al. 1999). Also the Southern Finland pelitic migmatite belt was metamorphosed at the same time but there the older metamorphism is overprinted by younger metamorphism, caused by a new, strong thermal pulse. This metamorphic event took place 1860 - 1810 Ma ago at the peak temperatures of 700 - 800°C and the pressures of 4 - 5 kb (Fig. 3.2-1), producing abundant potassium granite melts (microcline granites dated at 1840 - 1830 Ma). The peak metamorphic temperatures and pressures correspond the depth of 13 - 16 km (Korsman 1977, Hölttä 1988). In the psammitic migmatite belt, for example, in the Vammala area, the younger metamorphism is only locally visible (Kilpeläinen et al. 1994).

The Svecofennian high-temperature, low-pressure metamorphism is attributed to thinning of the tectonically thickened crust and subsequent intra- and underplating, which led to a strong increase in temperature (Korja et al. 1993, Lahtinen 1994, Nironen 1997, Korsman et al. 1999).

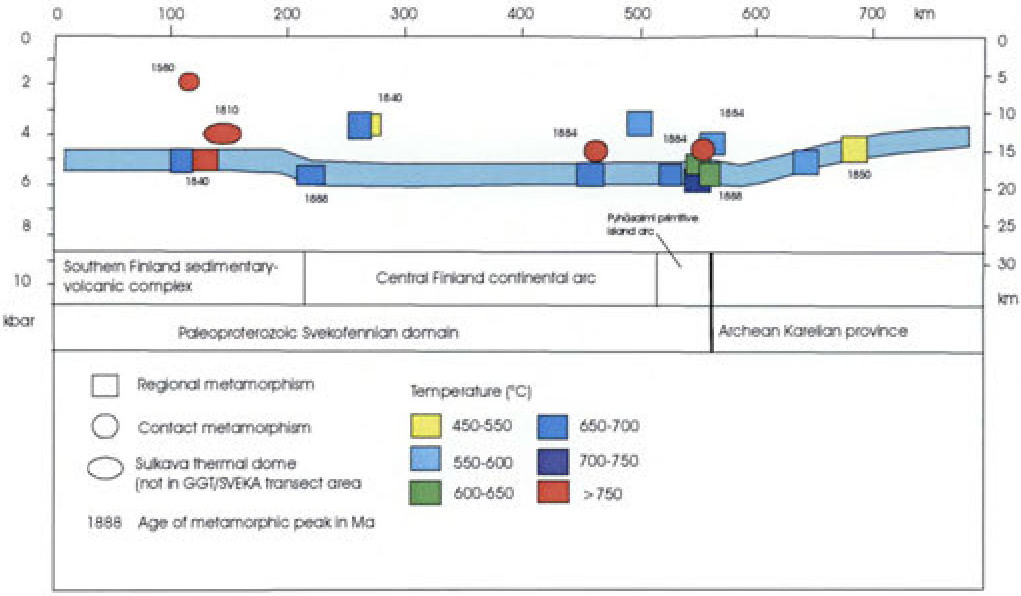


Figure 3.2-1. *Metamorphic peak conditions (pressure and temperature) along the GGT/SVEKA profile and the corresponding depth. Simplified after Korsman et al. (1999). The location of the profile is shown in Fig. 2.1-2.*

In the Rauma map-sheet area, southwest of the Satakunta sandstone, the typical metamorphic mineral assemblage of the pelitic migmatites is plagioclase-quartz-biotite-cordierite(±garnet and ±sillimanite), representing a medium-grade metamorphism (Suominen et al. 1997). Partial melting has taken place in the mica gneisses, as indicated by leucosome veins in mica gneiss paleosome.

In the surroundings of Uusikaupunki, south of the Rauma map-sheet area, the typical metamorphic mineral assemblage is cordierite-sillimanite-biotite-potassium feldspar, formed in a temperature of less than 650°C (Väisänen & Hölttä 1999). Most of the area south and southeast of the Laitila rapakivi batholith has been metamorphosed in temperatures of 650 - 780°C, the typical mineral assemblage being garnet+cordierite, without coexisting biotite and sillimanite. Metamorphism reached the conditions corresponding the upper amphibolite fasies (mineral assemblage garnet-potassium feldspar-quartz) during the deformation phase D_2 1890 - 1870 Ma ago (Väisänen & Hölttä 1999). The formation of granitic melts due to partial melting of the metapelites began already during D_2 but was strongly increased during D_3 deformation. The peak of the metamorphism was attained during D_3 1840 - 1830 Ma ago with a formation of granulite areas and potassium-rich granites. The granitic melt was formed *in situ* by dehydration melting, i.e., no water was added from outside (Väisänen & Hölttä, Mengel et al. 2001).

In the Kokemäki map-sheet area, the most common mineral assemblage of the mica gneisses is quartz-plagioclase-biotite-garnet corresponding the amphibolite fasies (Veräjämäki 1998). In the Kynsikangas shear zone, the degree of metamorphism is lower than usually in the area. In the centre of a large-scale fold structure east of Lake Köyliönjärvi the conditions have been almost anatectic, and the mica gneiss earlier migmatized by tonalite has altered to nebulitic granodiorite and granite. On the basis of the above mineral assemblages and the prevailing migmatizing granite the metamorphism has taken place in temperatures of 500 - 700°C and in relatively low pressures at a depth of about 15 km (Veräjämäki 1998).

Migmatitic gneisses of the psammitic migmatite belt in the surroundings of Lake Sääksjärvi are composed of garnet, cordierite, sillimanite and potassium feldspar. The migmatitic gneisses have amphibolite intercalations, consisting of diopside, hornblende and plagioclase. On the basis of structural relationships of the migmatites the present low pressure/high temperature mineral assemblages have formed by the following reactions (Mancini et al. 1996a):

(1) cordierite + orthoclase + H_2O= biotite + sillimanite + quartz

(2) staurolite + quartz = garnet + sillimanite + H_20

(3) muscovite + quartz = orthoclase + sillimanite + H_2O

The reactions indicate the peak metamorphic temperatures of 680 - 720°C and pressures of 4 - 5 kb (Mancini et al., op. cit., Mancini & Papunen 2000).

Also the mafic and ultramafic intrusions have been affected by the regional metamorphism. The mineral assemblage corresponding the peak of the metamorphism of the peridotite of the Sääksjärvi ultramafic intrusion is olivine+Ca-amphibole±ortopyroxene±chrome spine (Mancini et al. 1996a). The amphibole crystallised at least at 650°C and 5±1 kb. The peak metamorphic temperature was 700 - 750°C and the pressure 5 kb. The amphibolite of the intrusion is composed of Ca-amphibole with cummingtonite, plagioclase, quartz, albite and Fe-Mg-cummingtonite as exolution lamellae. The mineral composition of the amphibolite marks the metamorphic conditions after the peak of the metamorphism, when the temperature varied from 600°C - 620°C (Mancini et al. 1996a).

No signs of contact metamorphic effects caused by the Laitila rapakivi batholith have been observed in the study area. However, such effects are possible, since both at the southern margin of the Laitila batholith and in the surroundings of the Vehmaa rapakivi batholith, a contact metamorphic aureole extending up to 2 - 3 km from the contact has been observed (Väisänen et al. 1994, Väisänen & Hölttä 1999). The pelitic country rocks have melted at the immediate contact, but in the outer parts of the aureole the original structures have been preserved. In the innermost 100 m of the aureole, the mineral assemblage in pelitic country rocks is cordierite-biotite-orthopyroxene-potassium feldspar-quartz±garnet. The first sign of contact metamorphism is the appearance spinel inclusions in cordierite. Melting and recrystallisation are demonstrated by euhedral orthopyroxene, plagioclase and potassium feldspar with interstitial quartz. Thermobarometric measurements yield pressures of 0.5 - 2 kb and maximum temperature of about 800°C, implying that the crust was uplifted by about 10 km before the emplacement of the rapakivi granites (Väisänen et al. 1994).

Kahma (1951) has described the low-grade contact metamorphic phenomena caused by the Postjotnian diabase at Lake Kiperinjärvi in the Satakunta sandstone formation. Towards the contact the sandstone becomes harder, the reddish colour brighter and the lustre more vitreous. The quartz grains, which usually have smooth surfaces, become toothed on approaching the contact, while at the same time, the fine grained matrix between the quartz grains, consisting of fine-grained aggregate of quartz, pigmented feldspar, muscovite and chlorite, decreases. At the contact, the fine-grained quartz aggregate has almost vanished and instead the bigger quartz grains have become toothed, thus joining the constituents of the sandstone more firmly together. Sericite and chlorite are absent at the contact, but instead there are "dirty" shreds of biotite. Neither twinning of the feldspar nor perthitic growth of potash feldspar are observed at the contact, which also seems to be a result of a contact influence (Kahma 1951).

Low-grade metamorphism caused by the diabase dykes has also observed in Leistilänjärvi, Nakkila, where the sandstone has altered to purple and become more compact at the contact (Laitakari et al. 1986). The change in colour is due to oxidation of ferro-oxides into ferri-oxides. The sedimentary structures have been destroyed near the contact.

In the contact between the palingenic dykes (*see* p. 27) and the olivine diabase there is a narrow reaction zone, where olivine and augite of the diabase have altered into serpentine, amphibole, chlorite and iron-bearing opaque mineral, and plagioclase has become corroded. The reaction zone is strongest in the contact of the veins from the rapakivi granite.

3.3 Summary of ductile deformation and metamorphism

The ductile deformation and associated metamorphism and magmatism are summarised in Table 3.3-1. The earliest observed tectonic structure is the biotite foliation S_1 parallel to the bedding, identified only in the hinge zones of F_2 folds and as inclusion trails in porphyroblasts. No F_1 folds have been identified. The dominant foliation is usually penetrative S_2 with tectonic/metamorphic segregation. In the pelitic migmatite belt F_2 folding is recumbent and isoclinal to tight. The recumbent attitude of the folds suggests a layer-parallel shearing and over- or underthrusting during D_2, although no major thrust zones have been identified. Comparing to the pelitic migmatite belt, F_2 folding of the same age in the psammitic migmatite belt has a vertical axial plane.

Both D_1 and D_2 structures are deformed by the regional F_3 folding of the deformation phase D_3. The fold axes are generally horizontal or moderately plunging. Axial planes of folds are usually vertical but locally also overturned and recumbent folds exist. Fold limbs are often strongly sheared. Late-orogenic potassium granites were emplaced during D_3 in the pelitic migmatite belt. They occur both as large bodies and as leucosomes in metasedimentary rocks. The granite dykes are mainly parallel to the axial plane of the F_3 fold, but some dykes also cut the F_3 folds.

On the basis of structural analysis, the tonalite-granodiorite intrusions were emplaced before or during the deformation phase D_2 and were deformed during D_2 and later during D_3 deformation. The age of the D_2 deformation is, thus, close to the age of these granitoids, 1890 - 1880 Ma. In the psammitic migmatite belt the age of the D_3 deformation is about 1885 to 1870 Ma. The age of this deformation in the pelitic migmatite belt is close to the late orogenic microcline granites (1840 - 1830 Ma). The 1840 - 1830 Ma old granitic migmatization mainly concentrated in zones of intense D_3 at the southern part of the Vammala area is also found to be coeval with D_3. This does not have to mean that the deformation was long-lasting, since the D_3 structures are not necessarily contemporary everywhere. On the basis of structural geological studies in the coastal area of southern Finland, Hopgood et al. (1983) have suggested that the deformation and migmatization of the migmatites may have taken less than 20 Ma. On the other hand, Lindroos et al. (1996) have suggested that the D_3 deformation in southern Finland would have lasted until 1805 Ma.

In the psammitic migmatite belt the peak metamorphic temperature of 800 - 670°C and the pressure of 5 - 6 kb were attained at about 1885 Ma during a thermal pulse caused by the intrusion of the granitoids dated at 1889 - 1880 Ma, and the metamorphic evolution ceased soon after. The metamorphism took place during or after the D_2 deformation but before the D_3 deformation. Also the rocks of the pelitic migmatite belt were metamorphosed at the same time but the older metamorphism within the belt is overprinted by younger metamorphism, caused by a new, strong thermal pulse. This metamorphic event took place 1860 - 1810 Ma ago at the peak temperatures of 700 - 800°C and the pressures of 4 - 5 kb, producing abundant potassium granite melts (microcline granites dated at 1840 - 1830 Ma). In the psammitic migmatite belt, for example, in the Vammala area, the younger metamorphism is only locally visible.

Table 3.3-1. *Summary of ductile deformation and associated metamorphism and magmatism within the study area.*

Time (Ma)	Pelitic migmatite belt			Psammitic migmatite belt		
	Deformation phases	Metamorphism	Magmatism	Deformation phases	Metamorphism	Magmatism
<1910	D_1: S_1 foliation	Regional	-	D_1: S_1 foliation	Regional	-
1890-1885	D_2: penetrative S_2, F_2 folding, thrust tectonics?	Regional, T = 800-670°C P = 5-6 kb	Synorogenic granitoids, mafic and ultramafic rocks	D_2: penetrative S_2, F_2 folding	Regional T = 800-670°C, P = 5-6 kb	Synorogenic granitoids, mafic and ultramafic rocks
1885-1870				D_3: F_3 folding		Lavia porphyritic granite[1)] Pöytyä granodiorite[1)]
1850-1810	D_3: F_3 folds, shearing along axial plane Plastic to semi-brittle D_4	Regional T = 800-700° P = 4-5 kb	Late-orogenic granites	Late-D_3: the Kynsikangas shear zone Plastic to semi-brittle D_4	Only locally visible	Local granitic migmatization

1) located outside the study area

4 GEOPHYSICAL STUDIES

4.1 Petrophysical properties of rock types

Petrophysical properties of rocks, like density and magnetic susceptibility, are mainly affected by their mineral composition. Furthermore, porosity, weathering and alteration have their own influence on petrophysical parameters. Susceptibility-density diagrams of most rock types show typically two separate populations, called paramagnetic and ferrimagnetic ones. In the paramagnetic range (susceptibility at most $1000*10^{-6}$ SI) magnetisation (and increasing density) is effected by iron in silicate minerals. The more mafic minerals a certain rock type carries, the more strongly magnetised (in the paramagnetic range) and the denser it is. Within the ferrimagnetic range, (susceptibility over $1000*10^{-6}$ SI) magnetisation is dominated by ferrimagnetic minerals, mainly magnetite and pyrrhotite. Also remanent magnetisation is due to occurrence of ferrimagnetic minerals.

Density and magnetisation data for the most important rock types of the study area were collected from the petrophysical database of the Geological Survey of Finland (Table 4.1-1). According to the data, diabases and mica gneisses are most strongly magnetised. Also veined gneisses, granodiorites and gabbros are often ferrimagnetic. The highest remanence values are measured from mica gneisses, gabbros and diabases. Furthermore, veined gneisses have moderately high average remanence compared to susceptibility values. Most likely the magnetic anomalies observed at the study area are related to these rock types. Sandstones, rapakivi granites and other acid/intermediate intrusive rocks are usually paramagnetic. Of these, sandstones are most weakly magnetised.

The highest density values are measured from diabases, gabbros and veined gneisses (average densities 2946, 2935 and 2800 kg/m^3, respectively). The lightest rock types are sandstones, granites and rapakivi granites (average densities 2599, 2635 and 2639 kg/m^3, respectively).

Table 4.1-1. *Petrophysical properties of the most common rock types in southern Satakunta, map-sheets 1132, 1134, 1141 and 1143.*

Rock type	Number	% from all samples	Density average kg/m^3	Density std dev.	Suscept. average 10^{-6} SI	Suscept. std dev.	Remanence average mA/m	Remanence std dev.
GRANODIORITE	192	13,31	2727	77	1595	12553	265	1978
MICA GNEISS	171	11,85	2751	150	2562	19402	10857	136513
DIABASE/DOLERITE	154	10,67	2946	89	27052	18247	1623	1176
QUARTZ DIORITE	123	8,52	2752	88	463	457	83	186
GRANITE	94	6,51	2635	85	493	2671	105	303
TONALITE	69	4,78	2743	85	494	754	141	351
QUARTZ-DIORITE GNEISS	68	4,71	2732	49	346	149	29	42
VEIN GNEISS	50	3,47	2800	499	368	198	775	4684
GABBRO	43	2,98	2935	115	1793	5995	4499	24531
RAPAKIVI GRANITE	13	0,9	2639	46	376	438	51	37
SANDSTONE	12	0,83	2599	77	98	99	110	187

At the Olkiluoto site within the pelitic migmatite zone, numerous petrophysical measurements have been done on outcrop samples as well as on samples from drill cores (Lindberg & Paananen 1990, Heikkinen et al. 1992). In Figure 4.4-1, susceptibility-density diagram for the data from five boreholes (KR1 – KR5) is presented. According to the data, the migmatitic mica gneisses are most strongly magnetised, and their magnetisation is most likely due to occurrence of pyrrhotite. Tonalites and granites are usually paramagnetic.

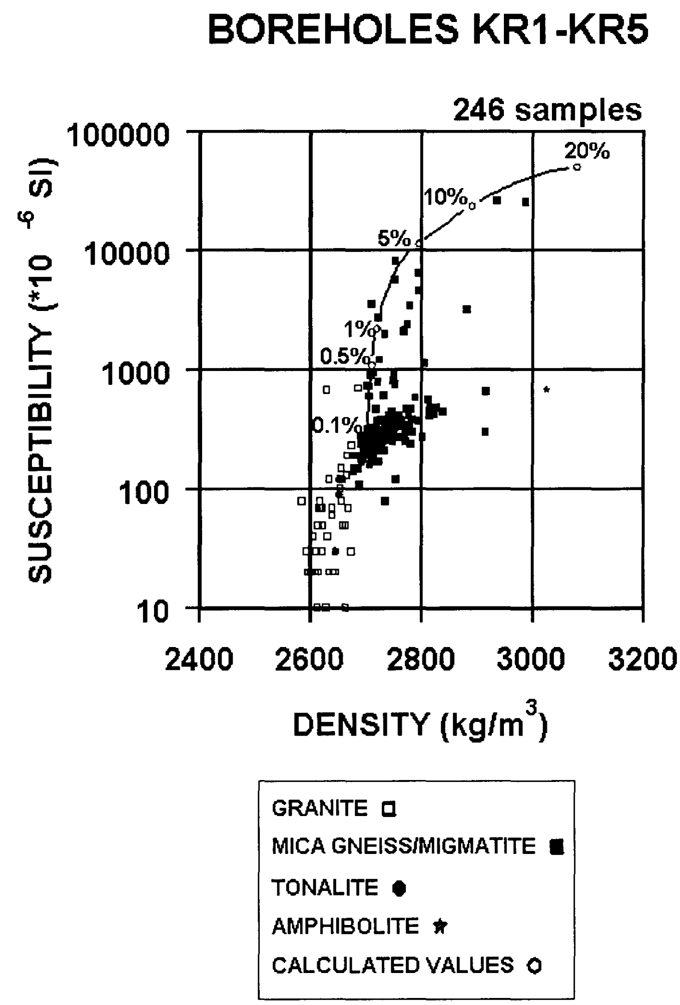


Figure 4.1-1. A susceptibility-density diagram for boreholes KR1 – KR5 at Olkiluoto and theoretically calculated values for different pyrrhotite contents (Lindberg & Paananen 1992).

In a petrophysical examination, related to a study of pre-Quaternary rocks of the Rauma map-sheet area (Suominen et al. 1997), magnetisation and density distributions of different rock types have been compared to the state-wide data. This examination reveals that in the Rauma area, rapakivi granites, granodiorites, tonalites, veined gneisses, migmatites and mafic plutonic rocks are denser than those in the comparison data. In granodiorites and tonalities, the reason is the abundance of hornblende. Furthermore, no strongly ferrimagnetic rocks appear to occur in the Rauma area. The paramagnetic values of sedimentary rocks as well as mafic supracrustal and plutonic rocks correspond the state-wide averages. Susceptibilities of granites, pegmatites and rapakivi granites are lower in the Rauma area than in the comparison data. Diabases in the Rauma area, like in the whole Satakunta region, are more strongly magnetised than on average, and their remanence is also relatively strong. This is mainly due to abundance of magnetite.

4.2 Aeromagnetic map

A magnetic map of the study area was composed at the Geological Survey of Finland using Geosoft software. The map covers low-altitude aeromagnetic data from four map sheets (Appendix 5). For the interpretation, the map was further processed by calculating horizontal and vertical gradients, by shading from different directions and by varying the colour scale. The fold structures and fracture zones interpreted from the magnetic map are outlined in Appendix 6.

In the pelitic migmatite zone in the western part of the study area, the magnetic map is dominated by two large olivine diabase sills. Elsewhere within the migmatitic zone, the magnetic anomalies are usually related to mica gneisses. According to petrophysical studies at Olkiluoto (Heikkinen et al. 1992), the mica gneisses are frequently ferrimagnetic, and their most important ferrimagnetic mineral is pyrrhotite. This is suggested by macroscopic examination of the drill core samples as well as by the susceptibility and density values calculated theoretically for certain pyrrhotite contents.

In the psammitic migmatite zone to the NE and E of the sandstone area, the aeromagnetic map is dominated by the Kynsikangas shear zone and concordant mica gneiss and tonalite units. Most frequently, the strongest magnetic anomalies are associated with mica gneisses. They are obviously caused by pyrrhotite, which also induces numerous electromagnetic anomalies. From the numerous small gabbro intrusions, located within the Pori block between the sandstone and Kynsikangas shear zone, only the most mafic ones can be observed magnetically. For example, the Hyvelä gabbro to the N of Pori can not be detected on the magnetic map.

According to petrophysical data, the Satakunta sandstone is paramagnetic and rapakivi granite at most weakly ferrimagnetic (Table 4.1-1), so the areas of these rock types are magnetically stable. In these areas, the strongest magnetic anomalies are related to dykes and sills of olivine diabase. In places, a weak NW-SE trending magnetic feature is associated with the contact of rapakivi granite and sandstone. This feature may be caused by a slight difference in magnetisation of these rock types or fracture zones related to the contacts.

The aeromagnetic map is discussed in more detail in next chapters, where the geophysical studies of the main lithological units and fracture zones are presented.

4.3 Rapakivi granites

Within the study area, rapakivi granite occurs in map-sheets 1132 Rauma and 1134 Kokemäki. The Laitila rapakivi batholith covers nearly half of the map-sheet 1134, and it is surrounded by the smaller Eurajoki and Kokemäki (Peipohja) rapakivi bodies. According to petrophysical data, the rapakivi granites are paramagnetic or weakly ferrimagnetic. However, some rapakivi types, like dark pyterlites, can be significantly ferrimagnetic (*cf.* Paananen & Paulamäki 1998). On the aeromagnetic map, the areas of rapakivi granite are magnetically stable, but in places different rapakivi phases can be distinguished from each other. For example in the Eurajoki stock, hornblende-bearing Tarkki granite can partly be separated from the more felsic Väkkärä granite.

The structure of the rapakivi granites in Satakunta was studied geophysically by Laurén (1970) and Elo (1981, 1982). Later, it was examined as a part of GGT/SVEKA project (Elo et al. 1999, Korsman et al. 1999) and most recently, near Olkiluoto by Elo (2001), under contract to Posiva Oy. According to the gravity interpretations, the southwestern contact of the Laitila rapakivi on the GGT profile (Fig. 5-1) is very gently dipping, ca. 10°. The rapakivi deepens towards the northeast, being over 10 km thick at the northeastern contact, which is interpreted to be nearly vertical (Elo et al. 1999). Further northeast, the rapakivi disappears under the Satakunta sandstone, although it may extend at the northeastern side of the sandstone near the ground surface. This idea is supported by the Kokemäki stock.

The interpreted dip of Väkkärä rapakivi at its southern and western contacts is, according to previous studies, moderately gentle to the S – W (Elo 1982). The most recent gravity interpretations (Elo 2001) suggest that the Väkkärä rapakivi granite extends downwards and outwards in every direction, and its depth exceeds 5 km. Two- and three-dimensional profile interpretations indicate that the western contact dips 50 - 58° to the west, eastern contact 40° to the east, northern contact 30° to the north and the southern contact 38 - 40° to the south (Fig. 4.3-1). A small local gravity maximum is located between Väkkärä and Laitila rapakivi granites, obviously caused by Tarkki granite and older (Svecofennian) rocks, which extend to the depth of 830 m at minimum. Deeper in the interpretation model, the Väkkärä and the Laitila rapakivi granites coalesce. Using a minimum dip value 50° for the western contact and a lateral distance 2.5 km from Tarkki granite to Olkiluoto site, the minimum depth of the rapakivi at Olkiluoto is at least ca. 3 km, supposing that the rapakivi extends linearly so far (*see* Section 4.7).

Anorthosite gabbros, typically related to rapakivi granites, have not been observed in the study area. However, farther away, the Kolinummi anorthosite/leuco-gabbro is located 0.5 km to the SE from Laitila rapakivi, and it has also been studied by Elo et al. (1996). The gabbro is paramagnetic or very weakly ferrimagnetic (susceptibility 340 – $850*10^{-6}$ SI) and slightly denser than rapakivi (2690 – 2880 kg/m^3). Magnetic and gravity anomalies related to it are insignificant, 75 nT and 2.5 mGal. The southeastern contact of the Laitila rapakivi dips to the southeast. It is faulted by a NE-SW trending fault zone, which is clearly discernible on an aeromagnetic map.

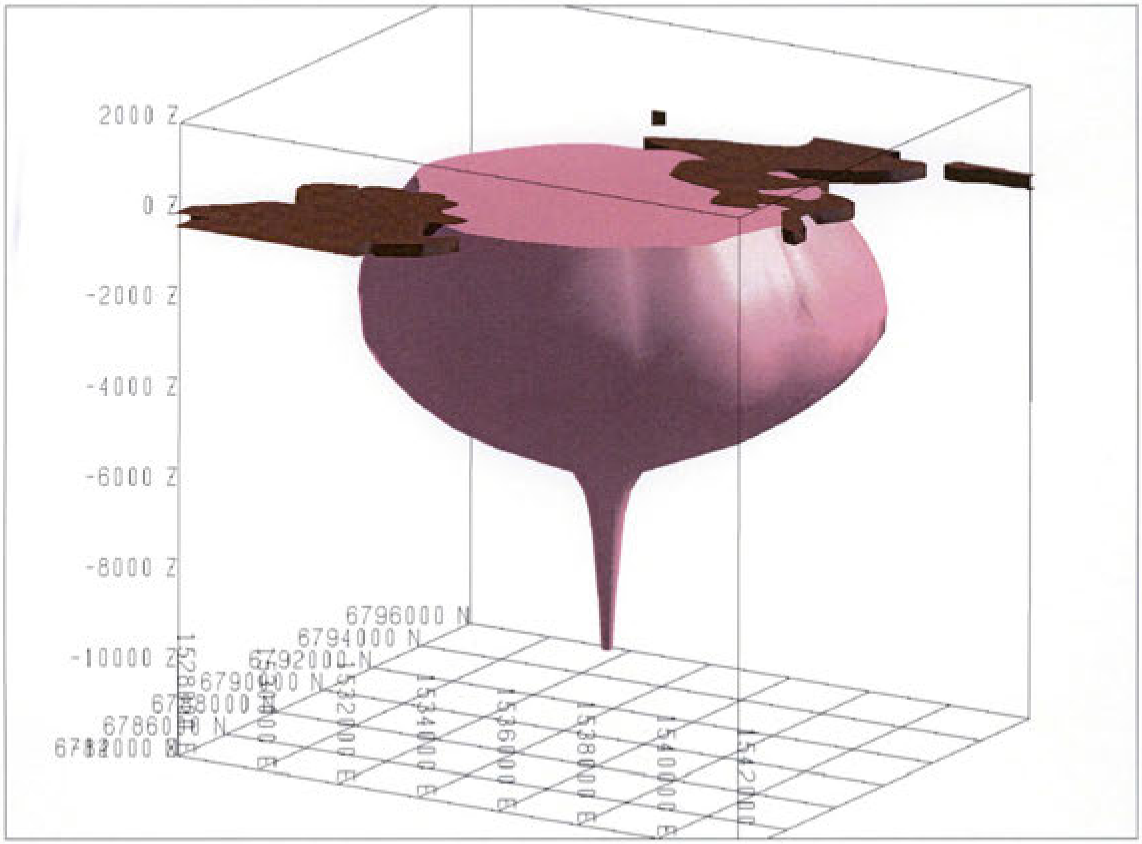


Figure 4.3-1. Interpreted three-dimensional structure of the Väkkärä rapakivi granite and cross-cutting olivine diabase sills, view from SSE (Paulamäki & Paananen 2001).

4.4 Satakunta sandstone

The Satakunta sandstone has been studied geophysically from 1960's using magnetic, seismic and gravity methods (Laurén 1970; Elo 1976, 1982; Elo et al. 1999). Locations of different survey lines are shown in Fig. 4.4-3. According to gravity interpretations, the Laitila rapakivi and Satakunta sandstone together induce a significant gravity low, which may also be effected by a major fracture zone below the sandstone. The interpretations suggest a depth of ca. 1 – 1.8 km for the base of the sandstone in the NW (Figs. 4.4-1 and 4.4-3). The sandstone thickens towards the NE, and it appears to have a steeply dipping, faulted contact with Svecofennian rocks. Near the northeastern contact, the sediments turn to moderately steep, ca. 35°, indicating activity of the bounding fault subsequent to the deposition of the formation. Towards the southwestern contact, the formation appears to shallow step-by-step (Laurén 1970). Three-dimensional structure of the southeastern part of the Satakunta sandstone is presented in Fig. 4.4-2.

On the GGT/SVEKA profile (Fig. 4.4-3), the structure of the sandstone was studied by integrating gravity interpretations, electromagnetic sounding results, refraction and reflection seismics and aerogeophysical data (Elo et al. 1999). An integration of different methods was needed, since the shallow sandstones in the SE are more porous and lighter than the deeper beds in the NW. According to the electromagnetic SAMPO soundings (Fig. 4.4-3), the lower boundary of the sandstone is at a depth of 500 ± 150 m. A seismic survey slightly to the SE from the GGT profile gave a depth of 210 m for the sandstone, nicely correlating with gravity results.

Since the density contrast between sandstone and rapakivi is rather low (*see* Table 4.1-1), the results of gravity interpretations are uncertain to some extent. Elo et al. (1993) studied the sandstone thickness between the lakes Köyliönjärvi and Pyhäjärvi using a gravity survey, and suggested a thickness of ca. 180 – 195 m. On the basis of regional gravity interpretations, the rapakivi extends below the sandstone. The deepest direct observation related to the depth of the sandstone is a 617 deep borehole, located 6 km to the south from Pori, which has not reached the bottom of the sandstone.

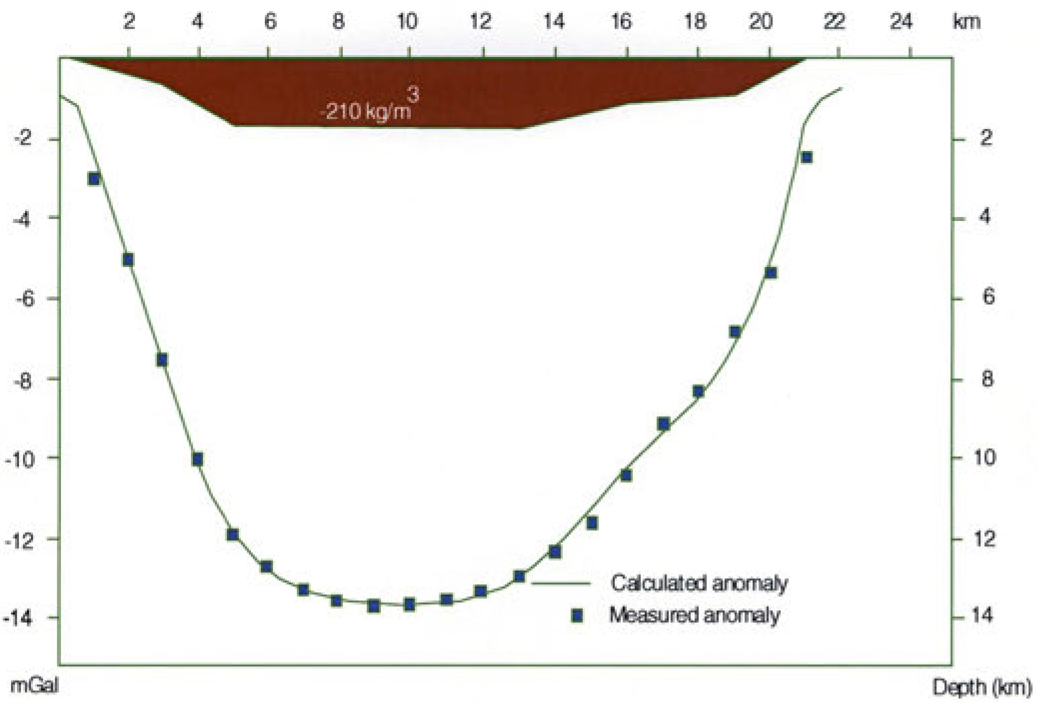


Figure 4.4-1. *The structure of the Satakunta sandstone on the basis of gravity interpretation (Elo 1976). Density contrast between the surroundings and the sandstone is –210 kg/m^3 in the interpretation. The profile location is shown on a map in Fig. 4.4-3.*

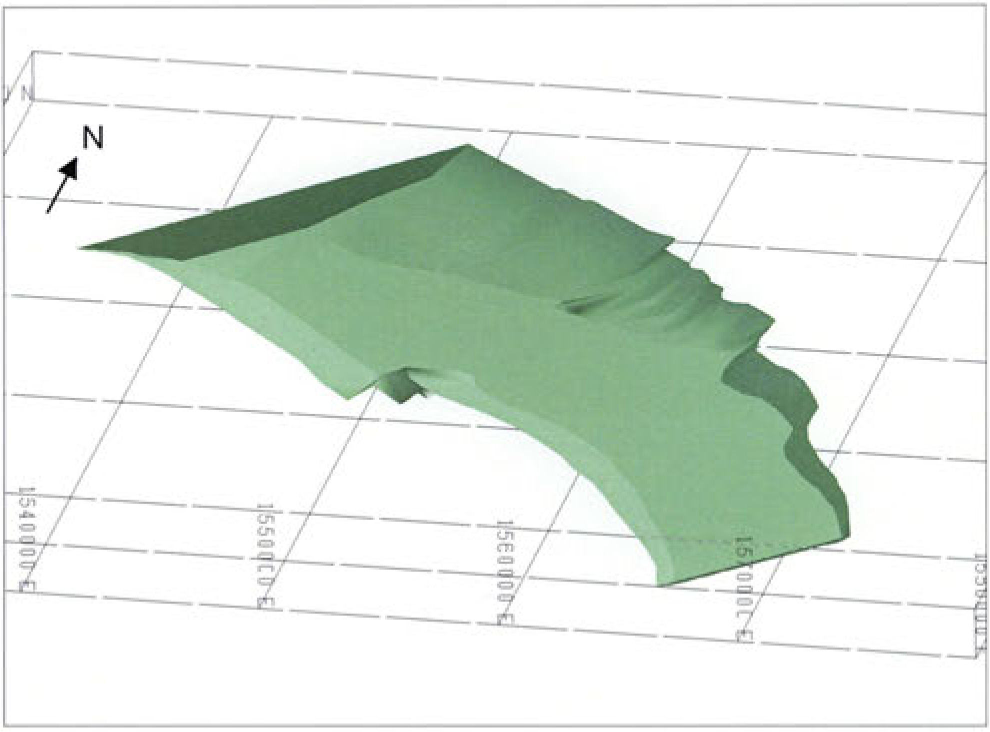


Figure 4.4-2. *Three-dimensional structure of the southeastern part of the Satakunta sandstone, view from SSE (Paulamäki & Paananen 2001). In the SE the thickness of the sandstone is ca. 200 m and in the NW ca. 1800 m.*

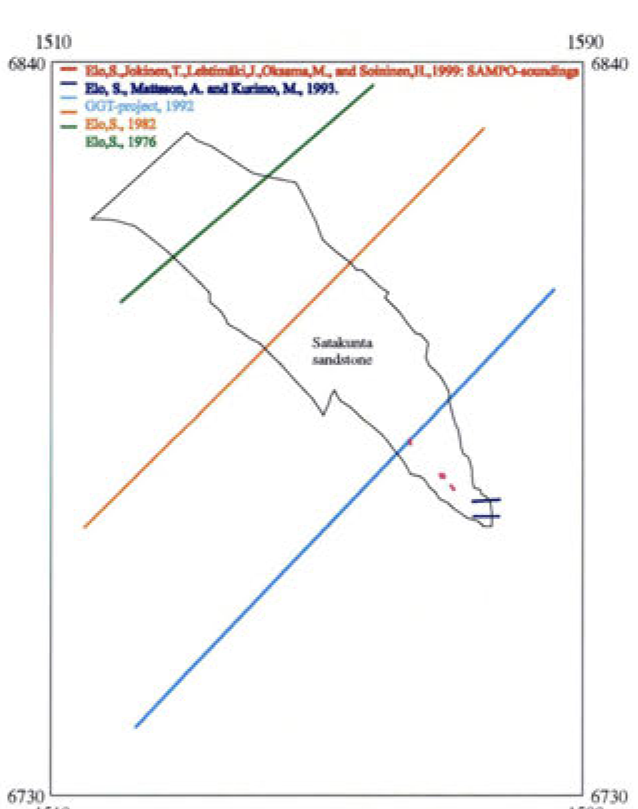


Figure 4.4-3. *Locations of different survey lines related to the Satakunta sandstone.*

4.5 Postjotnian diabases

In Satakunta, the Postjotnian olivine diabases occur as steep dykes and large horizontal sills. In the rapakivi areas, the diabases can be observed as cross-cutting, several hundreds of metres wide dykes, but closer to the sandstone basin, sills become more common (Hämäläinen 1987). In sandstone regions, olivine diabase covers large areas.

Petrophysically, the Satakunta diabases are clearly ferrimagnetic (Table 4.4-1), and they also appear to carry significant remanent magnetisation, differing from Earth's field (Neuvonen 1965, Pesonen 1992). Negative inclination is typical (I = -30°, D = 40°), resulting, together with induced magnetisation, in strong side minima at the northern – eastern sides of the maxima on the magnetic map. However, the direction of remanent magnetisation may vary spatially. On the magnetic map, the diabase dykes and sills can be detected as clear anomalous regions, and especially the edges of the diabase sills are well discernible. The average thickness of the diabase sills is ca. 300 m.

The geometry of diabases in the eastern part of the sandstone area (map sheet 1134) has been studied at the Geological Survey of Finland by a magnetic and a gravity survey (Kurimo et al. 1992a, b; Elo et al. 1993). The results suggest moderately gentle dips (averagely ca. 40°) to the west in the eastern part of the survey area, indicating possibly bowl-shaped structures.

Gravity interpretations on the northern side of the Eurajoki rapakivi stock gave depth estimations for two diabase sills (Elo 2001). According to the results, the depth of the northern and southern sills are 150 m and 300 m, respectively.

In the southern part of the study area, nearly north – south trending narrow anomaly zones, cross-cutting the olivine diabase sills, can be observed. These features are supposed to be diabase dykes younger than Postjotnian (Veräjämäki 1998).

4.6 The Sääksjärvi meteorite crater

The Sääksjärvi impact crater is characterised by a circular gravity low of – 6,5 mGal (Fig. 4.6-1). Furthermore, the structure is discernible on aerogeophysical maps as a magnetic and resistivity minimum. According to the petrophysics, rather low resistivity (10 – 50 ohmm) and density values, as well as high porosities, are attributed to the uppermost part of the structure (Elo et al. 1992).

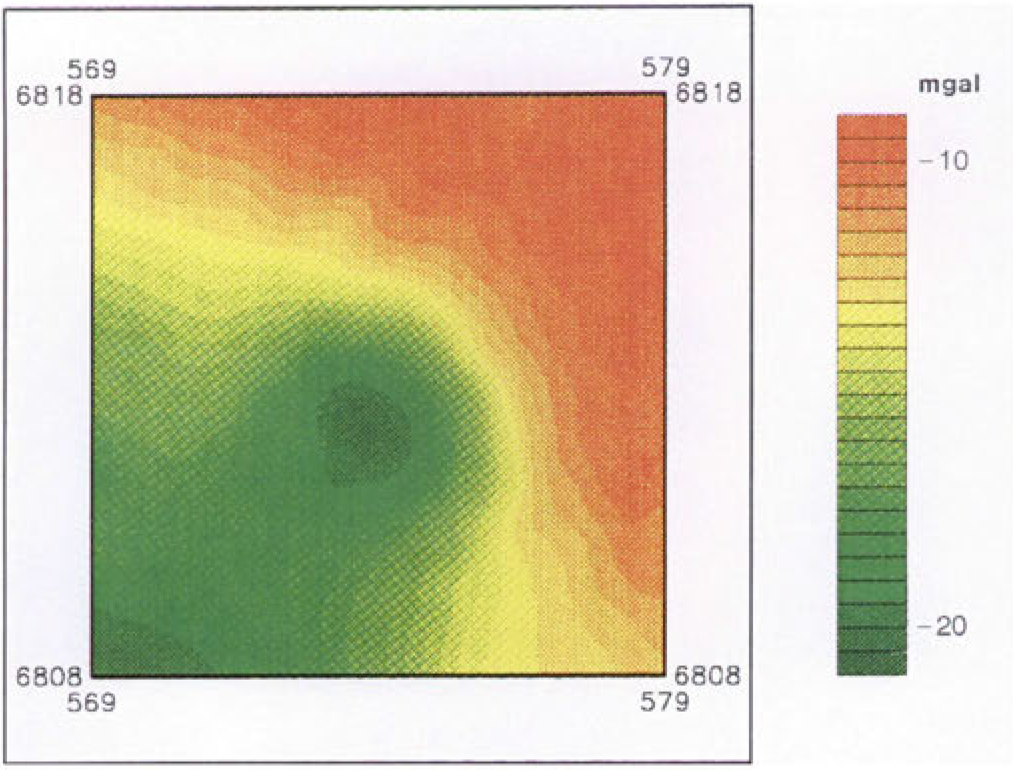


Figure 4.6-1. Bouguer anomaly map of the Sääksjärvi meteorite crater and its surroundings (Elo et al. 1992).

4.7 Gravimetric investigations in Eurajoki and Olkiluoto

During the summer 2000, a gravity survey was carried out in Eurajoki and Olkiluoto. The aims of the survey were to study the contacts of the Eurajoki rapakivi stock and its possible subsurface extension into the Olkiluoto island, to estimate the thickness of the diabase sills on the northern side of the Eurajoki rapakivi stock, and to obtain information on the nature of the granites and gneisses in Olkiluoto. This summary is based on the report by Elo (2001).

In total, 1906 gravity stations were measured along six separate profiles with a total length of 39.84 km. A new Bouguer anomaly map of Olkiluoto and Eurajoki (Fig. 4.7-2) was prepared by combining the new and old data (Fig. 4.7-1). Some auxiliary maps were prepared to help in the interpretation. The new data refines the earlier investigations (Elo 1981, 1982) and makes the interpretations less speculative.

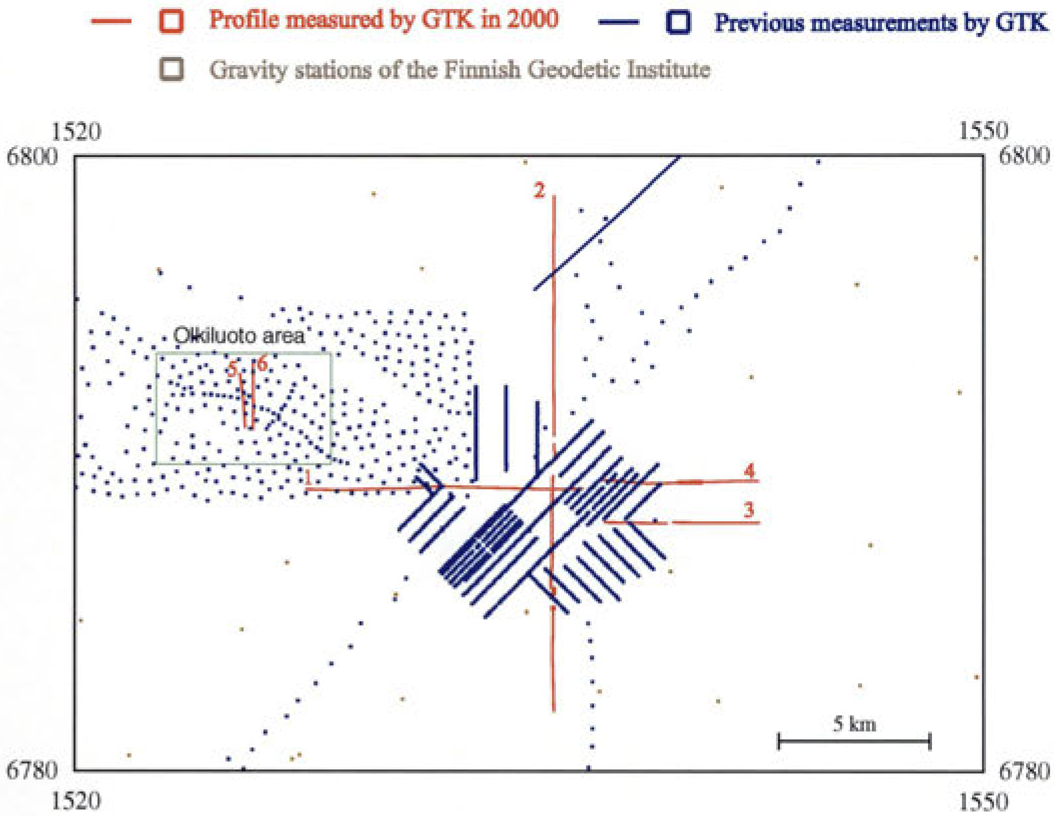


Figure 4.7-1. *Gravity measurements in Olkiluoto and Eurajoki.*

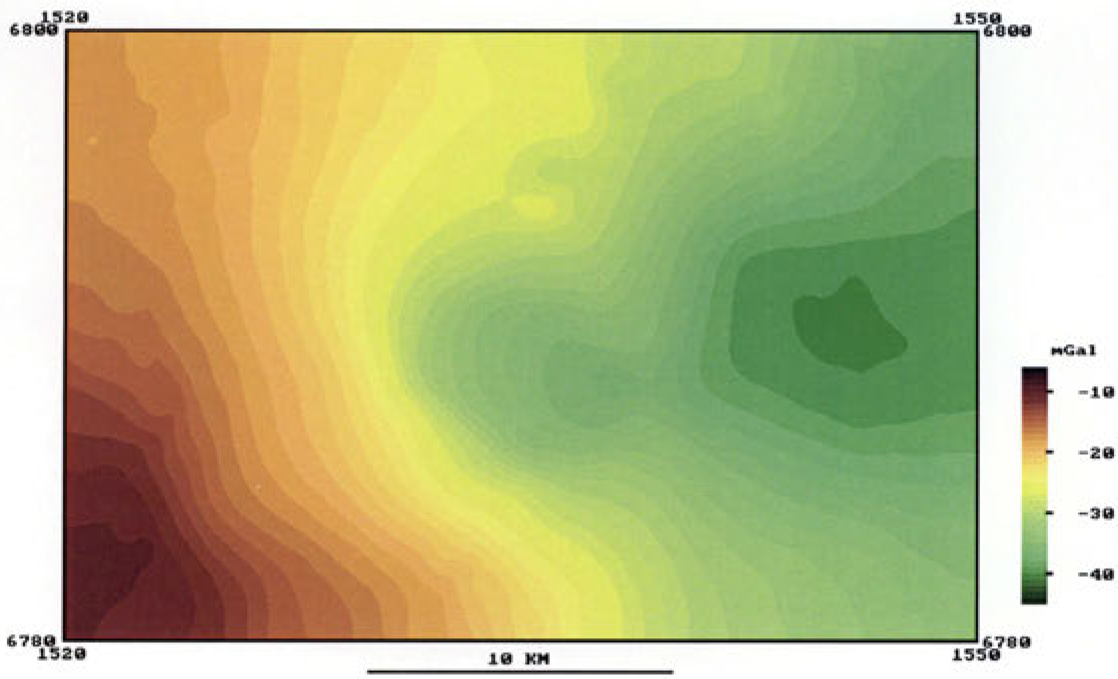


Figure 4.7-2. *Bouguer anomaly map of Olkiluoto and Eurajoki.*

Table 4.7-1 is a summary of the rock densities on which the interpretation of the gravity anomalies relies. In modelling the profile 1, the density contrast was among the parameters that were optimised. By adding the solved density contrast (93 kg/m^3) to the mean density of the Väkkärä granite a mean density of 2703 kg/m^3 is obtained for the surrounding bedrock. This value fits very well between the mean density of the Tarkki granite and the mean density of the drill cores from Olkiluoto.

Table 4.7-1. *Summary of rock densities (kg/m3) at Olkiluoto and Eurajoki.*

Rock type	**mean**	**st.dev.**	**number of specimens**	**reference**
Väkkärä granite	2610	20	63	Elo, 1981
Tarkki granite	2690	20	5	"
Diabase	3000	30	21	"
Laitila rapakivi granite	2640	10	12	Elo, 1982
OL-KR1...KR5 (Drill holes at Olkiluoto)				
All	2718	65	255	Lindberg & Paananen, 1990
granite	2635	30	53	"
mica gneiss/migmatite	2740	45	190	"

Based on two- and three-dimensional gravity modelling, the depth extent of the Väkkärä granite is at least 3 km and its contacts dip outwards. Between the Väkkärä and Laitila rapakivi massifs, there is a slice of the old bedrock, but apart from that, the two massifs are closely associated. At the Väkkärä location, there are rapakivi granites at deeper levels than 3 km from the surface. However, it is difficult to say, whether the Väkkärä granite extends to underneath the Laitila granite or vice versa or whether it is a question of a new unnamed massif underneath the both. According to more regional gravity interpretations, rapakivi granites occupy a much larger area, especially under the Satakunta sandstone formation, than can be delineated by means of the outcrops. The granite bodies are at some locations more than 8 km thick and extend at depth to underneath the Olkiluoto island. The Bouguer anomaly continuously increases from the Väkkärä granite to the Olkiluoto area, and there is no indication of rapakivi massifs close to the surface at Olkiluoto. The depth to the upper surface of possible rapakivi granite massifs at Olkiluoto is at least 3 km. The maximum thickness of the diabase sill on the northern side of the Väkkärä granite is at least 300 m. The modelling along the profile 2 is shown as an example in Figure 4.7-3.

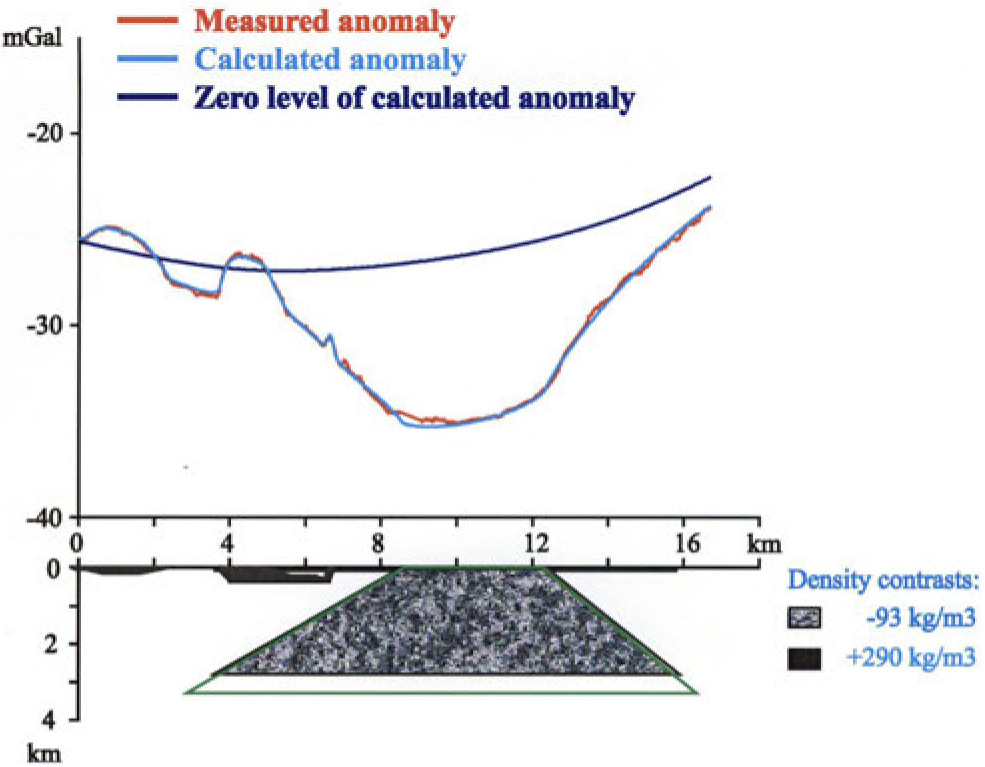


Figure 4.7-3. *Interpretation of the Bouguer anomaly profile 2 (Fig. 4.7-1). The two-dimensional model of the rapakivi granite is shown in grey, the corresponding three-dimensional model is outlined in green, and the diabases on the northern side of the rapakivi granite are shown in black. The profile starts at the map co-ordinates of x1=6798.67 and y1=1535.70 and ends at the map co-ordinates of x2=6781.98 and y2=1535.81 km.*

The residual anomalies along the profiles measured across Olkiluoto are very small. This means that at least the topmost part of the bedrock has undergone migmatisation and no larger homogeneous granite blocks exist along the profiles.

4.8 Fracture zones

Brittle as well as ductile deformation of the study area is already widely discussed in Chapter 3.1. This chapter summarizes the interpretations of aerogeophysical data of the area.

The northern part of the study area is already thoroughly interpreted by Pajunen et al. (2001). According to their results, five separate magnetic (and/or topographic) lineament trends can be found in the Pori area. The directions of these trends are as follows:

1) N 10 - 20° E : brittle/semiplastic (boundaries of the diabase sills)
2) N 40 - 50° E: brittle/semiplastic (boundaries of diabase sills, cuts the sandstone)
3) N 100 - 110° E: parallel to the sandstone graben
4) N 140 - 150° E: plastic, occasionally brittle, parallel to the Kynsikangas zone
5) N 80° E: brittle.

Interpretation of the southern part of the study area is done as a part of this work. For the interpretation procedure, the aeromagnetic map was further processed by calculating horizontal and vertical gradients, by shading from different directions and by varying the colour scale of the map. Furthermore, the digital elevation model of the area was utilized.

The results of interpretations are presented on a map in Appendix 6. The map shows that almost the same lineament directions, interpreted by Pajunen et al. (2001) in the northern part, can be found in the southern part and also in the near surroundings of Olkiluoto. The most dominating directions are ca. N 100 – 120° E (red colour), N 140 - 150° E (magenta), N 40 - 65°E (orange) and N 0 - 30° E (blue). However, they are different in character: especially on the Rauma map sheet in the SW, the NW-SE and NE-SW trending structures are interpreted to be related to brittle deformation, whereas the most significant structures of the ductile deformation occur in nearly E-W direction.

5 DEEP STRUCTURE OF THE CRUST IN SOUTHERN SATAKUNTA

In order to find out about the deep structure of the Fennoscandian shield, deep seismic refraction surveys were carried out already from the 1960's onwards, and with the aid of them it has been possible to gather a comprehensive data set on the thickness of the crust, correlating the results also with gravity data (Penttilä 1972, Luosto 1987a, 1987b). From the older seismic survey lines, the Sylen – Porvoo and the Trans-Scandinavian profiles are most clearly related to the study area (Luosto 1987a). Of these, line Sylen – Porvoo is in WNW – ESE direction so that the seismic sources could be located at sea and the receivers on land. The line runs via Pyhäranta across the Laitila rapakivi massif, and two observation stations are located in the rapakivi region. According to the interpretation (Luosto 1987a), the crustal thickness between Pyhäranta and Porvoo is rather constant, ca. 44 – 49 km. Below the Laitila rapakivi massif, the interpreted depth of the Moho is ca. 46 km.

The reflection seismic survey lines of the BABEL project, measured in 1989, are located in the Bothnian Sea to the west from the study area (lines 1, 6 and 7, Babel WG 1993). The structures and processes of the bedrock in southwestern Finland have also been studied in the international GGT/SVEKA project, covering also the Satakunta region (Fig. 5-1) The GGT/SVEKA transect is 160 km wide and 840 km long, covering the area from Ahvenanmaa to Kuhmo in a NE – SW direction. The project has integrated existing geophysical (seismic, electric, gravimetric, thermal and petrophysical) and geological (lithology, structural geology, metamorphism, isotopes and geochemistry) data (Figs. 5-3 to 5-5). On the basis of these results, the deformation history and structures of the bedrock within the transect have been studied by Korja et al. (1993), Korja & Heikkinen (1995), Korsman et al. (1999), Korja et al. (2001) and Heikkinen et al. (2000).

Along the GGT/SVEKA transect, the average crustal thickness is 52 km, according to seismic results, and beneath the Laitila rapakivi massif the thickness is 47 km (Korsman et al. 1999). The crust thickens gradually to the NE, reaching its maximum 60 km in a broad zone between lake Ladoga and Bothnian Bay (Fig. 5-2). The variations of crustal thickness are controlled by a layer of high seismic velocity ($V_p > 7$ km/s) in the lower crust, with a thickness between 7 and 25 km. In southwestern Finland, its thickness is interpreted to be ca. 14 – 22 km, and at Olkiluoto 16 km (Korja et al. 1993). In the regions of Svecofennian rocks and rapakivi granites, the layer of high seismic velocity in the crust is located at a depth of ca. 34 – 35 km. In Figure 5-3, the distribution of seismic P-wave velocity in southwestern Finland along GGT/SVEKA transect is presented. The profile covers the area from Åland to Tampere schist belt across the Laitila rapakivi massif and the Satakunta sandstone.

In the regions of thinner crust, the lower crust has a highly reflective lamellic structure and a strongly reflective, subhorizontal Moho (boundary between the crust and the mantle) (Korja et al. 1993). The lower crust (depth > 35 km) with its high P-wave velocity (>7 km/s) and high reflectance indicate increase of mafic material from the mantle to the crust (Korja 1995). The mechanism of this increase has been mafic underplating. Below the thinnest regions of the crust, a mafic layer of eclogite may occur.

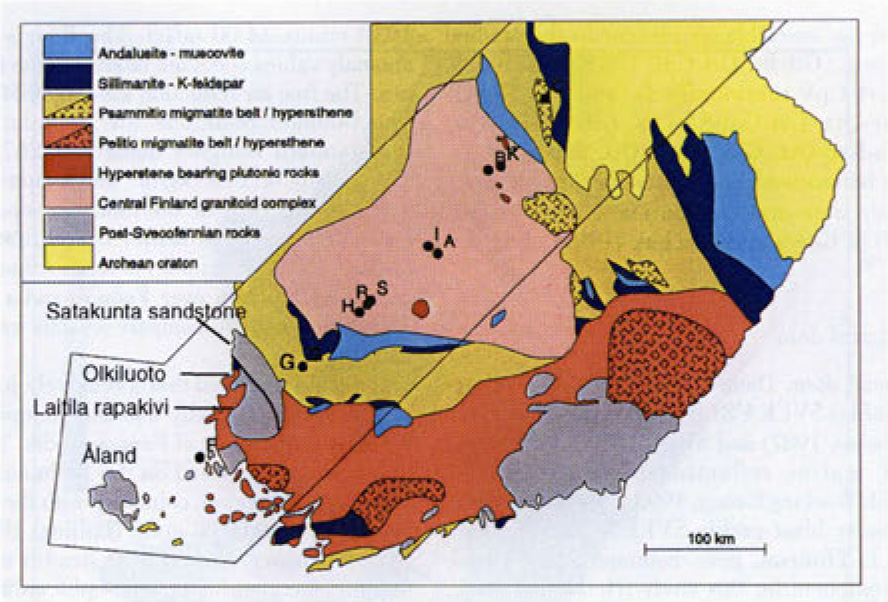


Figure 5-1. *Location of GGT/SVEKA transect (Korsman et al. 1999). Black dots mark the locations of seismic sources.*

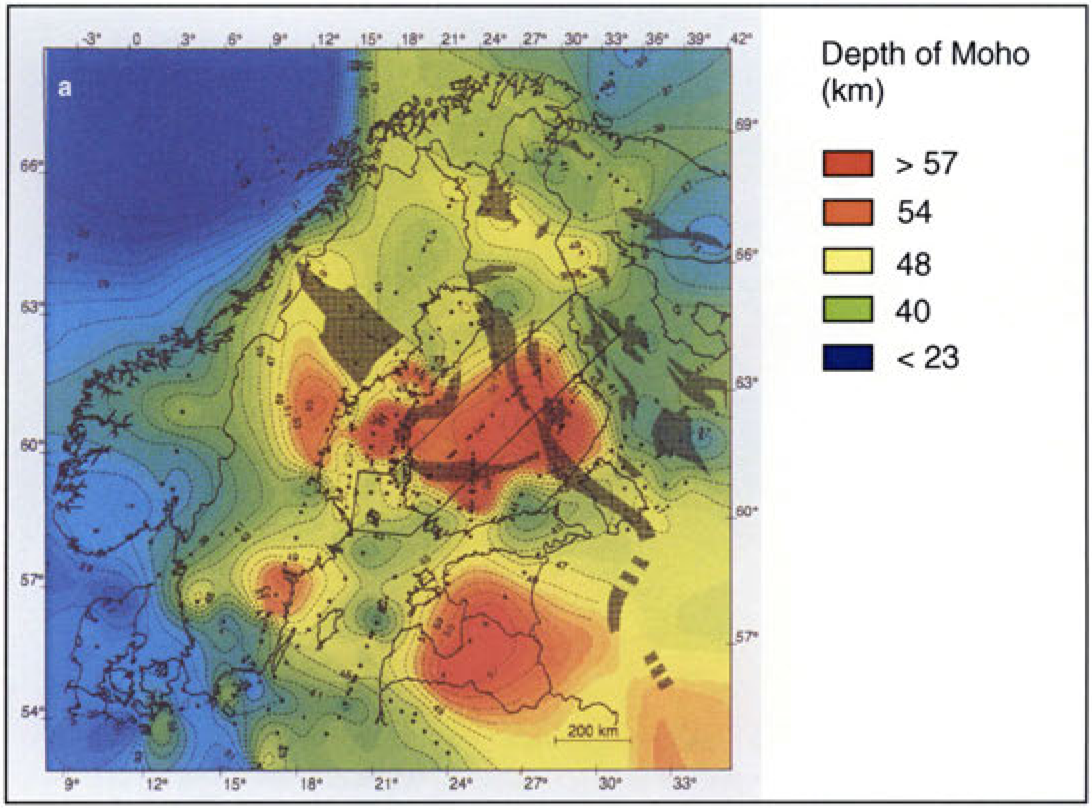


Figure 5-2. *Interpolated depth of Moho in Fennoscandia (colours and contours). Rasterised areas mark the locations of electric conductors in the crust (Korsman & Korja 1999). GGT/SVEKA transect is outlined by a polygon.*

In addition, the relation between crustal thickness and gravity anomalies has been studied by Elo (1994, 1997, 1999) and Korsman et al. (1999). The comparison indicates that variations in crustal thickness do not appear to induce corresponding gravity anomalies. The thickening of the crust by 10 – 20 km should cause a gravity minimum of 100 – 200 mGal. However, in the regions of thick crust, a significant gravity maximum has only been encountered occasionally. This may be due to thickening of the anomalous lower crust in relation to the upper crust, mafic intrusions or up-thrusting of blocks of higher metamorphism.

A significant feature on the gravity map of Fennoscandia is a large minimum around the Bothnian Bay and the Bothnian Sea, correlating with land uplift. It has a glacio-isostatic component of long wavelength (amplitude 20 mGal, half value width ca. 600 km) with a source located in upper mantle (Elo 1997, 1999). Wang (1998) has compiled a density model of the crust according to seismic velocities, and a computed geoid on the basis of the model. The calculated geoid minimum is about double that observed, which is interpreted to be caused by significant mass reduction in the crust and mass excess in the mantle. However, it is uncertain how accurately seismic velocities can be converted to densities.

Electrically, the crust consists of several electrically conducting or resistive homogeneous regions, bounded by conducting zones in the upper and middle crust. According to Korja et al. (1993), the old Archean crust has been thickened by emplacement of 2200 – 2100 Ma old mafic dykes and 2500 – 2000 Ma old, electrically conducting volcanites and metasediments. However, the most significant thickening is related to the Svecofennian orogeny, the collision of the Svecofennian island arc and the Archean continent, resulting in a collage of metasedimentary rocks squeezed between crustal blocks of the island arc. The boundaries of separate blocks can be detected as strong seismic reflections and/or vertical or tilting electric conductors, which extend at least to the middle crust. One of these conductive zones runs across southern Finland in an E-W direction to Pori (Fig. 5-2). On the coast, the zone makes a curve to the north towards Kokkola, from where it extends to Skellefteå area, Sweden. The zone is several tens of kilometres wide and ca. 500 km long, and it is characterised by numerous aerogeophysical anomalies related to surficial conductors. According to geophysical modelling, the conducting zone is nearly vertical and extends to the middle crust. On the basis of the magnetotelluric interpretations of the SVEKA profile (Korja & Koivukoski, 1990, 1994), the top of the conductor is located at a depth of 3 km. According to Korja (1990, 1993), Korja et al. (1993) and Korja et al. (1996), the conductor marks the Svecofennian collision zone between the continental arc in the central Finland and the island arc complex in southern Finland. On the basis of the correlation of airborne electromagnetic results and geological information, the conductor is related to graphite- and sulphide-bearing metasediments. It was formed when these rocks were tilted and deeply buried in a collision between two crustal terranes.

The thickness of the crust below the Bothnian Sea varies from 50 km to over 60 km along a survey line extending from Åland to Umeå, Sweden. Within this area, strong reflections occur in the upper crust, which are interpreted to be diabase sills (BABEL WG 1993). On the basis of outcrop observations in Sweden and the shape of the reflectors, these diabase sills are interpreted to form several bowl-shaped units related to each other. Their diameter may be up to 100 km. Above and below these strong reflectors, weaker reflections have been observed, indicating possibly Jotnian sandstones. Accordingly, sandstone formations with the thickness of several kilometres may exist within this region of the Bothnian Sea.

Heat flow in southwestern Finland along GGT/SVEKA transect is c. 50 mW/m^2, decreasing towards the NE. The temperature in the crust at a depth of 30 km is ca. 350 °C, whereas in the lower crust temperature is over 430°C at pressures over 12 kbar (Kukkonen 1994). At the interface of the crust and the mantle (Moho), the temperature is ca. 500°C (Fig. 5-4).

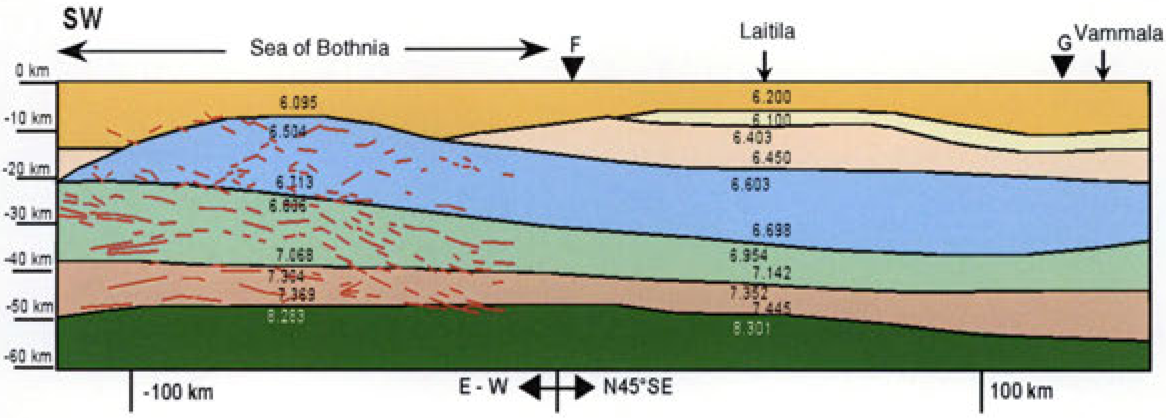


__Figure 5-3.__ Seismic P-wave velocity model (km/s) in the southwestern Finland, a portion of the GGT/SVEKA transect (Korsman et al. 1999). Red lines correspond seismic reflectors interpreted to represent fault zones. Location of the profile is shown in Fig. 5-1.

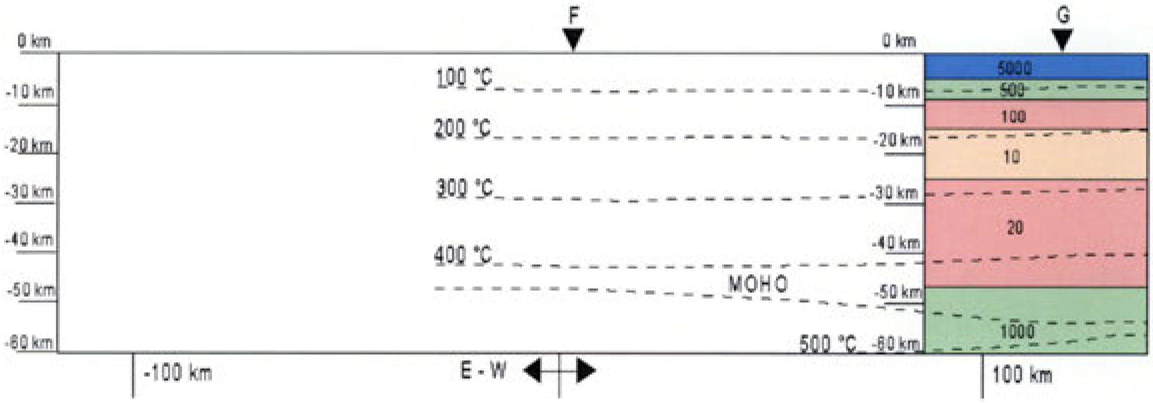


__Figure 5-4.__ Distribution of electric resistivity (ohmm) and temperature in the crust in southwestern Finland, a portion of the GGT/SVEKA transect (Korsman et al. 1999) Location of the profile is shown in Fig. 5-1.

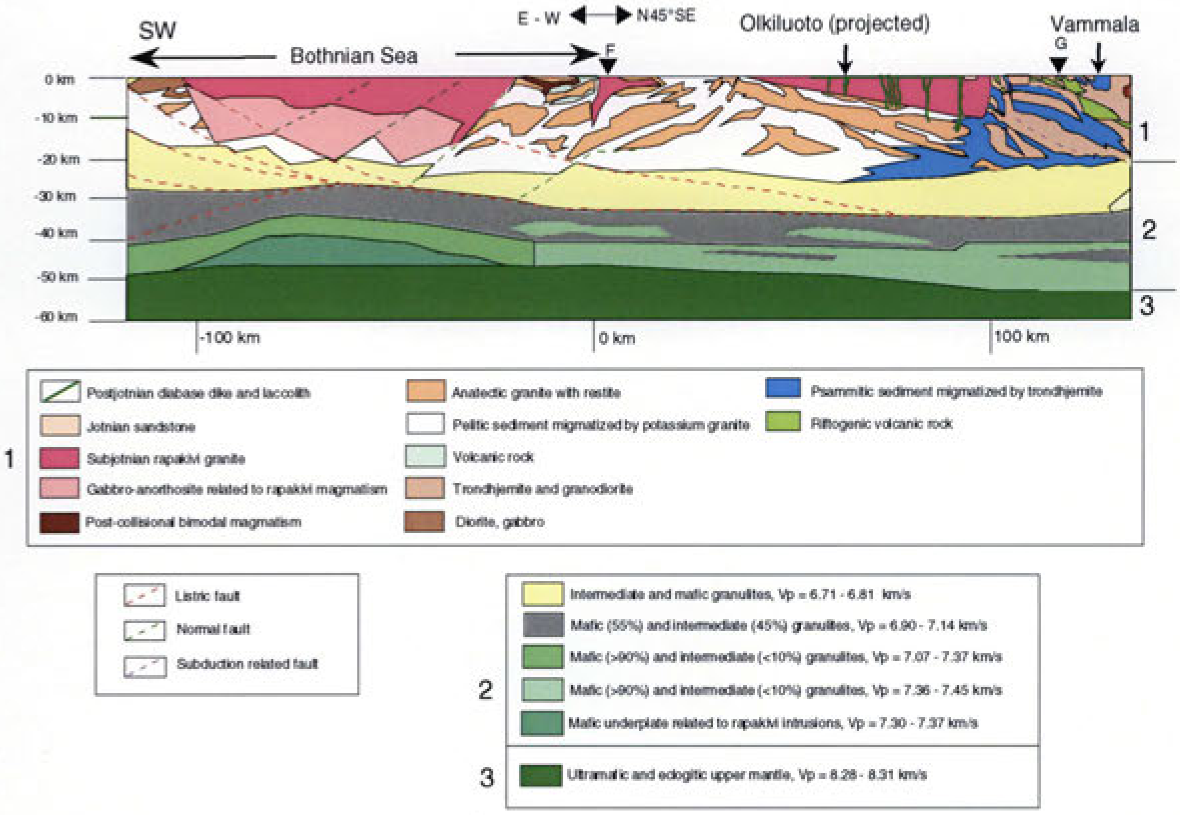


Figure 5-5. *Deep structure of the crust in southwestern Finland, a portion of the GGT/SVEKA transect (Korsman et al. 1999).* V_P *= seismic P-wave velocity. 1= upper crust, 2 = lower crust, 3 = mantle. Location of the profile is shown in Fig. 5-1.*

Magnetic character of the Tampere schist belt and the psammitic migmatite belts (*see* Figs. 2.1-2 and 6.2-1) is rather steady. The values vary usually within 200 nT, being typically -200…-400 nT (Korsman et al. 1999). However, two strongly anomalous zones with east-west direction can be observed across southern Finland. From these, the southern one (+400 nT) coincides with the southern Finland complex, and the northern (+350… +400 nT) with the northern part of the Tampere schist belt and, in particular, with the southern part of the central Finland granitoid complex. A regional minimum (-250 nT) is located between these anomalies in the southern part of the Tampere schist belt. This minimum coincides with graphite- and sulphide-bearing schists that form a continuous electrically conductive zone across southern Finland. According to Lahtinen and Korhonen (1996) the sources of magnetic anomalies are located rather deep in the crust. Magnetic profile interpretations (Pernu et al. 1989) suggest that volcanic rocks in the northern part of the Tampere schist belt are almost vertical, and they extend to the depth of several kilometres. Also the magnetotelluric results indicate that the contact between the Tampere schist belt and the central Finland granitoid complex is nearly vertical. According to petrophysics, magnetisation of the rocks is mainly related to magnetite.

The rapakivi granites in southern Finland, like the Åland and Wiborg massifs, appear to be characterised by a gravity minimum, surrounded by a broad maximum. It is likely that the maximum is caused by uplifted upper mantle deep below the rapakivi, whereas the more local minimum is associated with the low-density rapakivi in the upper crust (Korsman et al.1999).

6 GEOLOGICAL EVOLUTION OF THE SATAKUNTA AREA

6.1 Plate tectonic interpretation of the Precambrian in Finland

Since the pioneering work of Anna Hietanen (1975) the plate tectonic theory has been widely applied in explaining the evolution of the Proterozoic crust in Finland (for example, Gaál & Gorbatschev 1987, Gaál 1990, Ekdahl 1993, Lahtinen 1994, Korja 1995, Ruotoistenmäki 1996, Nironen 1997)

The Earth's crust and the upper mantle together form a lithosphere, which may be up to 70 km thick if composed of oceanic crust or 150 km incorporating continental crust. According to the plate tectonic model, the lithosphere today is divided into a series of relatively thin, but rigid plates, six large and a number of smaller ones, which are in constant motion relative to one another upon a plastic layer in the mantle, the asthenosphere. The driving force of the plate motion is provided by convection currents in the mantle, powered by heat generated deep inside the earth's core by radioactive decay. The surface layer of each plate is composed of oceanic crust, continental crust or a combination of both. The relative motion of the plates is a few centimetres per year.

Most of the Earth's tectonic, seismic and volcanic activity occurs preferentially at the boundaries of neighbouring plates. There are three types of plate boundaries: 1) divergent boundaries, where the plates pull away from each other, as new oceanic crust is generated at mid-ocean ridges from magma rising from the asthenosphere, 2) convergent boundaries, where crust is destroyed as one plate dives under or collides with another, and 3) transform boundaries, where crust is neither produced nor destroyed as the plates slide horizontally past each other (Fig. 6.1-1). The new oceanic crust is carried away from the ridge by convection cells in the mantle.

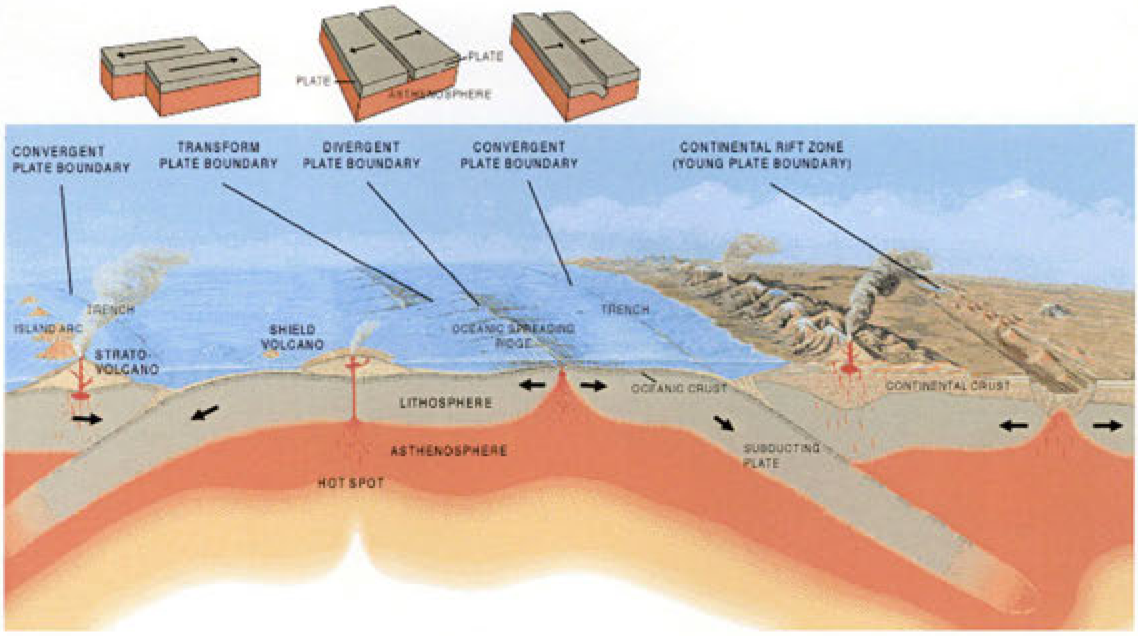


Figure 6.1-1. Types of plate boundaries (Kious & Tilling 1996).

The type of convergence that takes place between plates depends on the kind of lithosphere involved. Convergence can occur between an oceanic and a largely continental plate, or between two largely oceanic plates, or between two largely continental plates. In oceanic-oceanic and oceanic-continental convergence the oceanic plate is pushing into and being subducted under the other plate. The location where sinking of a plate occurs is called a *subduction zone* (Fig. 6.1-1). The area of subduction is marked by deep ocean trench. As one of the plates is subducted beneath the other it sinks back to the asthenosphere, and the resulting magma rises to the surface above the subduction zone to form a chain or arc of volcanoes (island arc in case of ocean-ocean convergence). When two plates with continents at their leading edge collide, neither is sunk because the continental rocks are relatively light. Instead, the crust deforms and pushes upward to form a range of mountains, such as the Alps or the Himalayas.

The plate tectonic cycle can be divided into three episodes: 1) rifting, 2) separation of the plates, and 3) subduction/collision. In the early stages of a rifting episode, the crust rises to form a dome (Burke & Dewey 1973). As the crust breaks three rifts develop at the crest of the dome. They come together at a central point called the triple junction. As such triple junctions develop, one of the three arms becomes inactive, while the remaining two continue to separate. The boundary between the two diverging plates develops into the mid-ocean ridge, where new oceanic crust is formed from the magma rising from the asthenosphere. The plates are constantly moving apart from the ridge, as the new sea floor is developed. In the last stage of the plate tectonic cycle the oceanic crust dives back into the mantle in the subduction zone. The sea closes and eventually the continents collide to form a mountain range.

The similarity between ancient orogenic belts and the young mountain ranges suggests that the same kinds of plate tectonic processes are involved in their development. According to the present conception, the Fennoscandian Shield was created during two major periods of crustal growth, the first in late Archaean times and the second during the culmination of the Svecofennian orogeny ca. 1900 Ma ago. The present model for the evolution of the Svecofennian crust involves the formation of two pre-Svecofennian island arcs. The first arc was accreted onto the Archaean craton, and the second one was accreted onto the first arc.

The oldest recognisable continental crust, characterised by tonalitic-trondhjemitic-granodioritic plutonics, was developed 3100 - 2900 Ma ago during the Saamian orogeny (Gaál & Gorbatschev 1987). Some kinds of plate tectonic processes are believed to be affected already at such an early stage of the crustal development (Gaál & Gorbatschev 1987, Ruotoistenmäki 1996). The Saamian crust may have been developed from a tholeitic protocrust.

The rocks of the main orogenic episode of the Archaean Karelian Domain, the Lopian orogeny, 2900 - 2660 Ma in age, include tonalitic to granodioritic gneisses and greenstone belts, which are intruded by granitoids with tonalitic to granitic compositions. The earliest magmatic and metamorphic event took place at around 2840 Ma (Luukkonen 1985). The greenstone sequences, containing tholeitic and komatiitic volcanics, were formed after this event, and were intruded by tonalitic to granitic granitoids between 2750 – 2690 Ma (Martin et al. 1984, Luukkonen 1988). The greenstone belts have been interpreted to represent modern island arcs or a back-arc rift (Gaál & Gorbatschev 1987, Luukkonen 1992).

The onset of a new rifting after the Lopian orogeny is represented in Lapland by bimodal komatiitic and felsic volcanics dated at around 2500 Ma, which unconformably overlie the Archaean basement (Lehtonen et al. 1998). About 2440 Ma ago, a global period of enhanced thermal activity resulted in voluminous mafic magmatism, recorded now by layered gabbro intrusions in northern and northeastern Finland (Alapieti 1982), and coeval mafic dykes (Vuollo 1994). In eastern Finland, the Sumian-Sariolan, Kainuan, Jatulian and Karelian platform to continental margin-type sediments were deposited, within the time period of 2500 - 2000 Ma, onto the weathered and peneplaned Archaean crust. In Lapland, the sediments corresponding to Jatulian to Lower Kalevian include epicontinental clastic sediments and deeper-water pelitic sediments (Hanski & Huhma 2000).

The Sumian-Sariolan metasedimentary and metavolcanic rocks were evidently deposited in narrow intracratonic rifts. The Kainuan metasediments have been interpreted as being deposited either in a narrow sea or in an intraplate sea, whereas the Jatulian rocks, lying unconformably on the Kainuan, are thought to represent fluvial and continental shelf-type sedimentary environment of the open sea stage (Laajoki 1991). The Jatulian metasediments and the diabase dykes cutting them have been interpreted as indicating a change from a cratonic environment to a passive continental margin in an opened pre-Svecofennian ocean (Gaál 1990, Lahtinen 1994). The continued spreading and rifting of the crust are demonstrated by mafic and ultramafic volcanics (e.g., Central Lapland greenstone belt) and the karjalitic and tholeitic diabase dykes, dated at 2200 - 1970 Ma (Vuollo 1994). According to Lahtinen (1994), the 2100 Ma old tholeitic diabase dykes represent the main rifting episode of the Archaean continent, 2100 Ma ago.

The depositional age and the sedimentary environment of the Kalevian group are uncertain, because the rocks are strongly deformed. However, they represent a change to a deeper water environment and seem to have been deposited in different environments, such as rift valley, passive margin and foredeep basin (Ward 1987, Korkiakoski & Laajoki 1988, Kohonen 1995). According to Kohonen (op. cit.) the deposition of the Kalevian group in Northern Karelia began 2100 Ma ago, when the Archaean continent was breaking.

The 1970 - 1950 Ma old Jormua and Outokumpu ophiolite complexes, closely associated with the Kalevian group, are interpreted as remnants of ancient, Palaeoproterozoic oceanic crust. Rifting of the Archaean crust culminated in the formation of oceanic crust, fragments of which were subsequently thrust onto the Karelian craton (Kontinen, 1987). Geochemical characteristics of the Jormua ophiolite suggest early stages of rifting, and show no indications of subduction (Peltonen et al. 1996). Consequently, both passive continental margin and intracontinental rift environment have been proposed as the tectonic environments for the Jormua and Outokumpu ophiolites (Kontinen 1987, Ward 1987, Nironen 1997, Korsman et al. 1999).

The break-up of the Archaean crust and opening of the pre-Svecofennian ocean seems, thus, to have taken place either in two separate stages, 2100 Ma and 1950 Ma ago, or the spreading related to the earlier rifting stage led to the formation of intracontinental sea near the passive margin, the ophiolites representing a marginal rift basin (Nironen 1997, Korsman et al. 1999). According to Korsman et al. (op. cit.), in the case of two separate stages of ocean opening, an unknown orogeny between the rifting stages would be required, which would represent the amalgation of the Palaeoprerozoic protocrust with the remnants of the Archaean continent. The second rifting phase would then indicate the opening of the new ocean at the same site.

After the rifting of the Archaean crust, 2100 Ma or 1950 Ma ago, a pre-Svecofennian ocean with island arcs was formed. The oldest Svecofennian rocks, 1930 - 1920 Ma old gneissic tonalites and volcanites, belonging to the Pyhäsalmi primitive island arc at the boundary between Archaean area and Svecofennides (see Fig. 2.1-3) were formed during a westward dipping subduction (Lahtinen 1994, Kousa et al. 1994). The chemical and isotope composition of the tonalites suggest a mantle origin, and that the rocks were formed in an island arc environment with relatively thin crust (Lahtinen 1994, Lahtinen & Huhma 1997).

The Central Finland continental arc and the Southern Finland sedimentary-volcanic complex (*see* Fig. 2.1-2) were formed when the early Svecofennian protocrust rifted before 1910 Ma and subduction zones developed at both the northern and southern margins of the opening ocean. The Svecofennian orogeny can be divided into two main collisional stages (Lahtinen 1994, Nironen 1997, Korsman et al. 1999). In the first stage, 1910 - 1885 Ma ago, the Pyhäsalmi primitive island arc and the Central Finland continental arc (*see* Fig. 2.1-2) amalgated and collided with Archaean Karelian craton (Lahtinen 1994, Nironen 1997, Korsman et al. 1999). During the collision the Proterozoic supracrustal rocks, slices of Archaean crustal blocks, and the craton margin ophiolites were thrust northeast onto the Archaean crust (Nironen 1997). In the collisional zone, the crust was strongly thickened both tectonically, as the colliding plates were thrust onto each other, and magmatically through magmatic underplating. The resulting thermal effect led to the melting of the lower crust, and the metamorphism and partial melting in the upper crust, 1885 Ma ago.

In the next stage, about 1890 to 1880 Ma ago, the Southern Finland sedimentary-volcanic complex and the Central Finland continental arc collided. This stage is discussed in more detail in the following sections. After the orogenic movements, the cooled and rigid upper crust was intruded by postorogenic granites, dated at 1815 - 1770 Ma. The cratonisation of the Svecofennian crust is represented by 1650 - 1540 Ma old rapakivi granites and associated mafic rocks, the development of the Jotnian sedimentary basins and intrusion of Postjotnian diabase dykes dated at 1270 - 1250 Ma.

6.2 Geological evolution of bedrock in southern and southwestern Finland

6.2.1 Svecofennian supracrustal belts

The Svecofennian supracrustal rocks in southern and southwestern Finland can be divided into several separate belts, which are from north to south: Tampere schist belt, the psammitic migmatite belt (also known as the Pori-Vammala-Mikkeli Migmatite Belt or the tonalite migmatite zone), Hämeenlinna-Somero volcanic belt and the Kemiö-Mäntsälä belt (Fig. 6.2-1). The first two belong to the Central Finland continental arc, whereas the latter two are part of the Southern Finland sedimentary-volcanic complex (also known as the Late Proterozoic potassium granite migmatite belt). In the study area, the area north and northeast of the Satakunta sandstone formation belong to the psammitic migmatite belt, whereas the area southwest of the sandstone is a part of the pelitic migmatite belt within the Southern Finland sedimentary-volcanic complex (Fig. 6.2-1, *see* also Fig. 2.1-2).

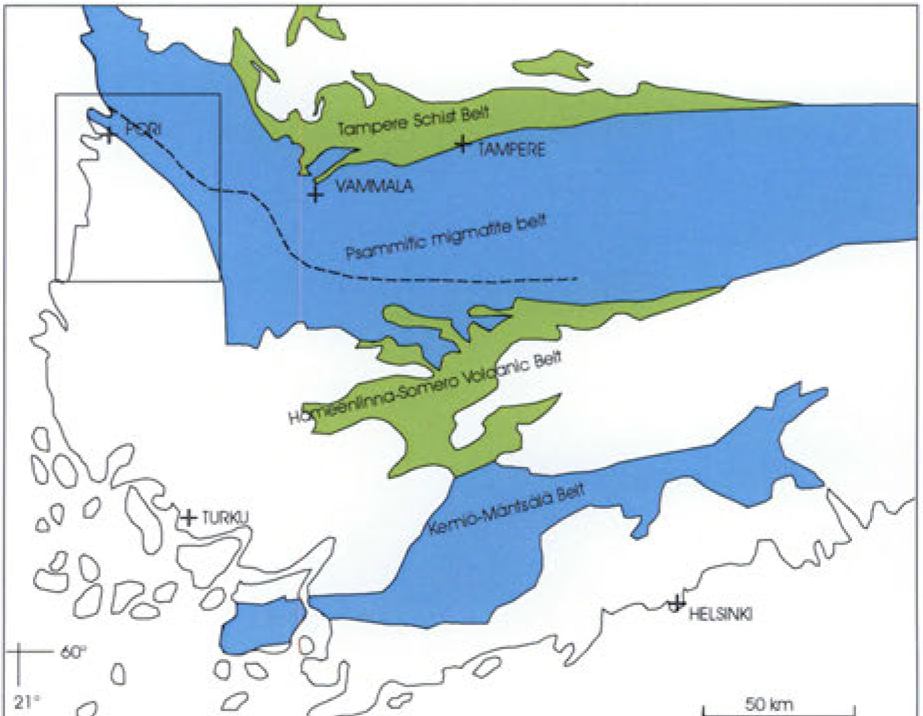


Figure 6.2-1. *The Svecofennian supracrustal belts of southern and southwestern Finland, and the location of the study area. Modified after Kähkönen et al. (1994) and Kilpeläinen (1998). The suture zone (dashed line) between the Central Finland and Southern Finland arc complexes has been drawn after Kähkönen et al. (op. cit) and Nironen (1997).The colours are just for the separation of the belts, and do not refer to any bedrock composition.*

The Tampere schist belt mainly consists of turbiditic metasediments (metagraywackes) interbedded with island arc-type pyroclastic metavolcanics (Kähkönen 1989, Kähkönen et al. 1989, Kähkönen & Leveinen 1994). They are underlain by pillow lavas and chemical metasediments dated at 1950 Ma (Kähkönen & Nironen 1994).

The psammitic migmatite belt is composed of intensely migmatized metagraywackes of turbiditic origin and tonalitic granitoids crosscutting them. Black schists and graphite-bearing schists are more common than in the Tampere schist belt. Mainly in the southern part of the zone, there are over 1910 Ma old ultramafic metavolcanics with MORB (Mid Ocean Ridge Basalt)- and WPB (Within-Plate Basalt)-affinities (Lahtinen 1996), associated spatially with mafic and ultramafic plutonites dated at 1890 Ma (Peltonen 1995).

The Hämeenlinna-Somero volcanic belt can be divided into the Forssa group and the stratigraphically younger Häme group (Hakkarainen 1994). The Forssa group consists of calc-alkaline, mainly andesitic metavolcanics, interbedded with minor metasedimentary rocks. The basement of the metavolcanics is unknown. The basaltic to andesitic metavolcanics of the Häme group have been intruded over the pelitic and turbiditic metasediments, which can be erosion products from the Forssa group. Fine-grained basaltic dykes are common. The Häme group is separated from the Forssa group by volcanic conglomerates and Fe-sulphide formations.

The Kemiö-Mäntsälä belt is mainly composed of acid metavolcanics and turbiditic metasediments, but sedimentary carbonates and banded iron formations are also rather common. Most of the metavolcanics are of island arc-type, but also WPB-type metavolcanics exist.

The metasedimentary surroundings of the Hämeenlinna-Somero and Kemiö-Mäntsälä volcanic belts are composed of gneisses and migmatites of turbiditic and pelitic origin. Abundant felsic schists and sedimentary carbonates occur south of the Kemiö-Mäntsälä belt. The felsic schists are originally acid volcanics and their erosion products.

6.2.2 Rifting of the Svecofennian protocrust and formation of island-arc complexes

The rifting of the Svecofennian protocrust began before 1910 Ma. Subduction zones, where oceanic crust was destroyed, developed both at the northern and southern margins of the opening ocean (Fig. 6.2-2). The subduction both to the north and south resulted in the formation of Central Finland and Southern Finland arcs, respectively (Lahtinen 1994, Nironen 1997).

The rifting stage and the incipient development of the island arc is represented by the tholeitic pillow lavas of the Haveri Formation, which is the lowermost unit in the Tampere schist belt (Mäkelä 1980, Kähkönen & Nironen 1994, Lahtinen 1996, Kähkönen 1999). The volcanics within the formation are characterised by MORB-affinities in the lower sections and island-arc magma features in the upper parts. Based on the similarities of the Pb isotopes, the Haveri volcanics may be correlated with the

Outokumpu ophiolites (Vaasjoki & Huhma 1999). The palaeoenvironment of the Haveri Formation with volcanics and chemical metasediments was a shallow, extensional ocean basin (Kähkönen & Nironen 1994).

The geochemical and isotope characteristics of the Tampere schist belt suggest the occurrence of thick continental crust already before 1910 Ma (Lahtinen & Huhma 1997). Th only evidence of this protocrust, which has served as a depositional basement and also mainly as provenance of the supracrustal rocks, are the detrital zircons of the turbidites dated at 1910 - 2000 Ma and the isotope compositions of some granitoids in the Central Finland granitoid complex.

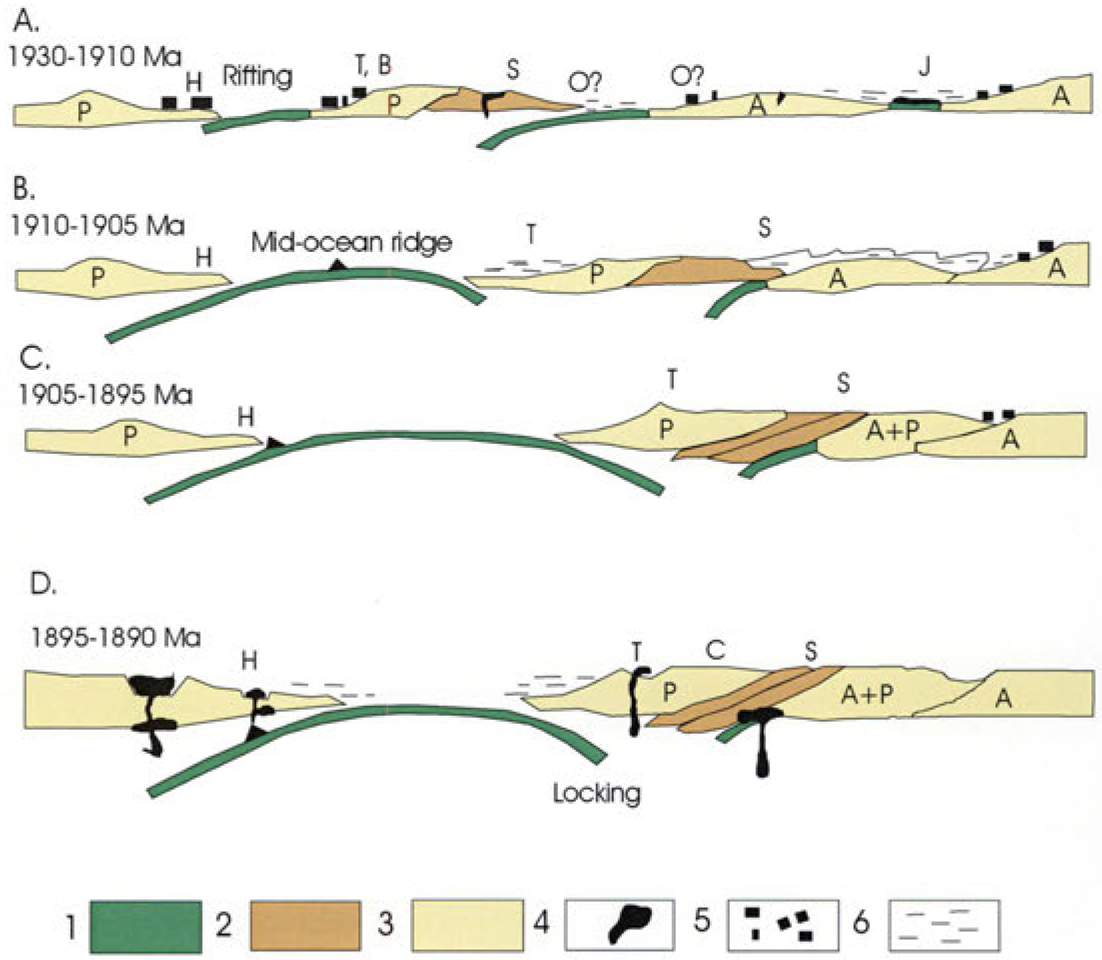


Figure 6.2-2. The evolution of the Svecofennian crust between 1930 - 1895 Ma. Modified after Lahtinen (1994). A) Rifting of the protocrust and formation of oceanic crust; B, C and D) subduction of the oceanic crust and subsequent deposition of turbiditic graywackes and intrusion of andesitic volcanics. Only selected magmatic episodes are shown and subduction related magmatism is excluded. 1. Oceanic crust, 2. primitive island arc, 3. continental crust, 4. mafic magmatism, 5. cratonic/passive margin/rift sedimentation, 6. turbidites. A = Archaean, P = Proterozoic, O = Outokumpu, J = Jormua, S = Savo schist belt, T = Tampere schist belt, B = Bothnian schist belt, H = Hämeenlinna-Somero volcanic belt, C = Central Finland granitoid complex.

In the Tampere schist belt, the rift volcanism (the Haveri Formation) was followed by deposition of a thick succession of turbiditic graywackes in the sinking basin and the intrusion of andesitic volcanics dated at 1904 - 1890 Ma. The chemical composition of the metavolcanics correspond with the volcanics of the modern island arcs or active margins, where the oceanic plate is being subducted under the continental plate. According to Lahtinen (1994, 1996), the subduction to the north under the Tampere schist belt may have lasted only 10 to 20 Ma.

The chemical composition of the turbiditic metasediments of the northern part of the psammitic migmatite belt resemble the lowermost turbiditic graywackes of the Tampere schist belt, suggesting that the two belts were originally formed close to each other (Kähkönen et al. 1994). Formerly, the migmatite area was considered to represent an older orogeny. Later studies, however, have shown that the difference in metamorphic grade and degree of migmatitisation in the two belts merely represent different depth sections in the crust. The Tampere schist belt was metamorphosed at greenschist/ amphibolite facies at a temperature of 350 – 530 °C and a pressure of 3 – 4 kb, while in the Vammala area of the psammitic migmatite belt, the highest temperature of 670 °C was attained at a pressure of 5kb.

The metasediments in the northern part of the psammitic migmatite belt have been interpreted to be deposited in the fore-arc basin before the eruption of the island arc volcanics, or maybe already during the rifting stage, as suggested by Lahtinen (1994, 1996). In the western part of the psammitic migmatite belt (in the Pori area), the structures of the deformation phase D_2 dip gently north (Pietikäinen 1994), indicating northward underthrusting of the sedimentary rocks (Korsman et al. 1999).

The Forssa group of the Hämeenlinna-Somero volcanic belt was formed in an island arc environment and the volcanics were erupted from a complex of stratovolcanoes (Hakkarainen 1994). The age of the Forssa group volcanics is uncertain, but according to Lahtinen (1996) they could be older than 1910 Ma and represent a subduction to the south. The Häme group, overlying the Forssa group, resulted from a short-time, internal rifting of the island arc, and the lavas erupted along an E-W trending fissure system (Hakkarainen 1994, Kähkönen et al. 1994). The age of the Häme group volcanics is about 1890 Ma (Vaasjoki 1994), which means that they are at least as old as the volcanics in the Tampere schist belt. The basaltic dykes, common in the Häme group, has been interpreted as feeder channels of the volcanics.

The main part of the Kemiö-Mäntsälä belt represents island arc environment, but the volcanics with WPB-affinities indicate an early rifting stage (Kähkönen et al. 1994, Väisänen 1998). The age of the volcanics in the zone is unknown, but they may be coeval with the volcanics of the Forssa group and the Tampere schist belt. In spite of the differences between the Kemiö-Mäntsälä and Hämeenlinna-Somero belts, Nironen (1997) considers them to be parts of the same island arc system. The differences can be due to different palaeotectonic settings or different stages in the evolution of a convergent plate margin (Kähkönen et al. 1994).

The calcareous sediments between the Hämeenlinna-Somero and Kemiö-Mäntsälä belts have been interpreted as having been deposited in a back-arc or an intra-arc basin (Nironen 1997, Väisänen 1998).

The Svecofennian system, with its two separate island arc complexes, has been presumed to resemble the modern Indonesian archipelago between the Australian and Eurasian continents (Ward 1987, Nironen 1997, 1998).

6.2.3 Collision of the arc complexes

About 1890 Ma ago the oceanic basin between the Central Finland continental arc complex (including the Tampere schist belt and the psammitic migmatite belt) and the Southern Finland arc complex (including the Hämeenlinna-Somero and Kemiö-Mäntsälä belts) began to close, which led to the collision of the complexes and commencing of the deformation (Lahtinen 1994, 1996, Ruotoistenmäki 1996, Nironen 1997). During the collision, the southern arc was emplaced over the Central Finland arc and the rocks were metamorphosed for the first time (Fig. 6.2-3). According to Lahtinen (1996), the subduction of the mid-oceanic ridge under the Hämeenlinna-Somero Belt caused the subduction under the Tampere schist belt to cease through locking of the subduction slab (Fig. 6.2-2C). At the same time, the northern plate, including the Tampere schist belt and the Central Finland Granitoid Complex, changed from overriding to passive underthrusting plate, which subducted and collided to the south. When the collision continued the southern arc complex was thrust upon the northern one.

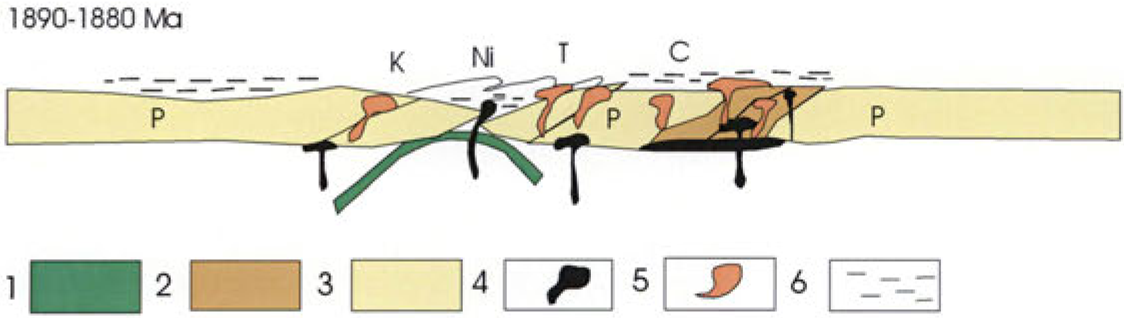


Figure 6.2-3. *Collision of the Central and Southern Finland arc complexes 1890 - 1880 Ma ago. Modified after Lahtinen (1994). Only selected magmatic episodes are shown. 1. Oceanic crust, 2. primitive island arc, 3. continental crust, 4. mafic magmatism, 5. felsic magmatism, 6. turbidites. P = Proterozoic, T = Tampere schist belt, C = Central Finland granitoid complex, K = Kemiö-Mäntsälä belt, Ni = Vammala Ni-belt.*

In the collisional event, the Svecofennian crust was thickened, both tectonically due to over- and underthrusting, and by magmatic underplating (Lahtinen 1994, Nironen 1997). The underplate was generated when the base of the tectonically thickened lithosphere was delaminated and compensated by hot asthenospheric magma, which then intruded to the crust-mantle boundary, causing remelting of lower crust and large-scale granitoid magmatism (Fig. 6.2-4). The remnants of this underplate can be seen in the lower crust as a seismic zone with high P-velocity (*see* Chapter 5).

The geometry of the Southern Finland E-W trending deep conductor, described in Chapter 5, seems to suggest overthrusting of the Southern Finland arc upon the Central Finland arc (Korsman et al. 1999). The deep conductor is associated with linear near-surface conductors, which, according to correlation of the geophysical data with the geological observations, are mostly graphite- and sulphide-bearing metasedimentary rocks. In the Tampere schist belt and the psammitic migmatite belt these near-surface conductors are directly above the deep conductor but no near-surface conductors occur in the Southern Finland sedimentary-volcanic complex above the deep conductor. However, the results of the airborne electromagnetic survey suggest that conducting rocks associated with the Tampere schist belt and psammitic migmatite belt may continue below the complex, suggesting the emplacement of the Southern Finland complex over the Central Finland arc. The geometry of the deep conductor also indicates underthrusting of rocks of the psammitic migmatite belt northward below the protocrust at deep crustal levels and overthrusting at upper crustal levels (Korsman et al. 1999)

A suture zone between the southern and northern complexes has been proposed to be located in the southern part of the psammitic migmatite belt between the Tampere schist belt and the Hämeenlinna-Somero volcanic belt, where MORB-affinity volcanics and younger, spatially associated, mafic to ultramafic plutonics may mark the occurrence of an inferred suture (Lahtinen 1994, 1996). The proposed suture is supported by the geochemical and isotope data of the Tampere and Hämeenlinna-Somero belts, which show that the rocks of the Tampere schist belt represent thicker and more mature crust compared to the Hämeenlinna-Somero belt (Lahtinen 1996). The detrital zircons in the Tampere schist belt and Kemiö-Mäntsälä belt show, at least at the Orijärvi area of the Kemiö-Mäntsälä Belt, different provenance, which suggests that there was an oceanic basin between the belts. The different ages of metamorphism in the pelitic migmatite belt and the psammitic migmatite belt (*see* Section 3.2) also support the proposed suture zone. The proposed suture partly corresponds the Vammala Ni-belt, which continues from the Vammala area northwest to the Pori area. Accordingly, it is possible that the suture zone continues to the Satakunta area, as suggested in Fig. 6.2-1. However, to this date, no direct evidence of the suture has been demonstrated in the Satakunta area.

The boundary between the Tampere schist belt and the psammitic migmatite belt is sharp and sheared. According to Nironen (1989), it is a south dipping thrust, possibly related to southward subduction. Along the shear zone the psammitic migmatite belt has been uplifted compared to the Tampere schist belt. Lahtinen (1994) has interpreted it as a trenchward dipping backstop or backthrust that subsequently took part in the northward vergent thrusting.

The collision of the arc complexes is characterised by an intense magmatic activity, which appears as synorogenic granitoids, mainly dated at 1890 - 1880 Ma. In the schist belts they are mainly granodiorites, tonalites and trondhjemites, whereas in the Central Finland granitoid complex coarse-grained granitic rocks are dominant. The Sm-Nd data of the synorogenic granitoids indicate that they consist mainly of juvenile material expelled from the mantle (Vaasjoki 1996). The intrusion of the mantle-derived magma to the lower crust/upper mantle boundary (mafic underplating) caused the partial melting of the lower crust (Fig. 6.2-4). The resulting melt, consisting of material both from the mantle and the crust, then ascended to the upper parts of the crust.

The synorogenic plutonic rocks also include a small number of mafic and ultramafic intrusions. In the study area, north and northeast of the Satakunta sandstone, these intrusions, described in Chapter 2.2.2, belong to the Vammala Ni-belt. Peltonen (1995) interpreted the intrusions of the Vammala area as remnants of synorogenic feeder dykes and sills of basaltic tholeitic magma that became progressively boudinaged and disrupted by tectonic processes.

During the collision of the arc complex the rocks were strongly deformed. In the main deformation phase (D_2), a strong foliation was developed and the bedrock was folded by subhorizontal folding. In the pelitic migmatite belt of the Southern Finland sedimentary-volcanic complex, the folds of the deformation phase D_2 are recumbent, whereas in the psammitic migmatite belt the axial planes are vertical. The recumbent attitude of the folds suggests thrust tectonics (Väisänen et al. 1994, Selonen & Ehlers 1998, Väisänen & Hölttä 1999). In the Central Finland continental arc, the initially subhorizontal D_2 structures have generally been folded into vertical position during D_3 at ca. 1885 - 1880 Ma. In the Southern Finland arc complex, the subhorizontal D_2 structure were folded by D_3 into upright position only in the boundary zone between the two complexes (Korsman et al. 1999 and references therein). The synorogenic tonalite-trondhjemite intrusions, as well as the mafic-ultramafic intrusions, were emplaced before or during the deformation phase D_2.

During the collision and the subsequent magmatic activity the sedimentary and volcanic rocks were metamorphosed for the first time. The high temperature/low pressure metamorphism was caused by mafic underplating (Fig. 6.2-4), which led to a strong increase in temperature and recrystallisation and partial remelting of the rocks in the upper crust. Both in the pelitic migmatite belt and the psammitic migmatite belt, the peak of the metamorphism was achieved during and after the deformation phase D_2 but before the deformation phase D_3 (Korsman et al. 1999). The age of the metamorphism is, thus, almost the same as for the 1890 - 1880 Ma granitoids.

In Southern Finland, the deformation characterised by compression continued until the emplacement of the granitoids dated at 1870 Ma (Väisänen & Hölttä 1999, Nironen 1999). In the Central Finland granitoid complex, however, the coeval granitoids are closer to extensional rapakivi granites than collisional granitoids in their geochemical character (Rämö & Nironen 1996, Nironen et al. 2000). Accordingly, compressional tectonism prevailed in Southern Finland, while Central Finland was already in the extensional or transtensional stage (Väisänen & Hölttä 1999, Nironen et al. 2000).

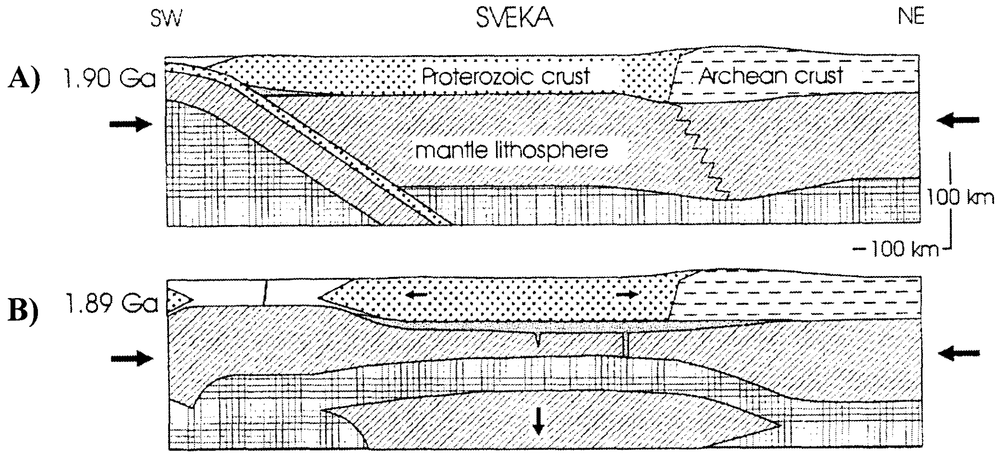


Figure 6.2-4. *A cross-sectional model for the development of the crust along the SVEKA profile (see Fig. 2.1-2) after Nironen (1997). A) Accretion of the Central Finland arc against the Archaean crust 1910 Ma ago caused thickening of the crust and the underlying mantle lithosphere. A small underplate (densely dotted area) was generated by subduction from SW. B) Collision of the Southern Finland and Central Finland arcs 1890 Ma ago resulted in thickened lithosphere, the base of which was subsequently delaminated and compensated by hot asthenospheric magma. The magma intruded to the crust-mantle boundary and generated a mafic underplate which caused remelting of lower crust and large-scale synorogenic granitoid magmatism.*

In Southern Finland the accretion of the Southern Finland sedimentary-volcanic complex against the Central Finland continental arc was followed ca. 20 Ma later by a new period of crustal shortening and thickening. This stage is characterised by a strong thermal pulse, which caused a high temperature (700 - 800°C) metamorphism and almost total melting of the sedimentary rocks. This stage of the Svecofennian orogeny is manifested as ca. 100 km wide and 500 km long potassium granite belt described by Ehlers et al. (1993). These late-orogenic, migmatite-forming potassium granites, dated at 1840 - 1830 Ma, are classified as so-called S-type granites, which have been formed by partial melting of the Svecofennian sedimentary rocks deep in the crust (Nurmi & Haapala 1986). They occur both as large bodies and as leucosomes in metasedimentary rocks. The metamorphism and the partial melting of the sedimentary rocks were caused by high heat flow associated with mafic underplating after lithosphere delamination (Korja et al. 1993, Lahtinen 1994, Nironen 1997). In some parts of the potassium granite belt, the generation of S-type granites continued until about 1830 - 1810 Ma as evidenced, for example, by 1815 Ma old potassium granites in the Turku granulite area (Väisänen et al. 2000). In the southern Satakunta area the granites, at least the granites, porphyritic granites and pegmatites of the Rauma map sheet (Fig. 2.1-3 and App. 1) are S-type granites (*see* Chapter 2.2.1).

According to structural geological studies made in southern and western Finland (for example, Ehlers et al. 1993, Selonen et al. 1996, Väisänen & Hölttä 1999, Stålfors & Ehlers 2000), the late-orogenic granites were emplaced during the deformation phase D_3 when the tectonic setting was a compressional crustal shortening. Ehlers et al. (1993)

suggest that the granites were emplaced within the dextral transpressional zone of strike-slip shears and horizontal shears in the middle crust (12 - 15 km) and that the zone of late-orogenic granites is located in a former extensional zone acting as an inherited zone of weakness (Fig. 6.2-5). The granites were emplaced subhorizontally parallel to the overturned axial planes of the deformation phase D_2 and folded during the deformation phase D_3 (Stålfors & Ehlers 2000). The geochemical studies suggest that the individual granite bodies are not formed by one major magmatic pulse but are composed of several subhorizontal layers at metre-scale or less, collected from small melt batches (Stålfors & Ehlers, op. cit.). A possible melt transport mechanism is strike-slip dilatancy pumping.

Also Väisänen and Hölttä (1999) regard the model of dextral transpression as most likely but according to them, the granite melt utilised, in addition to thrust zones, all existing structures such as F_3 fold hinges and axial planes, the D_3 shear zones and S_2 foliation planes. The high temperature metamorphism and associated granites are attributed to crustal thickening simultaneous with thinning of the mantle.

Korja & Heikkinen (1995) and Väisänen et al. (1994) have proposed that the potassium granites were emplaced during an extensional tectonic regime. Korja and Heikkinen (1995) suggest that the extension of the overthickened crust started in the late D_3 stage by reactivation of the old thrust planes to shear zones. The shear zones are represented by listric shear zones, identified in the deep seismic studies, which are detached at the middle to lower crustal boundary. In the deformation phase D_4, the extension continued, leading to the upwelling of the mantle and intrusion of the postorogenic granites along the shear zones.

According to Nironen (1997), intracontinental transpressional deformation started about 1860 Ma ago, and resulted in partial thrusting of the southern complex onto the northern one and in the development of a shear zone system. Nironen (op. cit.) combines the compressional structures by Ehlers et al. (1993) and the extensional model of Korja & Heikkinen (1995), and suggests that the potassium granites were intruded along transpressional structures during the extensional stage following transpression. However, on the basis of petrological studies, Lahtinen (1994) has demonstrated that the extension of the crust started only later and it is connected with the 1800 - 1770 Ma postorogenic magmatism. According to Lahtinen (1994, 1996), the emplacement of the late-orogenic potassium granites is connected with a new stage of intra-continental thrusting, which started 1860 - 1850 Ma ago, either due to a new collision further away or as a continuation of earlier collision after termination of lithospheric thinning.

According to Vaasjoki (1996), the absence of igneous rocks in the 1860-1850 Ma interval suggests that the Svecofennian orogeny terminated ca. 1870 Ma ago and that the ca. 1830 Ma migmatization in southern Finland is a separate tectonothermal event.

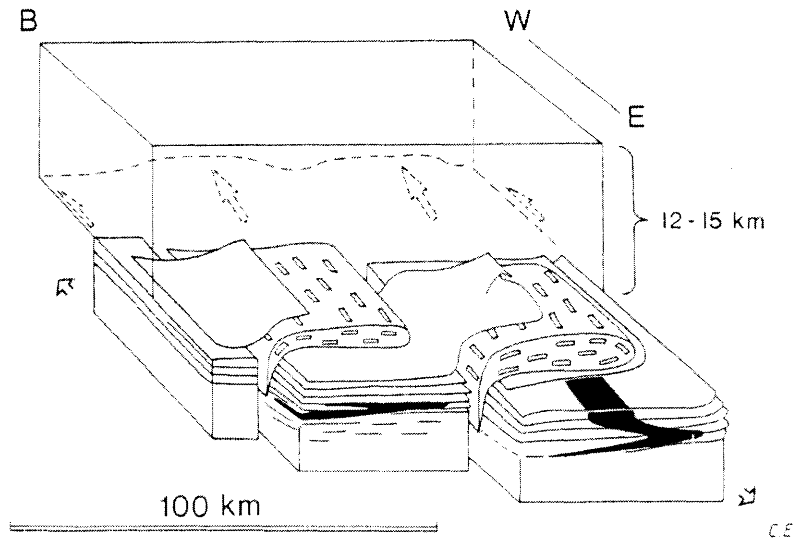


Figure 6.2-5. *The late-orogenic granites in southern and southwestern Finland could have been emplaced within a transpressional zone of strike-slip shears and horizontal shears in the middle crust (Ehlers et al. 1993) Early synorogenic granitoids and the folding of the deformation phase D_3 are omitted.*

The effects of the younger metamorphism are only weakly visible in the psammitic migmatite belt. However, in the Vammala area, the local granitic migmatization associated with the zones of intense D_3, could record the younger metamorphism at 1840 -1830 Ma (Kilpeläinen et al. 1994), although Kilpeläinen (1998) considers that it is more probable that the heat of the earlier metamorphic event (1890 - 1880 Ma) remained longer in those zones than elsewhere in the migmatite area.

Late-orogenic or post-collisional (*cf.* Väisänen et al. 2000) mafic and intermediate plutonic intrusions are scarce and none of them occur at the study area. South of the study area, however, a coarse-grained gabbro pegmatite, dated at 1815 ± 12 Ma (Patchett & Kouvo 1986), cuts the 1878 Ma old Tynki gabbro at Uusikaupunki. According to Suominen (1991), the dyke indicates mobilisation during the later metamorphism. In the Turku granulite area, there are 1815 Ma old shoshonitic monzodiorites and coeval, but not cogenetic, S-type granites (Väisänen et al. 2000). The monzodiorites are derived from enriched lithospheric mantle, whereas the granites were produced by crustal anatexis. Väisänen et al. (op. cit.) suggest that the mantle-derived mafic intrusions were the heat source for granulite facies metamorphism and late-orogenic crustal anatexis in the Turku area between ca. 1830 - 1810 Ma.

After the orogenic movements, the crust cooled and became more rigid. Fractures developed, along which postorogenic granites, dated at 1800 - 1770 Ma (Suominen 1991, Vaasjoki 1996), were intruded. According to Lahtinen (1994), the postorogenic granites were formed by fractional crystallisation of alkalibasaltic mantle magmas, which have probably assimilated crustal rocks. No such rocks occur at the study area, the nearest ones being located in the Åland archipelago (for example Åva and Lemland intrusions).

6.2.4 Subjotnian

The rapakivi granites and associated mafic rocks (diabases and gabbro-anorthosites) were emplaced 1650 - 1540 Ma ago, i.e., about 200 Ma after the collapse of the Svecofennian orogeny in a period before the deposition of the Jotnian sandstones (hence the name "Subjotnian"). The ages of all the large rapakivi batholiths in southwestern Finland (Åland, Vehmaa and Laitila) are very close to each other, 1575±10 Ma, and are thus 50 to 80 Ma younger than the rapakivi granites of southeastern Finland, for example the Wiborg rapakivi massif (Suominen 1991). The Laitila rapakivi batholith, the northern part of which lies in the study area, is 1583±3 Ma in age (Vaasjoki 1996). The Eurajoki rapakivi stock, east of Olkiluoto, is a satellite massif to the Laitila batholith, and consists of Tarkki granite and younger Väkkärä granite (Haapala 1977). Both are somewhat younger than the Laitila batholith, Tarkki 1571±3 Ma and Väkkärä 1548±3 Ma (Suominen et al. 1997). The Peipohja rapakivi, located northeast of the Satakunta sandstone, has the age of 1570±20 Ma (Suominen 1991). Additionally, the crust below the Bothnian Sea hosts several large rapakivi granite batholiths, 1500 - 1580 Ma in age, which are only partly exposed on land (Korja et al. 2001, p. 343).

The rapakivi granites are classified as anorogenic granites, which means that their generation and emplacement are not directly connected to the Svecofennian orogeny. The rapakivi magmatism is bimodal in nature, the mafic members being tholeitic diabases and minor gabbro-anorthosite bodies. On the basis of geochemical and isotope data, the rapakivi magmas were formed by partial melting of the lower to middle crust, caused by mafic intra- and underplating (Rämö 1991, Haapala & Rämö 1992) (Fig. 6.2-6). The underplating may have been caused by extensional collapse of the orogen, rifting, or deep mantle plumes, or it could have been triggered by compositionally unstable domains in the upper mantle (Haapala & Rämö 1992, 1999). The protolith of the rapakivi magma may have been intermediate to acid (granodioritic) plutonic rocks (Rämö 1991). The rapakivi granites have crystallised about 3 km below the Subjotnian erosional level and the present erosion level is close to the upper contact of the batholiths (Vorma 1975).

The intrusion of the rapakivi magma was polyphasic, as evidenced by several different rapakivi granite types in the Laitila rapakivi batholith and the Tarkki and Väkkärä granites of the Eurajoki stock (Vorma 1976, Haapala 1977). According to Haapala (op.cit.), the Tarkki granite had already cooled and fractured, when the Väkkärä granite was emplaced, and the crystallisation of the Väkkärä granite took place at higher crustal levels compared to the Tarkki granite.

Rapakivi magma also erupted on the surface as rhyolitic volcanics (Fig. 6.2-6). Because of the properties of acid magma, the eruptions were most likely explosive. In southwestern Finland, however, all these volcanics have been eroded away. Only a few places in the Fennoscandian Shield (for example in Suursaari in the Gulf of Finland) contain volcanic rocks connected with rapakivi magmatism.

The mafic rocks, diabases and gabbro-anorthosites, connected with the rapakivi intrusions are derived by partial melting of the upper mantle, caused by mafic underplating. On the basis of geochemical and isotope studies, the basic magma was contaminated by crustal components on its way through hot, partially molten crust (Rämö 1991). The gabbro-anorthosites are slightly older than the rapakivi granites, while the diabase dykes both intrude the rapakivi granites and are intruded by them. The diabase dykes, some of which are younger than the gabbros and anorthosites, are associated with stages when magmatic underplating had caused fracturing of the Svecofennian crust (Rämö 1991). They represent feeder dykes for continental flood basalts (Fig. 6.2-6), which, however, have been eroded away.

The anorogenic magmatism may be associated with the formation of an early Proterozoic supercontinent (Laurentia-Baltica), which promoted mantle upwellings, partial melting of the upper mantle, and subsequent partial melting of the lower crust (Rämö 1991).

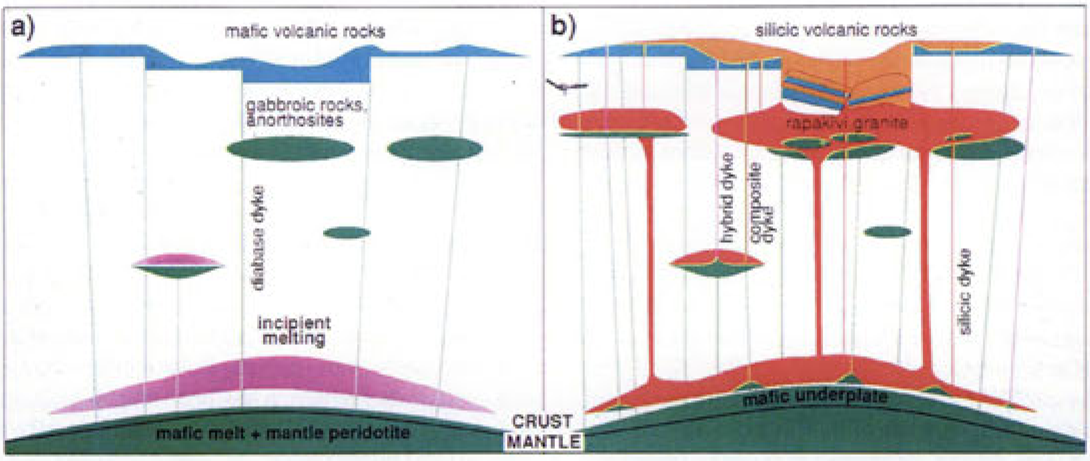


Figure 6.2-6. *Model of relations between rapakivi granites and associated mafic rocks after Rämö & Haapala (1990). The figure is from Rämö et al. (1994). a) Mantle-derived mafic magmas have been partly intruded into the lower and middle crust and partly extruded on the surface. b) Rapakivi granite magmas were formed by extensive melting of the lower crust, caused by the thermal effect of the mafic magmas. Both the felsic and the mafic magmas may have used the same pathways, forming composite dykes.*

Seismic soundings have shown that the rapakivi granite complexes and the associated gabbro-anorthosites and diabase dyke swarms are located in areas, where the crust is thinner than its surroundings (Korja et al. 1993). Under the Åland and Laitila rapakivi batholits the crust is ca. 47 km thick, while the average thickness of the Svecofennian crust is 55 km. In the Bothnian Sea, the crustal thickness gradually decreases from north to south (BABEL WG 1993). In the areas with thinner crust, the lower crust has a highly reflective lamellar structure and a strongly reflective subhorizontal boundary between the lower crust and the upper mantle (Korja et al. 1993). Strong, gently dipping seismic reflections in the lowermost crust are caused partly by crustal extension and partly by mafic sills associated with the Subjotnian diabases and rapakivi magmatism (Korja & Heikkinen 1995). In the upper crust, the rapakivi granites can be seen as seismically non-reflective areas, which are surrounded by reflections imitating horst- and graben structures.

On the basis of deep-seismic reflection data of the BABEL experiment in the Bothnian Sea, Korja and Heikkinen (1995) have composed a model which aims at explaining the changes in thickness and reflectivity of the crust, and the associated lithologies. The rapakivi batholiths were emplaced during an extensional tectonic regime. An isostatic imbalance due to the thinning of the upper crust caused the upwelling of the upper mantle and lower crust, which increased the geothermal gradient and temperature of the crust. Thinning of the crust took place by movements on two sets of listric shear zones. The older listric shear zones, detached at the lower to middle crustal boundary, may be inverted Svecofennian thrust zones, as suggested by Väisänen et al. (1994). The younger Subjotnian or Jotnian listric shear zones, detached at the lower crust to upper mantle boundary (Moho), developed due to continued extensional strain. The subsequent enhanced upwelling of the mantle caused the partial melting of the upper mantle, mafic underplating of the thinned crust, and partial melting of the lower crust. The younger listric shear zones, which are parallel to the main fracture zones associated with the rapakivi granites, gabbro-anorthosites and sandstones, provided pathways from the lower crust to the upper crust for the rapakivi granite and mafic magmas. Silicic-basic composite dykes demonstrate that both the felsic and the mafic magmas could use the same pathways (Rämö 1991) (Fig. 6.2-6).

According to Korja et al. (2001) thin crust with large crustal variations, large sedimentary basins (*see* Section 6.2.5) and large-scale bimodal magmatism suggest that the Gulf of Bothnia (including the Bothnian Sea) is an aborted rift, which is a part of a wide honeycomb-like palaeorift area extending from Lake Ladoga to Caledonides.

Åhäll et al. (1996, 2000) have contradicted the traditional anorogenic model for the rapakivi magmatism. They link the episodic rapakivi magmatism at 1650 - 1620 Ma, 1580 - 1560 Ma and 1550 - 1500 Ma to three stages of subduction-related magmatism of the Gothian orogen of southwestern Sweden at 1690 - 1650 Ma, 1620 - 1580 Ma and 1560 - 1550 Ma, and suggest that repeated westward-stepping subduction at the active margin provided the primary control on episodic mantle melting and consequent bimodal magmatism. Åhäll et al. (op.cit.) explain the relationship between subduction and the rapakivi magmatism by the changes in the regional stress conditions within the lithosphere after cessation each stage of subduction. These changes may then have controlled the lithospheric extension leading to rapakivi magmatism.

6.2.5 Jotnian and Postjotnian

After the rapakivi magmatism the geological evolution of the Satakunta area continued with the deposition of the Jotnian Satakunta sandstone. The upper parts of the sandstone were deposited ca. 1400 - 1300 Ma ago (Kouvo 1976), but the development of the sedimentation basin (graben) may have begun already during the Subjotnian rifting period, ca. 1650 Ma ago (Kohonen et al. 1993). On the basis of gravity and deep seismic studies the sandstone formation continues northwest to the Bothnian Sea as a V-shaped graben. The graben is bounded by steep normal faults, which are in connection with the Bothnian Sea sedimentary basin (Heikkinen et al. 1998). According to Axberg (1980) the fault line along the northeastern contact of the sandstone can be followed across to the Swedish coast.

The Bothnian Sea sedimentary basin, probably as much as 3 - 4 km in depth (Korja et a. 2001, p. 353), is composed of several faulted blocks, bounded by NE-, NNE- and NW-trending lineaments (Winterhalter 1972, Axberg 1980, Winterhalter et al. 1981). The basin is composed of Jotnian sandstone, overlain by Cambrian and Ordovician sedimentary rocks. Korja and Heikkinen (1995) and Heikkinen et al. (1998) connect the development of the Bothnian Sea basin and, thus, the Satakunta graben, with the process which resulted in the generation of rapakivi granites and associated mafic rocks (*see* Chapter 6.2.4). Thinning of the crust via listric shear zones during the extensional regime resulted in the development of a block structure in the upper part of the crust. The subsequent upwelling of the mantle caused the partial melting of the upper mantle, mafic underplating of the thinned crust, and partial melting of the lower crust. A thermal dome was developed in the upper crust due to the heat of the rapakivi magmas. With the cooling of the extended crust and mantle, and the rapakivi magma underlying the dome, the crust began to subside, and the sedimentary basin grew and filled, when the thermal dome was eroded.

The Satakunta sandstone has been interpreted to be a fluvial sediment formation deposited in an alluvial environment by poorly channelled braided streams (Fig. 6.2-7) before the outcropping of the rapakivi granites (Marttila 1969, Kohonen et al. 1993). However, block movements occurred even after the formation of the basin and the deposition of the sandstone, as evidenced by the tilted sandstone beds in the northeastern part of the graben. Faulting after the deposition is further supported by the coarseness of the upper part of the sandstone, indicating rapid deposition and high relief (Kohonen et al. 1993). These block movements were probably connected with subsidence caused by the intrusion of Postjotnian mafic magmas, represented at present by olivine diabase dykes and sills (Laitakari 1983). The Satakunta sandstone formation may have been much wider than today but all what is left is the part controlled by the Postjotnian movements (Fig. 6.2-7).

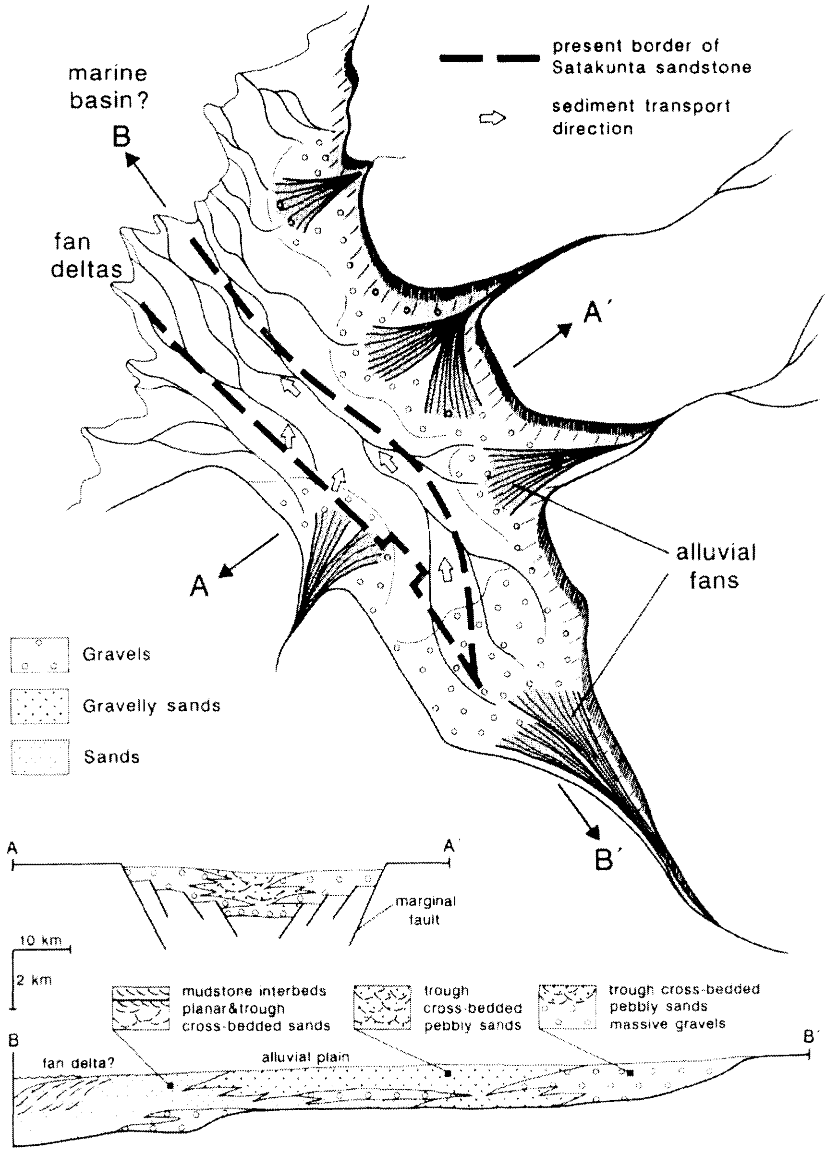


Figure 6.2-7. *The possible palaeogeography of the Satakunta sandstone basin 1400 Ma ago (Kohonen et al. 1993).*

The Postjotnian diabase dykes and sills represent the last phase of the Proterozoic evolution in the southern Satakunta area. The diabase dykes and sills can be connected with a world-wide thermal pulse, the results of which can be seen, for example, in Sweden, Greenland and Canada (Suominen 1991). Based on key palaeomagnetic poles and geological criteria, Buchan et al. (2000) have suggested that at the time of the ca. 1270 Ma diabase intrusions, the Laurentia craton (including North America and Greenland) and the Baltica craton (including Sweden and Finland) may have been joined. On the basis of parallel flow directions of the diabase sills in Sweden and Greenland, determined by a palaeomagnetic and anisotropy of magnetic susceptibility study, Elming & Mattsson (2001) suggest a common magma source, which would have been located in the reconstructed contact between Baltica and Laurentia. Consequently, Elming & Mattsson (op. cit.) suggest that tensional tectonics, indicated by the

Postjotnian magmatism, reflects the break up of the Baltica/Laurentia supercontinent. On the other hand, Korja et al. (2001), based on deep seismic studies, suggest that the diabase sills and dykes in Satakunta have intruded along shear zones and fault planes which were formed by the collapse of the cooling rapakivi magma chamber.

A large areal extent of the Postjotnian diabases in the surroundings of Satakunta is suggested by the seismic reflection surveys in the Bothnian Sea, which have revealed several strong reflectors, interpreted as diabase sills (BABEL WG 1993). The geochemical features of the Satakunta olivine diabases indicate that they could be feeder conduits to continental flood basalts. However, no such basalts have remained in Satakunta, although in central Sweden and to the north of Lake Ladoga, in Russian Karelia, coeval volcanic rocks in association with the Postjotnian diabases occur (Amantov et al. 1996).

6.2.6 Latest events affecting the bedrock

After the deposition of the Jotnian sedimentary rocks and the eruption of the basalts, represented today by the Postjotnian diabases, erosion was the main geological process in the Satakunta area. In the Fennoscandian Shield, two prolonged erosion phases can be distinguished: 1) prolonged Late Proterozoic erosion and planation, resulting in the peneplanation, and 2) Late Phanerozoic erosion, resulting in the present-day topography (Puura et al. 1996). During the first erosion period until the beginning of the Vendian, 650 Ma ago, the Fennoscandian Shield was located in the tropical zone (Pesonen et al. 1991), where a warm and moist climate favoured both physical and chemical weathering. The highest southern latitude of ca. 70°S was attained 930 Ma ago (Mertanen & Pesonen 1997). Kaolinitic regoliths from this period have remained, for example, in Virtasalmi, eastern Finland, where they are Middle Proterozoic in age (Sarapää 1996).

During the Late Vendian, large parts of the Fennoscandian Shield were covered by a shallow sea and a subsequent sedimentary basin was filled with sand and clay sediments. The Vendian sandstone occurs in Lauhanvuori, western Finland, where traces of crawling by annelids in sandstone are not more than 700 Ma in age (Tynni and Hokkanen 1982). In the Bothnian Sea basin, which existed already during the Jotnian, the Jotnian sandstone is in places overlain by over 200 metres of Cambrian and Lower Ordovician sandstones, siltstones and mudstones, which are covered by Middle and Upper Ordovician carbonate rocks (Winterhalter 1972, Thorslund & Axberg 1979, Axberg 1980). The maximum thickness of the Paleozoic sedimentary rocks in the basin is 375 m (Axberg 1980). According to van Balens and Heeremans (1998), these have once been overlain by ca. 230 m of younger sediments.

Although sedimentary rocks younger than Precambrian do not exist in the southern Satakunta area at present, it can be assumed that large parts of the Satakunta bedrock were once covered by them, as indicated by the Lauhanvuori sandstone, 60 km NNE of the study area, and presumably Lower Cambrian sandstone dykes in Åland, in the Turku archipelago, and in the Vehmaa rapakivi granite. Van Balens and Heeremans (1998) have estimated that the Finnish coast of the Bothnian Sea was once covered by hundreds of metres of Paleozoic and/or younger sediments but they were ablated during or after the Tertiary uplift (*cf.* also Puura et al. 1996, p. 124).

The bedrock movements after the deposition of the Paleozoic sediments are demonstrated by faults, cutting the Proterozoic basement and the overlying Paleozoic sedimentary cover in the Bothnian Sea basin (Winterhalter 1972). The faulting is presumably associated with the Late Paleozoic deformation, which, for example, created the Oslo graben and caused the subsidence of the Bothnian Sea basin (Puura et al. 1996).

7 SUMMARY

The summary of the magmatic, structural and metamorphic evolution of the southern Satakunta area is shown in Table 7-1. The oldest rocks in southern Satakunta are migmatized mica gneisses, metasedimentary in origin, being deformed and metamorphosed during the Svecofennian orogeny 1900 – 1800 Ma ago. In the northeastern part of the area they belong to the psammitic migmatite belt, whereas the southwestern part of the area belongs to the pelitic migmatite belt. The pelitic and psammitic migmatite belts can be distinguished on the basis of predominant granitic and trondhjemitic to granodioritic leucosomes, respectively.

The mica gneisses are intruded by 1890 - 1880 Ma old plutonic rocks, which consist of trondhjemites, tonalites and granodiorites, occurring conformably with the structures of the mica gneisses. Coarse-grained granites and pegmatites occur as large bodies and are present in older supracrustal and plutonic rocks, migmatizing them or occurring as cross-cutting veins.

The northern part of the anorogenic Laitila rapakivi batholith, about 1580 Ma in age, is located in the area. The Eurajoki rapakivi stock is a satellite massif to the Laitila batholith, and can be divided into two types, hornblende-bearing Tarkki granite and younger, topaz-bearing Väkkärä granite.

The Jotnian Satakunta sandstone, at least 1400 – 1300 Ma in age, has been preserved in a NW-SE trending graben structure. The sedimentary material of the sandstone derives from the Svecofennian supracrustal and plutonic rocks. The thickness of the sandstone is at least 600 m, and probably as much as 1800 m thick. The sandstone is cut by Postjotnian olivine diabase dykes, 1270 – 1250 Ma in age. Aeromagnetic mapping shows that these, in turn, are cut by a swarm of younger diabase dykes trending north-south. Lake Sääksjärvi in the northeastern part of the map area is an impact crater of early Cambrian age.

The rocks of the study area have undergone a polyphasic deformational history. The earliest observed tectonic structure is the biotite foliation S_1 parallel to the bedding, identified only in the hinge zones of F_2 folds and as inclusion trails in porphyroblasts. No F_1 folds have been identified. The dominant foliation is usually penetrative S_2 with tectonic/metamorphic segregation. In the pelitic migmatite belt F_2 folding is recumbent and isoclinal to tight. The recumbent attitude of the folds suggests a layer-parallel shearing and over- or underthrusting during D_2, although no major thrust zones have been identified. Compared to the pelitic migmatite belt, F_2 folding of the same age in the psammitic migmatite belt has a vertical axial plane. The synorogenic tonalite-granodiorite intrusions were emplaced before or during the deformation phase D_2.

Both D_1 and D_2 structures are deformed by the regional F_3 folding of the deformation phase D_3. The fold axes is generally horizontal or moderately plunging, and fold limbs are often strongly sheared. Late-orogenic potassium granites were emplaced during D_3 in the pelitic migmatite belt.

Table 7-1. *Summary of the magmatic, structural and metamorphic evolution of the southern Satakunta area.*

Age (Ma)	Magmatism and volcanism	Sedimentation and deformation	Metamorphism
<700		Development of fractures Block movements after the sedimentation Vendian, Cambrian and Ordovician sandstones, siltstones and carbonate rocks	
Planation of the Proterozoic bedrock			
1270-1260	Postjotnian diabase dykes (= feeder channels of the continental flood basalts)	Hydrothermal, illite-filled fractures in migmatites Subsidence of the Satakunta sandstone graben (block movements)	Low-grade contact metamorphic effects in the sandstone
1400-1300		Deposition of the upper parts of the Satakunta sandstone	
1580-1550	Subjotnian diabase dykes Anorogenic rapakivi granites and associated felsic volcanics Söörmarkku felsic dyke	Hydrothermal fracture in the Palaeoproterozoic rocks Development of the sedimentary basin (graben) of the Satakunta sandstone Extensional tectonic regime: rifting of the crust, faulting	Contact metamorphism T = 800°C P = 0.5-2 kb
1800-1770	Period of Postorogenic magmatism (not recorded in Satakunta)		
1850-1810	Late-orogenic potassium granites (1840-1830 Ma)	Pelitic migmatite belt: plastic deformation D_3 related with NW-SE-trending compression, overthrusting Plastic to semi-brittle, zonal deformation D_4 Late-D_3 within the psammitic migmatite belt: the Kynsikangas shear zone	Regional metamorphism within the pelitic migmatite belt; migmatization, anatectic melting T = 700 - 800°C P = 4 - 5 kb
1885-1870	Lavia porphyritic granite[1)] Pöytyä granodiorite[1)]	Plastic deformation D_3 within the psammitic migmatite belt	
1890-1885	synorogenic tonalites and granodiorites Ultramafic and mafic intrusions	N-S-trending compression; Plastic, complex deformation D_2, thrust tectonics within the pelitic migmatite belt?	Regional metamorphism; segregation, migmatization T = 800 - 670°C P = 5 - 6 kb
<1910	Basic and intermediate volcanics	D_1-deformation: S_1 -foliation Sedimentation: bedding S_0	Regional metamorphism; segregation
>1910	Unknown pre-Svecofennian protocrust		

[1)] located outside the study area

The age of the D_2 deformation is close to the age of the synorogenic granitoids, 1890 - 1880 Ma. In the psammitic migmatite belt, the age of the D_3 deformation is about 1885 to 1870 Ma. The age of this deformation in the pelitic migmatite belt is close to the late-orogenic microcline granites (1840 - 1830 Ma).

The Svecofennian Domain is characterised by a high-temperature/low-pressure metamorphism. Typical metamorphic minerals, porphyroblasts, in pelitic metasediments are andalusite, staurolite, garnet, cordierite, sillimanite and potassium feldspar. The temperature of the metamorphism is estimated at 650 - 700°C and the pressure 4 - 5 kb, corresponding the upper amphibolite facies. The peak temperature of 700 - 800°C was attained in the migmatite areas under pressures of 4 - 6 kb.

The supracrustal rocks of the study area represent sedimentation in two separate island arc environments, the rocks of the psammitic migmatite belt and the pelitic migmatite belt belonging to the Central Finland and Southern Finland continental arcs, respectively. The rifting of the Svecofennian protocrust begun before 1910 Ma. Subduction zones, where oceanic crust was destroyed, developed both to the northern and southern margins of the opening ocean. The subduction both to the north and south resulted in the formation of Central Finland and Southern Finland arcs, respectively

Collision of the Central and Southern Finland continental arcs took place 1890 - 1880 million years ago, during which the Southern Finland arc was emplaced over the Central Finland arc. A possible suture zone is located in the southern part of the psammitic migmatite belt.

The collision of the arc complexes is characterised by a intense magmatic activity, which appears as synorogenic granitoids, mainly dated at 1890 - 1880 Ma. In the study area they are mainly granodiorites, tonalites and trondhjemites. The synorogenic intrusions were emplaced before or during the deformation phase D_2.

During the collision and the subsequent magmatic activity the sedimentary and volcanic rocks were metamorphosed for the first time. The high temperature-low pressure metamorphism was caused by magmatic underplating, which led to a radical increase in temperature and recrystallisation and partly remelting of the rocks in the upper crust. Both in the pelitic migmatite belt and the psammitic migmatite belt, the peak of the metamorphism was achieved during and after the deformation phase D_2 but before the deformation phase D_3. The age of the metamorphism is thus almost the same as the 1890 - 1880 Ma granitoids.

In southern Finland, the accretion of the Southern Finland sedimentary-volcanic complex against the Central Finland continental arc was followed ca. 20 Ma later by a new period of crustal shortening and thickening. This stage is characterised by a strong thermal pulse, which caused a high temperature (700 - 800°C) metamorphism and almost total melting of the sedimentary rocks. This stage of the Svecofennian orogeny is manifested as ca. 100 km wide and 500 km long potassium granite belt. These late-orogenic potassium granites, dated at 1840 - 1830 Ma, are classified as so-called S-type granites, which have been formed by partial melting of the Svecofennian sedimentary rocks deep in the crust. The metamorphism and the partial melting were caused by mafic underplating after lithospheric delamination.

Subjotnian rapakivi granites associated with mafic rocks, Jotnian Satakunta sandstone formation and Postjotnian diabase dikes represent the cratonisation stages of the Svecofennides. The Laitila rapakivi batholith, dated at 1583±3 Ma, is composed of several different rapakivi granite types, the main type being the pyterlitic rapakivi granite. The two types of the Eurajoki rapakivi stock, Tarkki granite and Väkkärä granite, are somewhat younger than the Laitila batholith, Tarkki 1571±3 Ma and Väkkärä 1548±3 Ma.

The rapakivi granites are so-called anorogenic granites, meaning that their generation and emplacement are not directly connected to the Svecofennian orogeny. The rapakivi magmatism is bimodal in nature, the mafic members being tholeitic diabases and minor gabbro-anorthosite bodies. The rapakivi magmas were formed by partial melting of the lower to middle crust, caused by magmatic intra- and underplating. The rapakivi batholits were emplaced during an extensional tectonic regime. The listric shear zones, which are parallel to the main fracture zones associated with the rapakivi granites, gabbro-anorthosites and sandstones, provided pathways from the lower crust to the upper crust for the rapakivi granite and mafic magmas.

After the rapakivi magmatism the Jotnian Satakunta sandstone was deposited in a graben structure. The upper parts of the sandstone deposited ca. 1400 - 1300 Ma ago but the development of the sedimentary basin (graben) may have begun already during the Subjotnian rifting period ca. 1650 Ma ago. The Satakunta sandstone has been interpreted to be a fluvial sediment formation deposited in an alluvial environment before the outcropping of the rapakivi granites. Block movements occurred even after the formation of the sedimentary basin and the deposition of the sandstone, as evidenced by the tilted sandstone beds in the northeastern part of the graben. These block movements were connected with subsidence, caused by the intrusion of Postjotnian diabase dykes.

The Postjotnian diabase dykes represent the last Proterozoic evolution phase at the southern Satakunta area. Their geochemical features suggest them to be feeder conduits to continental flood basalts. However, no such basalts have remained in the Satakunta area.

Although sedimentary rocks, younger than Precambrian, do not presently exist at the southern Satakunta area, it can be presumed that large parts of the Satakunta bedrock were once covered by them, as indicated by the Vendian sandstone in Lauhanvuori, western Finland, and presumably Lower Cambrian sandstone dykes in Åland, in the Turku archipelago, and in the Vehmaa rapakivi granite. In the Bothnian Sea basin, the Jotnian sandstone is overlain by Cambrian and Lower Ordovician sandstones, siltstones and mudstones, which are covered by Middle and Upper Ordovician carbonate rocks.

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APPENDICES

Appendix 1: Bedrock map with topograpic relief 1:250 000

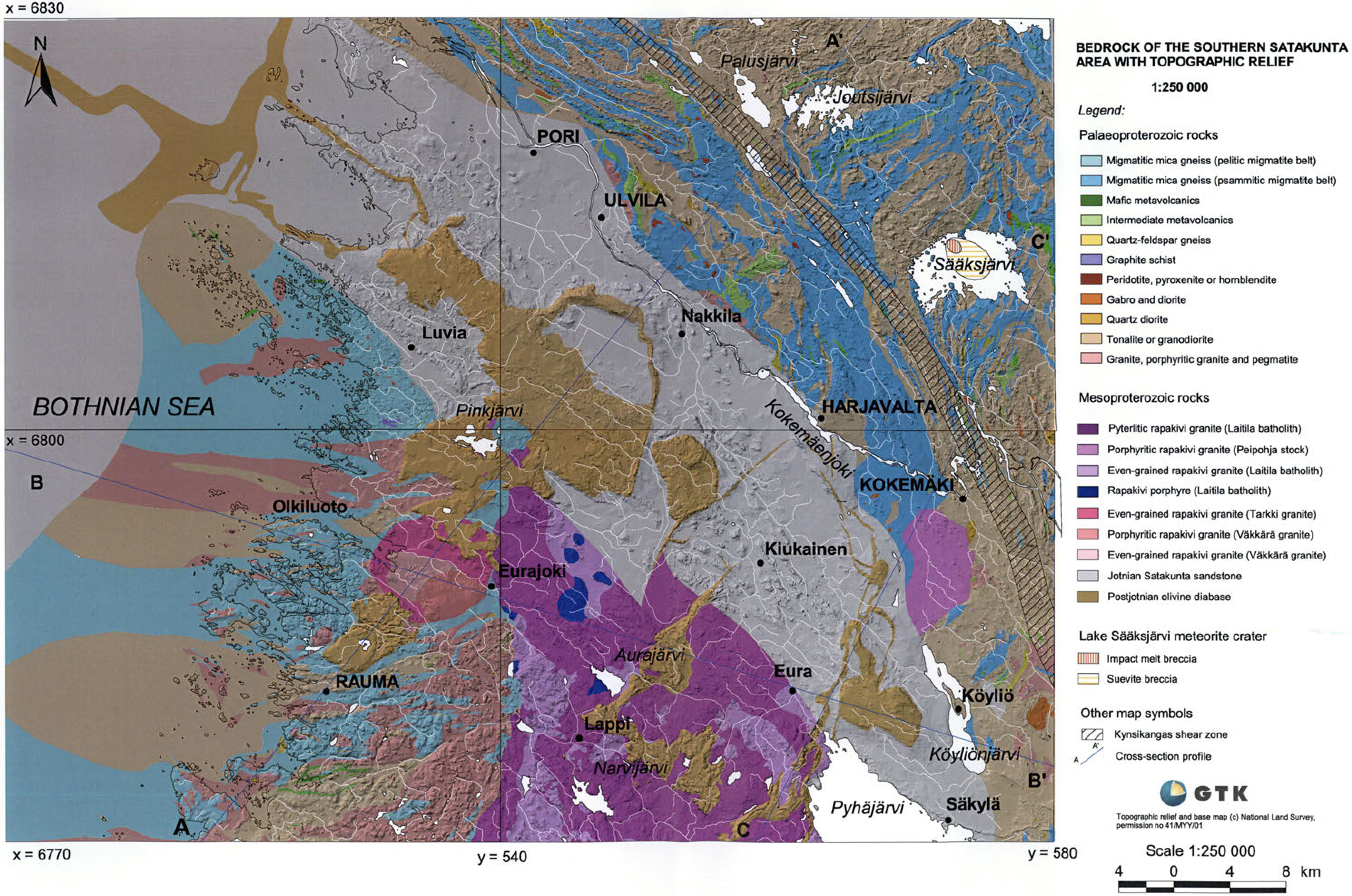
Appendix 2: Cross-section A - A'

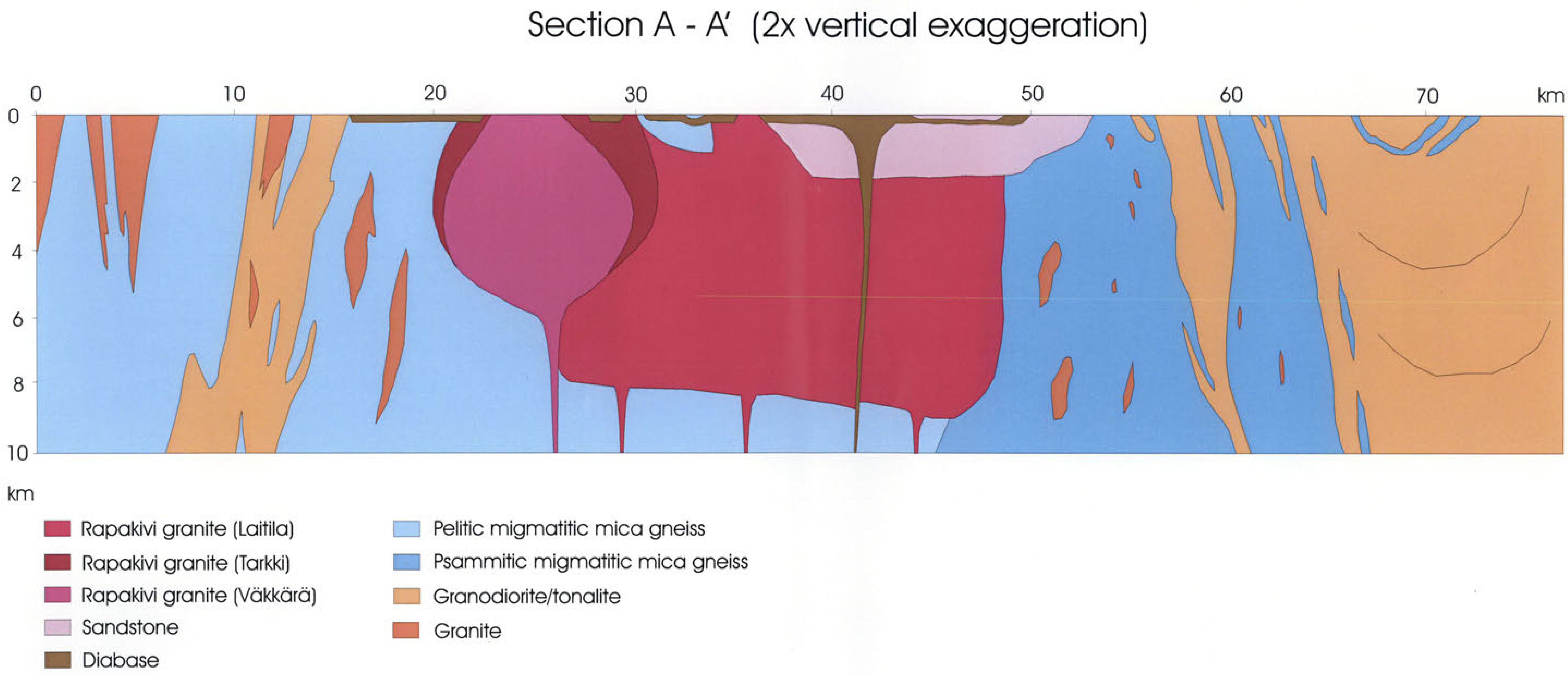
Appendix 3: Cross-section B - B'

Appendix 4: Cross-section C - C'

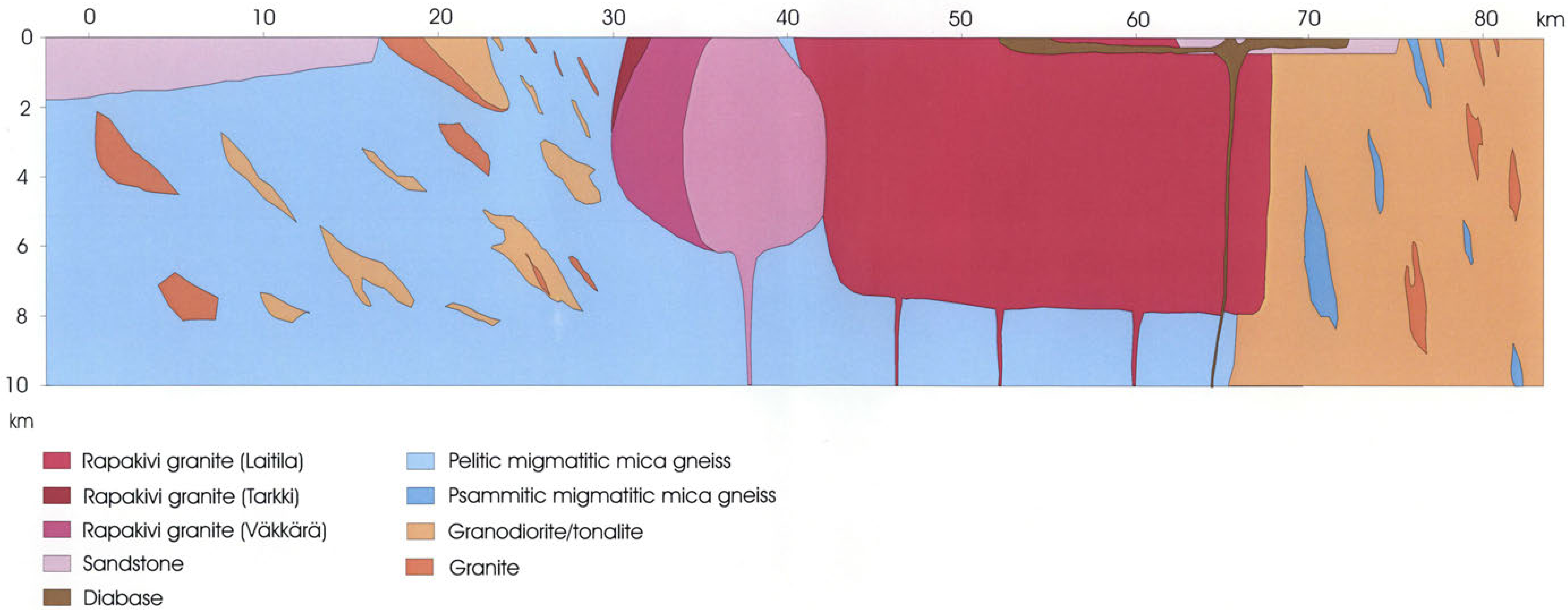
Appendix 5: Low-altitude aeromagnetic map 1:250 000

Appendix 6: Interpretation of low-altitude aeromagnetic map 1:250 000

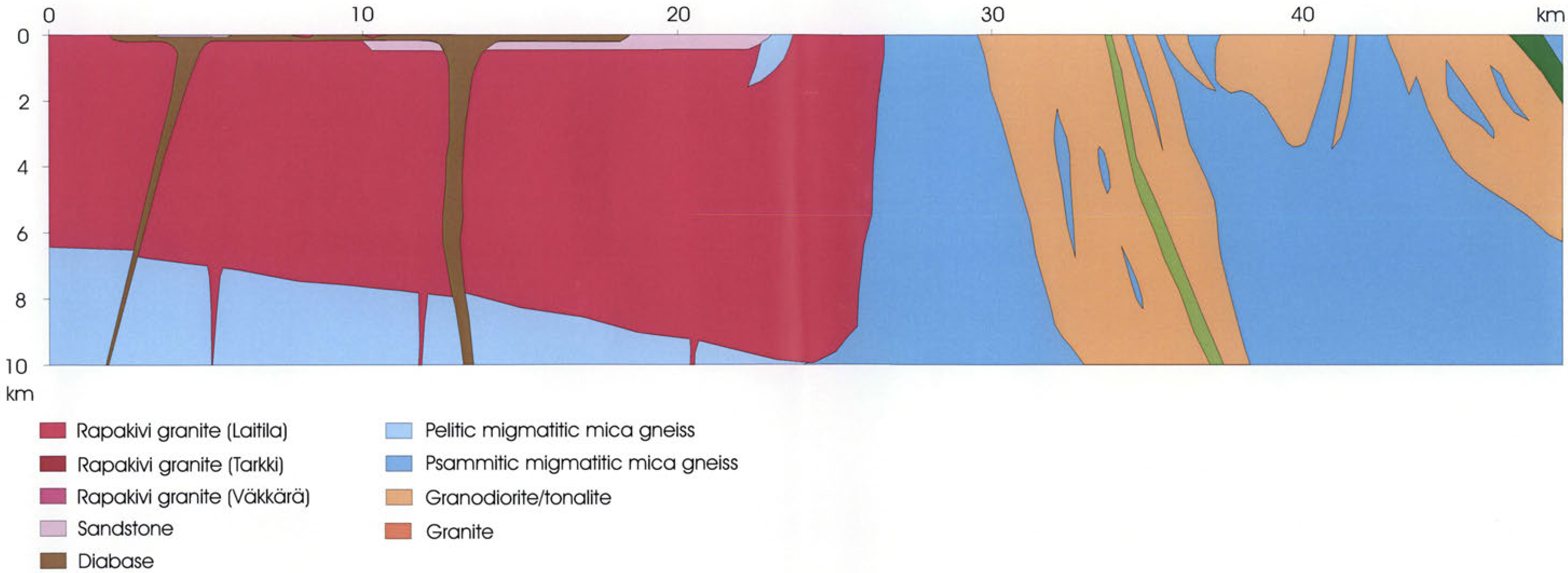

x = 6830
x = 6800
x = 6770
y = 540
y = 580
N
BOTHNIAN SEA
PORI
ULVILA
Nakkila
Luvia
HARJAVALTA
Kokemäenjoki
KOKEMÄKI
Olkiluoto
Eurajoki
Kiukainen
RAUMA
Eura
Lappi
Köyliö
Köyliönjärvi
Säkylä
Pyhäjärvi
Narvijärvi
Aurajärvi
Pinkjärvi
Palusjärvi
Joutsijärvi
Sääksjärvi
A
A'
B
B'
C
C'
BEDROCK OF THE SOUTHERN SATAKUNTA AREA WITH TOPOGRAPHIC RELIEF
1:250 000
Legend:
Palaeoproterozoic rocks
Migmatitic mica gneiss (pelitic migmatite belt)
Migmatitic mica gneiss (psammitic migmatite belt)
Mafic metavolcanics
Intermediate metavolcanics
Quartz-feldspar gneiss
Graphite schist
Peridotite, pyroxenite or hornblendite
Gabro and diorite
Quartz diorite
Tonalite or granodiorite
Granite, porphyritic granite and pegmatite
Mesoproterozoic rocks
Pyterlitic rapakivi granite (Laitila batholith)
Porphyritic rapakivi granite (Peipohja stock)
Even-grained rapakivi granite (Laitila batholith)
Rapakivi porphyre (Laitila batholith)
Even-grained rapakivi granite (Tarkki granite)
Porphyritic rapakivi granite (Väkkärä granite)
Even-grained rapakivi granite (Väkkärä granite)
Jotnian Satakunta sandstone
Postjotnian olivine diabase
Lake Sääksjärvi meteorite crater
Impact melt breccia
Suevite breccia
Other map symbols
Kynsikangas shear zone
Cross-section profile
GTK
Topographic relief and base map (c) National Land Survey, permission no 41/MYY/01
Scale 1:250 000
4 0 4 8 km

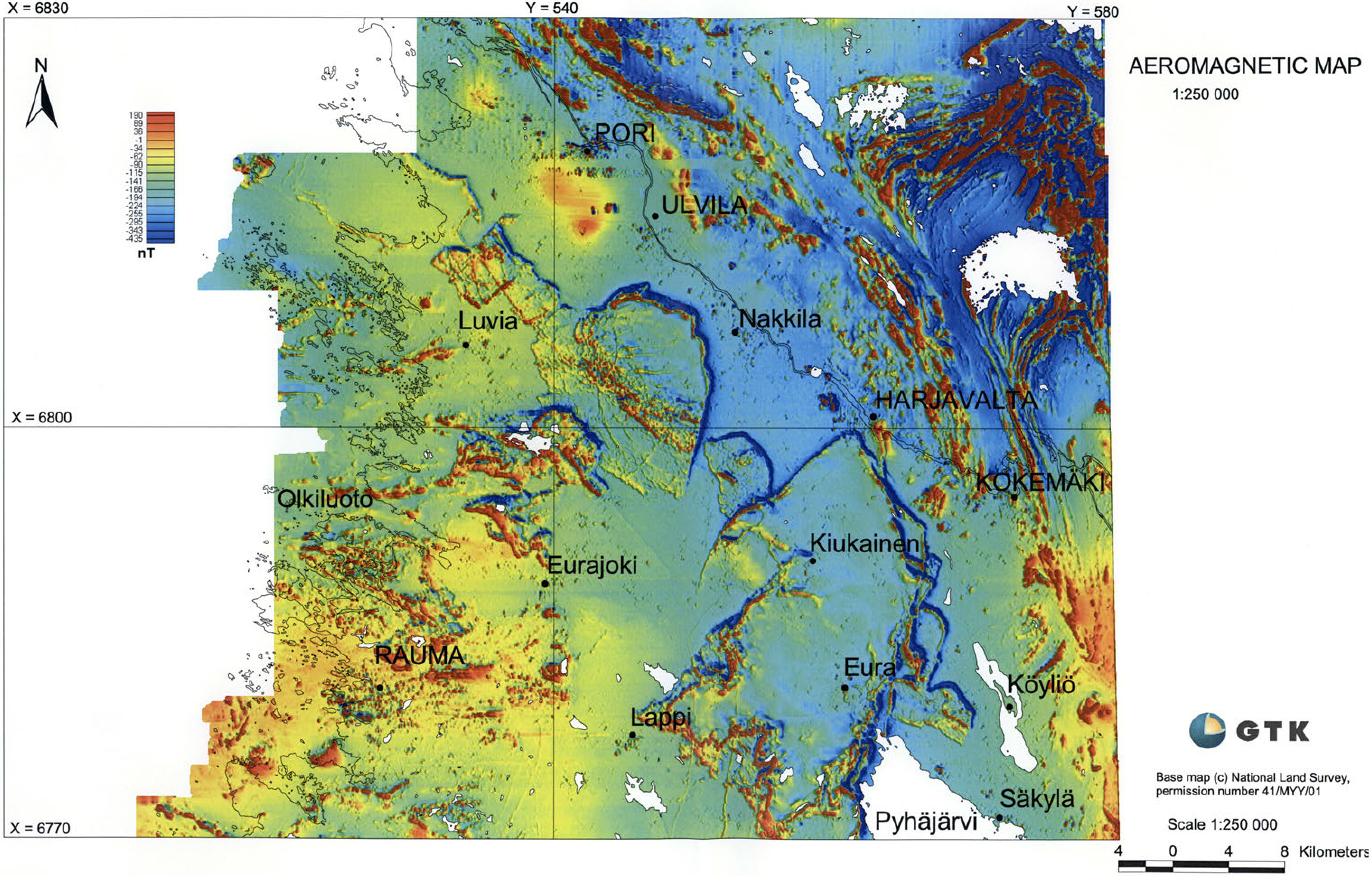

Section A - A' (2x vertical exaggeration)
0
10
20
30
40
50
60
70
km
0
2
4
6
8
10
km
Rapakivi granite (Laitila)
Rapakivi granite (Tarkki)
Rapakivi granite (Väkkärä)
Sandstone
Diabase
Pelitic migmatitic mica gneiss
Psammitic migmatitic mica gneiss
Granodiorite/tonalite
Granite

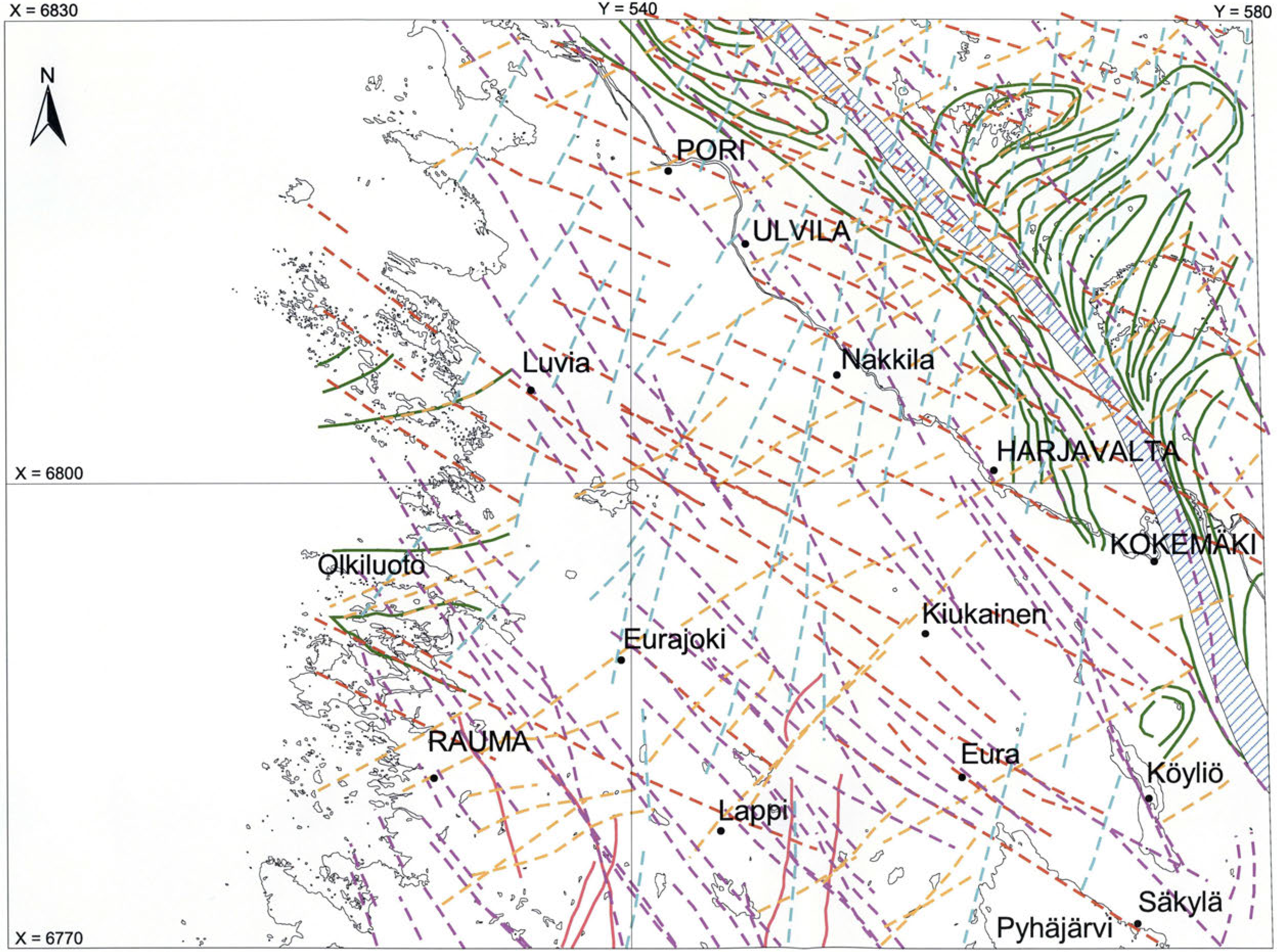
Section B - B′ (2x vertical exaggeration)



Section C - C′


0
10
20
30
40
km
0
2
4
6
8
10
km
Rapakivi granite (Laitila)
Rapakivi granite (Tarkki)
Rapakivi granite (Väkkärä)
Sandstone
Diabase
Pelitic migmatitic mica gneiss
Psammitic migmatitic mica gneiss
Granodiorite/tonalite
Granite


X = 6830
Y = 540
Y = 580
AEROMAGNETIC MAP
1:250 000
N
190
89
36
-1
-34
-62
-90
-115
-141
-168
-194
-224
-255
-295
-343
-435
nT
PORI
ULVILA
Luvia
Nakkila
HARJAVALTA
X = 6800
KOKEMÄKI
Olkiluoto
Kiukainen
Eurajoki
RAUMA
Eura
Köyliö
Lappi
GTK
Base map (c) National Land Survey,
permission number 41/MYY/01
Säkylä
X = 6770
Pyhäjärvi
Scale 1:250 000
4 0 4 8 Kilometers



GEOPHYSICAL INTERPRETATION

1:250 000

Legend

- Fracture zone, direction N 0-30° E
- Fracture zone, direction N 40-50° E
- Fracture zone, direction N 100-120° E
- Fracture zone, direction N 140-150° E
- Kynsikangas shear zone
- Folding/tectonic zone
- Diabase dykes

GTK

Base map (c) National Land Survey, permission number 41/MYY/01

Scale 1:250 000

4 0 4 8 Kilometers

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