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Depositional History and Tectonic Regimes within and in the Margins of the Fennoscandian Shield During the last 1300 Million years

Seppo Paulamäki
Aimo Kuivamäki

May 2006

POSIVA OY
FI-27160 OLKILUOTO, FINLAND
Tel +358-2-8372 31
Fax +358-2-8372 3709

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ABSTRACT

This report is a literature review, which describes the occurrences of Meso- and Neoproterozoic and Phanerozoic sedimentary rocks in Finland and their depositional history The report also summarizes the tectonic and magmatic events within and in the margins of the Fennoscandian Shield during the last 1200 - 1300 million years.

Mesoproterozoic (Middle Riphean) and Neoproterozoic (Late Vendian) sandstones and siltstones are presently found only in some tectonically protected basins, e.g., Satakunta and Muhos grabens, Bothnian Sea basin and Bothnian Bay basin, or in the impact structures, but they may have originally distributed over a much larger area. During the Lower and Middle Cambrian, arenaceous and argillaceous sediments were probably deposited in most of the western, southern and central Finland. During the Ordovician, carbonates were deposited in an epicontinental sea with repeated transgressions and regressions. In several occasions, the Bothnian Sea and Åland Sea basins, as well as southern and south-western Finland were covered by the sea, and in the Middle Ordovician, the sea reached as far as central Finland. No Late Palaeozoic (Silurian to Permian) sedimentary rocks have been recognized. The apatite fission track studies indicate, however, that extensive Silurian to Devonian deposits most likely covered large parts of the Fennoscandian Shield, with an estimated thickness of 3 - 4 km in Sweden and ca. 1 km in Finland. In eastern Lapland, the clay bed of Akanvaara was deposited in early Tertiary (Eocene) under marine conditions.

The Mesoproterozoic tectonics was dominated by the Sveconorwegian orogeny (ca. 1250 - 900 Ma), which mainly affected the present south-western Sweden and southern Norway. In Finland, the 'Postjotnian' olivine diabases, dated at 1270 - 1250 Ma, are considered to be connected to the initial rifting at the onset of the orogeny. After the Sveconorwegian orogeny, the Fennoscandian Shield was part of the worldwide supercontinent, Rodinia. At ca. 620 - 600 Ma, this supercontinent started to break-up. By Late Cambrian times, the plate convergence started again, leading to the Caledonian orogeny at the NW margin of the Fennoscandian Shield, the main phase of which occurred in Silurian, at ca. 425 Ma. The tectonics related to the late phase of the central European Variscan orogeny resulted in strike-slip faulting along the NW-SE trending Tornquist Zone in the south-western margin of the Fennoscandian Shield. The faulting was associated with volcanic activity and intrusion of dyke swarms, dated at 300 - 240 Ma. The Permian rifting and volcanism is demonstrated by ca. 400 km long Oslo Rift. During the Mesozoic rifting, the Tornquist Zone was repeatedly reactivated. In the Tertiary, the opening of the North Atlantic, the initiation of sea-floor spreading in the Norwegian-Greenland Sea and onset of the Alpine continent-continent collision dominated the evolution of north-western Europe. Mainly the margins of the Fennoscandian Shield were affected by the uplift of western Scandinavia. The ridge push forces from the Mid-Atlantic Ridge seem to be the major stress-generating mechanism in Fennoscandia today, but the current stress field is likely a combination of plate boundary forces with local sources (e.g. glacial rebound and local geology). Large postglacial faults with a length varying from a few kilometres to tens of kilometres and a scarp height from a few metres to tens of metres occur in northern Finland and elsewhere in northern Fennoscandia. Small post-glacial faults (scarp height 0-20 cm) located in ice polished bedrock outcrops have been found in southern Finland, but so far larger postglacial faults have not been recognised. Investigations on postglacial bedrock movements have revealed that the postglacial faults studied so far are situated in old, reactivated fracture zones.

Key words: sedimentary rocks, occurrence, depositional history, tectonic regimes, Meso- and, Neoproterozoic, Phanerozoic, Finland, Fennoscandian Shield

TIIVISTELMÄ

Raportti on kirjallisuustutkimus, jossa on esitellään tunnetut meso- ja neoproterotsooiset sekä fanerotsooiset sedimenttikiviesiintymät Suomessa, arvioidaan niiden kerrostumishistoriaa ja tehdään yhteenveto tektonisista ja magmaattisista tapahtumista Fennoskandian kilven sisällä ja reunaosissa viimeisten 1200 – 1300 miljoonan vuoden aikana.

Mesoproterotsooisia ja neoproterotsooisia hiekka- ja savikiviä esiintyy vain muutamissa suojaisissa altaissa, kuten Satakunnassa, Muhoksella, Selkämerellä ja Perämerellä sekä meteoriitti-impaktirakenteissa, mutta alun perin ne ovat kattaneet huomattavasti laajemman alueen. Ala- ja keskikambrikaudella hiekka- ja savisedimenttejä luultavasti kerrostui suurimmassa osassa Etelä-, Länsi- ja Keski-Suomea. Ordovikikaudella karbonaatti-sedimenttejä kerrostui epikontinentaalisessa meressä, joka ajoittain laajeni ja jälleen vetäytyi. Selkämeren ja Ahvenanmeren altaat sekä Etelä- ja Lounais-Suomi olivat useaan otteeseen meren peittämiä, ja laajimmillaan meri ulottui Keski-Suomeen ordovikikauden keskivaiheilla. Myöhäispaleotsooisia (siluurikaudesta permikauteen) sedimenttikiviä ei ole löydetty, mutta apatiitin fissiojälkitutkimukset viittaavat suuren osan Fennoskandian kilvestä olleen niiden peittämiä. Sedimenttikuoren paksuudeksi Ruotsissa on arvioitu 3 – 4 km ja Suomessa 1 km. Tertiäärikaudella Itä-Lapissa kerrostui savia merellisissä olosuhteissa.

Mesoproterotsooisella kaudella hallitseva magmaattinen ja tektoninen tapahtuma oli svekonorjalainen vuorijonomuodostus noin 1250 – 900 miljoonaa vuotta sitten. Vuorijonomuodostuksen alkuvaiheen repeämisvaihetta edustavat Suomessa Satakunnan oliviinidiabaasijuonet. Orogenian jälkeen mantereet olivat järjestyneet yhdeksi suureksi Rodiniaksi kutsuksi suurmantereeksi, johon Fennoskandian kilpikin kuului. Se alkoi repeillä noin 600 miljoonaa vuotta sitten, jolloin manteiden väliin muodostui laaja valtameri. Mantereet alkoivat lähestyä toisiaan myöhäiskambrikaudella, mikä johti niiden törmäämiseen siluurikaudella, jolloin muodostui Kaledonidien vuorijono Fennoskandian kilven luoteisreunalle. Myöhäisemmän permikautisen, Keski-Euroopassa vallinneen variskilaisen orogenian liikunnot heijastuivat Fennoskandian kilven lounaisreunalle (ns. Törnquistin vyöhyke), jossa tapahtui siirrostumista ja purkautui diabaasijuonia. Fennoskandian kilven sisällä muodostui Oslon repeämälaakso. Mesotsooisella kaudella Törnquistin vyöhykkeen siirrokset aktivoituivat useaan otteeseen. Tertiäärikaudella Luoteis-Euroopan tektonista kehitystä hallitsivat Pohjois-Atlantin aukeneminen ja alppilainen vuorijonomuodostus Keski-Euroopassa. Fennoskandian kilven länsi- ja luoteisreunalla tapahtui maankuoren kohoamista. Fennoskandian kilven jännityskentässä on nykyään päätekijänä Atlantin keskiselänteeltä tuleva työntö, mutta siihen vaikuttavat myös jääkauden jälkeinen maankuoren kohoaminen ja paikalliset geologiset tekijät.

Mannerjään sulamisen jälkeen tapahtuneet ns. postglasiaaliset liikunnot sekä maankuoren nykyliikunnot ovat todisteena kallioperän nuorista liikunnoista. Suuria postglasiaalisia siirroksia (pituus muutamasta kilometristä kymmeniin kilometreihin ja siirrostörmän korkeus muutamasta metristä kymmeniin metreihin) löydettiin ensin Pohjois-Suomesta

1960-luvulla ja sen jälkeen muualta Pohjois-Fennoskndiasta. Etelä-Suomessa on tähän mennessä havaittu vain pieniä postglasiaalisia siirroksia (siirrostörmän korkeus 0–20 cm). Postglasiaalisten liikuntojen tutkimukset ovat osoittaneet siirrosten olevan vanhoja uudelleenaktivoituneita ruhjevyöhykkeitä.

Avainsanat: Sedimenttikivet, esiintyminen, kerrostumishistoria, tektoniikka, mesoproterotsooinen, neoproterotsooinen, fanerotsooinen, Suomi, Fennoskandian kilpi

TABLE OF CONTENTS

PREFACE

This study was carried out under contract to Posiva Oy (order 9562/02/LIW). On behalf of the client the work has been steered by Liisa Wikström. Seppo Paulamäki was the contact person on behalf of the authors.

The aim of the report is to summarize the sedimentary development and tectonic regimes within and at the margins of the Fennoscandian Shield during the last 1300 million years. Seppo Paulamäki is responsible for Chapters 1 to 4.6.3 and Aimo Kuivamäki for Chapter 4.6.4. The authors have collectively compiled Chapter 5. Within the Geological Survey of Finland, the report has been reviewed by PhD Jarmo Kohonen. The authors wish to thank him for his valuable and critical comments and suggestions. On behalf of Posiva, Krister Sundblad and Timo Kilpeläinen from the University of Turku have made comments on the report.

1 INTRODUCTION

1.1 Scope of study

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. The Government made a positive decision at the end of 2000, 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.

In 2002, Posiva published a report by Paulamäki et al. (2002), which summarised the essential existing geological and geophysical data on the southern Satakunta area, in which the Olkiluoto site lies, until the intrusion olivine diabase dykes about 1270 million years ago. The report tried to compile and interpret all available geological and geophysical data relevant to understanding the regional geological setting of the Olkiluoto site. The present report, carried out under contract to Posiva, is a continuation to the previous report and its purpose is to describe the geological history of the southern Satakunta area during the last 1200 - 1300 million years. Because hardly any evidence of the time after the deposition of the Satakunta sandstone and the emplacement of the diabase dykes is left at the area, this literature review covers the whole of Finland and the Fennoscandian Shield. Chapters 2 and 3 describe the occurrences of Meso- and Neoproterozoic and Phanerozoic sedimentary rocks in Finland and their depositional history. Chapter 4 summarises the major magmatic and tectonic events within and at the margins of the Fennoscandian Shield during the last 1200 - 1300 million years, i.e. since the intrusion of the diabase dykes at the onset of the Sveconorwegian orogeny in the south-western part of the Fennoscandian Shield. Although published only now, the report was written already in 2002. Consequently, no literature published after 2002 is used.

1.2 Geological time units; terminology

The geological terminology of the geological time units used in the report is presented in Table 1-1. The Precambrian Eon is divided into Archaean and Proterozoic Eras. The Proterozoic Era is subdivided into three major subdivisions, the Palaeoproterozoic (ca. 2500 - 1600, the Mesoproterozoic (1600 - 1000 Ma) and the Neoproterozoic (1000 - 543 Ma). The Mesoproterozoic and Neoproterozoic are divided into Riphean and Vendian Periods. Early (1600 - 1400 Ma) and Middle Riphean (1400 - 1000 Ma) belong to the Mesoproterozoic Era, while the Late Riphean (1000 - 650 Ma) constitutes the early phase of the Neoproterozoic. The term 'Jotnian', used in the text, refers to Middle Riphean anorogenic igneous rocks ('Subjotnian'), unmetamorphosed or weakly metamorphosed sedimentary rocks ('Jotnian') and 1270 - 1250 Ma old olivine diabases ('Postjotnian') within the Fennoscandian Shield. The latest period of the Neoproterozoic Era is called Vendian (also called Ediacaran).

The Phanerozoic Eon is divided into the Palaeozoic, Mesozoic and Cenozoic Eras. The Palaeozoic Era is subdivided into six Periods: the Cambrian (543 - 490 Ma), the Ordovician (490 - 443 Ma), the Silurian (443 - 417 Ma), the Devonian (417 - 360 Ma), the Carboniferous (360 - 290 Ma) and the Permian (290 - 248 Ma). The Mesozoic Era is divided into three time periods: the Triassic (248-206 Ma), the Jurassic (206-144 Ma), and the Cretaceous (144-65 Ma).

The Cenozoic Era (from 65 Ma until present) is divided into two Periods, the Tertiary and the Quaternary. The Tertiary Period is further subdivided into the Palaeogene and the Neogene (Table 1-2). The Paleogene is subdivided into three epochs, the Palaeocene (65-54 Ma), the Eocene (54-38 Ma) and the Oligocene (38-23 Ma), while the Neogene contains two epochs, the Miocene (23-5 Ma) and the Pliocene (5-1.6 Ma). The Quaternary have two epochs, the Pleistocene and the Holocene, the latter comprising the last 11 000 years.

Table 1-1. *Geological time units (Koistinen et al. 2001 and references therein). The age (Ma) refers to lower boundary of each period.*

EON		ERA	PERIOD		AGE (Ma)
PHANEROZOIC		CENO-ZOIC	QUATERNARY		1.635
			TERTIARY		65
		MESOZOIC	CRETACEOUS		144
			JURASSIC		206
			TRIASSIC		248
		PALAEOZOIC	PERMIAN		290
			CARBONIFEROUS		360
			DEVONIAN		417
			SILURIAN		443
			ORDIVICIAN		490
			CAMBRIAN		543
PRECAMBRIAN	PROTEROZOIC	NEO	VENDIAN		650
			RIPHEAN	LATE	1000
		MESO		MIDDLE	1400
				EARLY	1600
		PALAEO			2500
	ARCHAEAN				4000

Table 1.2. *The subdivision of the Cenozoic Era (adapted from http://www.ucmp.berkeley.edu/cenozoic/cenozoic.html)*

ERA	PERIOD		EPOCH	AGE (Ma)
CENOZOIC	QUATERNARY		HOLOCENE	0.1
			PLEISTOCENE	1.6
	TERTIARY	NEOGENE	PLIOCENE	5
			MIOCENE	23
		PALAEOGENE	OLIGOCENE	38
			EOCENE	54
			PALAEOCENE	65

In Chapters 2 and 3 describing the Lower Palaeozoic rocks and the sedimentary development, Estonian terminology of the Cambrian and Ordovician systems is used. Table 1-3 shows the major subdivisions of the Cambrian and Ordovician Periods in Estonia and the comparable classification in the British Isles. In Estonian classification, the Cambrian consists of eight regional stages, while the Ordovician is subdived into 19 regional stages.

Table. 1-3. *Cambrian and Ordovician Estonian regional stages (Mens et al. 1987, Männil & Meidla 1994, Nõlvak 1997).*

Period	British Series	Estonian regional series	Estonian regional stages
UPPER ORDOVICIAN	ASHGILL	HARJU	PORKUNI
			PIRGU
			VORMSI
	CARADOC		NABALA
MIDDLE ORDOVICIAN		VIRU	RAKVERE
			OANDU
			KEILA
			JOHVI
			IDAVERE
			KUKRUSE
	LLANVIRN		UHAKU
			LASNAMÄGI
			ASERI
		OELAND	KUNDA
LOWER ORDOVICIAN	ARENIG		VOLKHOV
			BILLINGEN
	TREMADOC		HUNNEBERG
			VARANGU
			PAKERORT
UPPER CAMBRIAN			ULGASE
MIDDLE CAMBRIAN			PANERIAI
			KIBARTAI
			RAUSVE
LOWER CAMBRIAN			VERGALE
			TALSY
			LONTOVO
			ROVNO

2 SEDIMENTARY ROCK OCCURRENCES IN FINLAND

2.1 Introduction

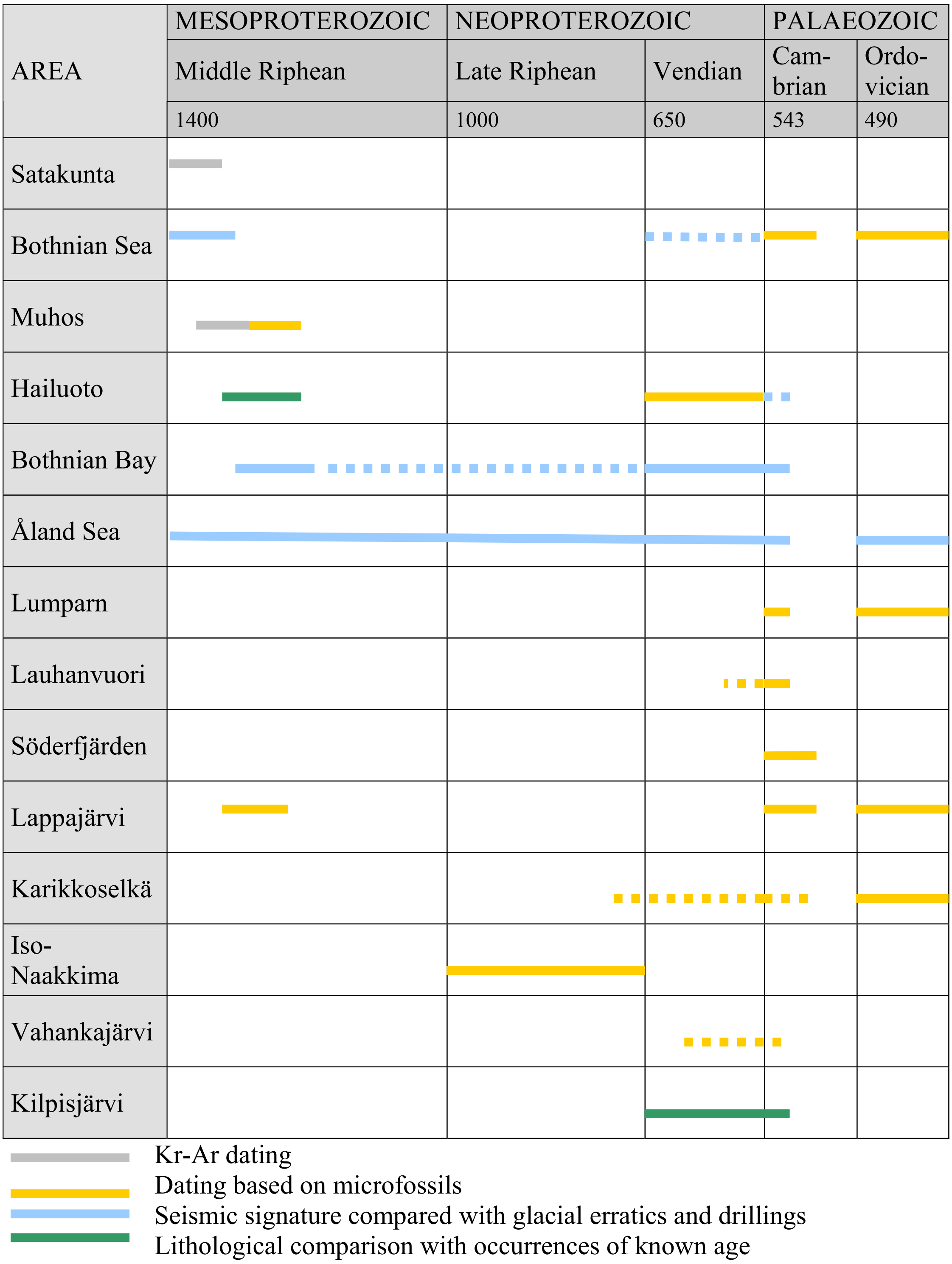
In Chapter 2, the main characteristics of the Meso-/Neoproterozoic and Phanerozoic (mainly Lower Palaeozoic) sedimentary rock occurrences in Finland are described. Table 2-1 lists the occurrences with their main dating criteria. Figure 2-1 shows the location of the main sedimentary rock occurrences in Finland.

In Satakunta and Muhos, the age of the sedimentary rocks are based on K-Ar datings giving ages of 1400 - 1300 Ma and ca. 1300 Ma, respectively (Simonen 1960; Kouvo 1976). In Satakunta, the sandstone is cut by 1270 - 1250 Ma old (Suominen 1991) olivine diabase dykes and sills.

In Hailuoto, Bothnian Sea, Lumparn, Lauhanvuori, Söderfjärden, Lappajärvi, Karikkoselkä, Iso-Naakkima and Vahankajärvi, the sedimentary rocks are dated by reference to microfossils in the drill core samples, which were compared with microfossils reported from known late Precambrian and Palaeozoic deposits, especially in Estonia and Sweden. Also lithological comparisons are used, as in the case of Middle Riphean rocks of Hailuoto, which are comparable with the rocks of Muhos. The stratigraphy of the Kilpisjärvi sedimentary sequence overlain by Caledonian nappes is comparable with the Vendian to Lower Cambrian Dividal Group in Norway and Sweden.

In the Bothnian Bay, Åland Sea and in the main part of the Bothnian Sea, no proved deposits of sedimentary rocks have been found but their occurrence are based on reflection and refraction seismic investigations and echo soundings. Based on the seismic velocities, the sedimentary bedrock has been divided into several seismic units. The lithology and stratigraphic age of each unit is based on comparison with sedimentary deposits onland or glacial erratics of sedimentary rocks.

Table 2-1. *Meso-/Neoproterozoic and Lower Palaeozoic sedimentary rock occurrences in Finland and their main dating criteria .*



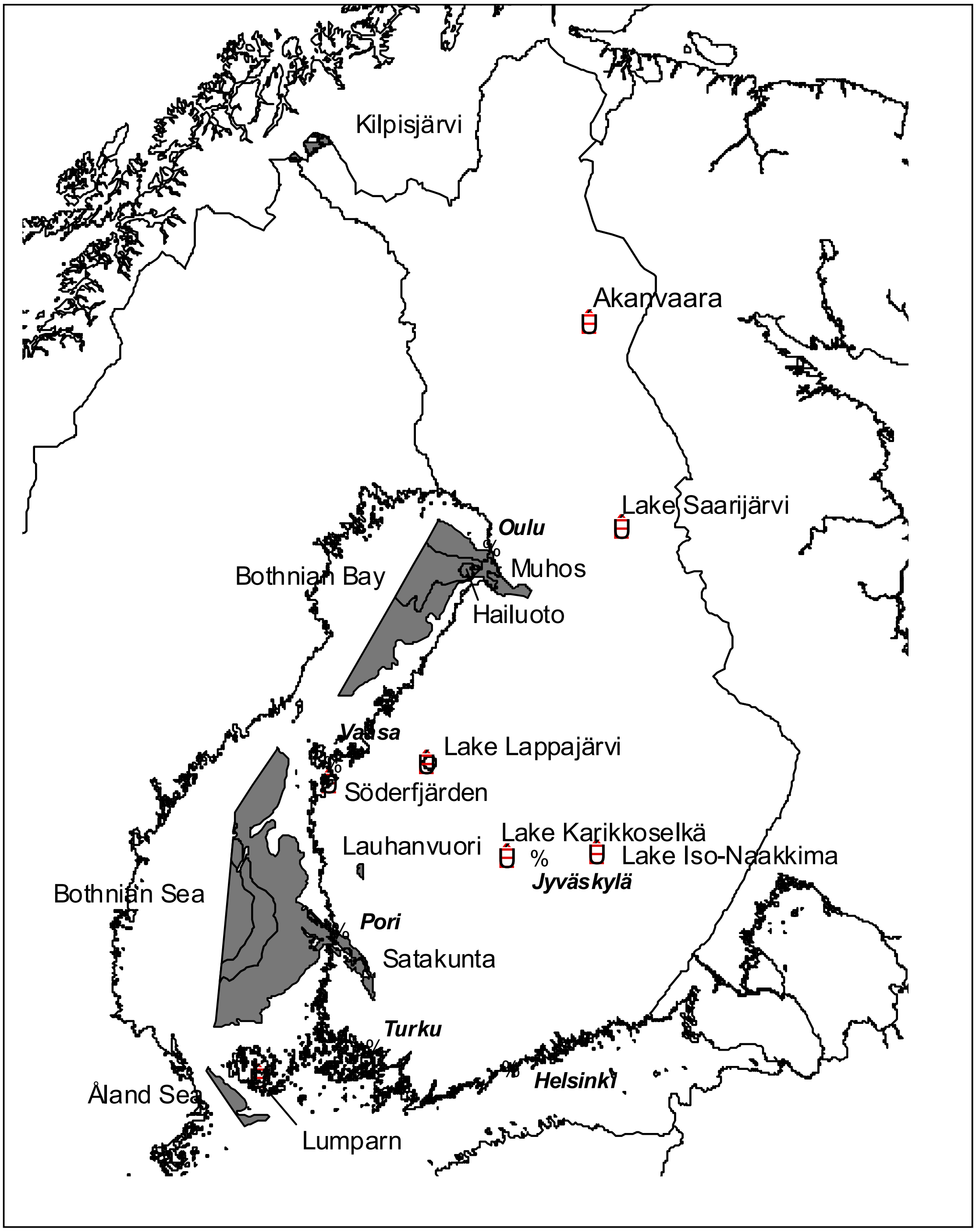


Figure 2-1. Location of sedimentary rock occurrences in Finland, excluding the erratics and the sandstone dykes. Compiled from Koistinen et al. (2001).

2.2 Occurrences

2.2.1 Bothnian Sea basin

Bothnian Sea

The Bothnian Sea marine basin is located in the southern part of the Gulf of Bothnia, the northern embayment of the Baltic Sea (Fig. 2-1). It is composed of several faulted blocks, bounded by NE-, NNE- and NW-trending lineaments (Winterhalter 1972, Axberg 1980, Winterhalter et al. 1981). The present depositional basin is filled with sedimentary rock ranging from Middle Proterozoic to Lower Palaeozoic (Cambrian and Ordovician) (Figs. 2-2 to 2-4).

The Middle Riphean ('Jotnian') sedimentary rocks are present below the Palaeozoic rocks as a submarine continuation of the 'Jotnian' sandstone formations in Satakunta (see below) and Gävle area in the east coast of Sweden. Seismic refraction survey indicates thickness of 900 m onshore the Gävle area in Sweden (Gorbatschev 1967) and over 1000 m offshore Pori (Winterhalter 1972). In the Sylen shoal in the south-western part of the Bothnian Sea, however, the thickness of the sandstone may only be about 100 m (Winterhalter, op. cit.), and in the Finngrundet area south of the Sylen shoal, the Palaeozoic sedimentary rock are lying directly on crystalline basement (Thorslund & Axberg 1979). On the other hand, the deep seismic surveys indicate that the northern part of the Bothnian Sea Basin (named the Strömminsbådan basin in Korja et al. 2001) is considerably deeper, up to 3 - 4 km (BABEL WG 1993, Korja et al. 2001). The areas consisting of 'Jotnian' sandstone are characterised by rather flat sea bottom topography, the sandstone dipping about 4° towards southwest (Winterhalter 1972; Axberg 1980).

The presence of Vendian rocks in the Bothnian Sea is uncertain. However, some weak, unidentified seismic reflectors in northern Bothnian Sea are, according to Axberg (1980), referable to beds of possible Neoproterozoic/Vendian age. Söderberg (1993) have found similarities in the seismic signatures between the inferred Vendian rocks in the Åland Sea (*see* Chapter 2.2.3) and the proposed Vendian beds in the Bothnian Sea. However, similar reflections have not been recorded in the southern part of the Bothnian Sea (Axberg 1980).

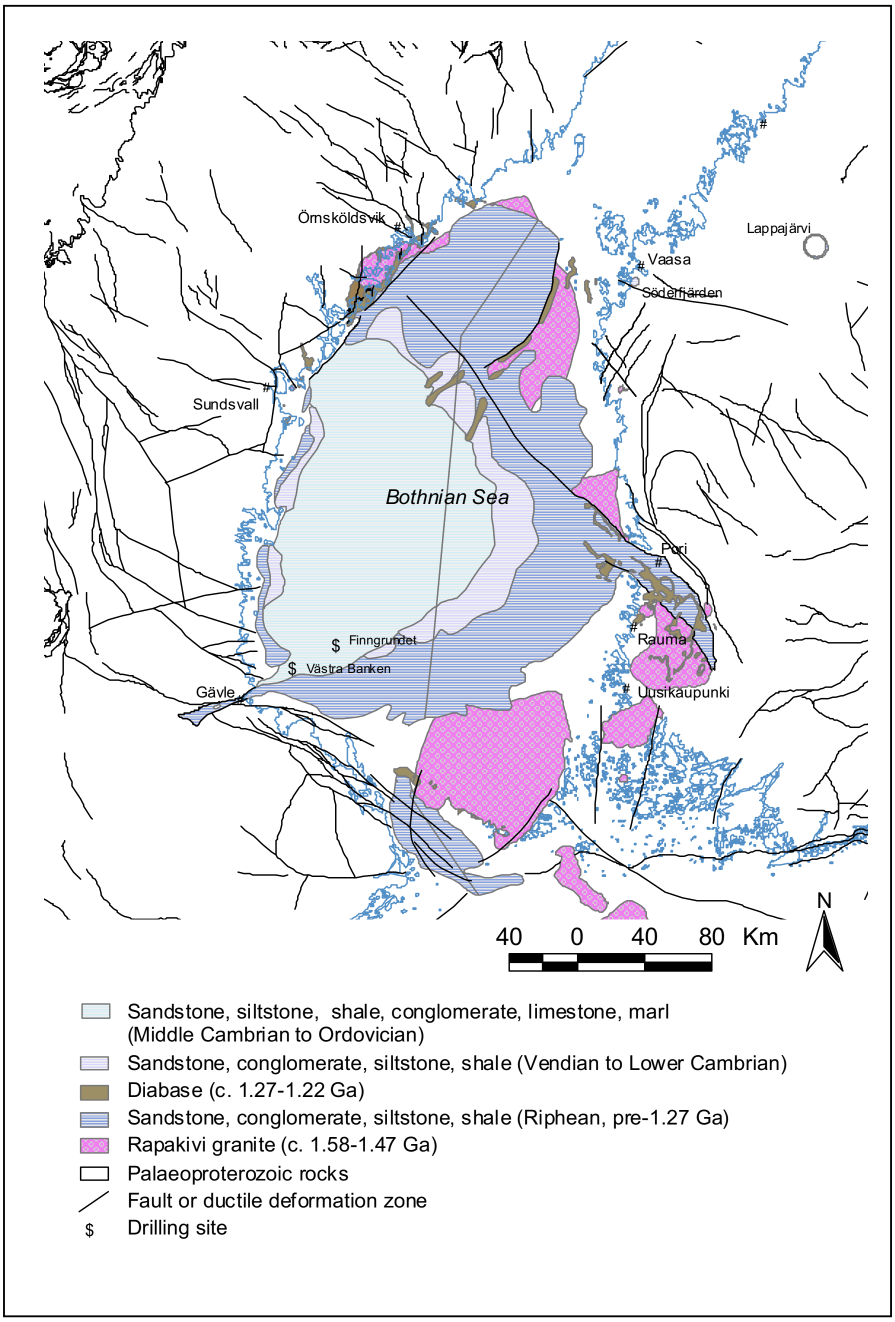


Figure 2-2. *Lithology of the Bothnian Sea. Simplified from Koistinen et al. (2001).*

Based on the drillings at Finngrundet and Västra Banken, in the Swedish part of the southern Bothnian Sea (Fig. 2-2), the lowermost parts of the Palaeozoic sedimentary sequence in the Bothnian Sea Basin is composed of Lower Cambrian clays, claystones, siltstones, sandstones, shales, arkoses and conglomerates overlain by Ordovician limestones Fig. 2-3) (Thorslund & Axberg 1979; Axberg 1980). Based on the seismic soundings, the thickest Lower Cambrian sequences, 165 m, occur in the central part of the Bothnian Sea, from where the preserved thickness rapidly decreases towards the south and southeast, (being about 41 m in the Finngrundet borehole) and less rapidly towards north and northeast (Axberg 1980).

Previously, no Middle or Upper Cambrian rocks were recognized in the Bothnian Sea Basin (*see* Axberg 1980). Later, however, Hagenfeldt (1988, 1989b) reported Middle Cambrian acritarch assemblages from the Finngrundet borehole. According to Hagenfeldt (1989b), the thickness of the Middle Cambrian sequence in the borehole is 37 m, consisting of siltstones and shales.

The Ordovician strata lies upon the Cambrian rocks without an angular discordance. The Ordovician sedimentary rocks have their interpreted maximum thickness (230 m) in the central part of the Bothnian Sea (Axberg 1980) (Fig. 2-4). According to the boreholes at Finngrundet and Västra Banken, the Lower Ordovician sequence mostly consists of limestones (Thorslund & Axberg 1979; Axberg 1980).

The lowermost part of the sequence is formed by ca. 1 m thick early Tremadokian (Lower Ordovician) bed, consisting of sandstone, alum shale and limestone (stinkstone). It is overlain by about 3 m of Lower Ordovician Billingen regional stage (*see* Table 1-3) limestones, 19 m of Lower Ordovician Volkhov regional stage limestones and 20.8 m limestones belonging to the Lower to Middle Ordovician Kunda regional stage (Tjernvik & Johansson 1979; Löfgren 1985). Based on the seismic studies, the maximum thickness of the Lower Ordovician in the Bothnian Sea Basin is 95 - 100 m (Axberg 1980).

Middle and Upper Ordovician sequences are missing in the Finngrundet and Västra Banken boreholes (cf. Thorslund & Axberg 1979). However, on the basis of seismic soundings, Axberg (1980) has interpreted that Middle Ordovician sequences composing of different kinds of limestones are present in the central Bothnian Sea with a thickness of 60 - 65 m. The confirmed deposit of fine-grained, green limestone in the Sylen shoal, in the southern part of the Bothnian Sea, is, according to microfossils, probably from the upper part of the Middle Ordovician (Winterhalter 1967; Tynni 1975).

According to seismic soundings, the Upper Ordovician sedimentary rocks can be divided into three units (Axberg 1980). The lowermost unit consists of limestone, commonly known as the Baltic limestone, and its thickness is 25 - 30 m. The middle part is about 30 m thick, and tentatively consists of shales. Less than 10 m thick uppermost unit has been interpreted to consist of marl or marly limestone tentatively referred to lower Ashgill of the British series (Axberg 1980), corresponding Nabala to Pirgu regional stages in Estonia (*see* Table 1-3).

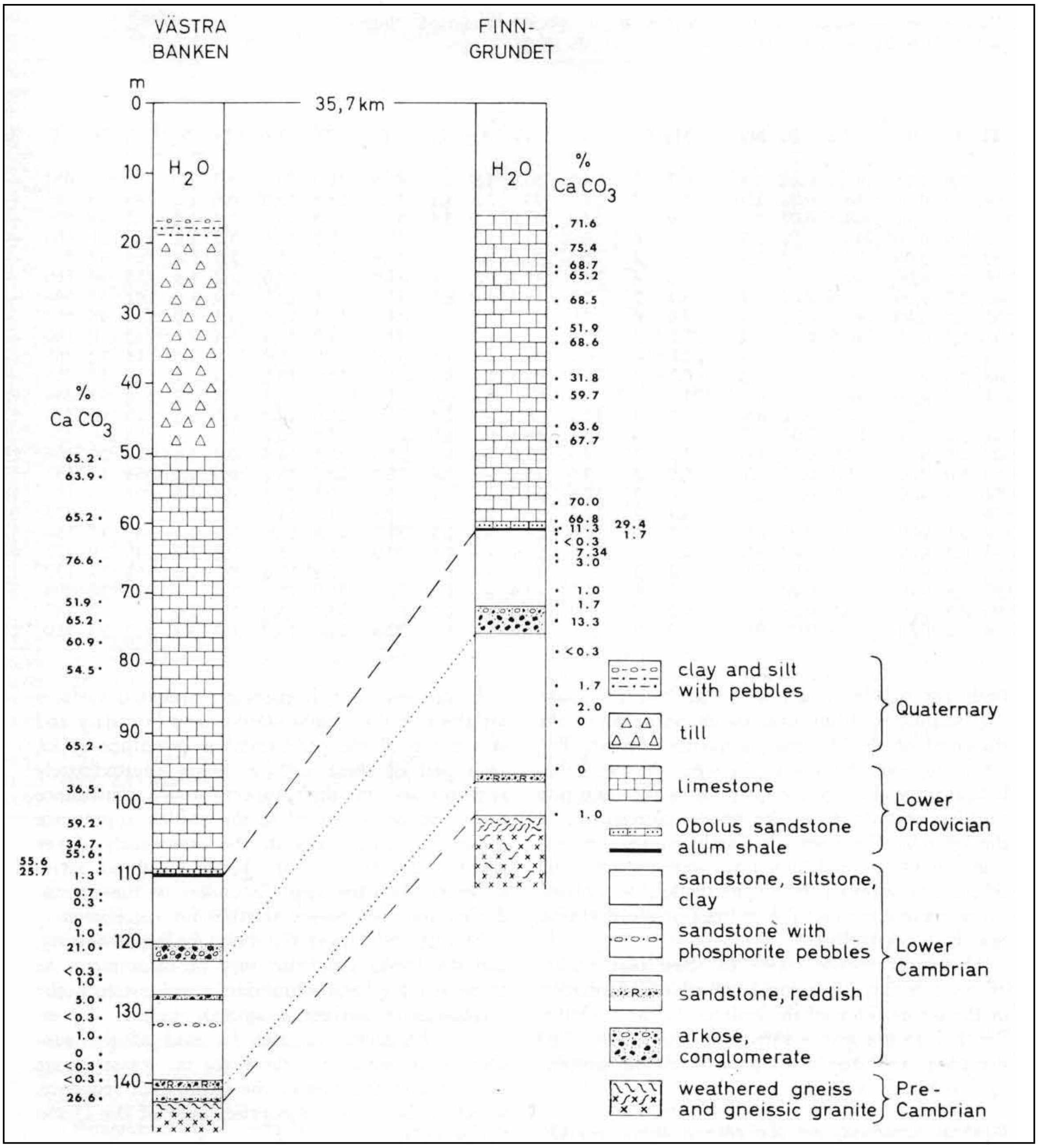


Figure 2-3. *Drill core sections in Finngrundet and Västra Banken and their correlation (Thorslund & Axberg 1979).*

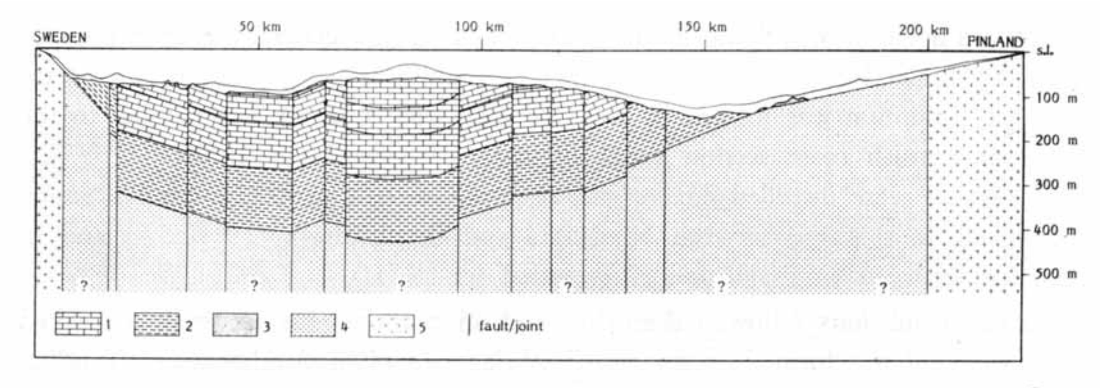


Figure 2-4. *Geological cross-section across the widest part of the Bothnian Sea (Axberg 1980). 1. Lower, Middle and Upper Ordovician, 2. Lower Cambrian, 3. outliers of Palaeozoic rocks, 4. Middle Riphean ('Jotnian'), 5. Proterozoic crystalline bedrock.*

Satakunta sandstone

The Mesoproterozoic (Middle Riphean) Satakunta sandstone is located in south-western Finland near the town of Pori (Fig. 2-1) and forms a 70 km long and 20 km wide, NW-SE-trending graben structure bounded by subvertical faults. The crystalline bedrock of the surrounding area is composed of Palaeoproterozoic (1900 - 1880 Ma) supracrustal and plutonic rocks. Anorogenic Laitila rapakivi batholith (age ca. 1580 Ma) is located south of the sandstone formation. The sandstone is cut by olivine diabase dykes and sills, dated at 1270 - 1250 Ma (Suominen 1991).

The thickness of the sandstone is, according to a borehole located south of Pori, at least 600 m, but the gravimetric studies indicate that the maximum thickness could be as much as about 1800 m (Elo 1982). In the south-eastern part of the sandstone, the thickness is probably less than 200 m (Elo et al. 1993). The sandstone is very poorly exposed, the only contact observations being from the contacts between the sandstone and the olivine diabase dykes and sills. The sandstone formation continues to the Bothnian Sea sedimentary basin, which is mainly composed of Mesoproterozoic ('Jotnian') sandstone (Axberg 1980; Winterhalter et al. 1981). Offshore, northwest of Pori, the sandstone is interpreted to be underlain by marine continuation of the Reposaari (Pori) rapakivi granite (Korja et al. 2001).

In the sandstone, the sedimentary materials range from massive conglomerates and pebbly sandstones over moderately sorted, medium to coarse-grained sandstones 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 south-western contact of the sandstone in Luvia (Kohonen et al. 1993).

The primary structures of the outcrops indicate deposition in a 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. The layering of the sandstone is predominantly subhorizontal, but in the north-eastern contact the sandstone layers are dipping about 35° to the southwest, towards the centre of the basin. Simonen and Kouvo (1955) interpret the red colour of the sandstone to be indication of a terrestrial, oxidising environment. However, according to Kohonen et al. (1993) the colouring may be partly epigenetic in origin.

The upper part of the Satakunta sandstone has been interpreted to be a fluvial sediment formation deposited in an alluvial environment by poorly channelled braided streams (Fig. 2-5) before the outcropping of the rapakivi granites (Marttila 1969, Kohonen et al. 1993). 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 Middle Riphean rifting period, ca. 1650 Ma ago (Kohonen et al. 1993).

It was formerly reported (Deutsch et al. 1998) that the Satakunta sandstone hosts the world's oldest known micrometeorites. Recently, however, the results of this study have been questioned and human failure cannot be excluded (Deutsch et al. 2002).

'Jotnian' sandstone erratics occur in Bromarv and south of the town of Salo in south-western Finland, ca. 150 km southeast from the Satakunta sandstone (Donner 1996). Donner (op. cit.) sees it most likely that these erratics come from a separate local source and not from Satakunta or from the bottom of the Bothnian Sea. Matisto (1964) discusses the possibility of a local source area for the sandstone erratics in Tyrvää, some 30 km east of Satakunta sandstone.

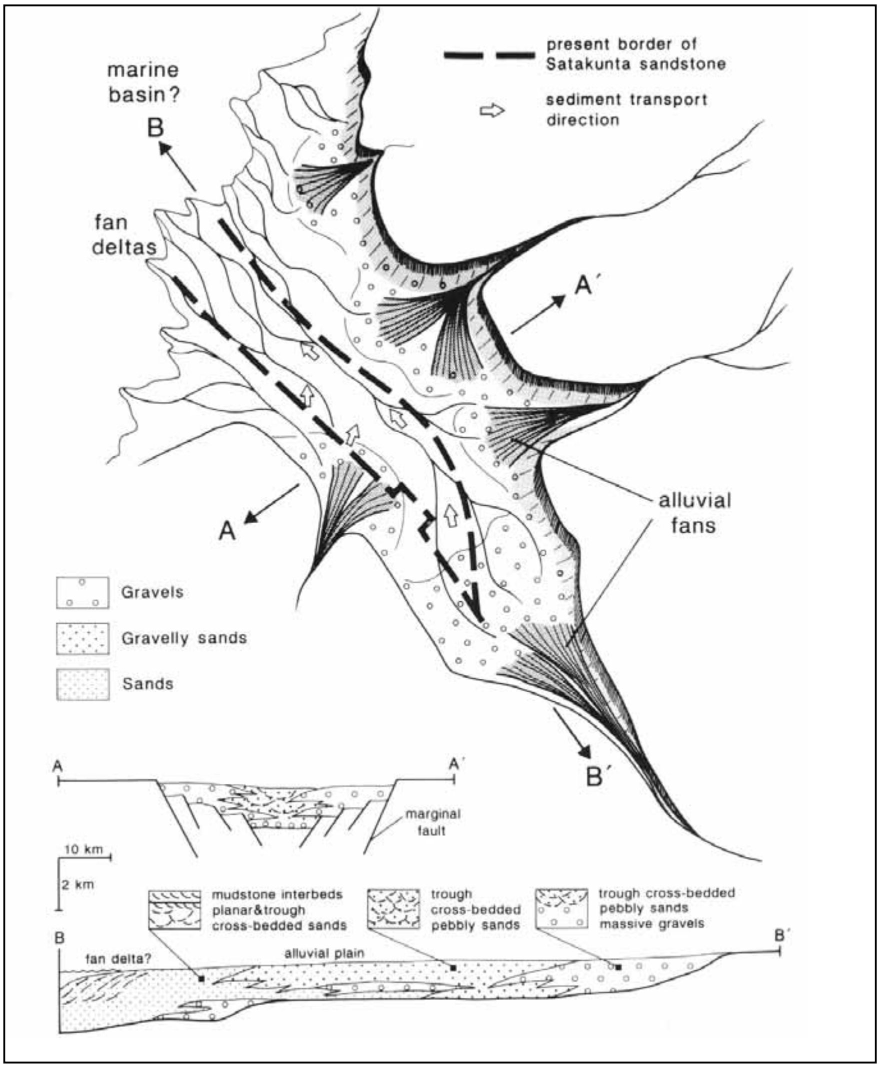


Figure 2-5. *The possible palaeogeography of the Satakunta sandstone basin 1400 Ma ago (Kohonen et al. 1993).*

2.2.2 Bothnian Bay

Muhos Formation

The unmetamorphosed sedimentary rocks of the Muhos formation occur in a 50 km long and 20 km wide graben-like basin, located south of Oulu (Fig. 2-1). The sedimentary rocks have been downfaulted along the northern boundary (Simonen & Kouvo 1955). The downfaulting of the deposits have protected the rocks from erosion.

Thick Quaternary deposits overlie the Muhos formation and only one outcrop is known. Three boreholes have been drilled in the formation, the deepest of them penetrating 894.50 m of sedimentary rocks and reaching the Palaeoproterozoic bedrock at ca. 977 m (Kalla 1960). The lowermost part of the sedimentary sequence is composed of about 20 m thick layer of basal conglomerate and coarse-grained arkosic sandstone, lying directly on the Palaeoproterozoic bedrock (Simonen & Kouvo 1955; Tynni & Uutela 1984) (Fig. 2-6). They are overlain by red-coloured siltstones and shales with arkosic sandstone interbeds. The only outcrop at Kieksi consists of less than 10 m thick bed of polymictic conglomerate, comparable to the basal conglomerate in the borehole. According to Simonen & Kouvo (op. cit.), the sedimentary formation was deposited in a floodplain environment in oxidizing, terrestrial conditions.

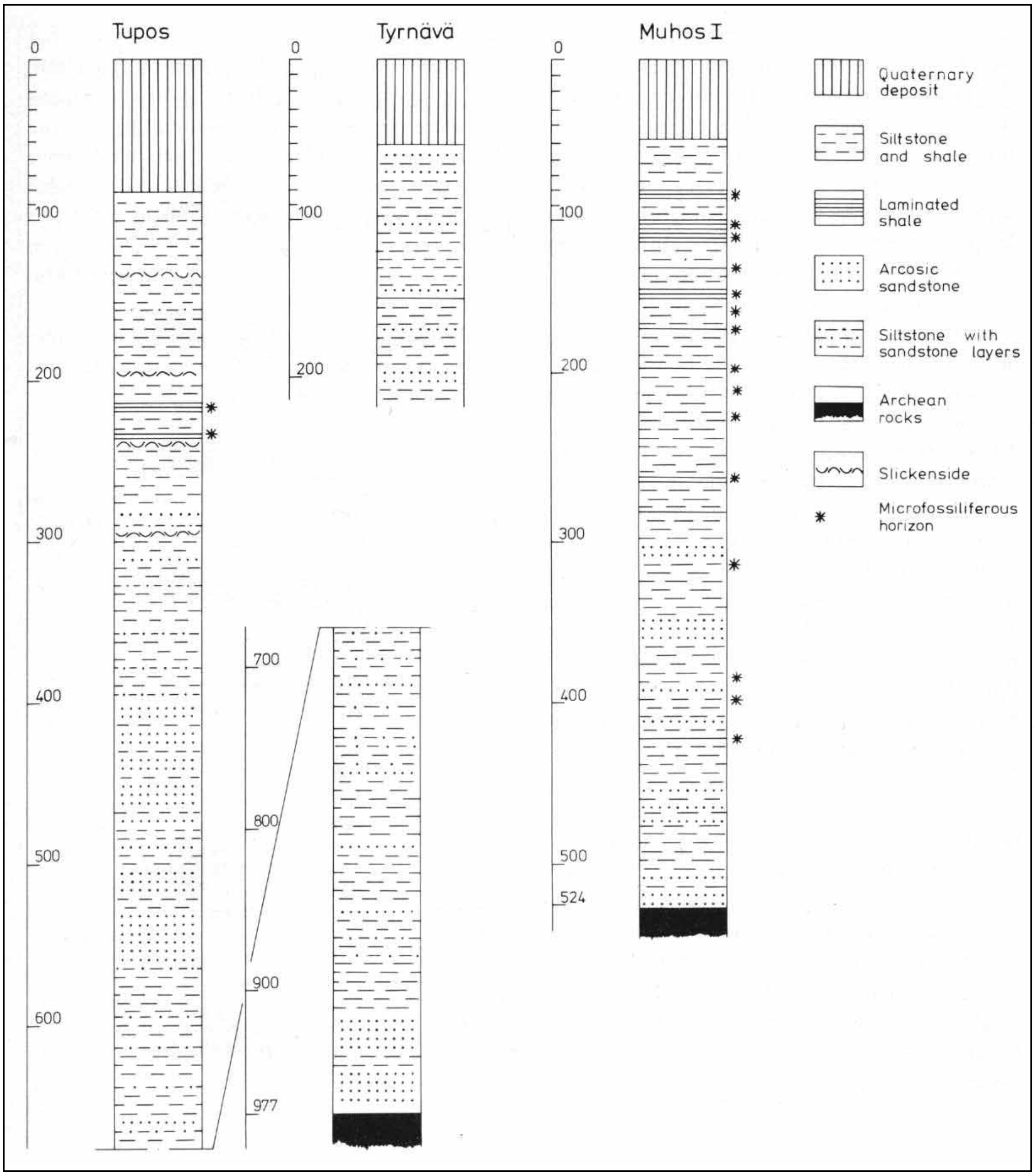


Figure 2-6. *Lithologies in the borehole sections of the Muhos formation (Tynni & Uutela 1984).*

The Muhos formation was formerly correlated in time with the Middle Riphean ('Jotnian') Satakunta sandstone, based on the corresponding petrographic properties (Simonen & Kouvo 1955) and on K-Ar-dating, which yield an age of ca. 1300 Ma for the diagenesis

(Simonen 1960). However, microfossil studies (Tynni 1978; Tynni & Uutela 1984) indicate that the formation may be about 100 Ma younger, ca. 1200 Ma.

Clastic dykes consisting of a mixture of arkose, red siltstone and granitic fragments have been found in granitic basement rocks at Oulu, Kempele and Tyrnävä, near the north-eastern margin of the Muhos formation (Kesola 1981). According to Kesola (op. cit.) the material of the dykes is composed of rocks of the Muhos Formation.

Hailuoto Formation

The island of Hailuoto is located in the northern part of the Bothnian Bay, 20 km west of the town of Oulu (Fig. 2-1). No outcrops are present on the island, but the whole island is covered by sand.

According to three boreholes drilled at Hailuoto, the sedimentary rocks can be divided into two units. The lower unit, comparable to the Middle Riphean ('Jotnian') Muhos formation, east-southeast of Hailuoto, has deposited directly on the Palaeoproterozoic Svecofennian bedrock, and is composed of basal conglomerate, overlain by red arkosic sandstone and red shale (Fig. 2-8) (Veltheim 1969; Tynni & Donner 1980; Kousa & Lundqvist 2000). The possible sedimentation environment was a shallow sea (Tynni & Donner, op. cit.).

The arkosic sandstone of about 60 - 100 m thick upper unit (the Hailuoto formation) has been deposited on the red shale, and is overlain by alternating beds of siltstone and shale (Fig. 2-8). On the basis of microfossil study, Tynni & Donner (1980) have suggested an Upper Vendian age (about 600 Ma) for the upper unit. The sediments were presumably deposited in a shallow sea.

The borehole at Ryydys, located offshore about 5 km east of Hailuoto, penetrates over 90 m of siltstones, without reaching the basement (Uutela 1986). Based on the low content of microfossils in the Ryydys borehole and the Tupos borehole in the Muhos formation (Tynni & Uutela 1984), compared to the rich Upper Vendian microfauna in the Hailuoto Formation, Uutela (op. cit.) suggest that the sedimentary rocks at Ryydys are Mesoproterozoic in age and belong to the Muhos formation.

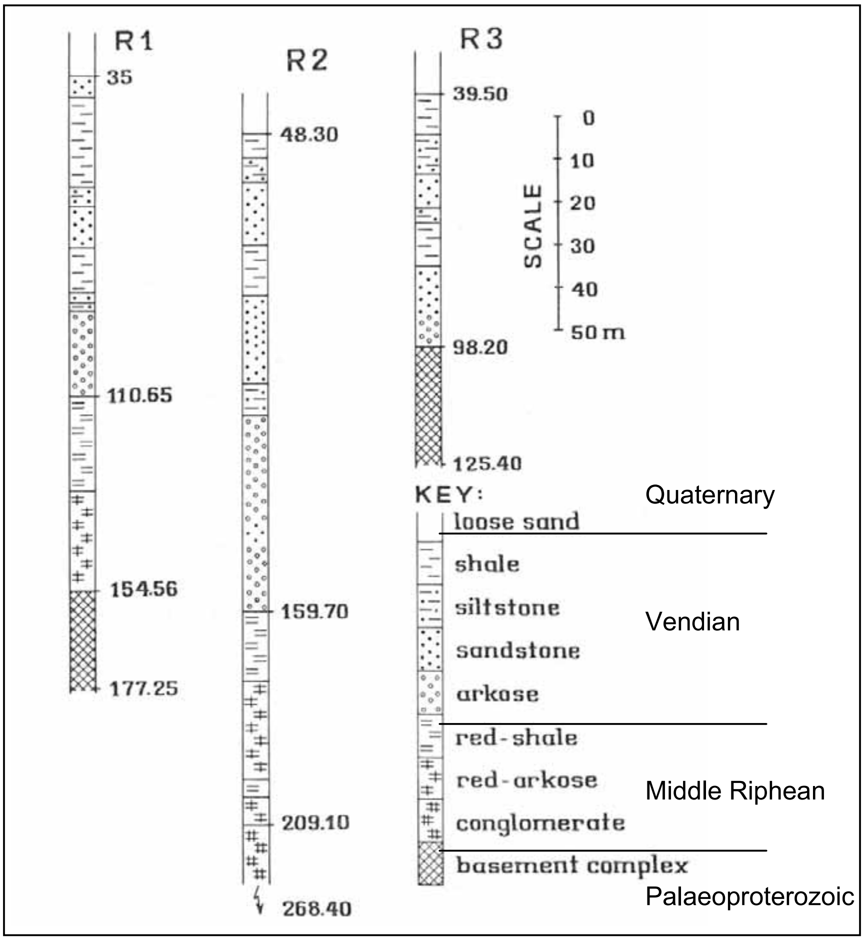


Figure 2-8. *Lithology of the boreholes at Hailuoto (modified after Veltheim 1969)*

Bothnian Bay area

The Bothnian Bay forms the northern part of the Gulf of Bothnia (Fig. 2-1). According to Wannäs (1989), the Bothnian Bay consists of several sediment-filled sub-basins, which have developed during several tectonic phases. No proved deposits of sedimentary rocks have been found but their occurrence is based on reflection and refraction seismic investigations.

Based on the seismic investigations, Wannäs (1989) subdivided the Bothnian Bay sedimentary rocks into four seismic units. The three lower units are interpreted as Meso- to Neoproterozoic, the uppermost unit being Cambrian in age (Fig. 2-9). The total thickness of the sedimentary rocks varies considerably in different parts of the Bay. The thickness of the sedimentary rocks is about 200 m in the main part of the Bay but in the northern part, in the Muhos basin, the thickness is up to 700 m.

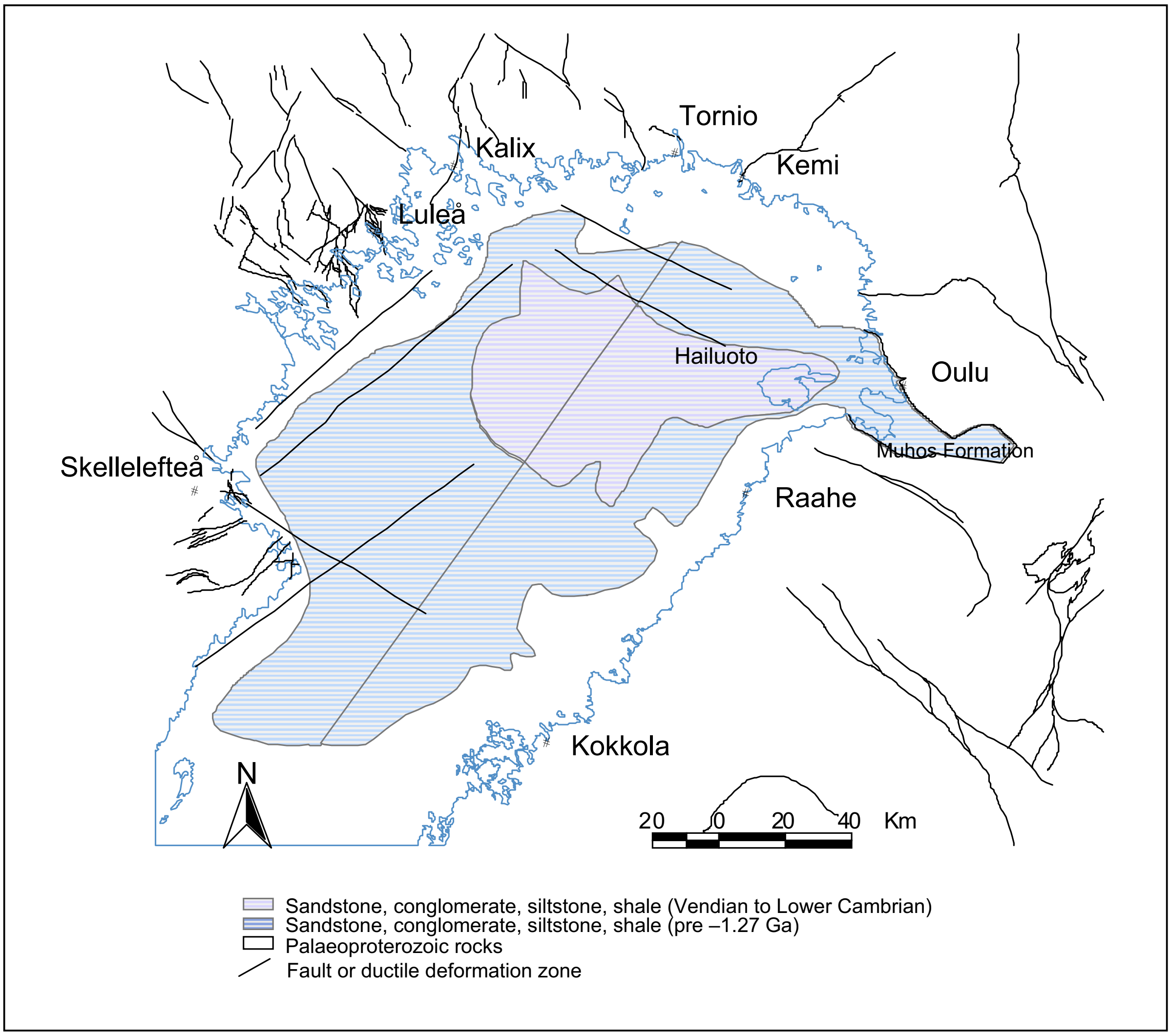


Figure 2-9. *Mesoproterozoic/Neoproterozoic (Middle/Upper Riphean) and Palaeozoic (Lower-Middle Cambrian) lithologies in the Bothnian Bay based on seismic reflection and refraction measurements (Wannäs 1989, Lundqvist et al. 1996).*

The 100 m thick lowermost unit (M1), interpreted to consist of sandstone, is comparable with the onshore Middle Riphean Muhos formation and extends across the entire Bothnian Bay (Wannäs 1989) (Fig. 2-10). The two overlying units (M2 and M3) form the main part of the so-called Muhos Basin, located northwest of the island of Hailuoto, and they are interpreted to be comparable with the Vendian rocks at Hailuoto. The unit M2 only occurs in the northern part of the Bothnian Bay, where it has a thickness of 150 - 350 m in the sub-basins and 0 - 50 m outside the basins. Wannäs (op. cit.) interprets the unit M2 consisting of conglomerate and sandstone with intervening siltstone and representing marginal wedges of coarse-grained submarine fans. The unit M3 forms the uppermost part of the Muhos basin, with a thickness of 150 - 200 m within the basin and 30 - 50 m outside the basin but also occurs over almost the entire Bothnian Bay area. It is interpreted to be composed of fine-grained sandstones with intervening siltstone and shale layers (Wannäs 1989).

Based on the similarities with the seismic reflections in the Bothnian Sea Wannäs (1989) has interpreted that Lower to Middle Cambrian sedimentary rocks occur in the north central part of the Bothnian Bay, with a maximum thickness of about 80 m. Seismically, the Cambrian rocks can be divided into two units, the upper one having a rather limited lateral extension. Based on the similar seismic velocities, Wannäs (1989) suggests that the uppermost rocks in the Vendian sedimentary sequence in northern part of the island of Hailuoto may be Cambrian in age.

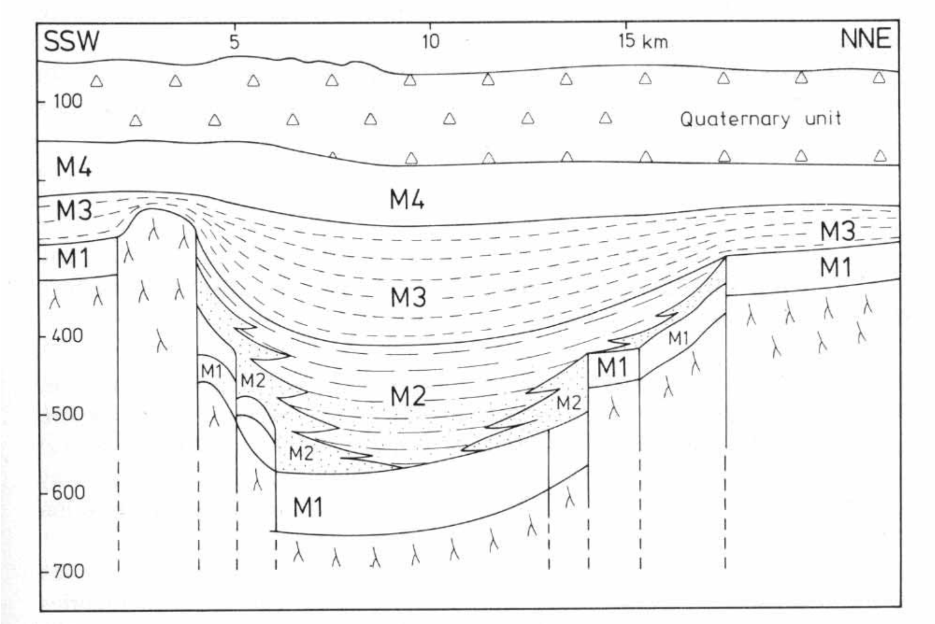


Figure 2-10. *NNE-SSW trending cross-sectional model of the Muhos basin (Wannäs 1989). Units M1-M4 are described in the text.*

No Ordovician sedimentary rocks have been found or interpreted from the seismic profiles in the Bothnian Bay. However, Uutela (1998) has studied the microfossils from some erratic limestone boulders in Ostrobothnia and suggests that the provenance of these Ordovician boulders is the Bothnian Bay, i.e. further north from the previously known Ordovician deposits in the Bothnian Sea. The boulders consist of reddish and greenish marls and yellowish grey Baltic limestone (calcilutite). On the basis of acritarchs (acritarchs = pollen-like remains of predecessors of present-day algae), the age of the yellowish grey limestone boulder is early Middle Ordovician (Lasnamägi regional stage).

No acritarchs exist in the marls, but the lithological features suggest them to be of Lower Ordovician age (Billingen, Volkhov and the beginning of the Kunda regional stage). The greenish marl is probably Middle to Upper Ordovician (Uutela 1998).

In Kauhajoki, western Finland, Ordovician acritarchs have been found in the till-covered Quaternary sand deposit (Kujansuu & Uutela 1997). Since the predominant transport direction of continental ice sheet in the area is from the northwest, the source of the sand is most likely the sedimentary rocks on the bottom of the Bothnian Bay (Kujansuu & Uutela, op. cit.). The occurrence of Ordovician sedimentary rocks in the Bothnian Bay is further supported by foundings of Eriksson et al. (1999) of reworked Ordovician acritarchs of the Lasnamägi regional stage in Eemian glacial deposits at Mertuanoja, Ostrobothnia.

2.2.3 Åland Sea basin

The Åland Sea marine basin between the Åland Islands and the mainland Sweden (Fig. 2-1) forms a semi-circular basin of an inferred Mesoproterozoic age (Söderberg 1993).

The occurrence of the sedimentary rocks in the Åland Sea basin is not confirmed but the marine seismic reflection and refraction studies and echo soundings, combined with frequency counts of glacial erratics, indicate that sedimentary rocks do occur, which rather continuously range from Middle Riphean to Upper Ordovician (Söderberg 1993, Hagenfeldt 1995). The interpreted sequence of the sedimentary rocks is: 1. Middle Riphean (‘Jotnian’) sandstone, 2. Upper Riphean to Vendian sandstone and siltstone, and 3. Palaeozoic rocks, composed of Lower Cambrian sandstones and Lower to Upper Ordovician limestones (Söderberg 1993; Hagenfeldt 1995) (Figs. 2-11 and 2-12).

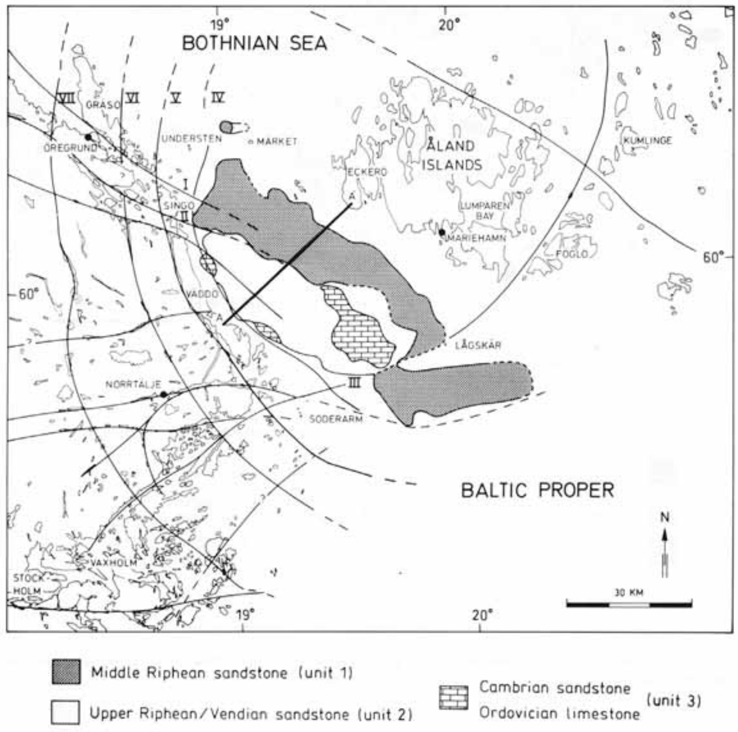


Figure 2-11. *Distribution of sedimentary rocks in the Åland Sea Basin (Söderberg 1993). I-VII = tectonic zones, faults and fracture valleys. Profile A-A' is shown in Fig. 2-12.*

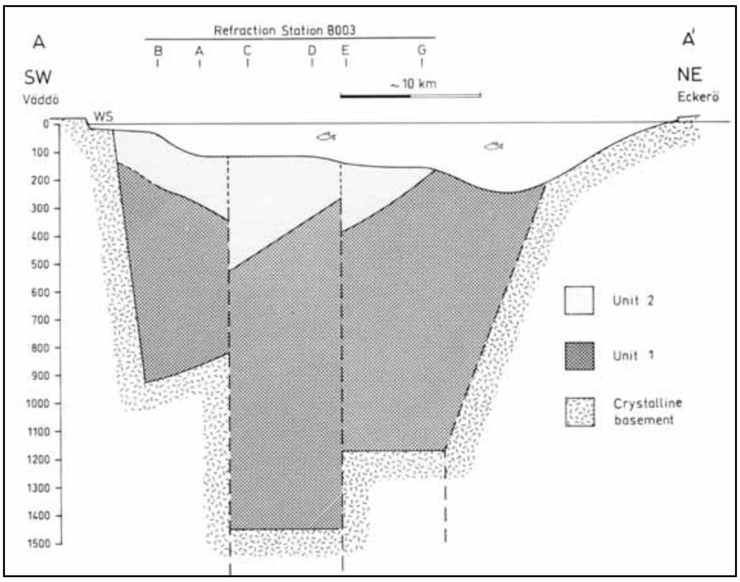


Fig. 2-12. *Geological cross-section across the Åland Sea based on refraction seismics (Söderberg 1993). Units 1 and 2 are block faulted.*

The reddish Middle Riphean sandstone makes up the basal part of the sedimentary bedrock sequence with an assumed maximum thickness of about 1200 m (Söderberg 1993). The sandstone is comparable with sandstone of Satakunta and sandstones of Dalarna, Gävle and Nordingrå in Sweden. However, the Åland sandstone contains less feldspar, which indicates that the erratics represent only the uppermost part of the sandstone (Hagenfeldt 1995).

The late Riphean to Vendian sandstones consist of feldspar-rich coarse-grained conglomerates and clasts of the 'Jotnian' sandstone and rock fragments of crystalline basement (Söderberg 1993). Also the dark shale erratics, found in the Stockholm Archipelago, are thought to belong the unit 2 (Hagenfeldt 1995). The distribution of the sandstones is restricted to the western part of the Åland Sea and has a maximum thickness of some 400 m, according to seismic investigations (Söderberg, op. cit.).

No Palaeozoic rocks have been detected either but the distribution of the rocks is solely based on seismic refractions and glacial boulders. The unit 3 has an inferred maximum thickness of less than 350 m, and it is restricted to an area in the south-central part of the Åland Sea Basin. Based on the seismic survey, the thickness of the Lower Cambrian beds is about 120 m, consisting, according to glacial boulders, of sandstones and siltstones.

Seismic studies indicate that also Middle Cambrian sequences may be present in the Åland Sea, although none of such rocks occur in the glacial drift (Söderberg 1993). The reason for this may the softness of the Middle Cambrian shales of *Oelandicus* beds.

The Lower Ordovician is composed of *Orthoceratite* limestone, which is comparable to Lower Ordovician limestones in the Bothnian Sea (Söderberg 1993). Hagenfeldt (1995) have also found erratics of Lower Ordovocian (Tremadokian) *Obolus* conglomerate, and suggests that it is present *in situ* in the Åland Sea area. Corresponding rocks are found, for example, in the Finngrundet drilling in the Bothnian Sea (*see* p. 16).

On the basis of seismic soundings, the occurrence of Middle Ordovician limestones in the Åland Sea is uncertain, and they are only poorly represented by erratics. However, judging from the inferred total thickness of the Lower Palaeozoic in the Åland Sea (350 m), Söderberg (1993) believes that Middle Ordovician sequences are present in the area.

Large number of erratics of the Upper Ordovician dolomitic Baltic limestone occurs in the Stockholm Archipelago suggesting *in situ* occurrence of such limestone in the Åland Sea (Hagenfeldt & Söderberg 1994; Hagenfeldt 1995). The limestone is possibly comparable to the upper limestone of the Lumparn basin (*see* Chapter 2.2.4), and can be correlated with the Rakvere, Nabala and Pirgu regional stages of Estonia (Hagenfeldt 1995). Dolomitic limestone erratics in the northern part of the Åland Sea probably originate from the Bothnian Sea (Hagenfeldt 1995). The total thickness of Ordovician sequences is assumed to be about 230 m.

2.2.4 Lumparn impact structure

The Lumparn impact structure is located in the main island of the Åland archipelago about 60 km southwest of the Finnish coast (Fig. 2-1). The rhomb-shaped structure has an area of ca. 80 km^2. The country rock is 1590 - 1570 Ma old rapakivi granite.

The distribution of confirmed sedimentary rocks in the Lumparn impact structure is shown in Fig. 2-13. The sedimentary rock sequence is up to 120 m thick. According to drillings, an arkosic basal breccia (conglomerate) is present in the bottom of the Palaeozoic sequence (Asklund & Kulling 1926; Winterhalter 1982; Bergman 1982). It consists of weathered rapakivi granite loosely cemented by red iron oxides or hydroxides and reddish calcite. Asklund & Kulling (1926) and Winterhalter (1982) favour a Mesoproterozoic (Middle Riphean) age for the breccia but according to Bergman (1982), the loose consolidation indicates younger, evidently sub-Cambrian age.

The lower part of the Palaeozoic sequence consists of Lower Cambrian sandstones and fewer siltstones. The sandstone is only poorly lithified (i.e., it has undergone only a very weak diagenesis) compared to the clastic dykes in the Åland archipelago and it is redder closer to the rapakivi granite country rock (Bergman & Lindberg 1979; Lehtovaara 1982a).

The upper part of the sequence consists of tens of metres thick limestone, which can be divided into two types (Merril 1980; Tynni 1982b). The lower limestone is greenish grey, the lowermost parts being glauconitic in composition. The lower limestone contains microfossils, which place it to the Middle Ordovician Lasnamägi regional stage of Estonia (Uutela & Sarjeant 2000, p. 33). In the upper part, the limestone (calcilutite) is light grey, hematitic, fine-grained and densely packed, and it is comparable with the limestones of the Middle Ordovician Rakvere regional stage and Upper Ordovician Pirgu regional stage in Estonia.

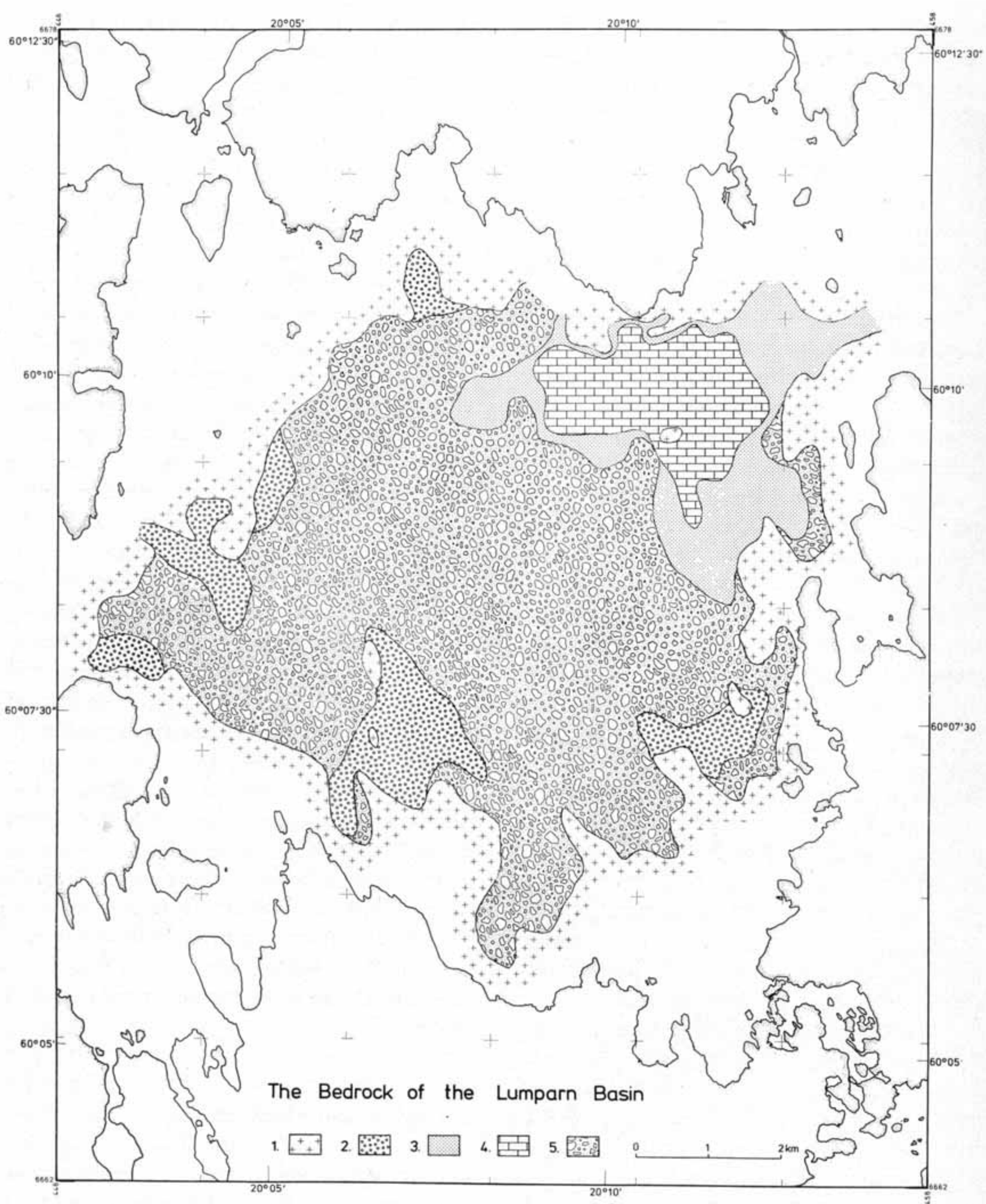


Figure 2-13. *Distribution of sedimentary rocks in the Lumparn impact structure (Winterhalter 1982). 1. rapakivi granite, 2. undifferentiated sedimentary rock (Cambrian sandstone, compact Quaternary sediments and/or weathered older sediments or weathered bedrock), 3. Cambrian sandstone and siltstone, 4.Ordovician limestone, 5.deeply weathered granite, possibly remains of sedimentary rock.*

2.2.5 Lauhanvuori

The sandstone of Lauhanvuori (Fig. 2-14), situated ca. 230 m above sea level in southern Ostrobothnia (Fig. 2-1) occupies an area of ca. 60 km^2. Only two outcrops have been found, but the area is covered with abundant sandstone blocks (Simonen & Kouvo 1955). The thickness of the sandstone is estimated to be some tens of metres. Drilling in the northern edge of the formation intersected ca. 10 m of sandstone above the porphyrite granite country rock (Lehtovaara 1982a).

The Lauhanvuori sandstone is pure quartz sandstone in mineral composition (Simonen & Kouvo 1955). The grain size is 0.2 - 0.5 mm and the mineral particles are well rounded and sorted. Based on the sandstone blocks, the sandstone is well bedded, being frequently cross-bedded (Sauramo 1916; Simonen & Kouvo 1955). Conglomeratic layers are abundant (Simonen & Kouvo 1955). Occurrence of a few boulders of strongly altered porphyritic granite in the conglomeratic layers indicates deposition of the sandstone on deeply weathered surface. Simonen & Kouvo (1955) suggest sedimentation in fluvial or, occasionally, eolian conditions.

On the basis of petrographic features, Simonen & Kouvo (1955) considered the Lauhanvuori sandstone to be Cambrian in age, corresponding to the Cambrian sandstone dykes in south-western Finland and Åland archipelago. The high P_2O_5 content of the clay balls in sandstone (Simonen & Kouvo 1955) resembles those of the phosphorous conglomerates in Cambrian Söderfjärden formation (Lehtovaara 1982a). Tynni & Hokkanen (1982) have found fossil traces of annelids in two sandstone boulders in Lauhanvuori, which suggest that the sandstone is younger than 700 Ma, possibly Vendian or, most probably, Lower Cambrian in age.

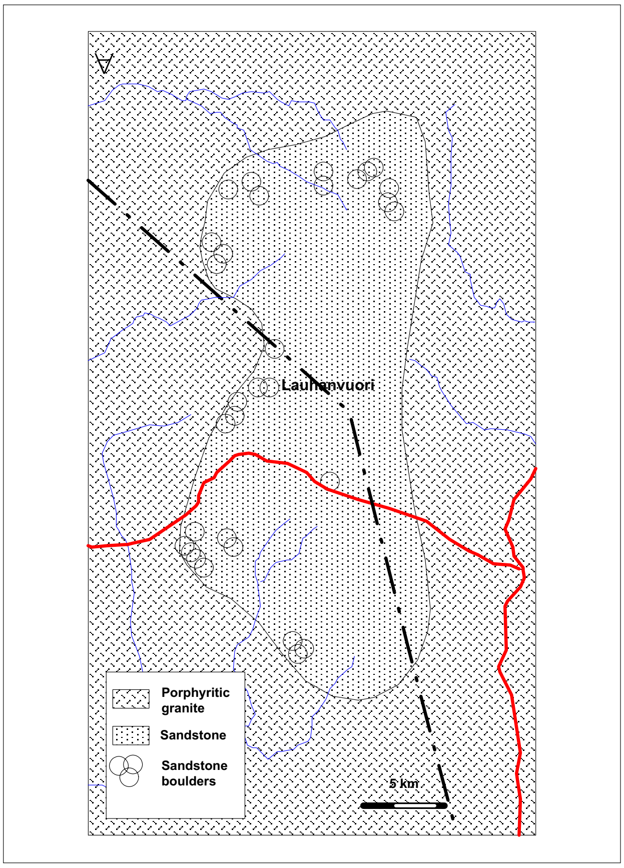


Figure 2-14. *Geological map of the Lauhanvuori sandstone area. Simplified after Simonen & Kouvo (1955). The location of the area is shown in Fig. 2-1.*

2.2.6 Söderfjärden impact structure

The Söderfjärden impact structure is located in western Finland, ca. 10 km south of the town of Vaasa (Fig. 2-1). It forms a plain hexagonal depression with a diameter of 5.5 km, surrounded by rims raising 20 - 40 m above the surface of the depression. The impact structure has a central uplift with a diameter of 1 km (Abels et al. 2000). The country rock consists of heterogeneous Palaeoproterozoic migmatites. The age of the meteorite impact is ca. 550 Ma (Lehtovaara 1992).

Several boreholes have been drilled in the central part of the structure, showing that the depression consists of sedimentary rocks (Fig. 2-15). The thickness of the sedimentary rocks in the deepest borehole, penetrating the whole sedimentary pile, is 244 m, and it is covered by 74 m of Quaternary deposits. The sedimentary rocks include sandstones, siltstones, shales, phosphoritic conglomerates and greywackes (Laurén et al. 1978;

Lehtovaara 1982b, 1984, 1992). No basal conglomerate exists, and the sedimentary pile rests directly upon the slightly altered migmatitic granodiorite. The lower half of the borehole from 215 m is clayey, whereas in the upper part sandy and silty beds are intercalating (Lehtovaara 1982, 1984).

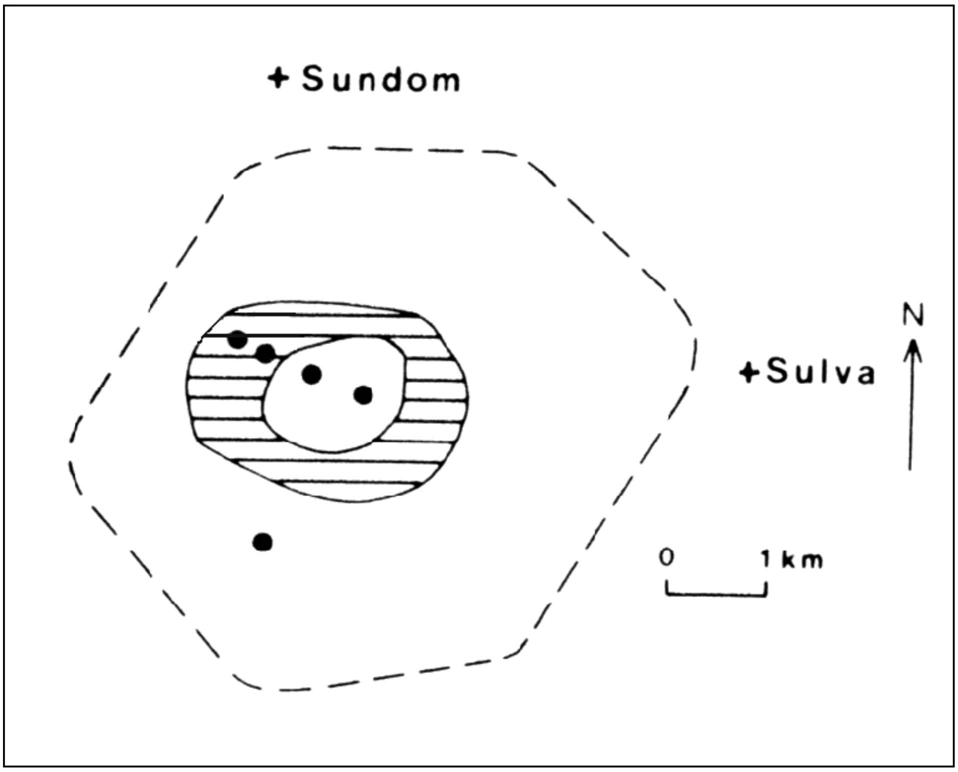


Figure 2-15. *Approximate contour of the Söderfjärden plain (dashed line) and possible distribution of Cambrian sedimentary rocks (hatched) (after Lehtovaara, 1984). Dots = drilling sites. The location of the area is shown in Fig. 2-1.*

The sedimentary rocks show well-preserved sedimentary structures including bedding and graded bedding with varying dips, lamination, small-scale slumping and minor faulting. The sedimentary structures and the abrupt fining of the sequence in the lower part of the borehole indicate very rapid and unsettled sedimentation in the deltaic environment (Lehtovaara 1982, 1984)

According to microfossil (acritarchs) studies by Tynni (1978, 1982a), the whole sedimentary sequence is Lower Cambrian in age, resembling the Lower Cambrian Finngrundet shale in the Bothnian Sea and the Lükati formation in Estonia. Hagenfeldt (1989b), however, has shown that the upper part of the Cambrian sequence is composed of Middle Cambrian siltstones and shales (*Oelandicus* beds). In borehole 3, the Middle

Cambrian rocks comprise about 14 m of a total of 53 m Cambrian sequence in the borehole (Hagenfeldt, op. cit.)

2.2.7 Lake Lappajärvi impact structure

Lake Lappajärvi impact structure is situated in western Finland (Fig. 2-1). The diameter of the structure is ca. 23 km, the difference between the topographic rim and the deepest troughs in the lake being about 140 m. The structure is located within the Svecofennian mica schist and pegmatitic granite, about 1900-1950 Ma in age. The age of the meteorite impact is 73 Ma, i.e. late Cretaceous (Mänttäri & Koivisto 2001).

Drilling in eastern part of the impact structure have penetrated ca. 18 m thick allochtonous sedimentary deposit, composing, from bottom to the top, of conglomerate, silty claystone, fine-grained sandstone and siltstone (Uutela 1990; Vaarma & Pipping 1997) (Fig. 2-16). The deposit is lying within the annular graben in the terrace zone of the impact structure. The sedimentary rocks have discordantly deposited on a ca. 40 m thick strongly weathered Proterozoic mica schist saprolith. According to acritarchs studies, the sedimentary deposit is probably contemporaneous with the ca. 1200 Ma old Muhos Formation but older than the ca. 600 Ma old Hailuoto Formation (Uutela 1990).

Uutela (1998) have found reworked Cambrian and Ordovician acritarchs from the mudstone filled fracture in one of the drill cores, suggesting that the Mesoproterozoic sedimentary rocks were once covered by Palaeozoic sediments. The Cambrian acritarchs range from Lower Cambrian (Vergale regional stage) to early Middle Cambrian (Kibartai regional stage). The Ordovician acritarchs are from early Middle Ordovician transition from Aseri to Lasnamägi regional stage and from middle of Middle Ordovician transition from Idavere to Jõhvi regional stage. Also two types of unknown spores were found, which, according to Uutela (op. cit.), indicate sediments and /or a terrestrial fauna younger than the Ordovician.

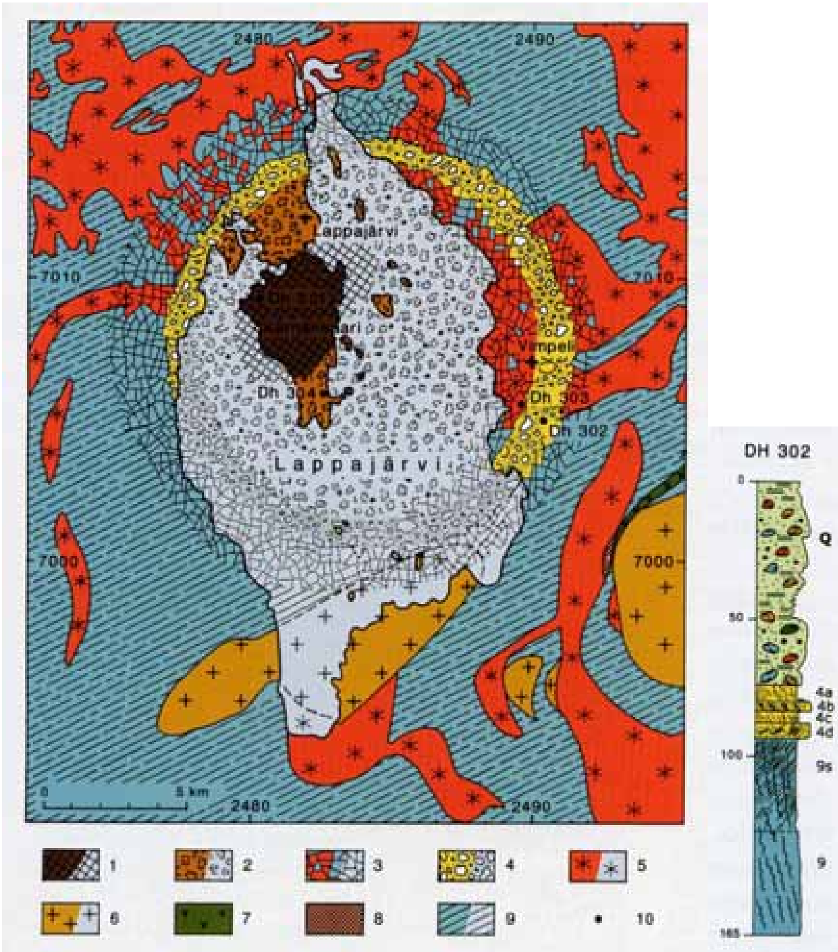


Figure 2-16. *Lithological map of the Lake Lappajärvi impact structure area and the drill hole profile DH 302 (Pipping 2000 after Vaarma & Pipping 1997). 1. Impact melt rock or kärnaite, 2. Suevite and impact breccia, 3. Autochthonous breccia zone (the terrace zone of the impact structure), 4. Mesoproterozoic sedimentary rock; 4a. siltstone, 4b. Silty sandstone, 4c. Muddy siltstone, 4d. Sandstone, 5. Granite pegmatite, 6. Granodiorite and tonalite, 7. Mafic and intermediate metavolcanic rocks, 8. Limestone, 9. Metagreywacke and mica gneiss, 10. Drill hole locations. Q = Quaternary cover deposits, mostly till.*

2.2.8 Lake Iso-Naakkima impact structure

The Lake Iso-Naakkima impact structure is located in the east central Finland (Fig. 2-1). The geophysical survey indicates that the diameter of the structure is about 2 km and the maximum depth about 160 m. The country rock consists of Palaeoproterozoic mica schists.

Within impact structure, four boreholes have penetrated ca. 100 m thick sequence of sedimentary rocks. The sedimentary sequence is lying on the Palaeoproterozoic migmatitic mica gneiss, which gradually passes upward to disintegrated saprolith (Elo et al. 1993b) (Fig. 2-17). The sedimentary sequence begins with conglomeratic sandstone, the clasts being predominantly of quartz. It is overlain by quartz sandstone with kaolinitic clay interbeds, siltstone with sandy and clay interbeds, quartz sandstone, red shale and violet shale. The sediments were probably deposited in alluvial or lacustrine environment (Elo et al. 1993b). The sedimentary beds are tilted, distorted and brecciated, demonstrating subsidence of the sedimentary basin during or after the sedimentation.

Microfossil (acritarchs) assemblage indicates that the sedimentary rocks are Neoproterozoic/Late Riphean (1000 - 650 Ma) in age (Elo et al. 1993b). Elo et al. (op. cit.) interpret the sedimentary rocks considerably younger than the impact.

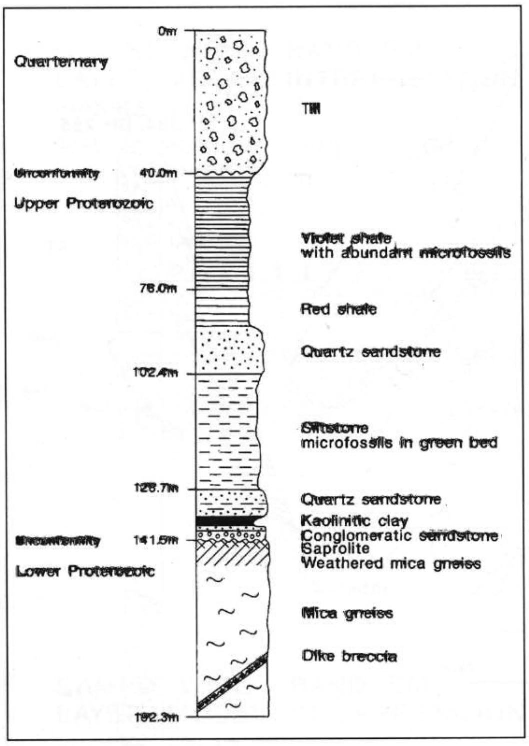


Figure 2-17. *Stratigraphy of the sedimentary deposit of the Iso-Naakkima impact structure based on drill holes DH 384 and DH 385 (Elo et al. 1993b).*

2.2.9 Lake Saarijärvi impact structure

The Lake Saarijärvi impact structure is situated in north-eastern Finland about 30 km south of the municipality of Taivalkoski (Fig. 2-1). The structure forms a roundish depression with a diameter of ca. 1.5 km. The surroundings of the structure consist of Archaean tonalitic gneiss granite, cut by Palaeoproterozoic diabase dykes.

Sandstones and shales with a thickness of 156 m have been preserved in the impact structure. Microfossils of the sedimentary rocks are comparable to those in the upper unit of the Hailuoto formation, suggesting a Vendian (ca. 600 Ma) age (Tynni & Uutela 1985).

According to recent microfossil studies also Cambrian sediments are present in the drill core in the centre of the impact structure (Öhman et al. 2000). Unfortunately, no details are available at resent.

2.2.10 Lake Karikkoselkä impact structure

Lake Karikkoselkä impact structure is located in central Finland, about 25 km west of the town of Jyväskylä (Fig. 2-1). The diameter of the circular lake is about 1.3 km. The structure is situated within the Palaeoproterozoic (1890 - 1880 Ma) Central Finland Granitoid Complex, here consisting of porphyritic granite.

Ca. 130 m of chaotic crater fill, composing of mixed and brecciated sedimentary rocks were found in the deep drillings of impact structure (Arkonsuo 2000). The sedimentary rocks have originally been intercalating beds of sandstone, siltstone and shale, which, after the impact, have re-deposited in the impact crater. On the basis of microfossil (acritarchs) studies, the sedimentary material represent Neoproterozoic (and maybe Mesoproterozoic), Cambrian and Ordovician sedimentary rocks, but during the impact the material has totally mixed (Arkonsuo 2000; Uutela 2001). The Proterozoic microfossils found in Lake Karikkoselkä material are previously recorded in Meso- and Neoproterozoic deposits at Muhos, Hailuoto and Lake Saarijärvi (Uutela 2001). The Cambrian species are from Lower Cambrian or Lower to Middle Cambrian, the Ordovician microfossils ranging from Lower Ordovician Kunda regional stage to Middle Ordovician Kukruse regional stage but also a microfossil from upper Middle Ordovician Keila regional stage has been found (Uutela 2001). Based on the palaeomagnetic measurement the crater itself is probably Late Permian (230 - 260 Ma) in age (Pesonen et al. 2000). No sedimentary rocks post-dating the impact have been found.

2.2.11 Kilpisjärvi

Autochthonous sedimentary rocks have been preserved under the Caledonian nappes in a small area in north-western arm of Finland (Fig. 2-1). The sedimentary rocks belonging to Vendian to Lower Cambrian Dividal Group consist of basal conglomerate overlain by intercalating sandstones and shales (Lehtovaara 1988, 1995). The Dividal Group with a maximum thickness of 200 m in Finnish Lapland has been deposited on the sub-Caledonian unconformity and is overlain by paraautochthonous rocks of the Jertta Nappe, which have been overthrust a short distance to its present position (Fig. 2-18). The overlying Kalak Nappe has been overthrust during late Cambrian to early Ordovician (Lehtovaara 1995).

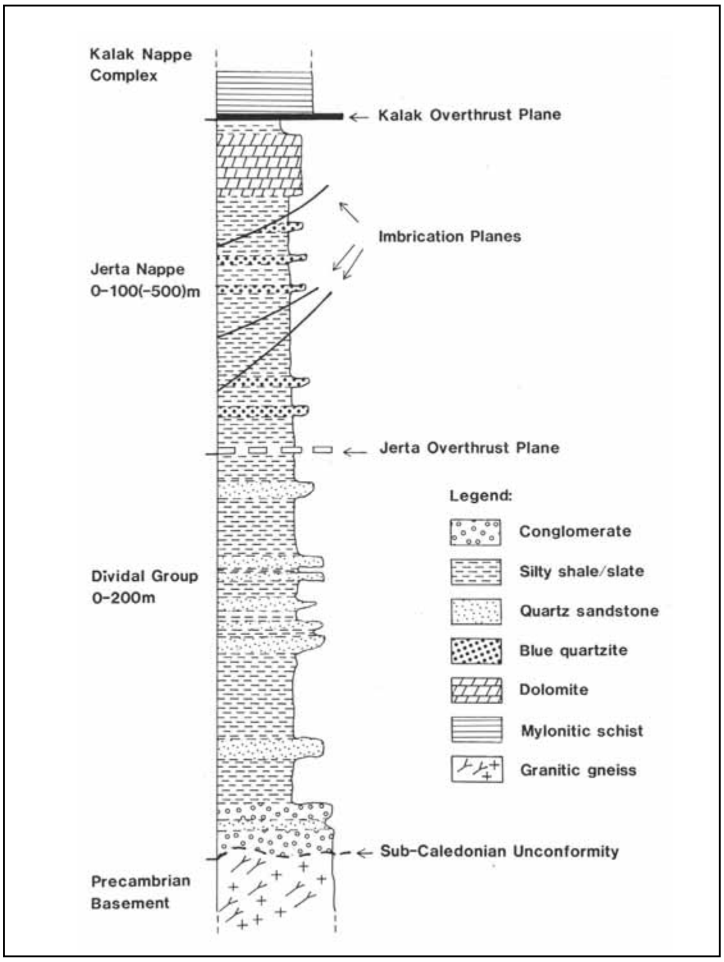


Figure 2-18. *Stratigraphy of the sedimentary sequence in the Finnish Caledonides (Lehtovaara 1988).*

2.2.12 Clastic dykes

Hundreds of clastic dykes occur mainly in the rapakivi granite area of the Åland archipelago, but some are also found in the coastal area of south-western Finland (Tanner 1911; Martinsson 1968; Bergman 1982). The dykes are mainly vertical and dominantly striking 0 - 20° (Bergman 1982). Based on the microfossil studies, the dykes are mainly from Lower Cambrian, the minority being of Lower Ordovician age (Tynni 1982b). Within the Lower Cambrian group of dykes, there are considerable age differences between some of the dykes. In the northern part of the Åland Islands some of the fracture fillings have been interpreted (on the basis of occurrence of brachiopod *Ceratreta tanneri*) as Upper Cambrian (Martinsson 1968; Holmer & Popov 1990), rather than Lower Ordovician, as suggested by Tynni (op. cit.). No Upper Cambrian rocks, however, occur in the glacial erratics in the Åland Sea area (Hagenfeldt 1995).

The clastic dykes can be divided into three main groups: breccia dykes, siltstone dykes and sandstone dykes (Bergman 1982). The sandstone dykes vary from pure quartz sandstones to arkosic sandstones, and the cementing material is mostly siliceous, rarely calcite (Bergman 1982, Simonen & Kouvo 1955). The dykes are usually well lithified. The grain size of the dykes varies from silt to coarse sand, and also the sorting varies considerable. The monoclinic symmetry of the feldspar indicates that the orthoclase originated from the rapakivi granite (Simonen & Kouvo 1955). Also the larger polymictic fragments consist almost solely of rapakivi granites. According to Simonen & Kouvo (op. cit.), the orthoclase and the fragments are not primary constituents of the sediments but are derived both from the walls of the dykes and from the weathered parts of the sub-Cambrian peneplain, since the clastic dykes appear in unweathered rapakivi.

Bergman (1982) has explained the clastic dykes to be formed by injection of clastic material from above. This mechanism presupposes that the fractures have opened after the sedimentation.

Narrow galena veins occur alone or in association of the sandstone dykes in the Åland rapakivi area (Bergman & Lindberg 1979). Bergman & Lindberg (op. cit.) propose that the galena veins were formed when lead-bearing solutions were pressed out from the Cambrian and Ordovician sediments during diagenetic processes and precipitated in fractures. Vaasjoki (1996) reports galena crystals with model ages of 300 – 400 Ma from fracture infillings in the Palmottu U-Th mineralisation at Nummi-Pusula in south-western Finland

2.2.13 Lake Vahankajärvi sandstone boulders

Lake Vahankajärvi sandstone, located in Karstula, central Finland, is only known as glacial boulders on the south-eastern shore of the lake. However, the boulder train indicates that the source of the boulders lies hidden on the bottom of the lake. The layered and cross-bedded sandstone is very pure quartz sandstone, where the secondary growth of silica is common (Sauramo 1916; Simonen & Kouvo 1955). Rounding and sorting of the mineral particles are very good.

Based on petrographic comparisons with the clastic dykes in the Åland archipelago, Sauramo (1916) and Simonen & Kouvo (1955) have tentatively interpreted the sandstone to be Cambrian in age. The microfossil studies by Tynni (1974) could not confirm the tentative age, because of lack of sufficient amount of microfossils. Tynni (op. cit.) has proposed a Vendian age, based on the resemblance to the Neoproterozoic microfossils from Norway.

2.2.14 Sandstone and limestone boulders in SW Finland

Uutela (1989) have studied a total of 2244 erratic boulders of Palaeozoic sedimentary rocks in south-western coast of Finland. The boulders consist of light-grey Cambrian sandstone, reddish brown, greenish grey marls and grey marls, and yellowish grey or reddish grey calcilutite/calcarenite of the Baltic limestone. Based on the microfossil determinations, the greenish grey and the grey marls were dated to the Lower/Middle to Middle Ordovician (Kunda to Kukruse regional stage of Estonia) and Middle to Middle/Upper Ordovician (Aseri-Lasnamägi transition to Nabala regional stage), respectively (Uutela 1989). The yellowish grey calcilutites are Lower/Middle to Upper Ordovician in age (Kunda to Pirga regional stage). Most of the specimens dated are from Kukruse to Keila stages.

The source of the boulders is uncertain. Arguments both in favour and against the short-distance and long-distance glacial (from the Bothnian Sea Basin) transport can be presented (cf. Uutela, op. cit.). However, no sedimentary rocks were found in echo-soundings and drillings offshore the town of Uusikaupunki (Alviola 1988), although the area is rich in glacial erratics.

Nõlvak et al. (1995) have studied the microfossils of four glacially transported limestone boulders from the south of Rauma, in south-western coast of Finland. On the basis of microfossils (ostracods, chitinizoans and acritarchs) and the resemblance to the Ordovician limestones of North Estonia, the limestone boulders are dated to Middle Ordovician (Uhaku and Rakvere regional stages of Estonia). The source of the boulders is probably a belt of Ordovician limestones, extending from northern Estonia to the archipelago of the south-western Finland.

The phosphoritic sandstone boulder, found in Susikari, 2 km west of Olkiluoto, is composed of rather coarse-grained, conglomeratic quartz sandstone, which is characterised by abundant brachiopod fossils (Lehtovaara & Tynni 1983). Based on the microfossils, the sandstone has been dated to the upper part of the Lower Cambrian, corresponding to the Lingulid sandstone facies in Västergötaland, Sweden (cf. Thorslund 1960). The provenance of the boulder is most likely the Bothnian Sea Basin. The fossiliferous sandstone boulder of

Pyhäjärvi, Säkylä, dates to Lower Cambrian, and probably represent older stratum than the Olkiluoto boulder (Lehtovaara & Tynni 1983).
Niemelä et al. (1985) describe pebbles of Cambrian and Ordovician sedimentary rocks from the potholes of Kattilaluoto, in the archipelago of SW Finland. The pebbles are composed of quartz arenite, pyrite concretions and red biomicritic and grey pelmicritic limestones. On the basis of microfossils (*Volborthella tenuis*), the quartz arenites have been dated to Lower Cambrian. The red and grey limestones petrographically correspond the Lower Ordovician orthoceratite limestone and Middle to Upper Ordovician Baltic limestone (e.g. in Lumparn), respectively. Stone count indicates a more local source than the Bothnian Sea, probably an unknown outcrop area on the bottom of the sea (Niemelä et al., op. cit.)

Edelman (1951) describes a local concentration of mostly Cambrian sandstone boulders in the archipelago of south-western Finland. Based on the boulder train, he suggests a local sandstone deposit hidden on the sea bottom south of the islands Borgön and Marskär in Hiittinen.

Northeast of Jomala in Åland, over 90% of the rocks in till are composed of Ordovician limestones (calcarenites, marls and micrites) but no *in situ* occurrence has been found in the area (Hokkanen 2002). The provenance of the limestones may be the Bothnian Sea deposits. Hokkanen (op. cit) discuss that the high percentage of limestone in till could be due to glacial drift of limestone as megablocks from its original sedimentation area.

2.2.15 Sediments with Tertiary microfossils in Finnish Lapland

Sediments with Tertiary microfossils are known from a few localities in central and eastern Lapland (Hirvas & Tynni 1976; Tynni 1982c). In the hill of Akanvaara in Savukoski, eastern Lapland (Fig. 2-1), there is about 80 m thick, possibly *in situ* clay bed at 205 m a.s.l consisting of very fine-grained clay with considerable amounts of kaolinite, montmorillonite and mixed-layer silicates (Hirvas & Tynni 1976). The clay bed is resting on Palaeoproterozoic bedrock, consisting of unweathered gabbro. According to Tynni (1982c), the microfossils indicate that the clay was deposited in Early Tertiary under marine conditions. Fenner (1988) have suggested a late Early Eocene age for the deposit.

About 50 cm thick deposit of diatomaceous earth containing late Tertiary microfossils is found in Naruskajärvi, eastern Lapland (Hirvas & Tynni 1976; Hirvas et al. 1976). It is, however, not certain, whether this deposit is primary or secondary (*see* Tynni 1982b). The pollen content of the diatomite suggests that it is from the Eemian Interglacial (Hirvas et al. 1976).

Redeposited Tertiary microfossils are found at several places in the Quaternary strata in eastern and central Lapland (Grönlund 1977; Tynni 1982c).

3 DEPOSITIONAL HISTORY

3.1 Mesoproterozoic (Middle Riphean)

At the onset of the Middle Riphean, a more or less even ‘Subjotnian’ peneplain existed in the Fennoscandian Shield. For example, in the Åland Sea the peneplain evidently dips approximately 2.5° towards the southwest (Winterhalter et al. 1981)

Remnants of Middle Riphean sedimentary rocks are found both onshore and offshore in several tectonically protected basins within the Fennoscandian Shield. The formation of the intracratonic basins is suggested to be due to rifting in connection of the emplacement of the rapakivi batholiths (e.g., Korja & Heikkinen 1995). The basins were successively filled with silty and sandy sediments.

In addition to the Finnish locations (see Chapter 2.2), Middle Riphean (‘Jotnian’) sandstones are found in Dalarna, Gävle, Nordingrå, Mälaren and Almesåkra areas in Sweden, in the Lake Ladoga area in Russia and in a tectonic basin between Gotland and the south-eastern Stockholm archipelago (Fig. 3-1). The Dala sandstone is an 800 m thick sandstone sequence intercalated with shales and conglomerates. The inferred thickness of the Gävle sandstone is 900 m (Gorbatschev 1967). In the south-western of the basin north of Gotland, the maximum thickness of the sandstone may exceed 900 m (Flodén 1980). The Almesåkra Group, consisting of sandstones, conglomerates and shales intruded by ca. 1000 Ma old diabase dykes, has a total thickness of about 1200 m (Rohde 1987). The ‘Jotnian’ rocks in central Sweden (Dalarna, Gävle) are characterised by prehnite-pumpellyite facies metamorphism, indicating a burial of several kilometres (Nyström & Levi 1980, Nyström 1983). In the Lake Ladoga area, unmetamorphosed ‘Jotnian’ sedimentary and volcanic rocks with a distinct unconformity overlay the rapakivi granites and Palaeoproterozoic igneous and metamorphic rocks (Amantov & Muller 2001).

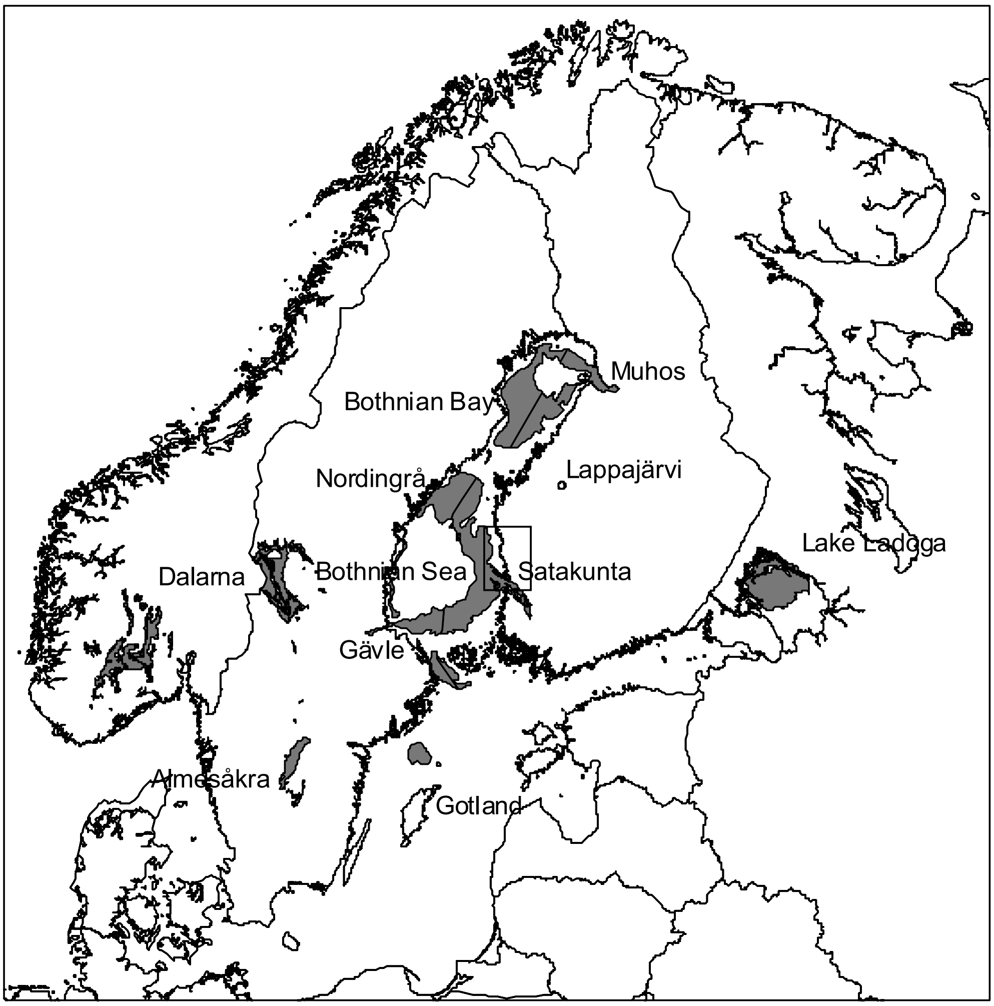


Figure 3-1. *Mesoproterozoic sedimentary rocks in the Fennoscandian Shield. Compiled from Koistinen et al. (2001). The southern Satakunta area is marked with a square.*

Although the Middle Riphean sandstones in the Fennoscandian Shield are presently only found in local basins, they were probably not solely restricted to basins but distributed over a much larger area (cf. Kohonen et al. 1993; Söderberg 1993; Andréassen 1994; Tullborg et al. 1996; Amantov 2001), as suggested by major rifted basins of Riphean age in the south-eastern and north-eastern parts of the East European Craton (Fig. 3-2). According to Weltheim (1972) it is possible that the Muhos Formation and the Dala sandstone in Sweden are comparable, and that the Bothnian Bay and central Sweden were connected by an epicontinental sea. Puura et al. (1996) have estimated that in places the erosion have removed ca. 500 - 2000 m of sedimentary rocks and 'Postjotnian' diabases from the peripheral parts of the sedimentary basins.

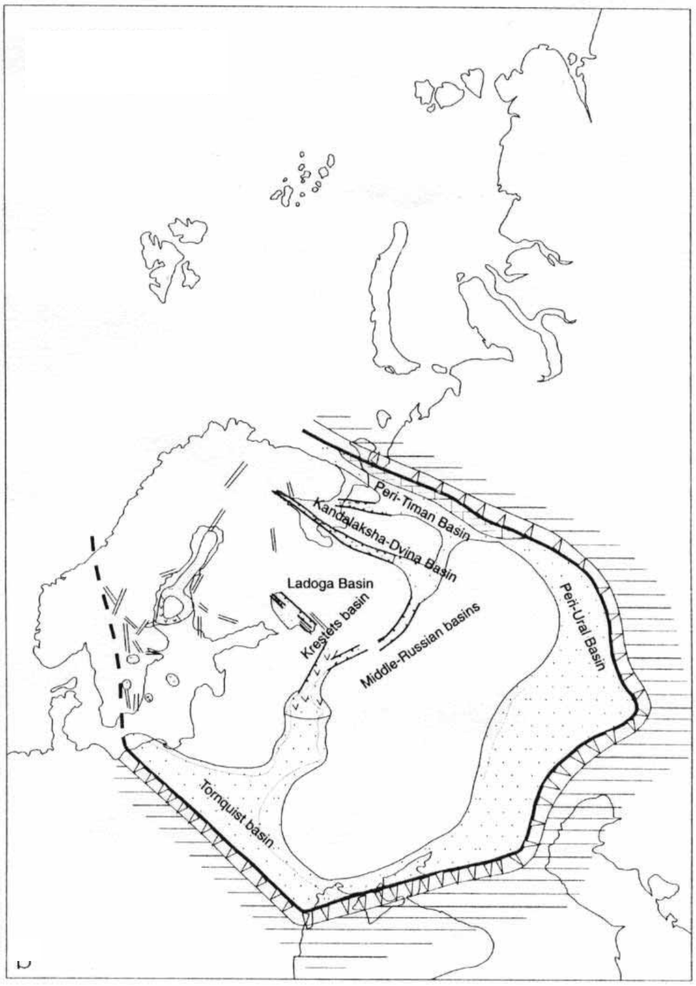


Figure 3-2. *Sedimentary basins within the East European Craton during Mesoproterozoic (Middle Riphean) (Nikishin et al. 1996). Strippled areas = alluvial-deltaic and shallow marine sand and shales, doupled hatch = dyke systems, black line with small protuberances = rifts, toothed line = continental slope, horizontal lines = oceanic basin.*

3.2 Neoproterozoic (Late Riphean - Vendian)

Middle Riphean sedimentation was followed by intense and prolonged erosion with the interpreted rate of 500 - 2000 m in places (Puura et al. 1996). In Virtasalmi, south-eastern Finland, tens of metres thick Mesoproterozoic kaolinitic saprolites occur with the age of 1180 Ma (Sarapää 1996), and in Iso-Naakkima impact structure there are kaolinite-bearing sediments, containing Neoproterozoic (1000 - 650 Ma) microfossils (Elo et al. 1993b). In Lauhanvuori, western Finland, presumably Lower Cambrian sedimentary rocks (Tynni & Hokkanen 1982) are underlain by kaolinised bedrock. In Lake Saarijärvi, north-eastern Finland, the bedrock underlying a Neoproterozoic sandstone-shale deposit is kaolinised to a depth of 10 - 20 m (Pekkala & Sarapää 1989). In Estonia, the upper part of the basement under the sedimentary cover is weathered at a depth 1 - 100 m (Puura et al. 1997). During the Neoproterozoic erosion period until the beginning of the Vendian, 650 Ma ago the Baltica craton, including the present-day 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). The denudation resulted in the formation of the sub-Vendian peneplain. In southern Finland, the sub-Vendian peneplain dips very gently (at less than 0.5°) southwards towards Estonia and the St. Petersburg region of Russia, where it is overlain by hundreds of metres of Vendian and Phanerozoic sedimentary rocks (Puura et al. 1996).

In the south-western part of the Fennoscandian Shield, the orogenic uplift of the area affected by Sveconorwegian orogeny (1200 - 900 Ma ago) resulted in the deposition of at least 8 km thick pile of foreland sediments ca. 100 km to the east of the orogen (Larson et al. 1999). Titanite and zircon fission track ages together with geological constraint indicate that foreland sediments were deposited ca. 950 Ma ago (Larson et al., op. cit.). Most of these sediments were eroded at the end of the Proterozoic when a sub-Cambrian peneplain was established.

During the Vendian, the Baltica craton drifted from an equatorial position toward the South Pole (Torsvik et al. 1996), resulting in the large-scale glaciation. Based on microfossils and geochemistry, the age range for this Vendian glaciation is from about 650 to 590 Ma (Knoll & Walther 1992; Torsvik et al. 1995). Within the present-day Fennoscandian Shield, Vendian-aged glaciogenic deposits (tillites) are only preserved in Norway and north-western Sweden (Fig. 3-3).

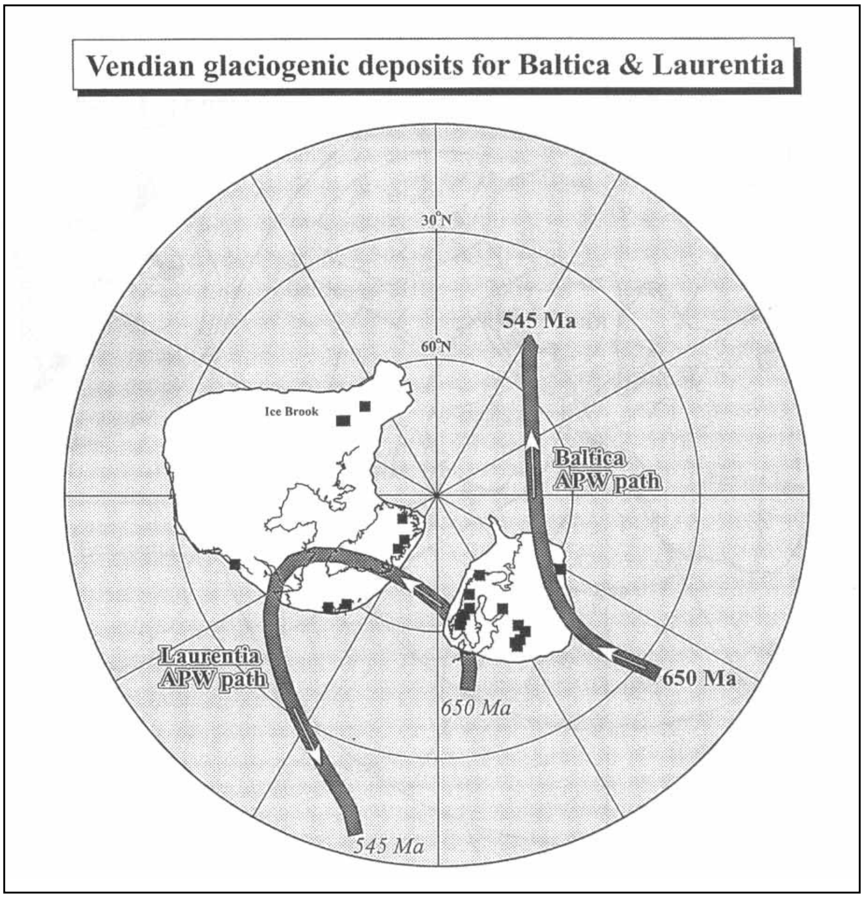


Figure 3-3. *Palaeogeography reconstruction and apparent polar wander paths of Baltica and Laurentia cratons, and the distribution of Vendian glaciogenic deposits (solid squares) (Torsvik et al. 1996).*

Main stage of the Neoproterozoic sedimentation took place during the Late Vendian when a shallow sea transgressed large parts of the present-day Fennoscandian Shield and the subsequent sedimentary basins were filled with arenaceous to argillaceous sediments (Nikishin et al. 1996; Puura et al. 1996). The remaining Neoproterozoic (Vendian) successions within the Fennoscandian Shield are present in isolated rift troughs, such as the Muhos-Hailuoto Basin (*see* Chapter 2.2) (Fig. 3-4). Additionally, Neoproterozoic sedimentary rocks occur in the Lake Vättern Graben in south-central Sweden, which is filled with up to 1000 m of Riphean to Vendian sediments of the Visingsö Group (Vidal 1982). Within the margins of the Fennoscandian Shield, autochthonous Neoproterozoic successions occur along the eastern border of the Scandinavian Caledonides (e.g. Laisvall and Torneträsk in Sweden), as well as allochthonous successions in the marginal areas of the Caledonides in southern Norway (Kumpulainen & Nystuen 1985; Vidal & Moczysłowska 1995). Neoproterozoic to Lower Palaeozoic successions occur also in the Varangerfjorden and Tanafjorden areas in East Finnmark, northern Norway and in the

White Sea area in Russia. Very thick Late Riphean to Vendian rock successions occur in the north-eastern half of the Varanger peninsula, Norway.

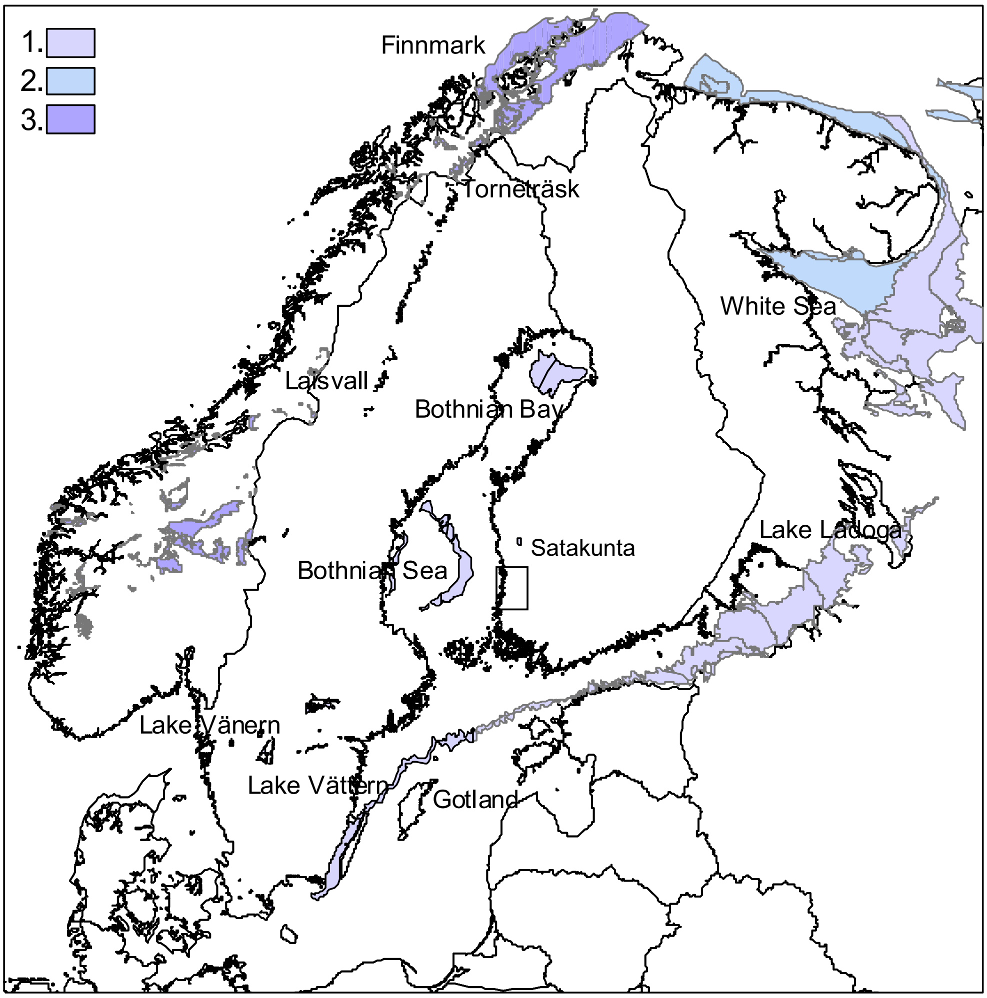


Figure 3-4. *Neoproterozoic-Early Palaeozoic sedimentary occurrences within and in the margins of the Fennoscandian Shield (compiled from Koistinen et al. 2001). Legend: 1. Sandstone, conglomerate, siltstone, shale (Vendian to Lower Cambrian), 2. Sandstone, conglomerate, siltstone, shale (Late Riphean and possibly older), 3. Neoproterozoic feldspathic metasandstone, meta-arkose, quartzite, metagreywacke, marble and tillite. The southern Satakunta area is marked with a square.*

Neoproterozoic (Late Vendian) sedimentary rocks occur also in North Baltic region under the Palaeozoic cover, in an area surrounding Lake Ladoga and east of St. Petersburg (Mens & Pirrus 1986; Puura et al. 1996). Because the present northern margin of Neoproterozoic successions in the Gulf of Finland and North Baltic, as well as in the Lake Ladoga-St.

Petersburg area, is erosional and the thickness of the successions does not show any decrease towards the north (Fig. 3-5), it is highly probable that the Late Vendian sedimentary rocks once were present also in southern Finland (Puura et al. 1996). Moreover, based on the occurrence of Neoproterozoic sedimentary deposits in Lake Saarijärvi, Lake Iso-Naakkima, Lake Karikkoselkä and possibly Lauhanvuori (*see* Chapter 2.2), it is probable that an extensive sedimentary cover once existed between Hailuoto and Lake Ladoga. Remains of that sedimentary cover were still present in places, when the Lake Lappajärvi impact structure formed in late Cretaceous, ca. 73 Ma ago (cf. Uutela 1990; Kohonen & Vaarma 2001).

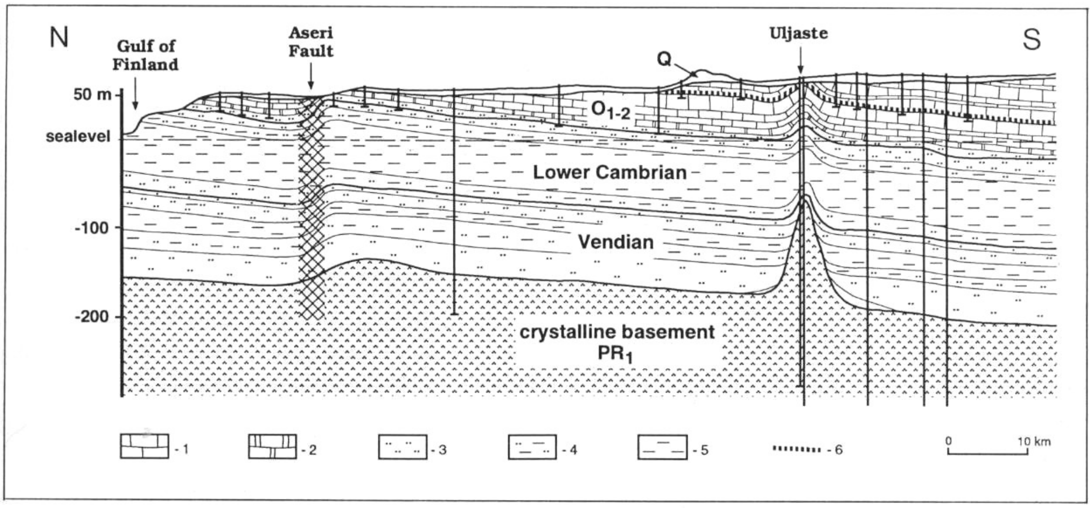


Figure 3-5. *North-south cross-section through Vendian to Ordovician sedimentary sequences in North Estonia (Puura et al. 1996). 1. Limestone, 2. Dolomite, 3. Sandstone, 4. Siltstone, 5. Claystone, 6. Oil shale unit. O = Ordovician, Q = Quaternary.*

Due to similarities between the Vendian sequences of the Åland Sea Basin and the Lake Ladoga area, Söderberg (1993) suggests a basinal connection between the Åland Sea Basin and the major Russian Basin in the east. Furthermore, according to Hagenfeldt (1995), the dark shales of the Åland Sea Basin may be compared with the Vättern Basin and the Bothnian Bay, indicating that the Late Riphean to Vendian marine transgression extended from southern Sweden to the Bothnian Bay region.

Based on biostratigraphy, Vidal & Moczydłowska (1995) tentatively correlate the Visingsö Group of the Lake Vättern Graben with successions within aulacogens in the East European Platform (e.g. the Volhyn Aulacogen) suggesting that rifting, formation of basins in Western Baltica craton and infilling of rift basins within the core of the Baltica craton might be largely contemporaneous.

3.3 Cambrian

Before the beginning of the Cambrian period, a new extensive phase of erosion and weathering resulted in the formation of the sub-Cambrian peneplain. Within the Fennoscandian Shield the exhumed and well preserved sub-Cambrian peneplain is a widespread feature occurring in the Bothnian Sea area, in south central Sweden and in southern Norway, west of the Oslo region (Lidmar-Bergström 1996; Riis 1996) (Fig. 3-6). In its contact with the cover rocks it is weathered to a maximum depth of 5 m (Elvhage & Lidmar-Bergström 1987). In the Bothnian Sea, the Palaeozoic sedimentary rocks are lying on a sub-Cambrian peneplain, the main part of which is formed by 'Jotnian' sandstone (Winterhalter 1972). The peneplain is dipping gently (0.24 - 0.26°) towards the central parts of the Bothnian Sea basin (Axberg 1980). Outside the Bothnian Sea basin the peneplain is very gently (0.1°) dipping westward (Veltheim 1962). In the Åland Sea Basin, this almost horizontal peneplain, situated along the Swedish coast about some 90 m below sea level, is formed by Upper Riphean to Vendian sedimentary rocks (Söderberg 1993). Similar erosional surface exists also in the Bothnian Bay (Wannäs (1989). Clastic dykes (sandstone-filled fractures) in the Åland archipelago indicate that the sub-Cambrian peneplain is close to the present erosional level.

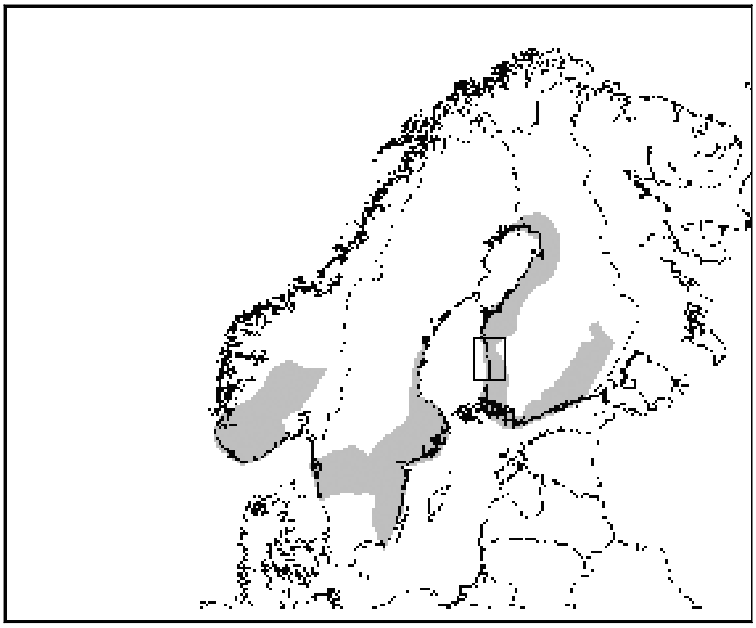


Figure 3-6. *Distribution sub-Cambrian peneplain within the Fennoscandian Shield (after Riis 1996). The southern Satakunta area is marked with a square.*

The Lower Cambrian and lower Middle Cambrian rocks in the Fennoscandian Shield are dominated by coarse clastic deposits, while the upper part of the Middle Cambrian and the Upper Cambrian is composed of dark and kerogen-bearing argillaceous sediments, the so-called alum shales. In addition to the location in Finland and the Gulf of Bothnia (*see* Chapter 2.2), Cambrian deposits are preserved in south-eastern (Kalmar, Öland) and south central Sweden (Västergötland, Östergötland, Närke), along the Caledonian front from Oslo region in the south to Finnmark in the north, in the northern Baltic Sea and northern Estonia (Fig. 3-7). In Gotland and in Estonia, Cambrian sequences are also present under the Ordovician to Silurian cover. The present thickness of the Cambrian sediments ranges from a few tens of metres up to 300 m (south of Gotland). It has been suggested that at the end of the Cambrian the region covering the Gulf of Finland, northern Baltic, Bothnian Sea and probably also southern Finland, was covered by a thin blanket (100 - 350 m) of clastic sediments (sands and silts) and clays deposited in shallow-water basins (cf. Puura et al. 1996).

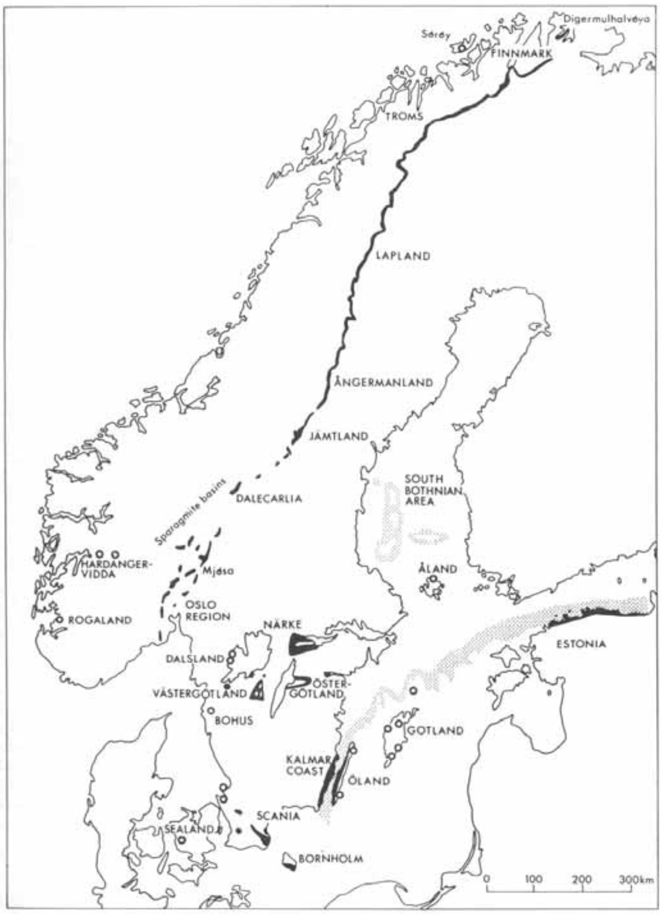


Figure 3-7. *Distribution of Cambrian deposits in Fennoscandia and Estonia (Martinsson 1974).*

During the Lower Cambrian, the sea transgressed the Baltic Sea region, probably as a consequence of rifting of the Rodinia supercontinent and subsequent opening of the Iapetus

Ocean (*see* Chapter 4.2.2). At its maximum the sea reached as far as the Bothnian Sea and Bothnian Bay at the beginning of the Lower Cambrian and also covered most of the western, southern and central Finland (Öpik 1956; Thorslund 1960; Jaeger 1984; Hagenfeldt 1989a; Uutela 1998, 2001) (Fig. 3-8), as evidenced by the Lower Cambrian sandstones in the Bothnian Sea, Bothnian Bay, Söderfjärden, Lake Lappajärvi and Lake Karikkoselkä (*see* Chapter 2.2). The Baltic Basin, comprising the present Baltic and southern part of the Baltic Sea, was connected towards the north with the Gulf of Bothnia (Flodén 1980). The early Cambrian transgression occurred during the Vergale regional stage and at the beginning of the Rausve regional stage (late early Cambrian) (Hagenfeldt 1989a; Uutela 1998).

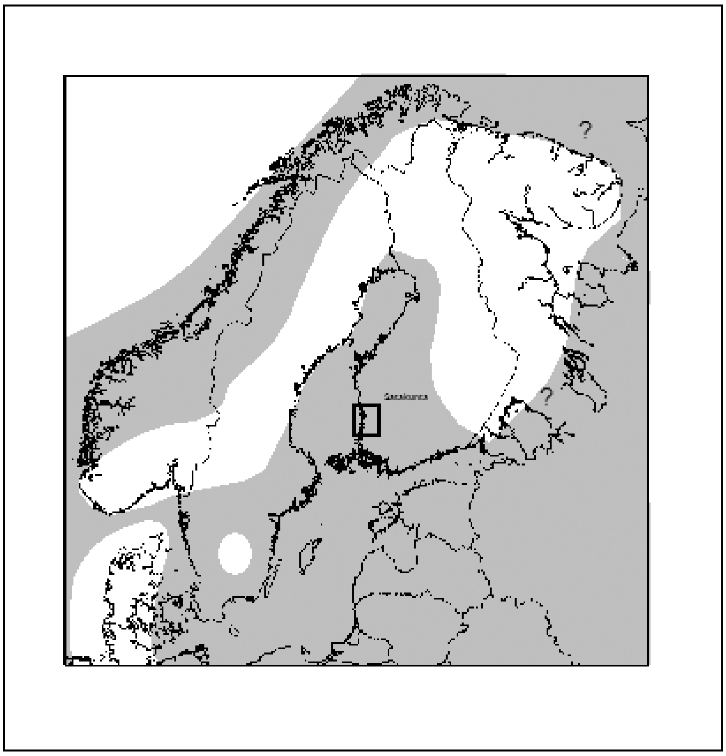


Figure 3-8. *Interpreted extent of the Lower Cambrian marine transgression (grey area) in the Fennoscandian Shield and in the marginal areas. Compiled after Thorslund (1960) and Jaeger (1984). The southern Satakunta area is marked with a square.*

It was formerly thought that during the Middle Cambrian, the sea did not reach more north than the Gotland basin in northern part of the Baltic Sea (*see*, e.g., Thorslund 1960). For example, Bergman (1982) did not find any Middle or Upper Cambrian sandstone dykes in the Åland area, which he interprets as indicating a gap in the sedimentation. According to Jaeger (1984), however, the sea reached Åland at the end of the Middle Cambrian. Moreover, the Middle Cambrian sea reached up to the Bothnian Sea, as evidenced by the microfossils in the Finngrundet borehole (Hagenfeldt 1989b). Middle Cambrian sandstones were also deposited in the Söderjärden area in western Finland (Hagenfeldt op. cit.). The study of Uutela (1998) indicates that the sea also covered the Lappajärvi area in western Finland during the early Middle Cambrian Kibartai regional stage, probably also the Lake Karikkoselkä area in central Finland (Uutela 2001). On the other hand, the lack of late Middle Cambrian acritarchs in the Lappajärvi and Lake Karikkoselkä material, suggests that both areas were probably dry land during the late Middle Cambrian (Uutela 1998, 2001). According to Hagenfeldt (1997), no sediments deposited in the Bay of Bothnia area during the late Middle Cambrian and Upper Cambrian.

Some of the clastic dykes in the Åland islands, interpreted by Tynni (1982b) as Ordovician are most likely Upper Cambrian in age. On the basis of occurrence of *Ceratreta tanneri* brachiopod in the clastic dykes, the Upper Cambrian transgression reached the area of Åland and also the Siljan district in south-central Sweden (Holmer & Popov 1990). No Upper Cambrian microfossils are found elsewhere in Finland. According to Hagenfeldt (1997), no deposition occurred in the Bay of Bothnia as well as in most of the interior of the Fennoscandian Shield during the late Middle Cambrian and Upper Cambrian.

A major regression probably happened at the end of the Cambrian. Seismic studies in the Stockholm archipelago indicate deposits of Ordovician limestones directly on the crystalline basement (Söderberg & Hagenfeldt 1995), suggesting a major erosion of Cambrian sequences before the Ordovician transgression.

3.4 Ordovician

Probably the most complete Ordovician sequences worldwide are found in Estonia and St. Petersburg area in Russia. Within the Fennoscandian area, outcrops of Ordovician rocks are present in southern and south central Sweden, as well as along the Caledonian front in southern Norway and central Sweden. The Ordovician sequences are rather thin, varying from a few metres to commonly less than 100 m. This indicates very low deposition rates, only a few millimetres per 1000 years (Jaanusson 1976).

The Ordovician lithofacies and faunas in Baltoscandia (Scandinavia and the Baltic) show large-scale zonation. Consequently, they are classified into so-called Confacies Belts, which are from northeast to southwest: the North Estonian Confacies Belt, the Central Baltoscandian Confacies Belt and the Scanian Confacies Belt (Männil 1966; Jaanusson 1976) (Fig. 3-9). In the North Estonian Confacies Belt, shallow-water limestones formed in an open shelf conditions, while deeper-water argillaceous carbonate sediments are typical for the Central Baltoscandian Confacies Belt.

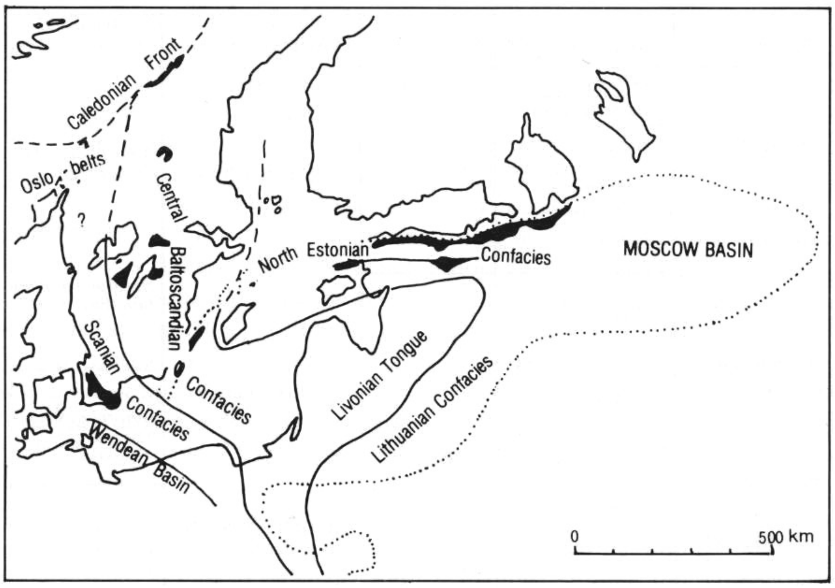


Figure 3-9. *Approximate boundaries of the Baltoscandian Confacies Belts and the location of major outcrop areas of Ordovician sedimentary rocks (Bruton et al. 1985 after Jaanusson 1976).*

In the Ordovician, an epicontinental sea with repeated transgressions and regressions covered most of the southern part of the Fennoscandian Shield. The sedimentation was predominantly governed by carbonates. In the onset of the Ordovician, the Baltica palaeocontinent, including the present-day Fennoscandian Shield, was lying in southerly latitudes (ca. 50°) and was moving northward to ca. 20° during the Ordovician (Torsvik et al. 1996). This led to change of climate from temperate to tropical, which resulted in growth of the sedimentation rate of carbonates and developing of deposits characteristic of the arid and tropical climate (Nestor & Einasto 1997). Such deposits can be seen in the sedimentary sequences from the middle of the Middle Ordovician onwards (Oandu regional stage), and by the end of the Ordovician (Porkuni regional stage) they became dominant.

Lower Ordovician

The first Ordovician transgression occurred at early Lower Ordovician (Early Tremadokian). Mainly non-carbonate terrigenous sediments were deposited: in the west chiefly graptolitic muds, in the east either only quartz sands or sands and graptolitic muds (Männil 1966). During this transgression, the sea is interpreted to have reached southern Finland, the Åland Sea basin and the Bothnian Sea basin (Thorslund 1960; Männil 1966). This is supported by occurrence of Early Tremadocian *Obolus* sandstone, alum shale and limestone (stinkstone) in Finngrundet borehole in the Bothnian Sea (Thorslund & Axberg 1979), and glacial erratics of *Obolus* conglomerate in the Åland Sea region (Hagenfeldt 1995). Alum shales are mostly black shales rich in organic matter, and they range from

Middle Cambrian to Lower Ordovician Tremadocian time in some places (Andersson et al. 1985). The alum shale represents long and very slow deposition under dominantly anoxic or euxinic conditions in an epicontinental sea (Andersson et al. 1985; Bergström & Gee 1985; Ahlberg et al. 1996). According to interpretation by Männil (1966), the transgression was followed by regression at the Late Tremadocian time, during which the whole of Finland was dry land.

Conceptions of the extent of the following transgression during the Billingen and Volkhov regional stages of the Lower Ordovician are variable. According to Thorslund (1960), most of Scandinavia and Finland were covered by sea during the Billingen stage with an exception of area around island of Gotland in Sweden and the north-eastern corner of Finland, whereas Röömusoks (1960), Jaanusson (1963), Männil (1966) and Uutela (1998) suggest that most of Finland was dry land (Fig. 3-10A). Microfossil studies of the Finngrundet borehole in the Bothnian Sea (Tjernvik & Johansson 1979, Löfgren 1985) suggest that the Lower Ordovician sea reached at least to the Bothnian Sea region. According to Uutela (1998), the glacial erratics of reddish marl in Ostrobothnia, in Finland, deposited during the Billingen regional stage and their source is the Bothnian Bay, suggesting that the transgression reached out up to the Bothnian Bay at that time (Fig. 3-10A). Transgression during the successive Lower Ordovician Volkhov regional stage extended at least to the Bothnian Sea, as evidenced by the Volkhov stage limestone in the Finngrundet borehole (Tjernvik & Johansson 1979).

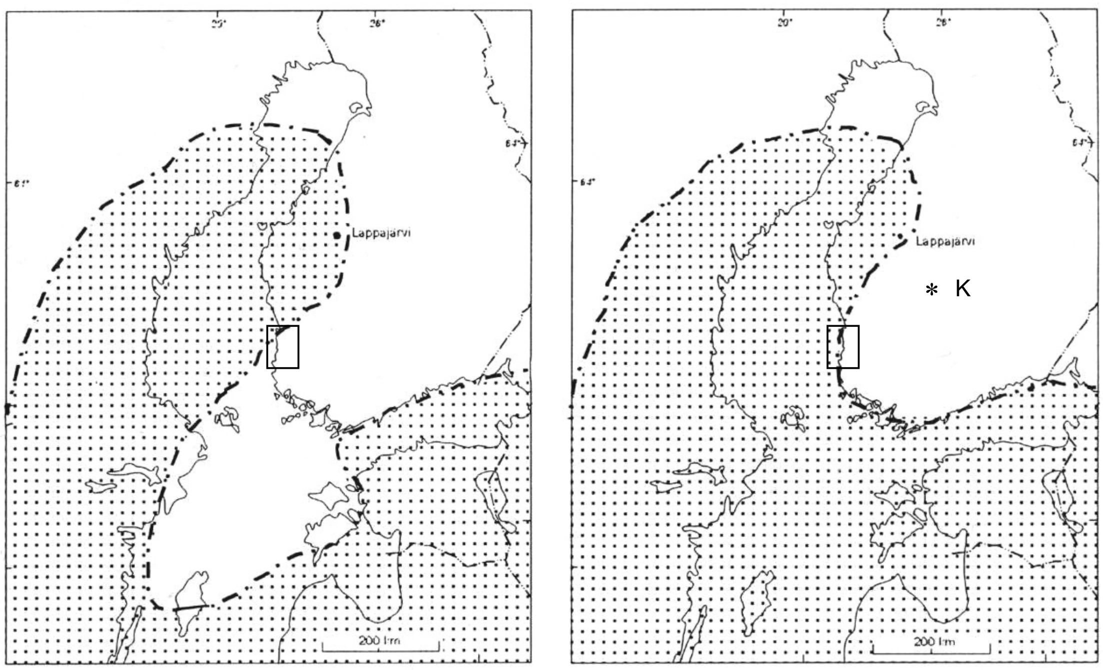
According to Nielsen (1992), Volkhov transgression can be divided into two cycles, a maximum transgression occurring during the early Volkhov regional stage, a major regression during early middle Volkhov, and a new transgression during the late middle Volkhov stage. The transgression was followed by a regression during the late Volkhov age, the sea receding from the whole North Estonian Confacies Belt (Nestor & Einasto 1997).

Middle Ordovician

Following the Lower Ordovician regression, a new phase of transgression began at the Lower to Middle Ordovician Kunda regional stage, during which the transgression reached its maximum spatial extent (Saadre 1992). The Finngrundet and Lake Lappajärvi drill core studies (Tjernvik & Johansson 1979; Löfgren 1985; Uutela 1998, 2001) suggest that during the Kunda regional stage and the successive Middle Ordovician Aseri, Lasnamägi and Uhaku regional stages, the Bothnian Sea and the western coastal area of Finland were covered by sea (Fig. 3-10B).

According to Männil (1966), the Åland Sea and Bothnian Sea belonged to the Central Baltoscandian Confacies Belt during the Lasnamägi-Uhaku regional stages, while the Åland archipelago was part of the North Estonian Confacies Belt. In the Åland Islands, grey-coloured detrital sediments deposited, while detrital sediments with oolites deposited the Åland Sea and Bothnian Sea during the Lasnamägi regional stage. Grey lime muds deposited in western part of the Bothnian Sea during the Uhaku regional stage. Also Hagenfeldt (1995) considers that the Ordovician limestone in Lumparn (*see* Chapter 2.2.4) represent the North Estonian Confacies Belt.

Recently, Middle Ordovician microfossils ranging from the Lasnamägi to Kukruse regional stages have been found in the Lake Karikkoselkä impact structure in Central Finland (Arkonsuo 2000; Uutela 2001). This indicates that the Middle Ordovician transgression reached far more east than what has been previously thought, and that carbonates were deposited also at the southern Satakunta area.



A) **B)**

Figure 3-10. *A.) Interpreted extent of transgression during the Lower Ordovician Billingen regional stage. B) Interpreted extent of transgression during Middle Ordovician Aseri/Lasnamägi and Idavere/Jõhvi transitions (Uutela 1998). The location of Lake Karikkoselkä impact structure is added to figure B and is shown with an asterisk. The southern Satakunta area is marked with a square.*

In Estonia, the period comprising the Uhaku, Kukruse, Idavere, Jõhvi and Keila regional stages, was rather stable in terms of tectonics and change of sea-level (Nestor & Einasto 1997). In northern Estonia, a short erosional sedimentation break exists at the end of the Kukruse regional stage. The period including the Idavere, Jõhvi and Keila regional stages is characterised by the occurrence of the most argillaceous sediments in the whole Ordovician sequence (Nestor & Einasto 1997). A short transgressive phase occurred in the beginning of the period, followed by longer stable period in the middle and drastic regression at the end.

Numerous bentonite (volcanic ash) intercalations deposited during the Idavere, Jõhvi and Keila regional stages, indicating intense volcanic activity in adjacent region (Rõõmusoks 1960, Nestor & Einasto 1997). According to Huff et al. (1992) the bentonites of Estonia can be correlated with the Millbrig bentonite in North America, indicating one of the largest recorded ash falls in the Phanerozoic time.

According interpretations by Männil (1966) and Nestor & Einasto (1997) carbonate sediments were deposited in the Åland Sea and Bothnian Sea areas during the Kukruse regional stage, whereas the mainland Finland was probably dry land. During the Idavere and Keila regional stages sedimentation continued in the Åland Sea and Bothnian Sea areas, probably also in the western and south-western Finland (Männil, op. cit.). According to Uutela (1989), the sea covered the Gulf of Finland and extended to the Bothnian Sea coast during Uhaku, Kukruse and Idavere regional stages. Also the Lappajärvi area was once again covered by sea (Uutela 1998) (Fig. 3-10B).

Late Middle Ordovician to Upper Ordovician

The transition from generally transgressive phase to a phase generally characterised by regressions occurred on the boundary between the Keila and Oandu regional stages (Hints et al. 1989; Nestor & Einasto 1997). Three phases of regression can be recognized: the maximum regression at the boundary between the Keila and Oandu stages, followed by two other phases (Ainsaar et al. 1996). The boundary can be seen as a stratigraphic gap in northern Estonia (Männil 1996) and in Gotland and Västergötaland in Sweden (Jaanusson 1973). In northern Estonia, the boundary between the Keila and Oandu stages also marks the change in the carbonate sedimentation from the microcrystalline limestone rich in skeletal debris, to dominantly cryptocrystalline limestones with variable clay content (Hints et al. 1989). A major change in fauna with remarkable reduction of species happened at the end of the Keilan stage (Hints et al., op. cit.). The Keila-Oandu regression also marks the change of climate from humid to warmer (Jaanusson 1973) or more arid (Hints et al. 1989; Hints 1998).

Männil (1966) suggests that during the Keila-Oandu regression event, Finland was most likely dry land, possibly also the Åland Sea and Åland Islands and most of the Bothnian Sea. Recently, however, a microfossil representing the Keila regional stage has been identified from the Lake Karikkoselkä impact structure in central Finland (Arkonsuo 2000; Uutela 2001), suggesting that the regression must have proceeded more slowly than previously thought.

The Keila-Oandu regression was followed in the Baltic Basin by cyclic alteration of transgressions and regressions, which can be seen as cyclical alteration of the deposition of pure and argillaceous lime muds (Nestor & Einasto 1997). According to Nestor & Einasto (op. cit.), this may possible be due to an interchange of the arid and humid climate periods characteristic of the Late Ordovician-Early Silurian glacial epoch. The regressive arid period caused a relatively low influx of terrigenous material and, consequently, deposition of purer lime muds, while during the transgressive humid periods more argillaceous muds were deposited. Nine of such cycles of alternating low- and high-influx phases have been recognized in the Baltic Basin during Late Ordovician to earliest Silurian times (Nestor & Einasto 1997).

During the late Middle to Upper Ordovician Rakvere, Nabala, Vormsi and to Pirgu regional stages, southernmost and south-westernmost Finland, the Åland Sea basin and the Bothnian Sea basin are interpreted to have been covered by water (Rõõmusoks 1960; Männil 1966; Axberg 1980; Hagenfeldt 1995; Nõlvak et al. 1995) (Fig. 3-11).

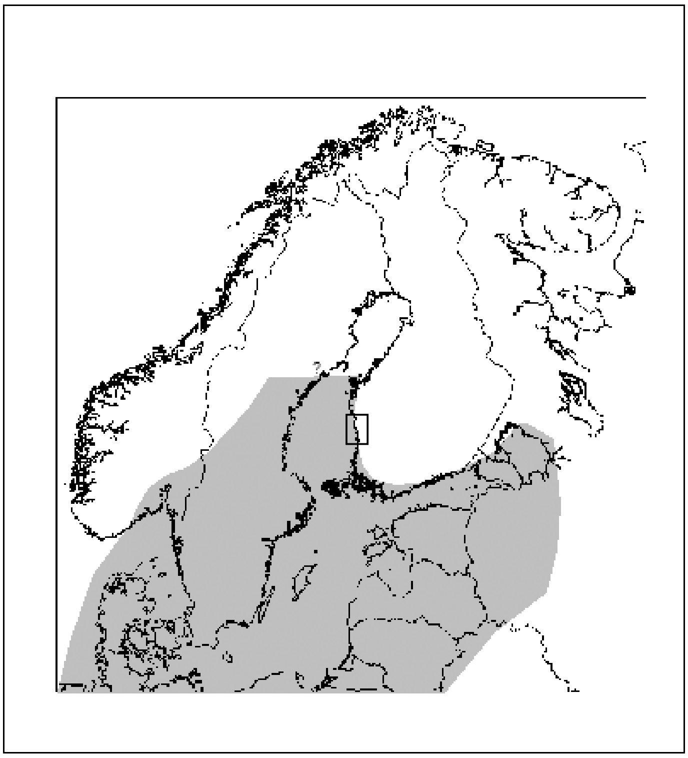


Figure 3-11. *Interpreted extent of the Upper Ordovician sea (grey area) during the Vormsi regional stage (after Männil 1966). The southern Satakunta area is marked with a square.*

Although beginning with a short transgressive event, the Upper Ordovician Porkuni regional stage is characterised by a drastic regression within the Baltic Basin and its surroundings (Rõõmusoks 1960; Männil 1966; Nestor & Einasto 1997). Whole of Finland, including the Åland Islands, the Åland Sea basin and Bothnian Sea basin were dry land (Männil 1966), and subaerial conditions prevailed also in northern Estonia (Nestor & Einasto 1997). According to Nestor & Einasto (op. cit.) this event was directly connected with drop of the ocean level. This regression was global, and due to a major Late

Ordovician glaciation centred in present-day Africa. This glaciation caused a major mass extinction at the end of the Ordovician.

From the late Upper Ordovician onwards until the Pleistocene glaciations, most of Finland including the Åland Islands, the Åland Sea basin and Bothnian Sea basin, is considered to have been in a terrestrial position.

3.5 Silurian-Permian

No Late Palaeozoic sedimentary rocks have been recognised neither in Finland nor in Sweden. However, the fission track studies made in the 1990s (Tullborg et al. 1995; Tullborg et al. 1996; Cederbom 1997; Cederbom et al. 2000; Cederbom 2001) have suggested that extensive Silurian to Devonian deposits most likely covered large parts of the present Fennoscandian Shield. These deposits of the "Old Red Sandstones" type were derived from the eroded Caledonian mountain chain in the north-western margin of the shield. The red colour of the sandstone indicates deposition under oxidizing conditions in continental or non-marine environments. Later, these deposits were eroded and subsequently re-deposited in the southern and south-eastern margins of the Fennoscandian Shield. For example, in Latvia, the maximum thickness of Silurian to Devonian sediments is 800 m.

Thermal indicators, such as $\delta^{18}O/\delta^{13}C$, conodont alteration indices (CAI), illite/smectite ratios and illite crystallinity together with apatite fission track studies including track-length distribution analyses show that during the period from Silurian to Devonian, the basement rocks in south central Sweden were heated to more than 120° resulting in total annealing of apatite (Tullborg et al. 1995; Tullborg et al. 1996; Cederbom 1997; Cederbom et al. 2000; Cederbom 2001). In the Åland Islands, the temperature was somewhat lower (ca. 60°) causing only partial annealing (Cederbom 1997). Also the Phanerozoic lower intercept U-Pb zircon ages recorded in the crystalline basement rocks of the Fennoscandian Shield are interpreted to be due to thick sediment cover (Larson & Tullborg 1998). Because no thermal event related to Silurian or Devonian is known in the Fennoscandian Shield, the annealing is suggested to be due to thick sedimentary cover derived from the Caledonides. Tullborg et al. (1995) propose that this sedimentary cover was largely composed of molasses, which were deposited into a Caledonian foreland basin.

Foreland basins are regions of subsidence, which are associated with regions of compressional tectonics. They are formed primarily as a result of the downward flexing of the lithosphere in response to the weight of the adjacent mountain belt. According to Samuelsson & Middleton (1998), this foreland basin was 6.5 km deep and 350 km wide, while Larson et al. (1999) proposed a much wider basin (more than 600 km).

The thickness of the sedimentary sequences in the Caledonian foreland basin may have been 3 - 4 km in southern and western Sweden thinning to about 1 km in the Åland Islands (Tullborg et al. 1995; Tullborg et al. 1996; Cederbom 1997; Cederbom et al. 2000) (Fig. 3-12). Based on apatite fission track results from the samples from Åland, Olkiluoto, Kivetty and Romuvaara, Larson et al. (1999) see it likely that most of Finland was buried by Upper Palaeozoic sediments that thinned to the east (Fig. 3-12). It can be inferred from Figure 3-12 that also the southern Satakunta area was covered by ca. 1 km thick pile of sedimentary rocks. According to Larson & Tullborg (1998) and Larson et al. (1999), the Upper Palaeozoic sedimentary deposits covered the Fennoscandian basement at least 200 million years.

Preliminary results of the apatite fission track study in Finland by Murrel & Andriessen (2000) show apparent ages of 350 to 450 Ma (Silurian-Devonian) in central southern Finland and either significantly lower (270 Ma) or higher (850 Ma) ages at the margins. The reason for this is not known but it may be due, for example, to differential movements in the crust (Murrel & Andriessen, op. cit).

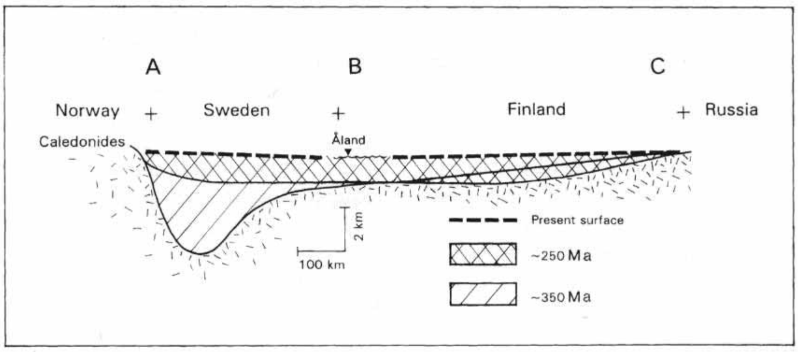


Figure 3-12. *Tentative cross-section of the Caledonide foreland basin. Thickness of the deposits is shown at ca. 350 Ma and ca. 250 Ma (Larson et al. 1999).*

Based on the lithofacies distributions in Estonia, the margin of the epicontinental shallow sea is interpreted to have been located in the southern part of Estonia during the Devonian (Kleesment 1997). In Estonian part of the Baltic Basin, the continental siliciclastic sedimentation in response to the uplift of the Caledonides started in the Early Devonian, followed by the main phase of sedimentation from the late Early Devonian to the end of the Middle Devonian (Nestor & Einasto 1997; Plink-Björklund & Björklund 1999). The siliciclastic deposition was dominated by deltaic deposition with the sediment influx from the northwest (Plink-Björklund & Björklund 1999).

Upper Carboniferous to Lower Permian sedimentary rocks are present in the fault-bounded Oslo Graben (Neuman et al. 1992) but are absent from the rest of the Fennoscandian Shield. Their original extent is unknown, but Bergström et al. (1985) have suggested that the Upper Carboniferous limestones in the Oslo Graben may be connected to the Moscow basin sediments southeast of the Fennoscandian Shield. If ever present, the Carboniferous sediments were probably restricted to southern Scandinavia, since during that time the Fennoscandian Shield was a relatively low-relief highland (Ziegler 1989).

3.6 Mesozoic (Jurassic, Triassic, Cretaceous)-Cenozoic (Tertiary)

The formation of the Mesozoic peneplain (or the surface defined by the summits of the highest mountains) of the Caledonian orogeny (cf. Riis 1996) probably took place already in Carboniferous or Permian (Peulvast 1985), and by the Mid-Jurassic, deeply weathered peneplain existed in western Fennoscandia, whereas a platform covered with Palaeozoic and Mesozoic sedimentary rocks was present in eastern Fennoscandia (Riis 1996). The Palaeozoic (Silurian to Devonian) sedimentary cover was significantly reduced due to late Palaeozoic to Mesozoic (Carboniferous to Triassic) uplift and erosion (Tullborg et al. 1995; Tullborg 1997; Larson et al. 1999; Cederbom 2001). Southern and central Sweden may have been covered by Palaeozoic and Mesozoic sediments until Tertiary time (Lidmar-Bergström 1991, Cederbom et al. 2000), and in Finland the occurrence of Cambrian and Ordovician microfossils in the Lappajärvi impact structure (Uutela 1998) bears evidence of remnants of the Lower Palaeozoic and possibly younger sedimentary cover still in present there in Late Cretaceous time (*see* Chapter 2.2.7). Jurassic cooling and erosion is suggested for south-eastern Sweden and north-western part of South Sweden (Cederbom 2001).

In South and Central Norway, the sub-Mesozoic peneplain was transgressed in the Late Jurassic to Early Cretaceous. In east-central Sweden, east of the southern part of the Caledonides (the Southern Scandes), the undulating hilly relief is interpreted to be a sub-Cretaceous erosional surface, which have probably been protected by Cretaceous sedimentary cover (Lidmar-Bergström 1999). At least in southern Sweden, the late Palaeozoic and Mesozoic uplift declined or possibly ceased during the Cretaceous, and the transgression caused deposition of thin marine sediments that have subsequently eroded (Tullborg et al. 1996; Larsson et al. 1999).

Hardly any Mesozoic or Tertiary sedimentary rocks are preserved within the Fennoscandian Shield, the Mesozoic to Tertiary strata restricting to the Fennoscandian Border Zone in the southernmost province of Scania in Sweden and adjacent offshore areas (Norling & Bergström 1986; Bergström & Kornfält 1998). In southern Sweden, the Mesozoic rocks are lying directly on the Precambrian basement indicating hundreds of metres of early Mesozoic uplift and erosion (Lidmar-Bergström 1991).

On the basis of microfossils, the clay of the Akanvaara deposit in eastern Lapland (*see* Chapter 2.2.14) was deposited in early Tertiary (Eocene) under marine conditions (Hirvas & Tynni 1976; Tynni 1982c). According to Tynni (op. cit.) the Akanvaara clay is possibly a relict of a more extensive marine clay bed, deposited in an arm of the Arctic Ocean (Fig. 3-13).

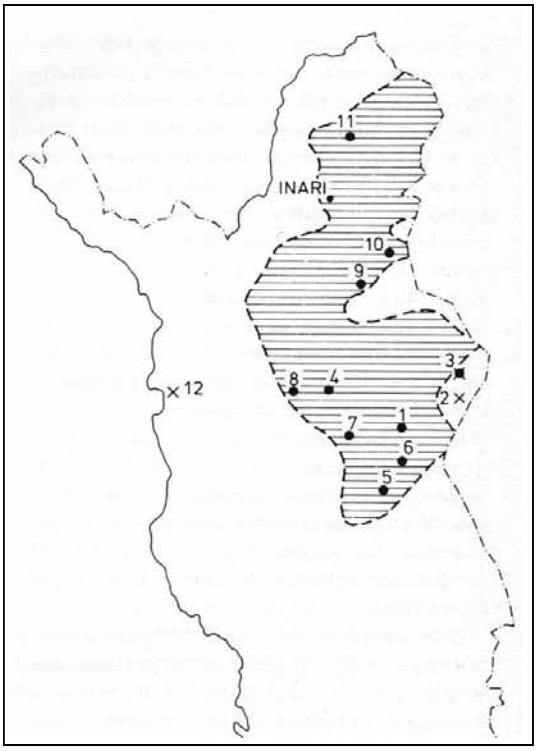


Figure 3-13. *Discovered early Tertiary marine diatoms (1, 3-11) and the interpreted extent of the submarine area (Tynni 1982c). Numbers 2 and 12 refer to locations of late Tertiary fresh-water diatoms.*

The Tertiary erosion resulted in morphology characterised by almost horizontal plains with residual hills in northern Finland and Sweden (Lidmar-Bergström 1999). Riis (1996) has interpreted the deeply weathered basement underlying the Quaternary cover and the remnants of Tertiary (Eocene) microfossils in Lapland as representing the Palaeocene-Eocene erosional surface. Fig. 3-14 shows the inferred extent of that surface of erosion. Based on offshore data, the peneplanation may have begun already in the late Cretaceous (Riis 1996).

The almost complete lack of Mesozoic and Tertiary sediments within the Fennoscandian Shield is considered to be due to Tertiary uplift (*see* Chapter 4.6).

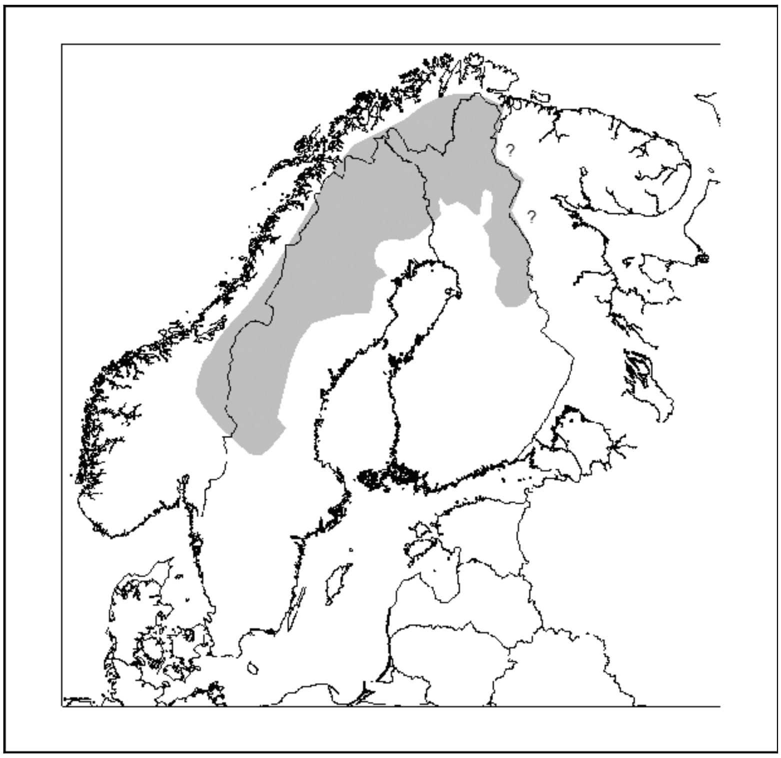


Figure 3-14. *Distribution of inferred Tertiary peneplain in Finland, Norway and Sweden (after Riis 1996).*

4 TECTONICS AND IGNEOUS ACTIVITY WITHIN AND IN THE MARGINS OF THE FENNOSCANDIAN SHIELD DURING THE LAST 1300 Ma

4.1 Sveconorwegian orogeny (ca. 1300 – 800 Ma)

4.1.1 Laurentia-Baltica supercontinent; palaeomagnetic and geological constraints

Recent palaeomagnetic and geological reconstructions (Åhäll & Gover 1997; Åhäll & Connelly 1998; Buchan et al. 2000; Connelly et al. 2001) suggest that at ca. 1265 Ma (and at least 200 Ma before that) Laurentia craton and Baltica craton formed a coherent continent with present-day northern Fennoscandia facing eastern Greenland. In North America, the Laurentia craton consists of a number of Archaean cratons that have been welded together by collisional orogenies. From the Late Archaean to the late Mesozoic or early Cenozoic, Greenland was attached to the North American portion of Laurentia (Buchan et al. 2000 and references therein). The northern British Isles are often assumed to be attached to Laurentia from at least Mesoproterozoic time until the Cretaceous opening of the Atlantic Ocean. The Baltica craton consists of the present-day Fennoscandian Shield and the East-European Platform that stretches from the White Sea area in north-eastern Russia to the Baltic countries and Poland and towards the Ukraine and Moldova.

The presence of the Mesoproterozoic Baltica-Laurentia supercontinent are geologically supported by bimodal intrusions in the SW margin of the Fennoscandian Shield, dated at 1500 to 1200 Ma, which are typical of intracratonic magmatism and have correlative intrusions in eastern Laurentia (Åhäll & Connelly 1996, 1998). Lack of any evidence of a continental margin suggests that the 1460 Ma old, N-S trending doleritic Trond-Göta dykes represent rifting event, which, however did not led to crustal separation. The dykes can be correlated with 1460 - 1430 Ma gabbros in eastern Laurentia, representing a failed rift. Åhäll & Connelly (1996, 1998) argue that this large scale extensional event affecting a strike-length of over 2000 km would require the presence of a large, coherent cratonic mass, and that a third unknown continent existed on each side of the extensional zone at ca. 1460 Ma (Fig. 4-1).

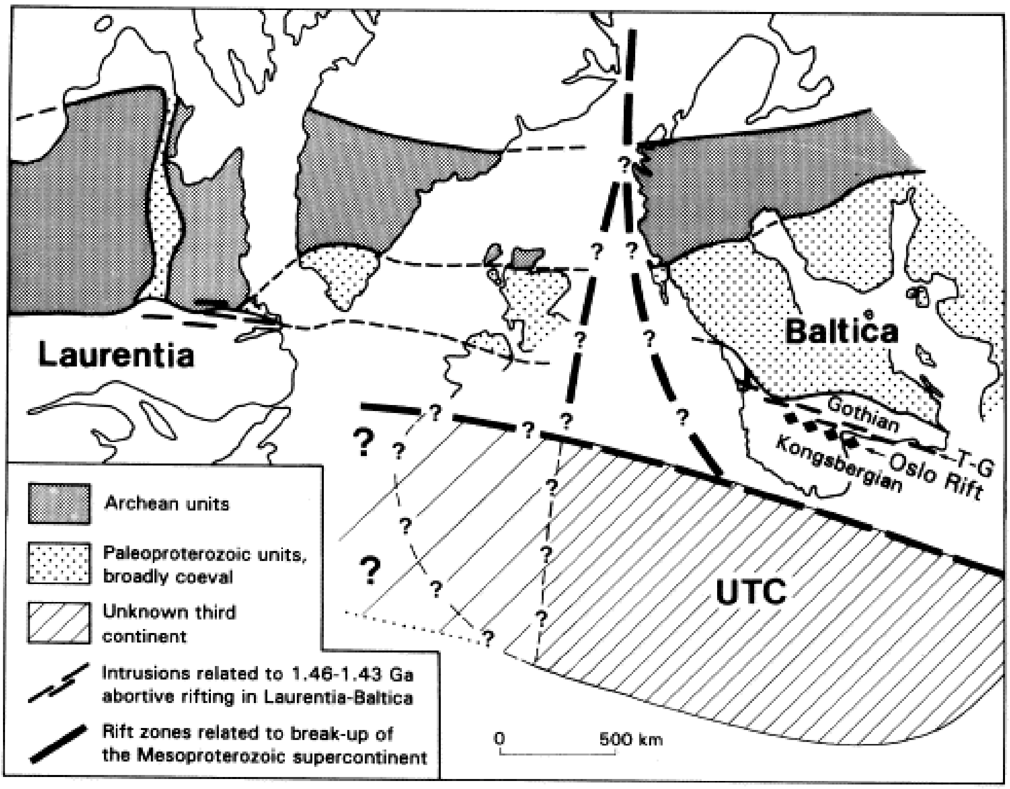


Figure 4-1. *Suggested plate reconstruction in the north Atlantic region at ca. 1.46 Ga (Åhäll & Connelly 1998). The Oslo rift separates an eastern and western part of the present southeast Scandinavia characterised by rocks of Gothian (1.75-1.55 Ga) and Kongsbergian (ca. 1.6 Ga) orogenies, respectively. T-G = 1.46 Ga Trond-Göta dykes, UTC = Unknown Third Continent.*

The margin of the supercontinent was variably reworked during the main phase of the Sveconorwegian orogeny (in North America the Grenvillian orogeny) ca. 1100 - 900 Ma ago, which, in the periphery of the Fennoscandian Shield, mainly affected present south-western Sweden and southern Norway. The area was earlier affected by the Gothian orogeny between 1750 and 1550 Ma (roughly equivalent to the Labradorian orogeny in Laurentia).

On the basis of palaeomagnetic data, Mertanen et al. (1996) suggest that Baltica and Laurentia separated from each other sometime between 1270 - 1050 Ma and they rotated about 90°. During the rifting phase Baltica was located at latitude of about 20°S and drifted further south at the beginning of the Sveconorwegian orogeny. The highest southern latitude of ca. 70°S was attained 930 Ma ago.

Romer (1996), however, argues that the Laurentia and Baltica cratons did not separate until *after* the Sveconorwegian orogeny. The Sveconorwegian (and Grenvillian) orogen is interpreted by Romer (op. cit.) to have formed during the collision of coherent Laurentia-Baltica with an unknown craton, which is in contrast to earlier explanations that favour a late-Mesoproterozoic rotation between Laurentia and Baltica followed by collision of these two cratons with each other.

Pesonen et al. (2001) have a different view on the Sveconorwegian orogeny. According to palaeomagnetic reconstruction they suggest that Laurentia and Baltica were still united at 1250 Ma, when Congo/Sao Francisco craton began to approach Baltica. At ca. 1200 Ma ago, Baltica and Laurentia were separated by Congo Sea. Pesonen et al. (op. cit) suggest that the Sveconorwegian orogeny was caused by the collision between Congo craton with Baltica at about 1080 Ma ago, while at about the same time Amazonia craton collided with Laurentia to form the Grenvillian orogenic belt in Laurentia. At about 1050 Ma ago, all these continents with other cratons were together and form the supercontinent Rodinia. In the last stage, Baltica collided with the north-eastern margin of Greenland causing late-Sveconorwegian overprints in north-eastern Greenland and Baltica. In the following chapters, however, the Sveconorwegian orogeny is described on the basis of more common assumption of Laurentia-Baltica collision.

4.1.2 Break-up of the supercontinent; initial rifting and magmatism

According to studies made in southern Norway, the Sveconorwegian orogeny can be divided into three main phases (Starmer 1993): (1) an early Sveconorwegian thrusting phase between ca. 1340 and 1250 Ma, (2) extensional period between ca. 1250 and 1100 Ma, and (3) the main phase of the Sveconorwegian orogeny (the collision between Baltica and Laurentia) between ca. 1150 - 950 Ma

In an early Sveconorwegian phase between ca. 1340 and 1250 Ma, NW thrusting took place and early Sveconorwegian granitoids were emplaced in SW Sweden and southern Norway (Åhäll et al. 1990; Lindh & Person 1990; Starmer 1993) (Fig. 4-2). The Mesoproterozoic, 'Postjotnian' olivine diabases, dated at 1270 - 1250 Ma, in Sweden (Central Scandinavian Dolerite Group), Finland (Satakunta and Vaasa olivine diabases) (Fig. 4-3), and related diabases in Greenland are considered to represent the initial rifting between the Baltica and Laurentia cratons. 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 for the diabases, which would have been located in the reconstructed contact between Baltica and Laurentia. Consequently, Elming & Mattsson (op. cit.) suggest that the 'Postjotnian' magmatism reflects the tensional tectonics during the break up of the coherent Baltica/Laurentia continent. 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.

Extensional period between ca. 1250 and 1100 Ma followed the early Sveconorwegian collisional phase. Gabbroic and granitic rocks related to this early Sveconorwegian extension are found in the Bamble-Kongsberg sector in SE Norway (Starmer 1991, 1993). The main phase of the orthopyroxene gabbro (hyperite) intrusions was at ca. 1250 - 1220 Ma, the emplacement of the granitic intrusions culminating at ca. 1140 Ma (Starmer 1991).

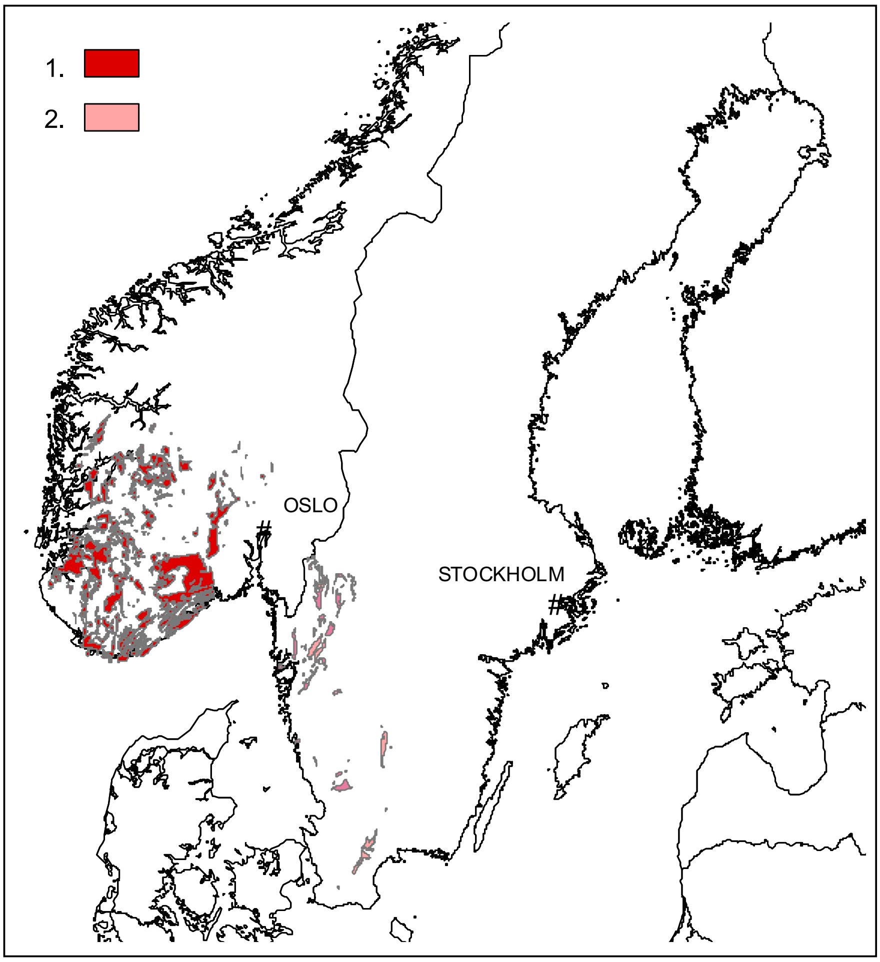


Figure 4-2. *Sveconorwegian intrusives in southern Norway and south-western Sweden. Legend: 1. granites, granodiorites, trondhjemites, monzonites, monzodiorites, diorites and metamorphic equivalents (ca. 1270 - 1000 Ma), 2. granites, syenites and metamorphic equivalents (ca. 1250 - 1200 Ma). Compiled from Koistinen et al.(2001).*

Nearly coeval with the 'Postjotnian' diabase dykes are ca. 1180 Ma old mafic dyke swarms associated with the Protogine Zone in southern Sweden (Johansson & Johansson 1990) (Fig. 4-3). Besides alkaline mafic dykes, coeval syenites and granitic rocks occur. The dykes may indicate continental rifting at the onset of the Sveconorwegian Orogeny (Solyom et al. 1992).

In Finnish Lapland, the NW-SE trending, over 100 km long and 60 - 100 m wide Salla diabase dyke swarm (Fig. 4-3) is 1130 - 1170 Ma in age (Lauerma 1987). The E-W striking diabase dyke swarm in central Sweden (Fig. 4-3) could be of the same age (Larson & Tullborg 1994). In Luleå and Kalix archipelagos, in the west coast of the Bothnian Bay, about 100 kimberlite dykes, mainly striking N-S occur (Ödman 1957). They are classified as alnöites, silicocarbonatites and beforsites, and have ages around 1150 Ma, thus coinciding with the Sveconorwegian orogeny (Kresten et a. 1977; Welin 1979). According to Wannäs (1989), however, the dykes may have intruded during several events connected to lateral and/or vertical movements, as indicated by variation of the petrographic composition of the dykes.

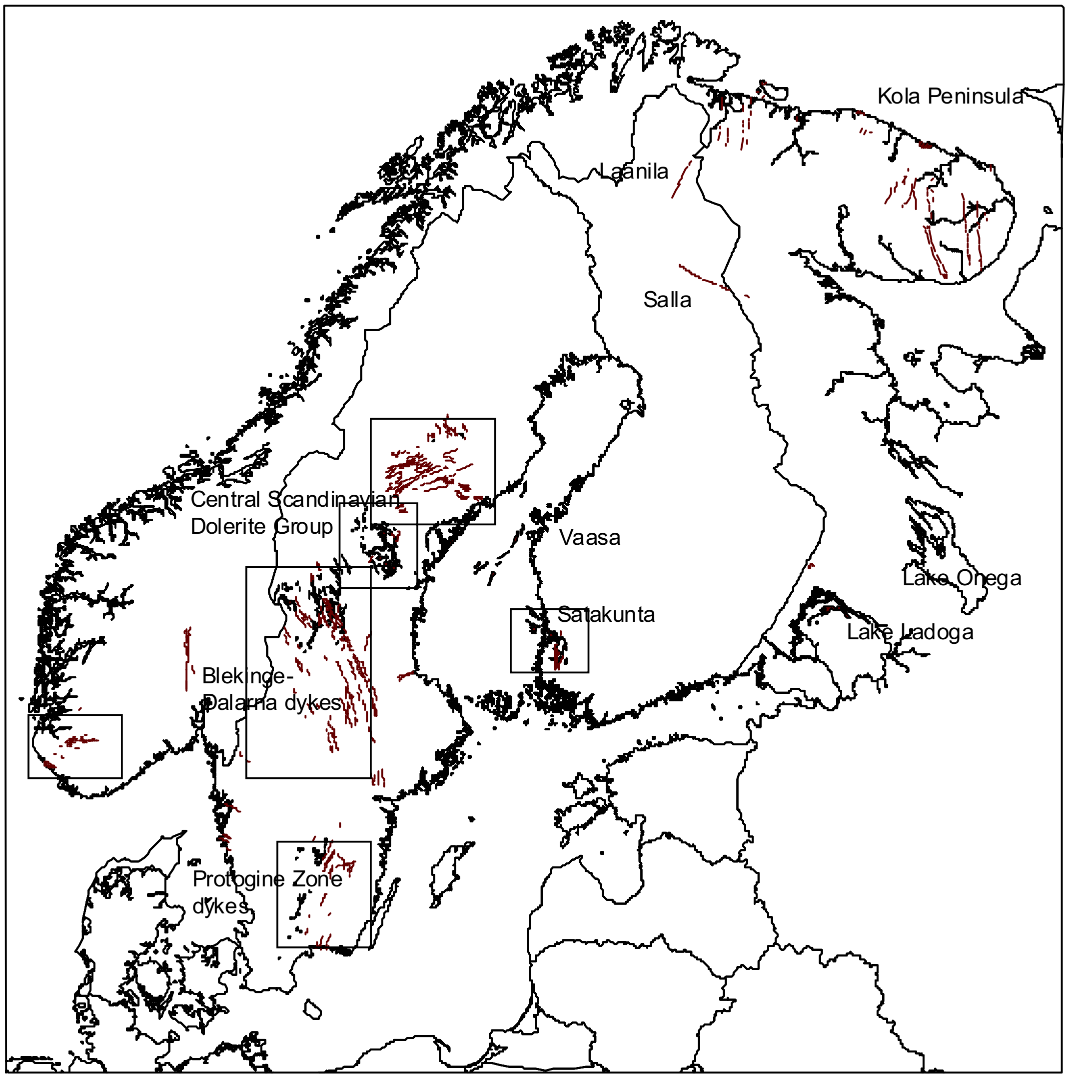


Figure 4-3. *Mesoproterozoic and Neoproterozoic (see text) diabase dykes and sills in the Fennoscandian Shield. Compiled from Koistinen et al.(2001).*

The rifting gave rise to a system of rifts and aulacogens both along the present-day western and eastern margins of Baltica and in the interior of the craton (Nikishin et al. 1996). According to Nikishin et al. (op. cit.), the Gulf of Bothnia rift, including the Satakunta and Muhos Grabens, Bothnian Sea Basin and the Bothnian Bay Basin, started at that time but its development may have started already earlier due to rifting in connection with the emplacement of the rapakivi batholiths (e.g., Korja & Heikkinen 1995). According to Korja et al. (2001) thin crust with large crustal variations in the Gulf of Bothnia region, large sedimentary basins 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. However, subsidence occurred in the Satakunta graben in connection with rifting and the intrusion of ‘Postjotnian’ diabases (Laitakari 1983).

According to Wannäs (1989) the initial development of the Bothnian Bay sedimentary basin (*see* Chapter 2.2.2) is related to the rifting at the onset of the Sveconorwegian orogeny (1250 - 900 Ma). In the Muhos Graben, the sedimentary pile has downfaulted hundreds of metres after the sedimentation ca. 1200 Ma ago, based on the similarity between the outcropping Kieksi conglomerate and the basal conglomerate in the borehole.

4.1.3 Sveconorwegian orogeny; convergence, thrusting and extension

The main phase of the Sveconorwegian orogeny (the collision between Baltica and Laurentia) occurred ca. 1150 - 950 Ma ago (Starmer 1993). In the Bamble segment, SW Norway, early compression and partial melting is dated at 1130 Ma, and it was preceded by subduction of the continental margin at around 1200 Ma (Knudsen and Andersen 1999). The oblique convergence during the collision resulted in south-eastward and eastward thrusting in SW Sweden (Park et al. 1991) with thickening of the crust (Johansson et al. 1991). In southern Norway, coarse-grained and medium-grained granites were emplaced at ca. 1060 - 1050, 1025 Ma and 1000 - 990 Ma (Starmer 1991) (Fig. 4-2). The Sveconorwegian metamorphism resulted in widespread granulite facies assemblages in the southern distal (eastern) part of the orogeny indicating burial depth of 30 - 45 km (Johansson et al. 1991), while amphibolite facies assemblages were dominant elsewhere (e.g. Wahlgren et al. 1994). The Sveconorwegian deformation and metamorphism is bracketed between ca. 1030 and 920 Ma (Johansson & Johansson 1993; Connelly et al. 1996; Page et al. 1996; Wang et al. 1996; Möller & Söderlund 1997). The older ages (until ca. 960 Ma) have been assigned to crustal thickening (Page et al. 1996; Möller & Söderlund 1997; Andersson et al. 2001), while the younger ages may be associated with extensional uplift (Page et al. 1996). According to Romer & Smeds (1996), the eastward thrusting ceased by ca. 980 Ma in the entire orogen.

During the Sveconorwegian orogeny, major deformation occurred in south-western Sweden along N-S striking major shear zones, including from west to east the Dalsland Boundary Thrust, the Göta Älv Zone, the Mylonite Zone and the Protogine Zone (Larson et al. 1990;

Park et al. 1991; Wahlgren et al. 1994). These shear zones divide the area into several crustal segments (Western, Median and Eastern), between which the timing, character and intensity of deformation and metamorphism may vary considerably. These differences indicate large-scale displacements along the shear zones (Andersson 2001; Andersson et al. 2001).

The Mylonite Zone (Fig. 4-4) is a N-S striking, ductile deformation zone, which can be followed for over 400 km from the west coast of Sweden into Norway (Wahlgren et al. 1993). It is dominated by sinistral, strike-slip motion with associated compressive duplex and imbricate structures, indicating that the Mylonite Zone is transpressional in character (Wahlgren et al. 1993, Stephens et al. 1996). Studies of the kinematics of the Mylonite Zone suggest oblique NW-SE convergence between Baltica and Laurentia during Sveconorwegian collision and eastward or south-eastward thrusting along the Mylonite Zone (Park et al. 1991). Compressional deformation with eastward thrusting was followed by sinistral transpressional deformation and finally by extensional uplift between ca. 934 and 915 Ma (Page et al. 1996; Cornell et al. 1996; Åhäll & Schöberg 1999; Möller & Söderlund 1997; Connely & Åhäll 1996; Åhäll 1995). According to Åhäll & Gower (1997) the Mylonite zone most likely developed over a pre-existing Gothian suture, the view that is not shared by Andersson (2001).

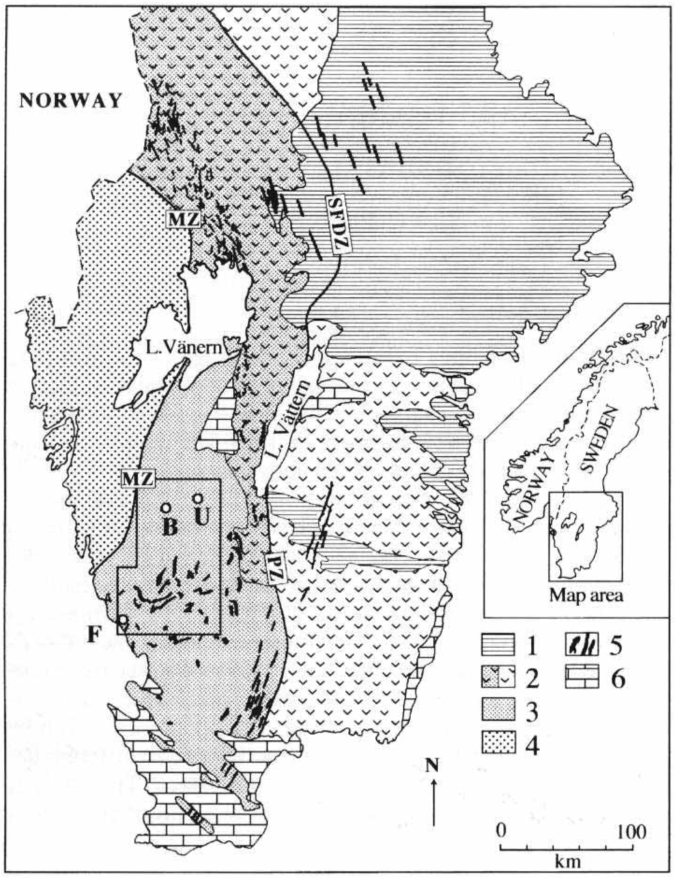


Figure 4-4. *Major lithological and tectonic units of southern Sweden (after Möller & Söderlund 1997). 1. Svecofennian Province, 2. Transscandinavian Granite-Porphyre Belt (shaded part affected by Sveconorwegian deformation), 3. Eastern Segment, 4. Western Segment, 5. Undifferentiated mafic rocks including amphibolites, gabbros, diabase dyke swarms, etc., 6. Phanerozoic cover. MZ, Mylonite Zone; PZ, Protogine Zone; SFDZ, Sveconorwegian Frontal Deformation Zone.*

The Protogine Zone or the Sveconorwegian Frontal Deformation Zone (Fig. 4-4) is a ca. 20 km wide and 700 km long fan-like tectonic belt of nearly vertical to subvertical faults or shear zones and steep fractures (e.g., Andréassen & Rohde 1990, Larson et al. 1990, Wahlgren et al. 1994). Oblique sinistral compression along the zone between about 1205 Ma and 1010 Ma (Hansen & Lindh 1991; Page et al. 1996) was followed by dextral transpressional deformation at ca. 940 Ma (Johansson 1990), and an extensional uplift at 930 - 920 Ma (Johansson and Johansson 1990, Eliasson and Schöberg, 1991). According to Larson (1996), the southern part of the Protogine Zone may originally be a Gothian suture.

In the Eastern segment, extensional uplift along the Mylonite Zone and the Protogine Zone resulted in exhumation of high-grade rocks at 960 - 920 Ma (Söderlund 1999).

Several authors (e.g. Larson et al. 1986, Åhäll 1995; Åhäll et al. 1995; Connelly et al. 1996) have considered that the Sveconorwegian deformation and migmatisation outside the major shear zones was subordinate compared to the earlier Gothian deformation. Möller & Södelund (1997), Andersson et al. (1999) and Andersson (2001) have, however, demonstrated that in the eastern part of the orogeny (the Eastern segment) high-grade Sveconorwegian ductile deformation with associated partial melting took place outside the shear zones.

The late Sveconorwegian extension resulted in an extensive rifting. The 700 km long and 150 km wide ca. NNE-SSSW striking Blekinge-Dalarna dyke swarm (Fig. 4-3), dated at ca. 930 Ma, is related to the late Sveconorwegian extensional uplift at the area west of the Protogine Zone (Johansson & Johansson 1990). Andréassen (1994), however, argues that the dykes are not related to Sveconorwegian orogeny but indicate initial stages of rift magmatism at the onset of a new continental break-up after the Sveconorwegian orogeny.

At Södermanland, east central Sweden, alkali-basaltic dolerite dykes, belonging to the Blekinge-Dalarna dyke swarm, are predominantly striking N-S and have an Rb-Sr age of ca. 1000 Ma (Risku-Norja 1992).

In SW Finland low-altitude aeromagnetic maps suggest that the 1250 - 1270 Ma old olivine diabase dykes are cut by a swarm of younger diabase dykes striking almost north-south (Vorma & Niemelä 1994, Veräjämäki 1998). These dykes are not exposed but according to drilling observations they are 1 to 5 m wide and, contrary to the olivine diabase 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). Whether these dykes are comparable to the Sveconorwegian dykes in Sweden or are they related to some other event, remains to be solved.

In the Kola Peninsula in Russia, the 700 km long tholeiitic Kola-Lake Onega dyke swarm (Fig. 4-3) has been dated at ca. 1000 Ma (Gorbatschev et al. 1987). The NE-SW trending Laanila dyke swarm in Finnish Lapland (Fig. 4-3) is around 1000 Ma (Pihlaja 1987), and may belong to the Kola swarm. The geochemical characteristics of the diabase dykes indicate continental within-plate rift volcanism.

Post-compressional granitic intrusions of the Sveconorwegian orogeny include the 920 Ma Bohus granite in Sweden and the coeval granites in south-western Norway (Fig. 4-5) (Eliasson & Schöberg 1991, Zheng 1996 and references therein) and the ca. 960 Ma old plagioclase porphyritic Vinga intrusion (Åhäll & Schöberg 1999). Extensional conditions have been inferred for the emplacement of the intrusions (Zheng 1996, Åhäll & Schöberg 1999). Post-compressional movements with a sinistral strike-slip component along the N-S striking shear zones (the Mylonite Zone) resulted in WNW trending fracture system, which provided an emplacement mechanism for the Vinga intrusion (Åhäll & Schöberg 1999).

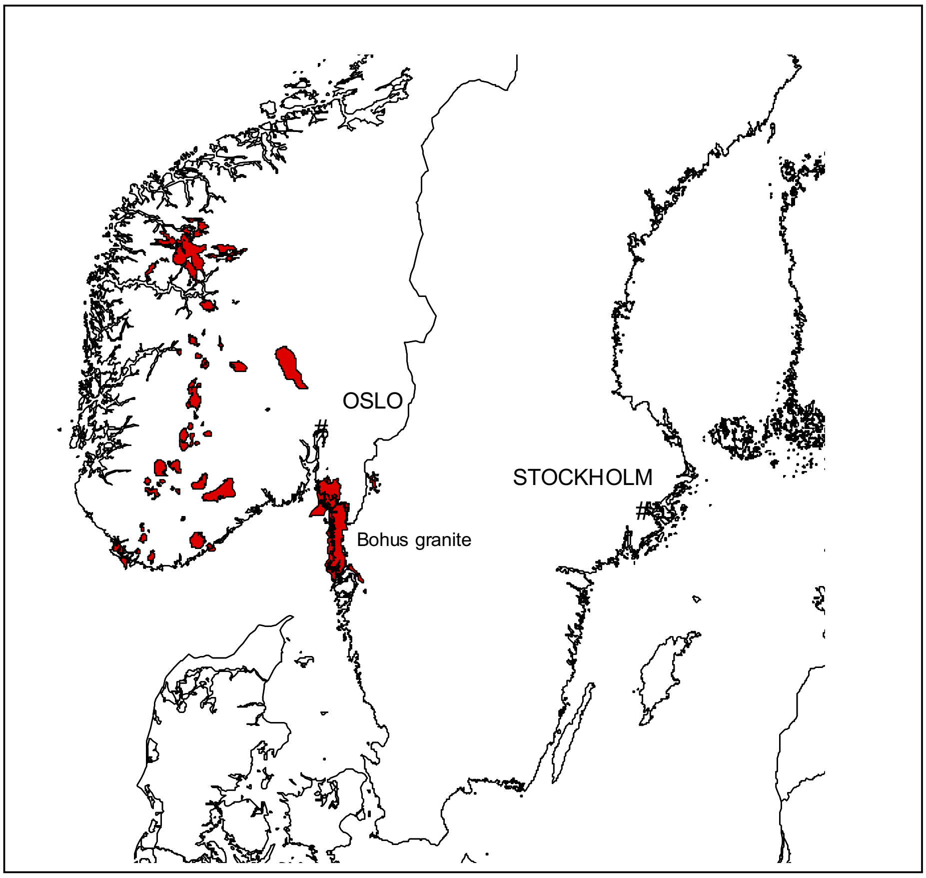


Figure 4-5. *Ca. 1000 - 920 Ma old late Sveconorwegian granites and pegmatites in SW Sweden and Norway. Compiled from Koistinen et al.(2001).*

Approximately E-W striking swarms of mafic dykes, ca. 900 - 800 Ma in age, in southern Norway (Fig. 4-3) and adjoining parts of Sweden probably represent a final Sveconorwegian fracturing event (Gorbatschev et al. 1987).

The late Sveconorwegian orogenic uplift caused 30 - 45 km of erosion in the south-western part of the Fennoscandian Shield (Johansson et al. 1991). Sphene and zircon fission track studies together with geological constraints, indicate that 7 to 8 km of sediments were deposited in the foreland, east of the Sveconorwegian orogen (Zeck et al. 1988, Larson et al. 1999). The subsequent uplift and erosion removed most of the sediments until the end of the Proterozoic.

4.2 Neoproterozoic break-up of Rodinia supercontinent (ca. 800 – 540 Ma)

After the Sveconorwegian orogeny, Laurentia and Baltica together with all the other continents formed one single worldwide supercontinent, Rodinia (Fig. 4-6). According to palaeomagnetic reconstructions by Torsvik et al. (1996) and Dalziel (1997), the north-western margin of Baltica was facing East Greenland, while the Amazonian continent was the conjugate margin of the south-western margin of Baltica. This supercontinent started to break up at about 750 - 725 Ma (Dalziel et al. 1994; Dalziel 1997), when East Gondwana (Australia and Antarctica) rifted off the western margin of Laurentia (Torsvik et al. 1996).

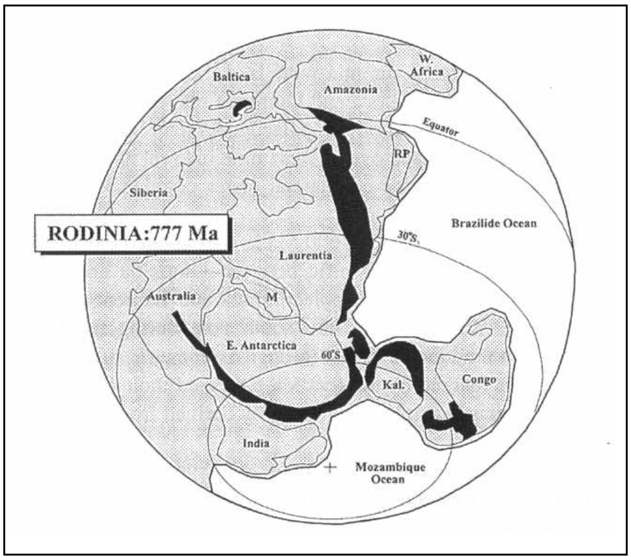


Figure 4-6. *The Rodinia supercontinent at c. 777 Ma (Torsvik et al. 1996).*

4.2.1 Intracratonic rifting and basin evolution within Baltica craton

Within Baltica craton, extensional tectonics and intracratonic basin evolution was initiated ca. 800 - 750 Ma ago (Kumpulainen & Nystuen 1985; Vidal & Moczydłowska 1995). Aulacogens, rift basins and continental depressions formed within the craton from Ukraine to Scandinavia, characterized by predominantly continental sediments (Fig. 4-7) (Kumpulainen & Nystuen 1985; Vidal & Moczydłowska 1995; Nikishin et al. 1996). These include, e.g., the Vättern Graben in SW Sweden, the Ljublin basin in eastern Poland, the Volyn aulacogen in Belarus and Ukraine, and the Moscow rift in Russia. At the same time, Middle Riphean Kandalaksha and Ladoga aulacogens were tensionally reactivated (Nikishin et al. 1996). Within the Caledonides Upper Proterozoic to Cambrian sedimentary rocks occur within number of thrust sheets (Kumpulainen & Nystuen 1985). The sedimentary rocks were originally deposited in fault basins west of the present Caledonides, from where they were thrusted to their present locations during the subsequent Caledonian orogeny. In the eastern margin of the Baltica craton, rifting resulted in separation of a crustal block from the craton and formation of Peri-Timan and Peri-Ural pericratonic sedimentary basins with marine sediments (Nikishin et al. 1996).

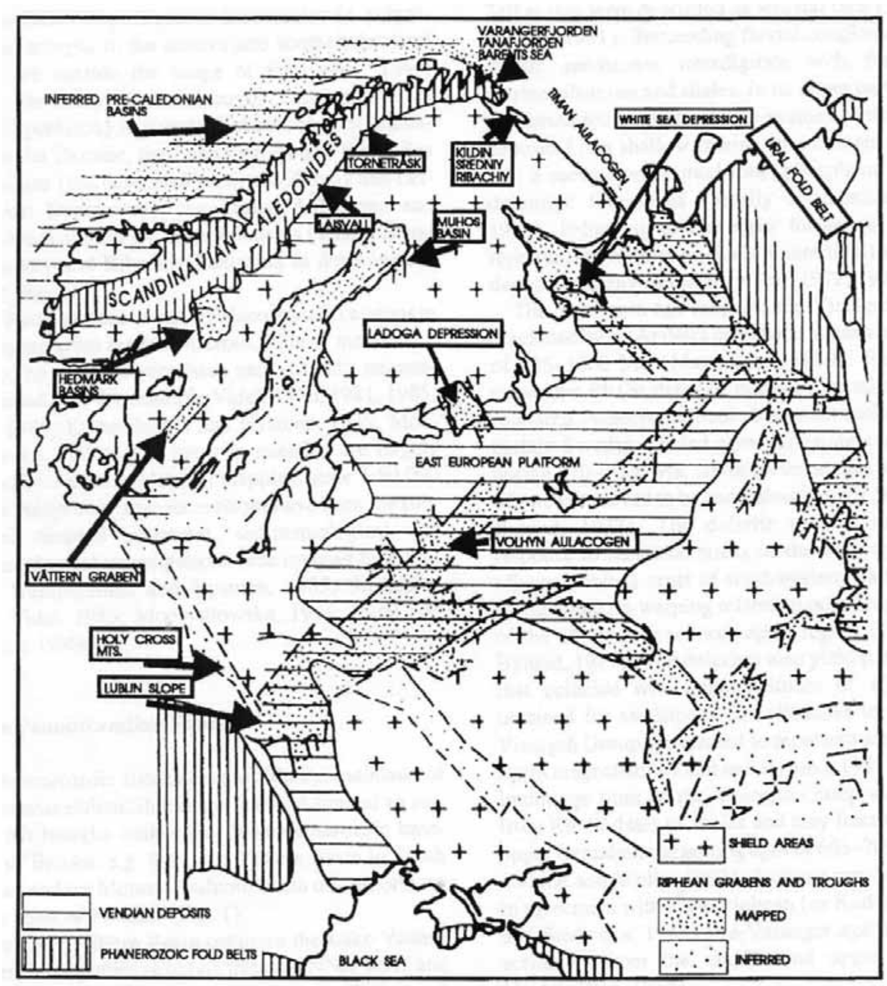


Figure 4-7. *Generalized geological map of Baltica craton with subsurface aulacogen basins (grabens and througs) and the approximate distribution of shield areas and Meso- and Neoproterozoic (Riphean and Vendian) deposits (largely subsurface) (Vidal & Moczydłowska 1995 mainly after Kumpulainen & Nystuen 1985).*

The Lake Vättern fault basin was initially an erosional depression where the lower unit of the Visingsö Group was deposited ca. 800 Ma ago (Månsson 1996). The initial stage of the Vättern faulting, resulting from a regional extension, took place ca. 750 Ma ago syntectonically with the deposition of the middle unit of the Visingsö Group.

In the Bothnian Bay, reactivation of the NW-SE faults (Senja-Oulunjoki Tectonic Zone) is interpreted to have taken place ca. 650 - 600 Ma ago, related to the rifting and opening of the Iapetus Ocean between Baltica and Laurentia, particularly affecting the development Muhos basin (Wannäs 1989). The development of the Muhos basin is interpreted to have been controlled by sinistral strike-slip movements along these faults. According to Wannäs (1989) the Muhos basin is comparable with the strike-slip basin of the Gulf of Eilat in the Red Sea.

The further development of the Bothnian Sea basin, which, according to Korja & Heikkinen (1995) initially originated from Mesoproterozoic rifting associated with the emplacement of the rapakivi batholiths, is suggested by van Balen & Heeremans (1998) to be related to Neoproterozoic development of intracratonic basins. According to van Balen & Heeremans (op. cit.), the cause of the basin development is asthenospheric upwelling.

4.2.2 Break-up up Rodinia; rifting and associated magmatism

Palaeomagnetic data indicates that Baltica and Laurentia were together until at least 620 - 630 Ma (Torsvik et al. 1996). Basaltic dyke swarms and sheeted dykes related to the rifting of the two cratons are common and locally voluminous in the north-western margin of Baltica, and are preserved in the Caledonian nappes of Middle and Upper Allochtons (Andréasson 1994; Svenningsen 1994; Andréasson et al. 1998) (Fig. 4-8). Normal faults, pull-apart structures and folds record the extensional deformation prior the emplacement of the dykes (Svenningsen 1995). The rift magmatism was most intense and voluminous during a period of 30 - 40 Ma and broadly coeval with rift magmatism along the Appalachian margin (615 - 550 Ma) (Andréasson et al. 1998). The Egersund basaltic dyke swarm in south-western Norway (Fig. 4-8), dated at 616 Ma, has been interpreted as relating to rifting along the western margin of Baltica that resulted in opening of Iapetus Ocean between Baltica and Laurentia (Bingen et al. 1998). The Sarek dyke swarm in Sweden (Fig. 4-8), dated at 608 Ma, is inferred to date the onset of seafloor spreading (transition from rift to drift) in the Iapetus Ocean (Svenningsen 1994, 2001).

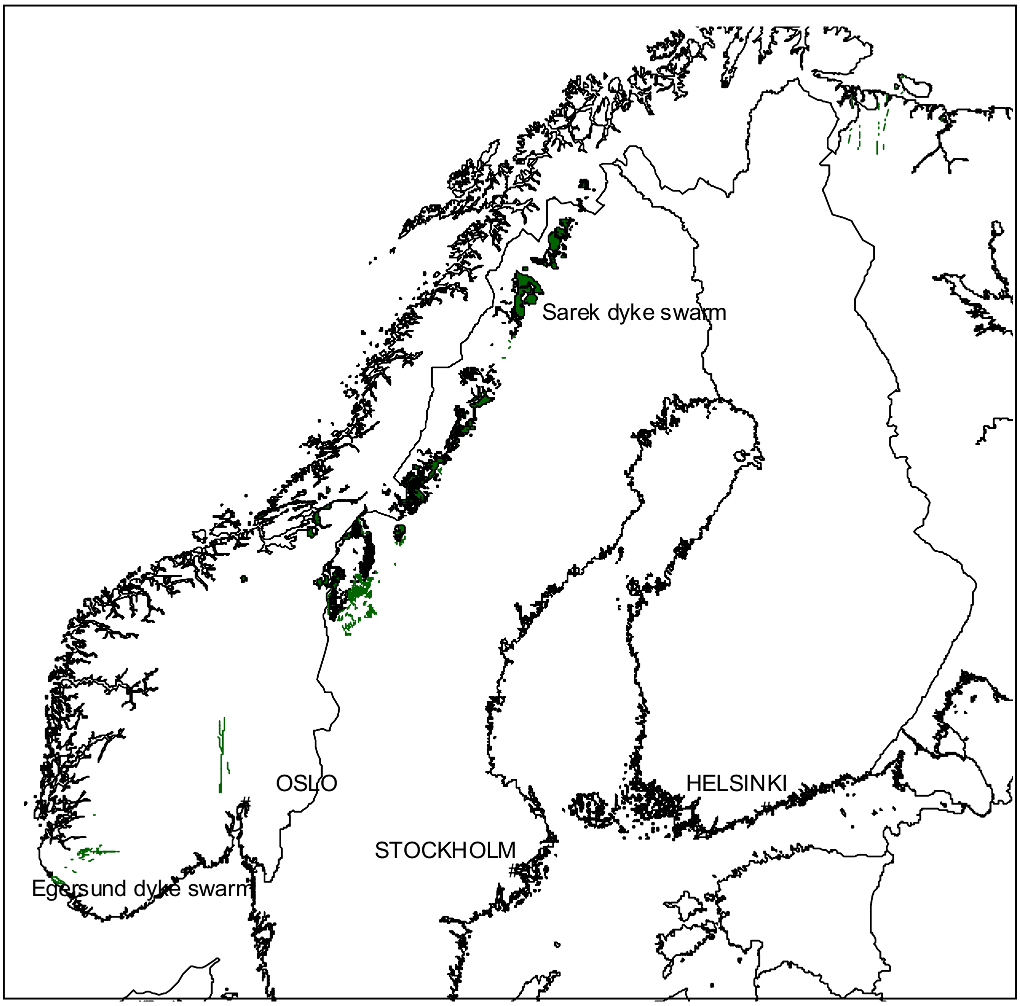


Figure 4-8. *Neoproterozoic diabases and metadiabases including sheeted dyke complex, amphibolite, gabbro, eclogite and ultramafic rock. Compiled from Koistinen et al. (2001).*

Also in the south-western margin of Baltica, volcanism was intense, including the 650 - 570 Ma old Volyn flood basalts in the south-western Tornquist margin of Baltica (Nikishin et al. 1996). The Volyn basalts are interpreted to evidence the final phase of rifting leading to the opening of the Tornquist Sea, an arm of the Iapetus Ocean, separating Baltica from Gondwana craton. The development of the Baltic Depression, i.e. the NE-SW trending central and eastern part of the Baltic Basin, is related to this tectonic evolution of the south-western margin (Flodén 1980; Nikishin et al. 1996).

Extensional tectonic regime after the initial opening of the Iapetus Ocean is also manifested by alkaline magmatism, represented by the Alnön complex in the western shore of the Bothnian Sea in Sweden and Fen complex in southern Norway (Meert et al. 1998), both dated at ca. 580 Ma. The Alnön complex is located at the intersections of two major NW-

SE lineaments on the western shore of the Bothnian Sea (Axberg 1980), being probably remnants of an aborted rift (Larson & Tullborg 1994). Bergström & Gee (1985) regard it possible that these complexes are distal features of rift magmatism related to the opening of the Iapetus Ocean, and note that a line connecting Fen and Alnön coincides with an axis of inferred Cambrian uplift.

In the Kaavi and Kuopio areas in Finland, the emplacement of the kimberlites has been dated at ca. 600 Ma (Peltonen et al. 1999) but its relationship with the break-up of Rodinia is not known.

4.3 Cambrian-Silurian

4.3.1 Palaeomagnetic constraints

Resulting from the break-up of the supercontinent Rodinia during late Precambrian to early Cambrian, possibly up to 5000 km wide Iapetus Ocean separated Baltica from Laurentia, while Tornquist Sea was located between Baltica and Gondwana. Baltica, lying at southerly latitudes (ca. 30-60°), was 180° geographically inverted, present-day northern Baltica facing the NW margin of Gondwana, which covered the South Pole (Torsvik & Rehnström 2001) (Fig. 4-9). The present-day western margin of Baltica (the Scandinavian Caledonides) was facing Siberia during the Late Cambrian to Early Ordovician with oceanic separation (Ægir Sea by Torsvik & Rehnström 2001) in the order of 1200-1500 km (Torsvik et al. 1996).

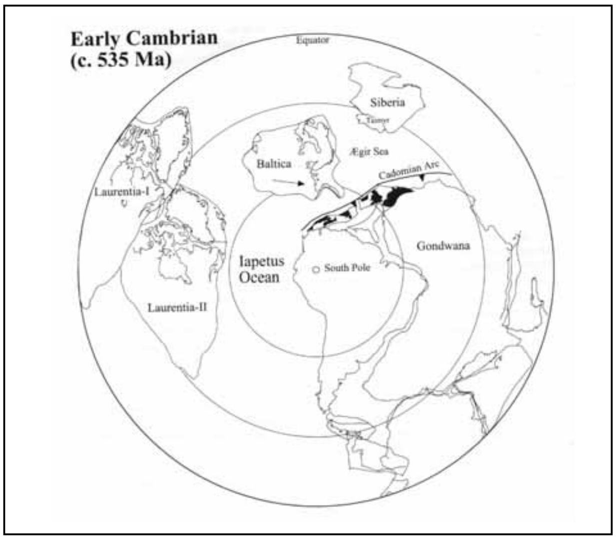


Figure 4-9. *Early Cambrian reconstruction of the cratons based on palaeomagnetic data (Torsvik & Rehnström 2001). Laurentia I and Laurentia II refer to two possible locations based on alternative palaeomagnetic poles.*

4.3.2 Caledonian orogeny

By Late Cambrian times, the plate convergence started again. This ultimately led to the collision of north-western Baltica margin with an assumed island-arc complex, resulting in the early stage of the Caledonian orogeny, referred to as the Finnmarkian orogeny, in Late Cambrian-Early Ordovician (Stephens 1988, Stephens & Gee 1989). The Finnmarkian event involves deep (at least 50 km) subduction northward and formation of subduction-related eclogites, followed by post-metamorphic uplift and obduction of Early Ordovician ophiolites (Andréasson 1994; Andrésson & Albrecht 1995; Sturt & Ramsay 1999). According to palaeomagnetic studies, Baltica underwent 55° counter-clockwise rotation between ca. 500 and 478 Ma, which Torsvik & Rehnström (2001) link with the Finnmarkian orogeny.

By Late Ordovician-Early Silurian (ca. 440 Ma) the Tornquist Sea between Baltica and micro-continent Avalonia (including the present-day southern British Isles), derived from Gondwana, closed (Torsvik et al. 1993). The convergence of eastern Avalonia to Baltica led to collision and development of the North German-Polish Caledonides (Katzung et al. 1993). South-dipping reflections in the basement close to the southern Horn Graben in the North Sea, is suggested to bear evidence for thrusting of Avalonian crust onto Baltica crust (Vejbæk 1990).

Due to counter-clockwise rotation of Baltica, the Caledonian margin was facing Laurentia instead of Siberia by Silurian time (Torsvik et al. 1996). Baltica, together with Avalonia, approached Laurentia, resulting in narrowing of the Iapetus Ocean. On the basis of faunal evidence, Iapetus Ocean had essentially closed by Late Ordovician, ca. 450 Ma (Stephens & Gee 1989). At ca. 425 Ma Baltica and Avalonia, collectively known as Balonia, collided with Laurentia, which resulted in closure the Iapetus Ocean and the main stage of the Caledonian orogeny, referred to as the Scandian orogeny (Fig. 4-10). During the continent-continent collision, Baltica subducted deep beneath Laurentia, giving rise to associated eastward thrusting over the western Baltoscandian margin. Several terranes of exotic origin relative to Baltica were subsequently attached to Baltica and are now present as the nappes of the Scandinavian Caledonides (Stephens 1988). The crustal shortening during the Scandian event is estimated to be at least 400 km (Gee 1975; Stephens 1988). During the collision, Laurentia was nearly stationary at the equator, while Baltica moved rapidly (8 - 10 cm/year) northwards (Torsvik et al. 1996; Torsvik 1998) (Fig. 4-10). As a result of the collision of the two continent, a new "supercontinent" Laurussia was formed. Ziegler (1990) states that there may have been up to 1500 km of dextral strike-slip shear movement involved in the collision.

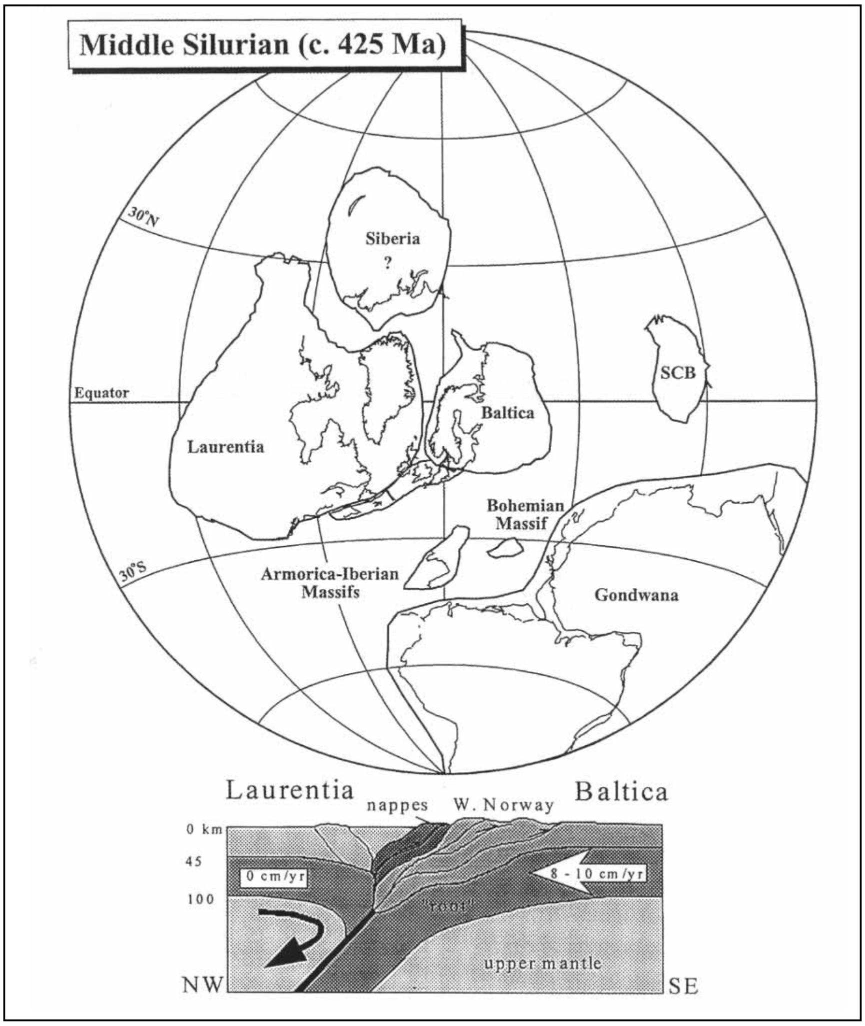


Figure 4-10. *Mid-Silurian (ca. 425 Ma) palaeogeographic reconstructions and the schematic cross-section of the collision between Baltica and Laurentia (Torsvik et al. 1996).*

4.3.3 Intracratonic tectonics

The clastic dykes found in various places in Finland and Sweden (Fig. 4-11) are considered to mark the intracratonic tectonic activity under extensional regime during the Cambrian. Bergman (1982) has studied the fractures filled with Lower Cambrian sandstone in the rapakivi area in the Åland Islands. The older Lower Cambrian dykes are mainly vertical and dominantly strike 0- 20°, while the younger Lower Cambrian dykes strike 40 - 160°. Also Upper Cambrian dykes occur (Martinsson 1968).

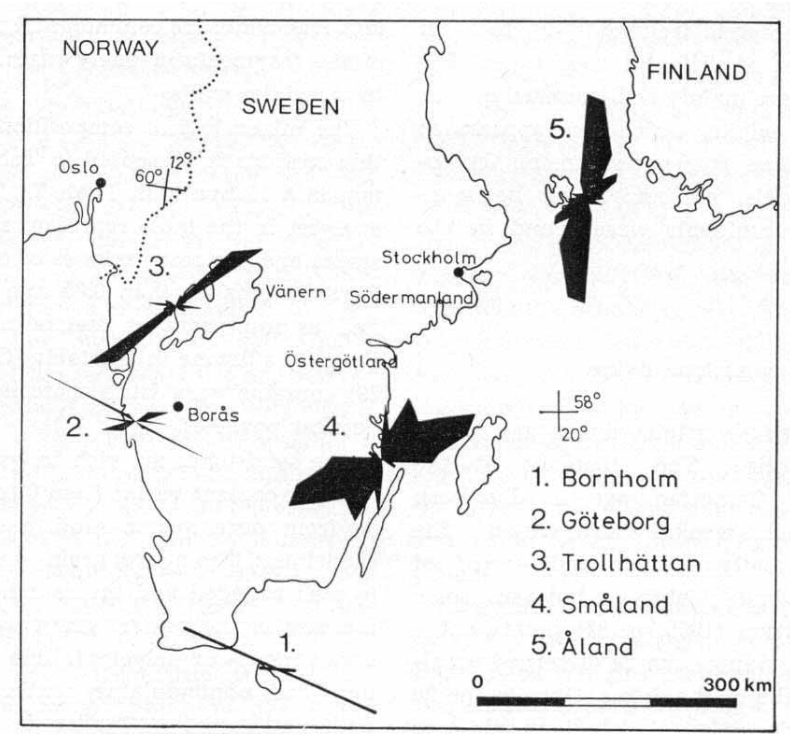


Figure 4-11. *Orientation of clastic dykes in Åland and southern Sweden (Bergman 1982).*

Bergman (1982) explains the clastic dykes to have been formed by injection of clastic material from above. This mechanism presupposes that the fractures have opened after the sedimentation. The sedimentation in the Åland area started during the early Lower Cambrian and at the same time tectonic activity opened fractures. The lack of Middle Cambrian dykes indicate that the main tectonic events took place in Lower and Upper Cambrian, since the sedimentation at the area continued also during Middle Cambrian (cf. Hagenfeldt 1989b). Weathered rapakivi granite found in the drillings in Lumparn (*see* Chapter 2.2.4), indicates that the clastic dykes occurring today in the unweathered rapakivi also penetrated the weathered crust and hence were originally very deep (Bergman 1982).

According to Poprawa et al. (1999), the sandstone-filled fractures may be related to the same extensional processes that resulted in development of the Baltic Basin. On the basis of subsidence analysis, the Baltic Basin originated during late Vendian to Middle Cambrian times as a result of rift-related extensional processes occurring along the south-western margin of Baltica (Teisseyre-Tornquist Zone). In Cambrian and Ordovician times, the Baltic Basin formed a large epicontinental sea.

Bergström & Gee (1985) suggest that various episodes of transgressions and regressions in the Lower Cambrian sandstones and the Middle and Upper Cambrian shales may be related to local crustal instability. However, at least some of these transgression and regression may be of eustatic origin, perhaps related to periods of faster and slower sea-floor spreading of the Iapetus Ocean (Bergström & Gee, op. cit.). According to Hagenfeldt (1994, 1997), development of the subduction zone along the western margin of the Fennoscandian Shield resulted in subsidence at the margin of the Shield in Middle Cambrian, while doming took place in the Central Baltic and in the Bay of Bothnia.

According to Artyuskov et al. (2000), the Early to Late Cambrian transgressions in southern Sweden and the Baltic area due to regional tectonic movements rather than eustatic sea-level changes. Based on relative stability of global sea level in the late Early Cambrian to Early Ordovician and using relative sea-level curves and sediment thickness, the authors argue that vertical crustal movements with an amplitude up to several hundreds of metres may have taken place in the East Baltic and southern Sweden in the Cambrian and early Ordovician. According to Artyuskov et al. (op. cit.) these tectonic movements are not related to foredeep subsidence of the Caledonides but they can be caused by changes in the forces in the lithospheric layer with a laterally variable thickness, and by phase transitions in the mafic lower crust.

Towards the Upper Cambrian and early Lower Ordovician the environments became more stable within the Fennoscandian Shield. The thin deposits of the Middle Cambrian to Lower Ordovician (Tremadoc) alum shales represent long and very slow deposition, indicating great stability of the bedrock (Andersson et al. 1985; Bergström & Gee 1985). Based on the distribution of Lower Ordovician sedimentary rocks, Hagenfeldt (1997) suggest that tectonic activity increased in transition from early Lower Ordovician Tremadoc to late Lower Ordovician Arenigian time due to convergence between the Laurentia and Baltica cratons. In Estonia, however, hardly any indications of structurally induced landforms are present in the Ordovician and Silurian sedimentary sequences before the Caledonian compressional tectonics in the north-western and southern margins of the Baltica craton (Puura et al. 1999).

The extent of influence of the Scandinavian Caledonides east of the orogen is poorly understood and speculative. This influence may have been rather modest, since the thick crust of the Fennoscandian Shield may have prevented the propagation of movements to the inner parts of the shield (cf. Muir Wood 1995). A forebulge axis migrating with time and developed in the periphery of the Caledonian foreland basin by flexing of the lithosphere, may have been located along the Gulf of Bothnia during the late Silurian-Early Devonian (Fig. 4-12) (Tullborg et al. 1995). In the Bothnian Sea Basin, the Cambrian to Ordovician sedimentary sequences are rather undisturbed, however, block faulting and associated flexural folding can be noticed in some of the reflection profiles (Fig. 4-13) (Winterhalter 1972; Axberg 1980). Three major tectonic zones are found both in the Bothnian Sea and onshore; two NW-SE trending zones and one, the Bothnian Zone, extending northward along the Swedish coast to the northernmost part of the Gulf of Bothnia (Axberg 1980). The Bothnian zone is the most prominent tectonic lineament in the Bothnian Sea, consisting of several lineaments parallel to the Swedish coast. Considerable vertical displacements have happened along the zone; it separates the onland crystalline bedrock from the offshore sequences of sedimentary rocks by at least 1200 - 1300 m. Offshore on the Swedish coast, northwards from Gävle, two NE-SW trending lineaments separate the 'Jotnian' Gävle sandstone and the overlying Palaeozoic sequence from the crystalline basement, the east side of the lineaments being repeatedly downfaulted (Axberg 1980; Flodén 1980).

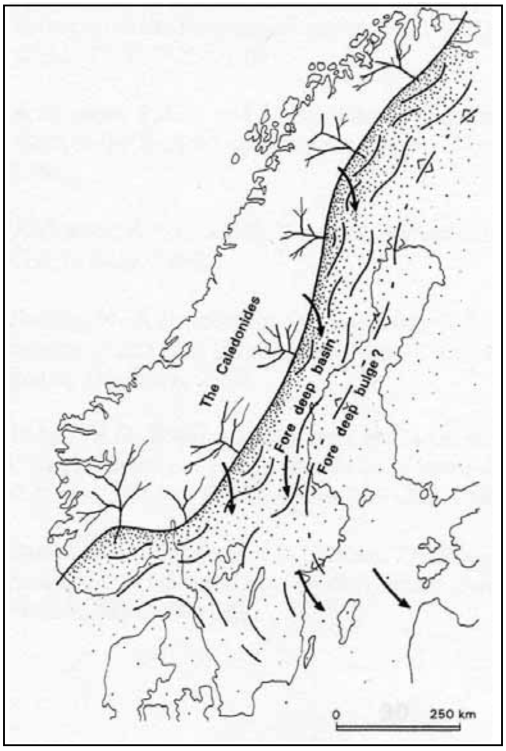


Figure 4-12. *Tentative sketch of a Caledonian foredeep basin and foredeep bulge (Tullborg et al. 1995).*

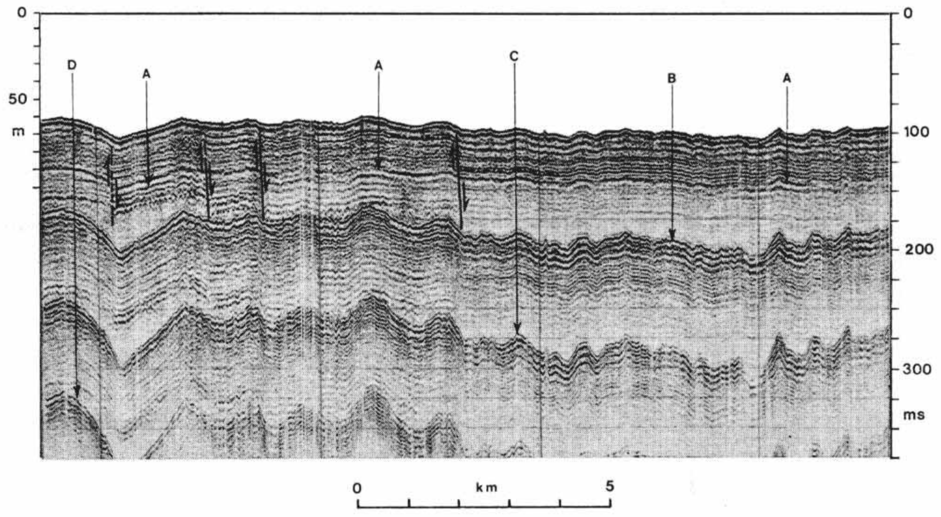


Figure 4-13. *Part of the SE-NW reflection profile east of the Finngrundet Banks showing strong block faulting. The total vertical displacement of the blocks is on the order of 100 metres (Winterhalter 1972).*

The block faulting in the Bothnian Sea basin is probably due to post-depositional (i.e. post-Ordovician) tectonic movements of the underlying crystalline basement (Winterhalter 1972). The lineaments, along which the faulting has happened are, however, originally Precambrian in age. The re-activation of the faults may be related to the Caledonian orogeny in Silurian time or, as suggested by Puura et al. (1996), the faulting occurred in an extensional deformation field coeval with the Permian Oslo Rift (*see* p. 88).

In the Bothnian Bay, downfaulting of the Cambrian sequence, which caused it to be protected from erosion, was most likely related to movements in the Caledonian orogen (Wannäs 1989).

In the Åland Sea and the Stockholm archipelago, only remnants of Cambrian sequences are preserved on the sub-Cambrian peneplain. According to Söderberg (1993) and Söderberg & Hagenfeldt (1995), this may be due to erosion during the latest Cambrian, or earliest Ordovician (Tremadoc), which was probably caused by regional tectonics, such as the Finnmarkian event of the Caledonian orogeny in northern Norway. There is only minor evidence for syn-depositional tectonic movements in the Åland Sea area (Söderberg & Hagenfeldt 1995).

Axberg (1980) reports a seismic structure in the western part of the Bothnian Sea, which he interpreted as diabase dyke cutting the Ordovician strata. Flodén (1980) has reported similar features near Gotland in the Baltic Sea. In the Åland Sea there are interpreted diabase dykes, which seem to cut the Upper Riphean to Vendian rocks, i.e. they are younger than ca. 550 Ma (Söderberg 1993). Whether these dykes are associated with the Caledonian orogeny is not known.

In the Baltic Basin, the late Middle Cambrian to Middle Ordovician development is governed by thermal subsidence (Poprawa et al. 1999). A convergent tectonic regime (the Caledonian orogeny) began to dominate from the late Ordovician, and differential tectonic movements and minor uplifts occurred in the eastern part of the basin. Silurian subsidence with especially high tectonic subsidence and sedimentation rates is interpreted by Poprawa et al. (op. cit.) to be related to the flexural bending of Baltica craton in front of the North German-Polish Caledonides in the south. Also the Scandinavian Caledonides may have had an influence to the development of the Baltic Basin, particularly in the development of the central part of the basin, the Baltic Depression, during Middle-Late Silurian.

Majority of the linear features in Estonia, Latvia and St. Petersburg region in Russia are considered to be from the Silurian-Devonian transition, and related to late Caledonian compressional movements in north-western Scandinavia and along the Tornquist Zone in the south and southwest (Puura et al. 1996). Due to the Caledonian compressional tectonics the area was uplifted, slightly tilted (ca. 0.1° in a SSE direction) and gently faulted (Puura et al. 1999). In the Estonian-Latvian border, 20 - 30 km wide and 200 km long Valmiera-Lokno Uplift, bordered by deep-seated faults, developed in the Late Silurian to Early Devonian resulting in erosion of Lower Palaeozoic rocks (Puura & Vaher 1997). Within the fault zones, there are evidences of fluid migration, including hydrothermal metasomatic dolomitisation of limestones, lead and zink mineralisation, and karst formation. The regional dolomitisation and ore mineralisations could be connected with the Permian event that produced the Oslo rift system (Puura et al. 1996). A number of Palaeozoic faults and other dislocations within the sedimentary cover south of the Gulf of Finland represent reactivation Precambrian structures. However, in many occasions, the faults cutting the boundary between the basement and the Phanerozoic sedimentary cover have no relation to structures in the basement (Puura et al. 1996).

4.4 Devonian-Permian

The Silurian continent-continent collision was followed in the Devonian by uplift and denudation of the Scandinavian Caledonides. Probably up to 4 km of sediments ("Old Red Sandstone") deposited on the Caledonian foreland basin east of the orogen (*see* Chapter 3.4). The extensional collapse of the orogeny caused the uplift of the foreland basin, and the subsequent erosion of the basin fill started during the mid-Devonian. In southern Scandinavia, this sedimentary cover was significantly reduced by the Carboniferous to Jurassic uplift and erosion (Tullborg et al. 1995; Tullborg 1997; Cederbom 2001).

4.4.1 Post-Caledonian extension and magmatism

The Devonian, post-Caledonian tectonics are characterized, especially in western and southern Norway, by major extensional detachments and extensional reactivation of major shear zones (Fossen & Rykkelid 1992; Hurich 1996; Milnes et al. 1997; Andersen 1998; Fossen & Dunlap 1998). Fossen & Rykkelid (1992) demonstrated that the intense post-collisional extension included the development or reactivation of major west-dipping shear zones, and 20 - 30 km backsliding of the nappes toward the central part of the Caledonides. The latter caused an additional uplift of 30 - 35 km of the Caledonian eclogites, formed at depths of 60 - 100 km or more. Rapid change from convergent to divergent plate movement between 408 - 402 Ma is suggested to be the cause for the extensional deformation (Fossen & Rykkelid 1992, Fossen & Dunlap 1998). Displacement along the shear zones varies from a few kilometres or tens of kilometres to more than 100 km (Fossen & Dunlap 1998 and references therein). Uplift along the fault zones resulted in erosion of the footwall rocks and the development of sedimentary basins along the hanging wall (cf. Norton 1987, Séranne 1992). In the eastern Caledonides, the Proterozoic basement was only little affected by extension (Andersen 1998).

In north-eastern part of the Fennoscandian Shield, large number Palaeozoic alkaline eruptive centres and basic dykes occur in the area between north-eastern Finland and eastern part of the Kola Peninsula in Russia. In north-eastern Finland, the Sokli carbonatite complex and Iivaara complex (*see* Fig. 2.1-1) belong to this Kola Alkaline Province. The intrusive rocks of the Province have been dated at 380 - 360 Ma, i.e. late Middle Devonian to Upper Devonian (Kramm et al. 1993). The alkaline complexes have temporal and spatial relationships with the Kontozero Graben of the Central Kola Peninsula, the formation of which coincides with the extensive Middle Devonian rifting in the eastern part of the East European Craton (*see* Nikishin et al. 1996).

4.4.2 Permian rifting and volcanism

During the Late Palaeozoic, the continents were gradually assembled into the supercontinent Pangaea. The final completion of Pangaea took place at Late Carboniferous when the northern margin of Gondwana collided with the combination of Laurentia and Baltica, known as Laurussia (Tait et al. 2000), resulting in Variscan orogeny in Europe. At the end of the Palaeozoic also Siberia collided with Laurussia, forming the Ural Mountains. By the Permian, all of the continents had collided and joined to form the supercontinent Pangaea**.** An ocean called Panthalassa surrounded Pangaea. Tethys Sea was located between Africa and Europe.

The Variscan fold belt originated from the collision, in Devonian and Carboniferous times, of Laurussia and Gondwana, a number of microplates (Avalonia, Armorica, and possibly others) being caught between the larger plates (Franke 1996). During the late Variscan orogeny, rifting and volcanism occurred throughout north-western Europe. The tectonics related to the late Variscan orogeny resulted in a dextral transpressional and/or transtensional strike-slip faulting along the NW-SE trending Tornquist Zone in the south-western margin of the Fennoscandian Shield (Fig. 4-15), and the early Palaeozoic sedimentary rocks were downfaulted as tilted blocks (Ziegler 1988; Ro et al. 1990, Mogensen 1994). The faulting was associated with volcanic activity and intrusion of dyke swarms (Fig. 4-14), the ages of which range from 300 to 240 Ma (Klingspor 1976). Diabase sills of similar age occur in southern Sweden northeast of the Tornquist Zone (Gorbatschev et al. 1987; Obst & Solyom 2000). The Tornquist Zone may originally have been a Precambrian block boundary, which has reactivated several times since the late Palaeozoic and acted as a "buffer zone" during the changes in the regional stress field (Mogensen 1994).

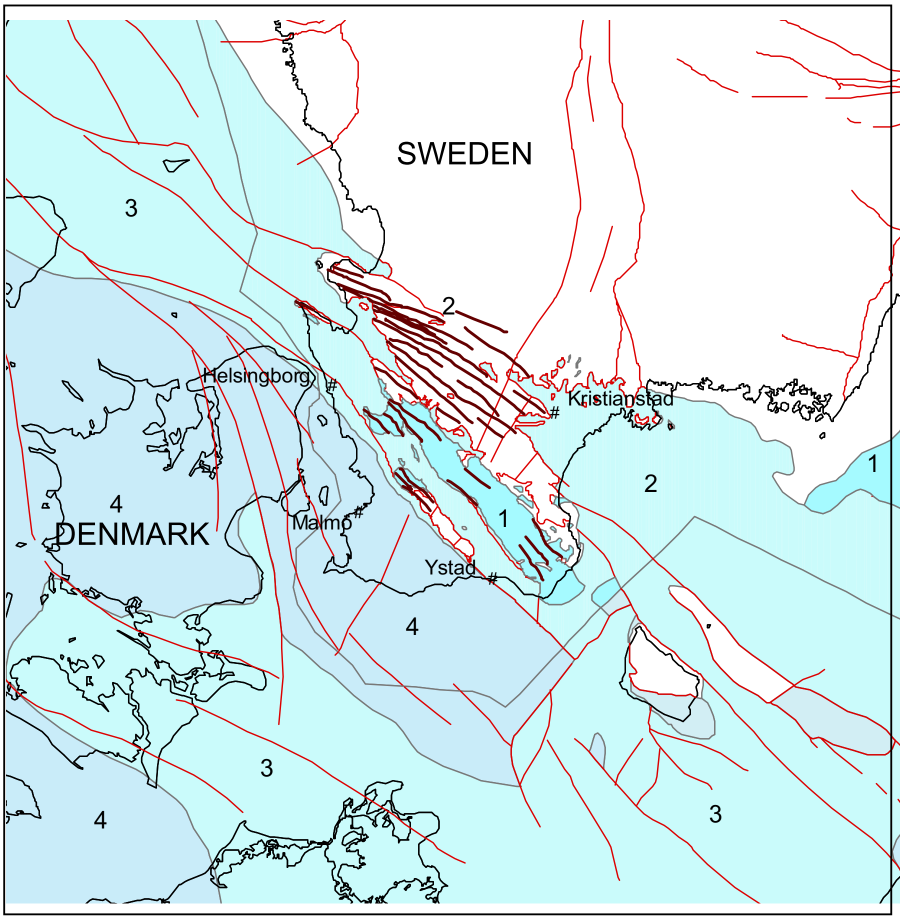


Figure 4-14. *Permo-Carboniferous diabase dykes in southernmost Sweden (compiled from Koistinen et al. 2001). 1. Middle Cambrian to Permian sedimentary rocks, 2. Permo-Carboniferous diabase dykes, 3. Mesozoic sedimentary rocks, 4. Cenozoic sedimentary rocks, red lines = faults or ductile deformation zones.*

The Permian rifting and volcanism is demonstrated by ca. 400 km long Oslo Rift (incl. the Oslo Graben on land and its offshore continuation the Skagerrak Graben) (Fig. 4-15) where mafic to silicic extrusive and intrusive rocks formed in several phases during the time period of 305 - 240 Ma (Sundvoll et.al. 1990; Ro et al. 1990; Neuman et al. 1992). The Oslo Rift is characterized by major N-S and NNE-SSW to NE-SW striking faults, which have been formed in several phases under both ductile and brittle tectonic regimes (Swensson 1990 and references therein, Neuman et al. 1992). Both the crust and lithosphere under the rift are thinned relative to the Precambrian basement on both sides of the rift.

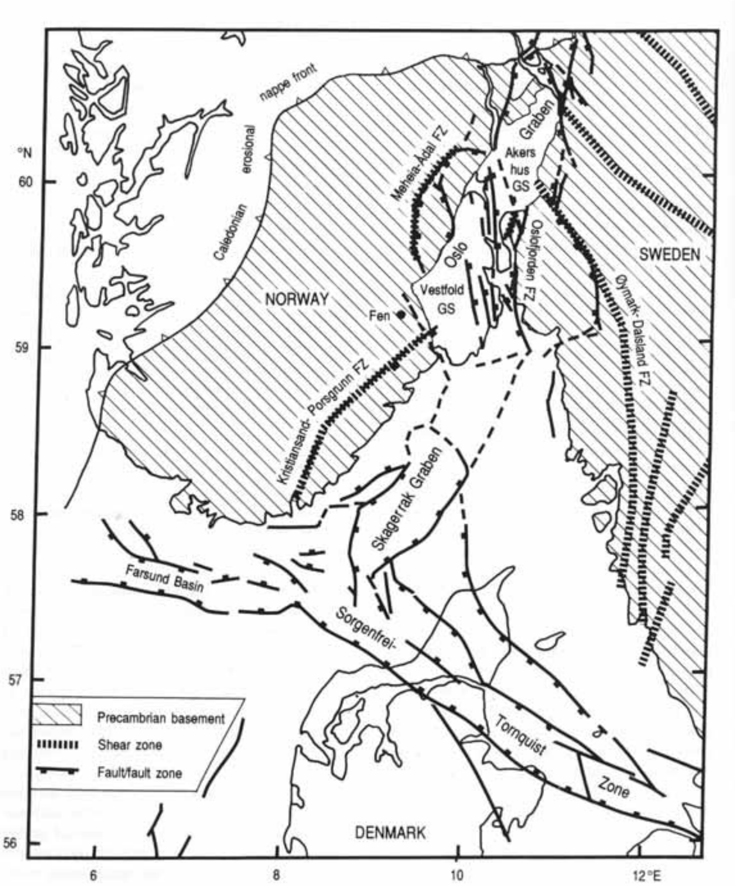


Figure 4-15. *Regional structural map of the Oslo Rift and adjacent areas (Neuman et al. 1992). FZ = fracture zone, GS = graben segment.*

The main faulting and graben formation took place at 295 - 275 Ma, with an extrusion of basaltic and intermediate, plateau-type lavas propagating from south to north (Sundvoll et al. 1990, Neuman et al. 1992). Block faulting was intense during this period. The total vertical displacement along the Oslofjorden fault zone is estimated to be about 3 km (Neuman et al. 1992). Precambrian shear zones occurring east and west of the rift were reactivated in several occasions during the Oslo Rift event (Ramberg & Larsen 1978;

Swensson 1990; Sundvoll et al. 1990; Ro et al. 1990). The rift itself is considered to have had a Sveconorwegian precursor (a zone of weakness), which has controlled the later rifting (Ro et al. 1990; Swensson 1990).

The formation of the Oslo Rift has been suggested to be caused by dextral wrench movements along the Tornquist Zone in response to the Variscan orogeny (Pegrum 1984; Ziegler 1988; Ro et al. 1990). According to Ziegler (1988) rifts and grabens in the North Sea are contemporaneous with the initial phase of the post-Variscan break-up of Pangaea supercontinent.

In western and central south Norway, fault breccias were formed and older faults reactivated in the late Permian as a result of extension (Eide et al. 1997; Andersen et al. 1999). In western Norway, this extension was associated with basaltic to alkaline magmatism (Andersen et al., op. cit.).

In the Vendian Vättern Graben, the main phase of faulting took place during the Permian, resulting in the formation of more than 100 m deep trench, which constitute the present day lake (Månsson 1996). The faulting is characterized by hydrothermal veining, coatings of chlorite, epidote and hematite, and occurrence of Mn-mineralizations in the fault zones. Månsson (op. cit) have tentatively proposed that the Vättern Graben was reactivated by a splay from the Tornquist Zone, as an analogue to the Oslo Rift.

The diabase sills in the Västergötland region in south-western Sweden are suggested to be early Permian (290 Ma) (Priem et al. 1968). The Särna alkaline complex in the Dalarna in Sweden, about 150 km NNE from the northern termination of the Oslo Graben (Lake Mjøsa), has an age of 282±14 Ma (Sundvoll et al. 1990) and may be related to the Oslo Graben volcanism.

According to Ahlin (1987) post-Silurian to pre-Triassic faulting with N-S, NE-SW, NNE-SSW and E-W directions is frequent in Västergötland. The faults have a dip-slip component, the displacement rarely exceeding 70 m. Ahlin (op. cit) has suggested that the faults have developed under extensional tectonic regime, many of the faults being probably located at previously existing fracture zones in the bedrock. Muir Wood (1995) considers, however, that this faulting is mainly mid-Devonian in age and related to Caledonian orogeny.

In Småland, south-eastern Sweden, the block faulting is regarded as post-Silurian in age (Milnes et al. 1998). The faults are normal faults of multiple orientations, suggesting extension in various directions. Slight tilting of the Småland mega-block and the faulting along its NW margin is suggested by Milnes et al. (op. cit.) to be probably late Carboniferous or younger in age.

Close to the Gotland coast and further towards the SE in the Baltic Sea, post-Silurian and post-Devonian tectonic movements have been postulated by Flodén (1980).

4.5 Triassic-Cretaceous

4.5.1 Breaking-up of the Pangaea supercontinent

At the beginning of the Mesozoic, all the continents were assembled into a supercontinent Pangaea, formed during the Palaeozoic. Major tectonic activity was concentrated to the break-up of Pangaea, which begun in the Triassic. During the Jurassic, the Tethys-Central Atlantic-Gulf of Mexico rift system developed into a principal fracture system along which the Central Atlantic opened by sea-floor spreading initiating the Laurasia-Gondwana separation during Middle Jurassic to Early Cretaceous (Ziegler 1990). Although rifting occurred in the Norwegian-Greenland Sea rift throughout the Jurassic and Cretaceous, the final crustal separation between Greenland and Europe took place only during the Early Tertiary.

4.5.2 Tectonics in the SW margin of the Fennoscandian Shield

There was hardly any deformation within the Fennoscandian Shield during the Mesozoic except for, probably, minor reactivation of the Oslo Graben (Muir Wood 1995). Instead, the south-western margin of the shield, the Tornquist Zone, was repeatedly reactivated during the Mesozoic rifting phases, which affected much of the northwest Europe (cf. Ziegler 1990). From Early Triassic to Early Cretaceous the stress regime in the Tornquist Zone was dominantly transtensional (Norling & Bergström 1987).

During Early Triassic, rifting and transtensional dextral movements within the Tornquist Zone occurred in the reactivated Early Permian faults, with the displacement of probably 3 - 4 km (Mogensen 1994, 1995; Erlström et al. 1997). During Jurassic to Early Cretaceous, the subsidence was primarily restricted to the area between two main faults in the Tornquist Zone, the Grenå-Helsingborg Fault and the Børglum Fault and minor dextral movements of the order of 1 - 2 km took place (Mogensen 1994).

The Middle Jurassic tectonic event in the Tornquist Zone, characterized by differential uplift of the blocks, was accompanied by basaltic volcanism along NW-SE trending faults and fracture zones (Norling & Bergström 1987). During Late Jurassic to Early Cretaceous, movements took place along pre-existing faults. In Scania, southern Sweden, the Late Triassic to Early Cretaceous tectonic events caused subsidence and uplift of the basement blocks (Wannäs & Flodén 1994).

Late Cretaceous to Early Tertiary transpressional, dextral strike-slip movements within the Tornquist Zone caused inversion of the zone (Norling & Bergström 1987; Mogensen 1994; Deek & Thomas 1995). Basin subsidence and uplift of the Tornquist Zone took place. Along the Børglum Fault a strike-slip displacement of the order of 1 - 2 km took place during the inversion (Mogensen 1994). In Scania, several basins and througs were formed and reactivated. The inversion probably took place as a response to the Alpine movements (Deeks & Thomas 1994).

Recently, Marek (2000) have recognised stratigraphic unconformity in the Sorgenfrei-Tornquist Zone, which contradicts the traditional opinion that the major part of the inversion took place in late Cretaceous-early Tertiary. The new seismic data point toward a Jurassic major inversion phase. The inversion took place along a few of the old extensional faults, which were rejuvenated.

In the late Jurassic to early Cretaceous times, extensional movements related to rifting of the Pangaea supercontinent caused fault breccias in central south Norway (Andersen et al. 1999). In Scania, southern Sweden, basaltic volcanics were erupted at Early and Late Cretaceous (Norling & Bergström 1987).

In the north-western part of southern Sweden, the apatite fission track ages exhibiting a wide range from Carboniferous to Jurassic are suggested to be due to pre-Cretaceous vertical block movements (Cederbom 2001). At the same area, sharp and straight linear structures occur, which Tirén & Beckholmen (1992) have interpreted to be the youngest structures in the southern Sweden.

4.6 Tertiary to Quaternary

4.6.1 Opening of the North Atlantic

In the Tertiary, the opening of the North Atlantic and the initiation of sea-floor spreading in the Norwegian-Greenland Sea, and the beginning of the Alpine continent-continent collision dominated the evolution of north-western Europe. The Late Cretaceous-Early Tertiary rifting phase that preceded the onset of sea-floor spreading in the Arctic-North Atlantic was accompanied by the Thulean bimodal alkaline volcanism in Greenland, Ireland and Scotland (plateau basalts, sills and dykes) (Ziegler 1990). The northernmost volcanics, dated at ca. 56 Ma, associated with this event occur off the western coast of Norway (Bugge et al 1980) and possibly also in a few dykes on the coast of southern Norway (Storetvedt 1968). The termination of the Thulean volcanism in the Early Eocene was followed by the onset of sea-floor spreading in the Iceland Sea and the Norwegian-Greenland Sea. During the Cenozoic, continued convergence between Eurasia and Africa gave rise to Palaeogene Alpine and Carpathian main orogenic phases.

Transform faults were created in the North Atlantic spreading ridge. According to Talbot & Slunga (1989), some of the transform faults, the Jan Mayen Transform Fault and the Iceland Transform Fault, may continue into zones of weakness on land. Furthermore, Talvitie (1979) suggests that the seismic activity along the Proterozoic Senja and Lapland fracture zone segments indicates the Mid-Atlantic Ridge push guided inside the Fennoscandian Shield along NW-trending strike-slip faults. Muir Wood (1995), however, does not believe that these transform faults, created in the North Atlantic, continue on land but argues that the rheological differences between the Fennoscandian Shield and its surroundings prevent such a continuation. According to Muir Wood (op. cit.), the Red Sea spreading ridge provide probably the most realistic present-day analogue of the impact of the North Atlantic spreading on tectonic deformation within the Fennoscandian Shield. In the Red Sea rift, the continental separation was initiated ca. 15 Ma ago. While there is some deformation around the rift, there is practically no seismicity nor active deformation inside the Arabian Shield, 300 - 500 km from the spreading ridge. Muir Wood (1995) assumes that a comparable situation applies to the Fennoscandian Shield.

4.6.2 Uplift of western Scandinavia

The latest phase of deformation mainly affecting the margins of the Fennoscandian Shield was the uplift of western Scandinavia (Jensen et al. 1992; Muir Wood 1995; Riis 1996; Stuevold & Eldholm 1996). The first phase of uplift, amounting close to 1500 m in northern Sweden (Fig. 4-16), occurred in the Palaeogene in connection with the opening of the North Atlantic and emplacement of mantle plume in Iceland. The second major episode of uplift occurred in Neogene starting from late Oligocene (Stuevold & Eldholm 1996), and had its centre in southern Norway (Fig 4-17). During Late Pliocene and Pleistocene, the tectonic uplift was amplified by isostatic rebound in response to the glaciation (Stuevold & Eldholm 1996), so that most of the Neogene uplift probably took place during that period of time (Riis 1996). In southern Scandinavian Caledonides, a Neogene uplift of about 1000 metres has been inferred (Riis 1996, Lidmar-Bergström & Näslund 2000). Resulting from uplift and increased erosion, thick sedimentary sequences were deposited on the shelf in the late Pliocene and Pleistocene. The Tertiary uplift of the Fennoscandian Shield does not include any significant faulting, suggesting a flexural uplift mechanism (Stuevold & Eldholm 1996).

In southern Sweden uplift during Late Cretaceous to Early Tertiary resulted in denudation of the surface to a peneplain (Larson & Tullborg 1994). Late Tertiary to recent vertical movements have affected this peneplain (Lidmar-Bergström 1991). In the south-eastern part of the Landsort Trench, north of Gotland, the sub-Cambrian peneplain is offset by about 100 m due to movements of an inferred Tertiary age (Flodén 1980). The tensional movements in the Baltic area may have been induced by the Tertiary uplift of the Scandinavian Caledonides (Flodén (op. cit.).

The formation of the Vänern depression in southern Sweden is interpreted to be dominantly Cenozoic, and there is evidence that this period of tectonic activity has not ceased; there are observations of Holocene faulting, causing displacements of Quaternary deposits (Ahlin 1987). The development of the tectonic basin of Lake Mjörn, near Lake Vänern, was probably similar to that of the Lake Vänern, the downfaulting occurring after the erosion of the Palaeozoic rocks (Ahlin 1980). Late Phanerozoic dip-slip faulting corresponds to an extensional tectonic regime, with an axis in NE-SW.

The occurrence of Tertiary clay bed 205 m a.s.l. in east-central Lapland (*see* Chapter 2.2.18) is suggested to be due to some 100 - 200 m uplift of the peneplain during the late Tertiary, possibly in connection with block movements (Mikkola 1932; cf. also Riis 1996).

Van Balen & Heeremans (1998) explain the morphological difference between the flat Finnish coast and the uneven Swedish coast of the Bothnian Sea to be due to Palaeocene-Neogene differential uplift of Fennoscandia. Based on the uplift analysis of Riis (1996), van Balen & Heeremans (op. cit.) suggest that the uplift of the Bothnian Sea area was about 500 m.

Within the Tornquist Zone, transtensional stress regime triggered by the Alpine orogeny continued from Late Cretaceous into Tertiary with dextral faulting and inversion movements (Norling & Bergström 1987). Regional uplift took place in the Neogene resulting in removal of much of the Late Cretaceous and pre-Neogene sedimentary cover in southern Sweden (Erlström et al. 1997). In the Skagerrak area in the North Sea, 500 - 1500 m of Mesozoic and Tertiary sediments were eroded following the Neogene uplift (Jensen & Schmied 1992). In the Skagerrak area, the uplift is interpreted to be due to broad-scale warping without major faulting.

Stuevold & Eldholm (1996) suggest a thermal origin for the Neogene uplift. Accordingly, the uplift would have caused by a small-scale convection and preferential volume expansion in the asthenosphere beneath the Caledonides-Fennoscandian Shield transition. Rohrman & van der Beek (1996) suggest a diapirism connected to the opening of the North Atlantic for the cause of the Neogene uplift.

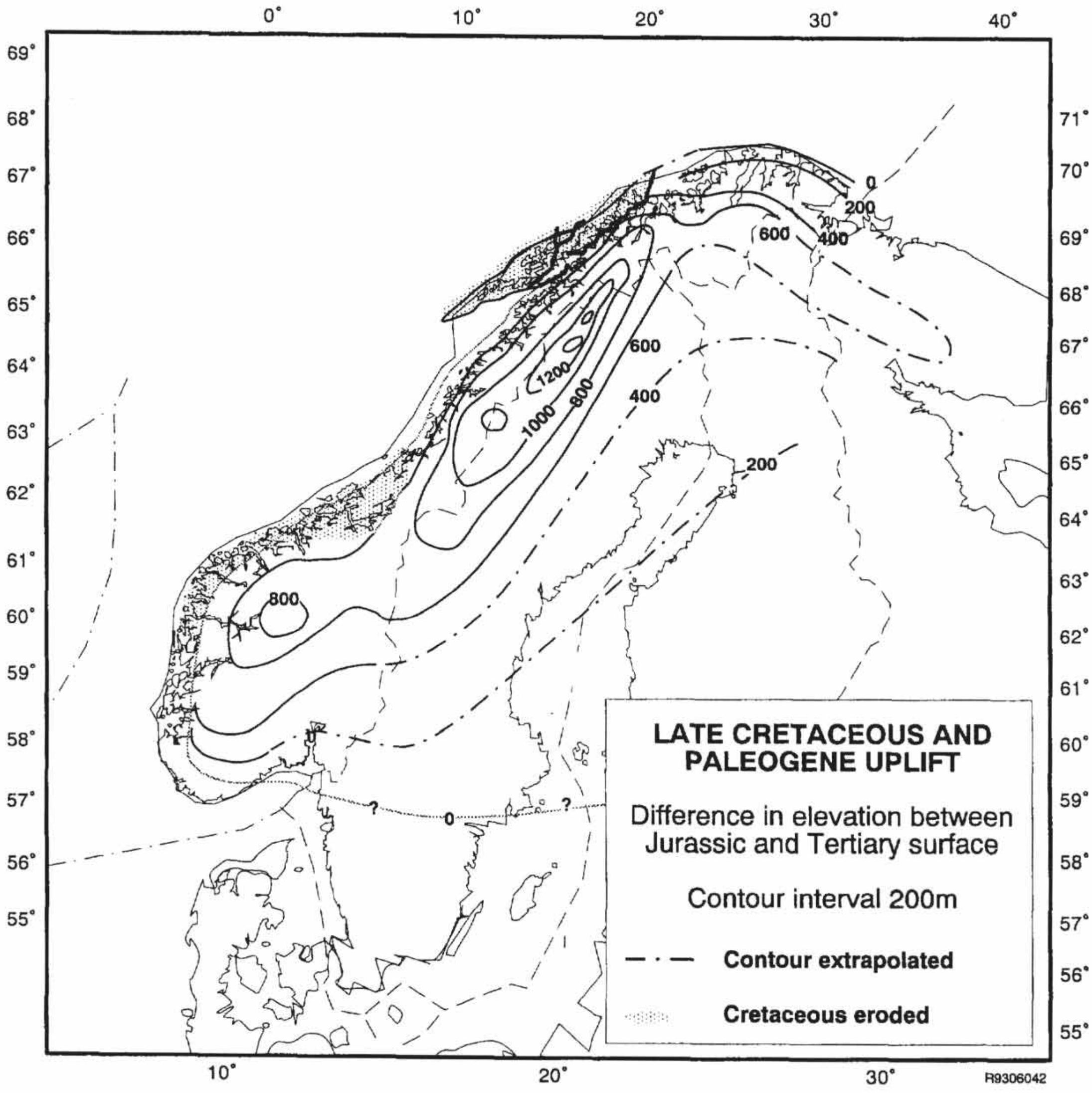


Figure 4-16. *The approximate amount of Late Cretaceous and Palaeogene uplift within and in the margin of the Fennoscandian Shield (Riis 1996).*

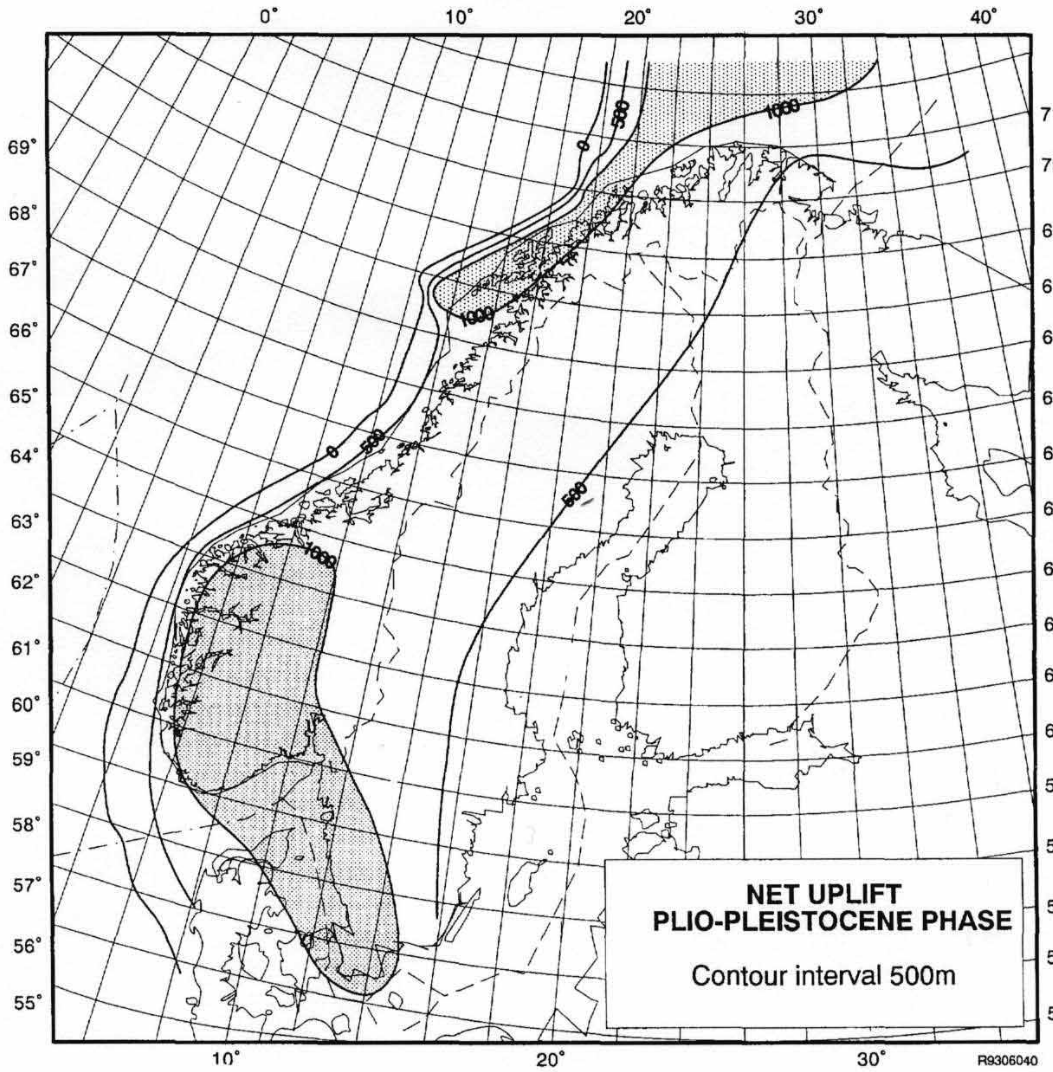


Figure 4-17. *The approximate amount of Neogene (mainly Pliocene and Pleistocene) uplift within and in the margin of the Fennoscandian Shield (Riis 1996).*

4.6.3 Present stress regime within the Fennoscandian Shield

Slunga (1991) has summarized the fault plane solutions of more than 200 Fennoscandian earthquakes. The strike-slip faulting is dominant at subvertical fault planes indicating that the dominant stress direction is horizontal. On the basis of Fennoscandian earthquakes, the orientation of maximum horizontal compression Fennoscandia is in the NW-SE direction Gregersen et al. (1991). The same orientation for the maximum horizontal stress can be seen in *in situ* stress measurements in Central Sweden and Central Finland at depths below

300 m (Stephansson et al. (1987). The same trend is also present in the horizontal strain pattern, based on geodetic observations in Finland (Chen 1991; Kakkuri & Chen 1992).

The orientation of the maximum horizontal stress in southern Sweden and in most of Finland is, however, a statistical result. In detail there are significant variations due to local bedrock structures. Fig. 4-18 shows the directions of the maximum horizontal stress within the Fennoscandian Shield based on the Fennoscandian Rock Stress Data Base. Compared to the orientation of maximum horizontal stress in Central Europe (rather constantly 135°), the stress directions in Fennoscandia are unevenly distributed. In Central Europe the stress field is considered to reflect the ongoing tectonic event, i.e. the Alpine orogeny, which is associated with high heat flow and thinning of the lithosphere. In the cold and stabile shield areas, the stresses field is irregular, due to irregular block movements (Clauss et al. 1989).

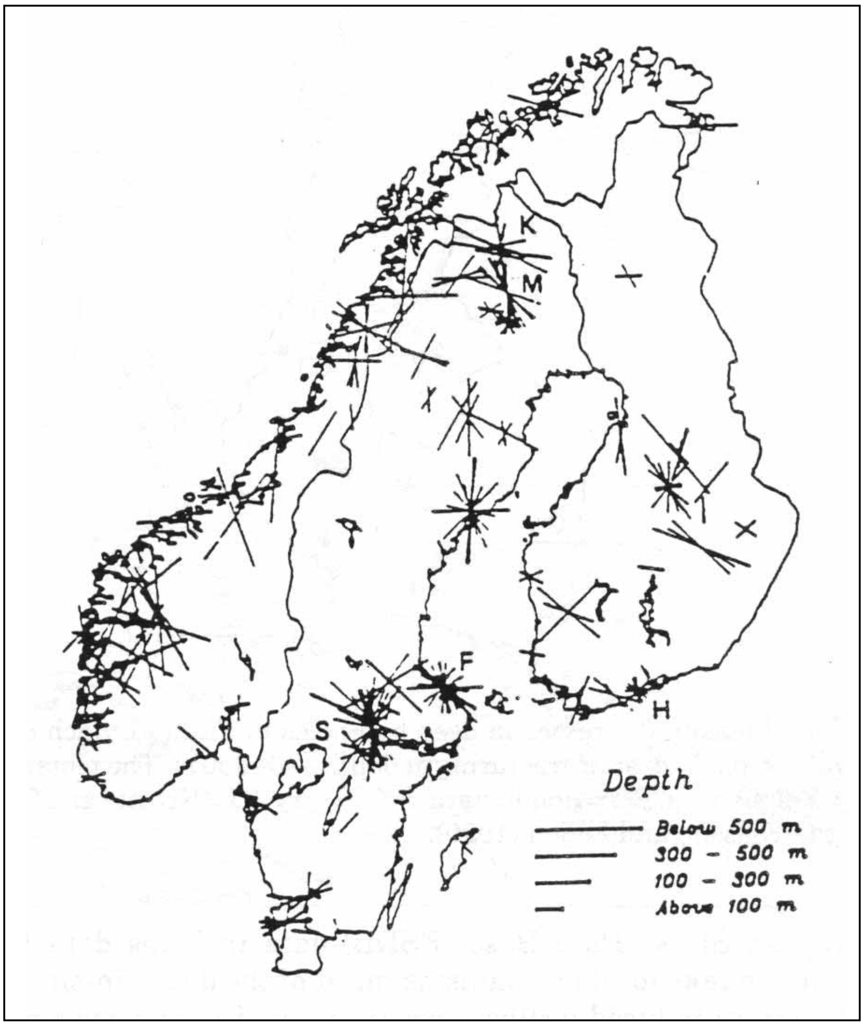


Figure 4-18. *Directions of the maximum horizontal stress within the Fennoscandian Shield at different depths (Stephansson et al. 1987).*

This orientation of maximum horizontal compression is in good accordance with the direction of ridge push forces from the Mid-Atlantic Ridge, which, accordingly, seems to be the major stress-generating mechanism. Skordas et al. (1991) have demonstrated that the seismic events in the Mid-Atlantic Ridge have a clear correlation with the Fennoscandian Shield. The best correlation comes from the events in the ridge between Jan Mayen and Iceland. However, the current stress field in Fennoscandia is likely a combination of plate boundary forces with local sources (e.g. glacial rebound and local geology) (Stephansson 1988; Saari 1992). According to Stephansson (1988) the stresses in the Fennoscandian Shield are mainly due to three mechanism: 1) residual stress left from the isostatic rebound after deglaciation, 2) concentration of the stress due to creep, and 3) stress caused by the spreading of the Mid-Atlantic Ridge.

4.6.4 Neotectonic movements

Neotectonic movements (postglacial and recent crustal movements) are clear evidences of young bedrock movements and offer therefore good possibilities for studying the behaviour of bedrock.

During the last Ice Age the mass of the Fennoscandian ice sheet made a depression to the Earth's crust as the slow viscosity material beneath the crust flowed aside. When the ice sheet melted, the crust began to rise back to its former level. The uplift was initially very rapid, but soon slowed down to its present level, with a maximum uplift of 9 mm per year in the northern part of the Gulf of Bothnia. The greatest uplift so far has occurred around the Vaasa – Umeå region, where it is estimated to be around 830 metres. The remaining potential for isostatic uplift around the current focus is believed to be 40-128 metres (Kakkuri 1985, Fjelskaar & Cathles 1991.

Large postglacial faults were first discovered in 1960s in northern Finland (Kujansuu 1964) and later all over the northern Fennoscandia (Lagerbäck 1979, Olesen 1984) (Fig. 4-19). In Finnish Lapland, the postglacial faults are 4-36 km long and the scarp height is 0-12 m. In Swedish Lapland up to 150 km long faults have been found and the maximum scarp height is 35 m. Small post-glacial faults located in ice polished bedrock outcrops and with a scarp height 0–20 cm have been found in southern Finland (Edelman 1949; Tynni 1965, 1966; Nenonen & Huhta 1993), but so far larger post-glacial faults have not been recognised.

Investigations on postglacial bedrock movements (Kukkonen & Kuivamäki 1985; Paananen 1987, 1989; Vuorela 1990; Veriö et al. 1993; Kuivamäki et al. 1998) have revealed that the postglacial faults studied so far are situated in old, reactivated fracture zones. The Finnish post-glacial faults, except for the Ruostejärvi fault, are reverse faults with the strike of SW–NE and dip to the SE. The drilling results at the Pasmajärvi postglacial fault have been interpreted to show a constant dip angle (47°) at least in the upper part (0–180 m) of the bedrock. Ruostejävi postglacial fault is a normal fault dipping to the SE, which together with the Venejärvi postglacial fault (reverse fault) borders an uplifted bedrock block. The strike direction of the postglacial faults in Finland is perpendicular to the direction of the prevailing maximum horizontal stress field (NW–SE) and evidently the same direction of stress field has existed for a longer time (Wahlström & Assinovskaya 1998).

Estimations of earthquake magnitudes connected with postglacial faults in the Finnish Lapland have varied from 5.3 to 7.5 (Kuivamäki et al. 1998). Those magnitudes are much greater than the ones of recent earthquakes in the same area. The epicentres of the recent earthquakes seem to be located mainly on the SE side of the postglacial fault lines and to form parallel zones with them. This possibly indicates that the postglacial fault lines are surface expressions of old, reactivated and still active regional fracture zones, which dip to the SE.

The preliminary results of the acoustic–seismic sounding of the lake Pyhäjärvi in south-eastern Finland (Kotilainen 2001) have revealed the existence of disturbed structures in the bottom part of sediments with the age 10700–10200 BP but the upper sediments are undisturbed. The seismic origin of disturbed structures is still unclear but the results indicate, that the possible high seismic activity is connected just with the beginning of postglacial period without any later repetition of high seismic activity. Contradictory results have been presented by Lukashov (1995) from Russian Karelia and by Mörner (2001) from Sweden.

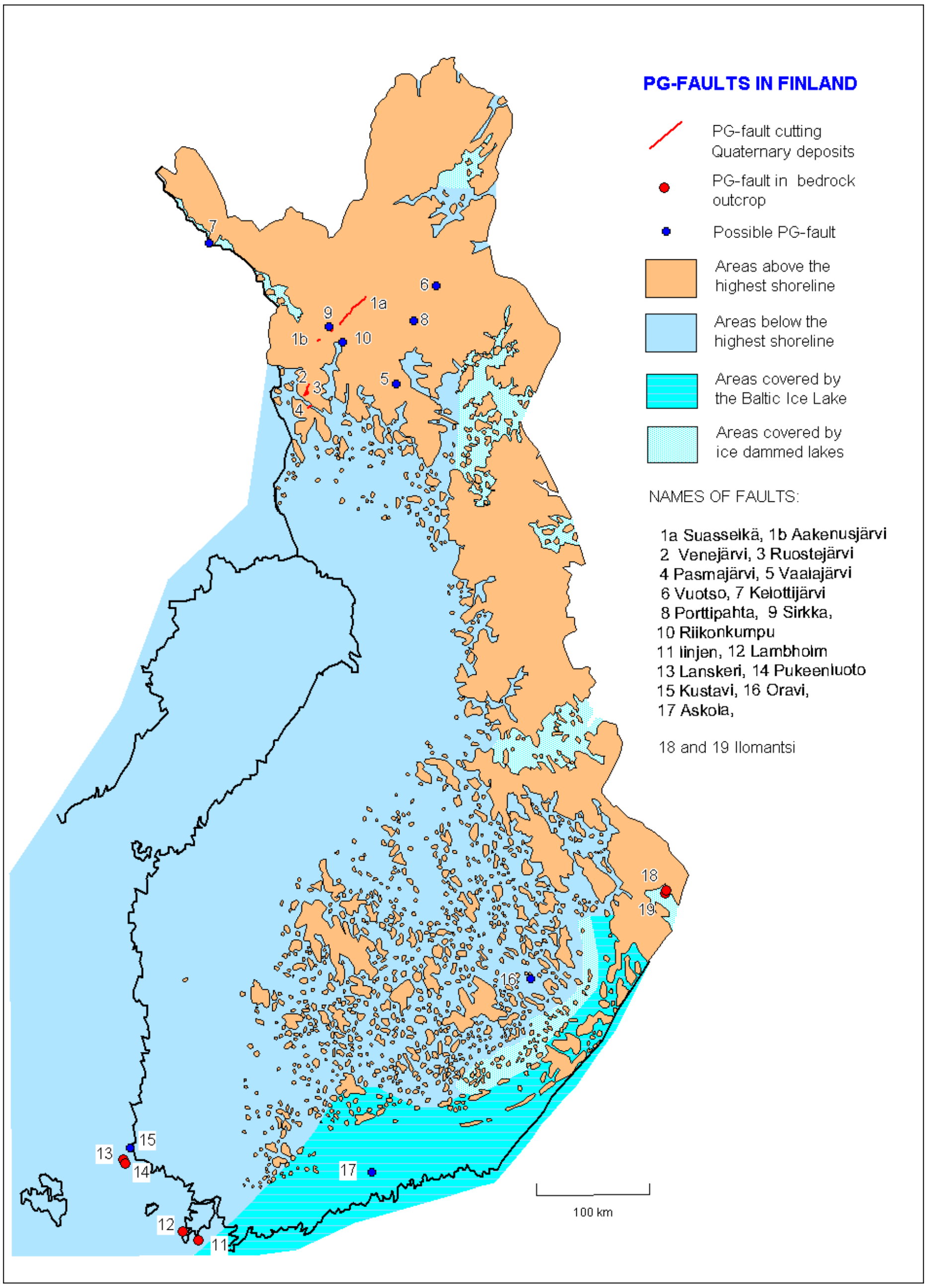


Figure 4-19. *Postglacial faults (PG faults) of Finland. Modified after Kuivamäki et al. 1998.*

According to Muir Wood (1989) the origin of the postglacial faults can be explained as a result of horizontal tectonic stress induced by Mid-Atlantic Ridge and quick isostatic land uplift combined with material flow beneath the crust toward the uplift centre. The friction of material flow caused extra shear stresses in the crust, and the stresses relaxed as reverse faulting dipping to the SE after the loading of the ice sheet disappeared. The ice loading was especially heavy in the area southwest of the Caledonian mountains and for this reason it seems logical that the longest postglacial faults are located in northern Fennoscandia.

The present vertical bedrock movements of Fennoscandia have been studied traditionally with precise levellings. In Finland the precise levelling has been carried out for three times (I 1892-1910, II 1935 –1975 and III started in 1978). Based on precise levellings, land uplift map for Fennoscandian area have been presented in Fig. 4-20.

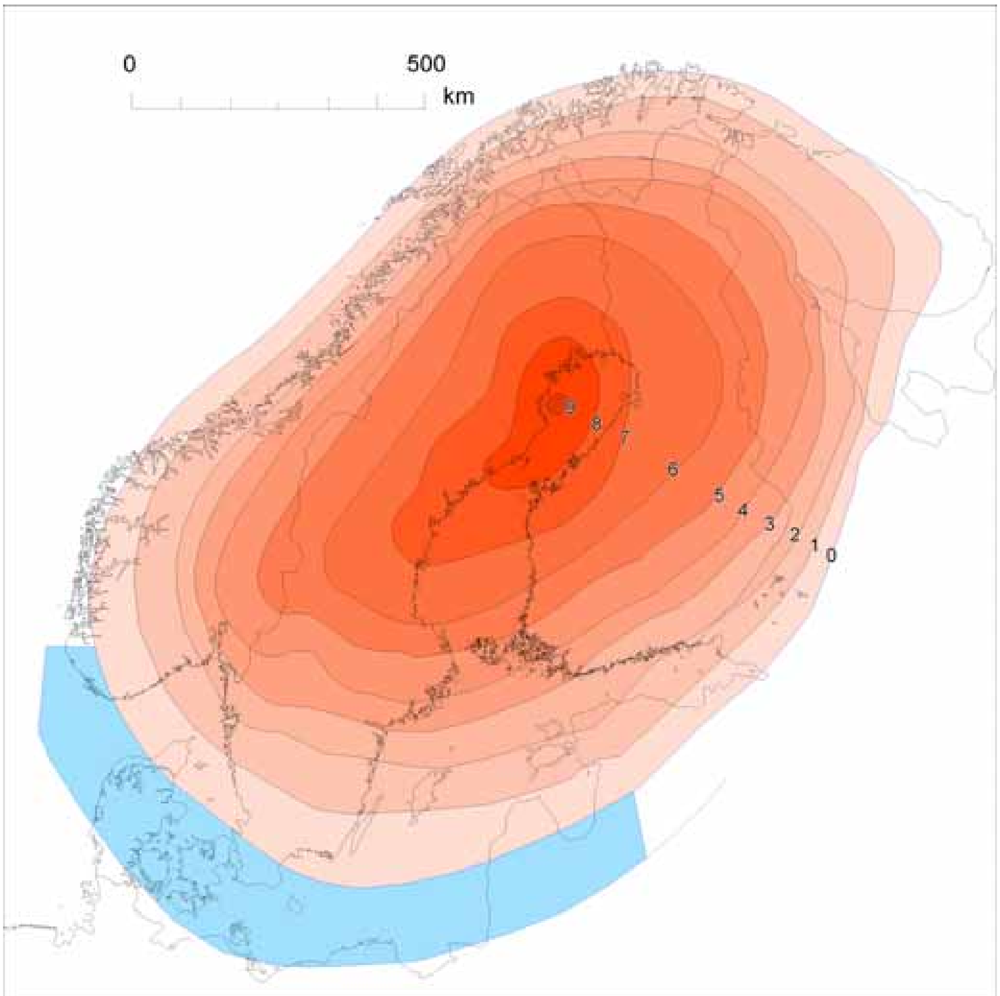


Figure 4-20. *Recent land uplift (uplift rate mm/year) in Fennoscandia. Redrawn after Ekman 1996).*

Levelling data has also been used for searching active faults in Finland. Several areas with levelling profiles crossing inferred fault zones of different size categories have been studied (Veriö et al. 1993). Of the 53 profiles levelled by the National Land Survey of Finland, 28 recorded statistically significant local changes in elevation, while 7 showed variations that deviate substantially from predictions based on uniform plastic uplift (Fig. 4-21). Magnitudes of uplift rate, classified statistically as significant to highly significant, varied from 0.24 mm/year to 1.52 mm/year.

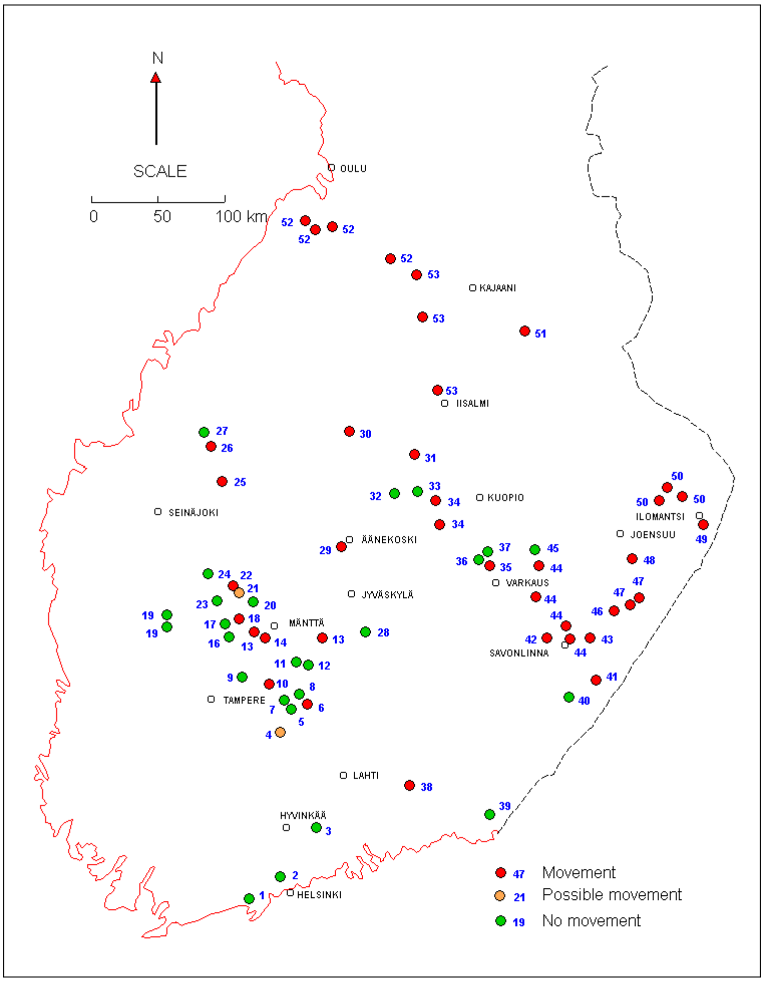


Figure 4-21. *Results of the fracture zone levellings carried out by the National Land Survey in 1974–1992 (Veriö et al. 1993). Figure from Kuivamäki et al. 1998.*

Precise levelling data of the Geodetic Institute of Finland has also been used for searching active faults in Finland (Lehmuskoski 1996). Lehmuskoski (op. cit.) found 34 places, where the uplift values between benchmarks (established in bedrock) were exceptional. These results support the notion that the present day land uplift is taking place on a regional scale plastically, but on a local scale as block movements. These block movements are preferentially concentrated within the old fracture zones. Preliminary knowledge of horizontal crustal deformations in Finland has been obtained by using measurements of the first order triangulation network. For the country as a whole, a maximum compression in the NW–SE direction is clearly visible (Chen 1991).

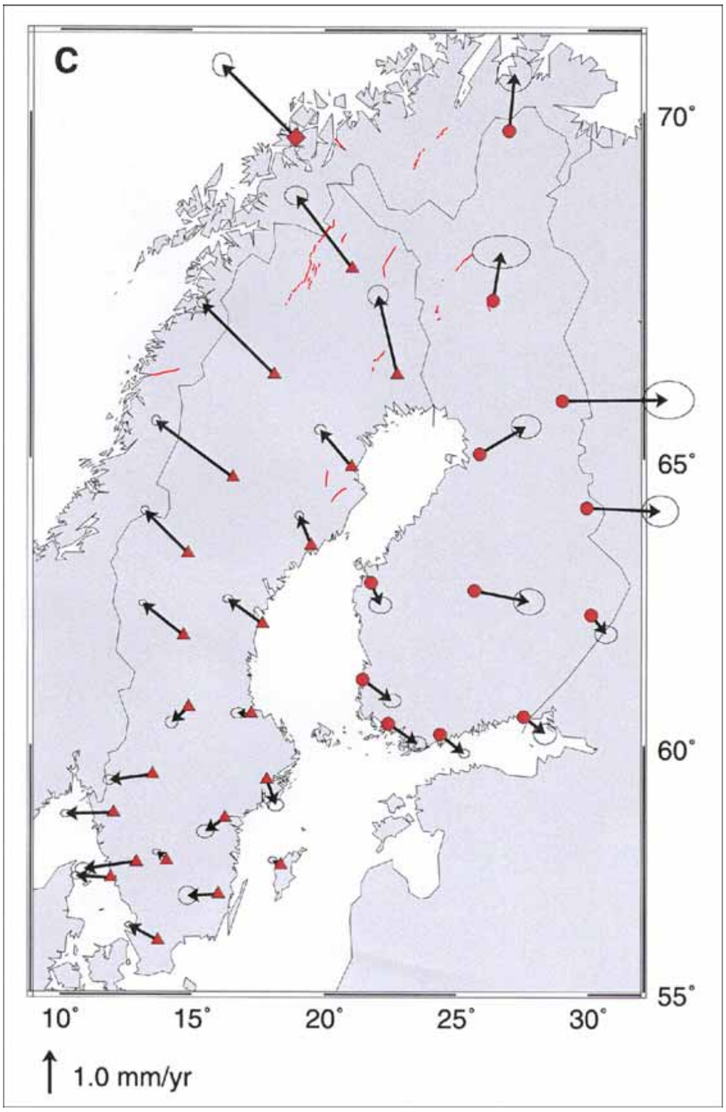


Figure 4-22. *Horizontal velocity vectors estimated at each of the BIFROST sites. The scale associated with the each of these vectors, as well as with the associated 1s error ellipses, is given at the base of the plot (Milne et al. 2001). The locations of known postglacial faults in Fennoscandia have been added (red lines) to the map (Kuivamäki & Vuorela 2001).*

5 SUMMARY

Depositional history

Meso- and Neoproterozoic sandstones and siltstones are present on land at Satakunta, Muhos and Hailuoto, as well as in the impact structures at Lappajärvi, Saarijärvi, Karikkoselkä and Iso-Naakkima. Thick deposits of Meso- and Neoproterozoic sedimentary deposits are also present in the Bothnian Sea, Bothnian Bay and Åland Sea. Early Palaeozoic, Cambrian and Ordovician, sandstones, siltstones, shales and limestones occur at the Söderfjärden, Lumparn and Saarijärvi impact structures, and at Lauhanvuori and the Bothnian Sea, presumably also at the Bothnian Bay and Åland Sea. Reworked Palaeozoic microfossils occur at the Lappajärvi and Karikkoselkä impact structures, indicating Palaeozoic strata there at the time of the impact. No middle or late Palaeozoic (Silurian to Permian) and Mesozoic (Trias to Cretaceous) sedimentary rocks are present in Finland. In eastern Lapland, Tertiary clay bed occurs at the hill of Akanvaara. Furthermore, redeposited Tertiary microfossils are found at several places in the Quaternary strata in eastern and central Lapland

At the onset of the Middle Riphean a more or less even ‘Subjotnian’ peneplain existed in the Fennoscandian Shield, where the sediments of Mesoproterozoic Middle Riphean age were deposited. Remnants of these sedimentary rocks are found both onshore and offshore in several tectonically protected basins within the Fennoscandian Shield. Although the Middle Riphean sandstones in the Fennoscandian Shield are presently only found in local basins, they were probably not solely restricted to basins but distributed over a much larger area.

Middle Riphean sedimentation was followed by intense and prolonged erosion. During the Neoproterozoic erosion period until the beginning of the Vendian, 650 Ma ago, the Fennoscandian Shield was located in the tropical zone, where both physical and chemical weathering were favoured by a warm and moist climate. Main stage of the Neoproterozoic sedimentation took place during the Late Vendian when a shallow sea transgressed large parts of the Fennoscandian Shield and the subsequent sedimentary basins were filled with arenaceous to argillaceous sediments. Although remaining Neoproterozoic (Vendian) successions within the Fennoscandian Shield presently occur only in isolated rift throughs, it is highly probable that an extensive Late Vendian sedimentary cover once existed between Hailuoto in the Bothnian Bay and Lake Ladoga in Russian Karelia. Remains of that sedimentary cover were still present in places, when the Lake Lappajärvi impact structure formed in late Cretaceous, ca. 73 Ma ago.

Before the beginning of the Cambrian period, a new extensive phase of erosion and weathering resulted in the formation of the sub-Cambrian peneplain. During the Lower Cambrian, the sea transgressed the Baltic Sea region. At its maximum the sea reached as far as the Bothnian Sea and Bothnian Bay at the beginning of the Lower Cambrian and probably covered most of the western, southern and central Finland. In the Middle Cambrian, the sea reached up to the Bothnian Sea, and Middle Cambrian sandstones were also deposited at the Söderjärden and Lake Lappajärvi areas in western Finland, and at the Lake Karikkoselkä area in central Finland. Upper Cambrian transgression reached the area of Åland islands, while the mainland Finland probably was dry land.

In the Ordovician, an epicontinental sea with repeated transgressions and regressions covered most of the southern part of the Fennoscandian Shield, including the Baltic Basin, the sedimentation being predominantly governed by carbonates. The first Ordovician transgression occurred at early Lower Ordovician (Early Tremadocian). During this transgression, the sea is interpreted to have reached southern Finland, the Åland Sea basin and the Bothnian Sea basin. The following transgression during the Billingen and Volkhov regional stages of the Lower Ordovician reached at least to the Bothnian Sea region, and probably the Bothnian Bay.

A new phase of transgression began at the Lower to Middle Ordovician Kunda regional stage, and during that stage and the successive Middle Ordovician Aseri, Lasnamägi and Uhaku regional stages, the Bothnian Sea and the western coastal area of Finland were covered by sea. During the Lasnamägi to Kukruse regional stages, the sea covered also the central Finland.

During the Middle Ordovician Keila-Oandu regression event, Finland was most likely dry land, possibly also the Åland Sea and Åland Islands and most of the Bothnian Sea, while during the successive late Middle to Upper Ordovician Rakvere, Nabala, Vormsi and to Pirgu transgression events, southernmost and south-westernmost Finland, the Åland Sea basin and the Bothnian Sea basin are interpreted to have been covered by water. The Upper Ordovician Porkuni regional stage is characterized by a drastic regression, and from that on until the Pleistocene glaciations, most of Finland including the Åland Islands, the Åland Sea basin and Bothnian Sea basin, is considered to have been in a terrestrial position.

No Late Palaeozoic sedimentary rocks have been recognized whether neither in Finland nor in Sweden, however, the fission track studies indicate that extensive Silurian to Devonian deposits most likely covered the Fennoscandian Shield. The thickness of the sedimentary sequences may have been 3 - 4 km in southern and western Sweden thinning to about 1 km in the Åland Islands. Hardly any Mesozoic or Tertiary sedimentary rocks are preserved within the Fennoscandian Shield, the Mesozoic to Tertiary strata restricting to the Fennoscandian Border Zone in the southernmost province of Scania in Sweden and adjacent offshore areas. In eastern Lapland, the clay bed of Akanvaara was deposited in early Tertiary (Eocene) under marine conditions.

Tectonic regimes

The Mesoproterozoic tectonics was dominated by the Sveconorwegian orogeny (in North America the Grenvillian orogeny), which, in the periphery of the Fennoscandian Shield, mainly affected present south-western Sweden and southern Norway. At ca. 1265 Ma (and at least 200 Ma before that) Laurentia craton and Baltica craton formed a coherent continent with present-day northern Baltica facing eastern Greenland. The Mesoproterozoic, 'Postotnian' olivine diabases, dated at 1270 - 1250 Ma, in Sweden (Central Scandinavian Dolerite Group) and Finland (Satakunta and Vaasa olivine diabases), and related diabases in Greenland are considered to represent the initial rifting between the Baltica and Laurentia cratons. Extensional period between ca. 1250 and 1100 Ma followed the early Sveconorwegian phase. Gabbroic and granitic rocks related to this early Sveconorwegian extension are found in the Bamble-Kongsberg sector in SE Norway. The rifting gave rise to a system of rifts and aulacogens both along the present-day western and eastern margins of Baltica and in the interior of the craton.

The main phase of the Sveconorwegian orogeny, i.e. the collision between Baltica and Laurentia (or some other), occurred ca. 1150 - 950 Ma ago. The oblique convergence during the collision resulted in south-eastward and eastward thrusting in SW Sweden with thickening of the crust. The Sveconorwegian metamorphism resulted in widespread granulite facies assemblages in the southern distal part of the orogeny, while amphibolite facies assemblages were dominant elsewhere. During the Sveconorwegian orogeny, major deformation occurred in south-western Sweden along N-S trending major shear zones, including the Mylonite Zone and the Protogine Zone. These shear zones divide the area into several crustal segments (Western, Median and Eastern), between which the timing, character and intensity of deformation and metamorphism may vary considerably. The late Sveconorwegian extension resulted in an extensive rifting, which is seen, for example, as N-S trending Blekinge-Dalarna dyke swarm, dated at ca. 930 Ma and related to the late Sveconorwegian extensional uplift at the area west of the Protogine Zone. Post-compressional granitic intrusions of the Sveconorwegian orogeny include the 920 Ma Bohus granite in Sweden and the coeval granites in south-western Norway.

After the Sveconorwegian orogeny, Laurentia and Baltica together with all the other continents formed one single worldwide supercontinent, Rodinia. Within Baltica craton, extensional tectonics and intracratonic basin evolution was initiated ca. 750 - 800 Ma ago and aulacogens, rift basins and continental depressions formed within the craton from Ukraine to Scandinavia, characterized by predominantly continental sediments.

Palaeomagnetic data indicates that Baltica and Laurentia were together until at least 620 - 630 Ma. Basaltic dyke swarms and sheeted dykes related to the rifting of the two cratons are common and locally voluminous in the north-western margin of Baltica, and are preserved in the Caledonian nappes. Egersund basaltic dyke swarm in south-western Norway, dated at 616 Ma, has been interpreted as relating to rifting along the western margin of Baltica that resulted in opening of Iapetus Ocean between Baltica and Laurentia.

The Sarek dyke swarm in Norway, dated at 608 Ma, is inferred to date the onset of seafloor spreading (transition from rift to drift) in the Iapetus Ocean.

Resulting from the break-up of the supercontinent Rodinia during late Precambrian to early Cambrian, possibly up to 5000 km wide Iapetus Ocean separated Baltica from Laurentia, while Tornquist Sea was located between Baltica and Gondwana cratons. By Late Cambrian times, the plate convergence started again, leading to collision of north-western Baltica margin with an assumed island-arc complex, which resulted in the early stage of the Caledonian orogeny, referred to as the Finnmarkian orogeny, in Late Cambrian-Early Ordovician. By Late Ordovician-Early Silurian (ca. 440 Ma) the Tornquist Sea between Baltica and micro-continent Avalonia closed, leading to development of the North German-Polish Caledonides.). In Silurian, at ca. 425 Ma, Baltica and Avalonia collided with Laurentia, which resulted in closure the Iapetus Ocean and the main stage of the Caledonian orogeny, referred to as the Scandian orogeny. During the continent-continent collision, Baltica subducted deep beneath Laurentia, giving rise to eastward thrusting over the western Baltoscandian margin. The crustal shortening during the Scandian event is estimated to be at least 400 km.

The Silurian continent-continent collision was followed in the Devonian by uplift and denudation of the Scandinavian Caledonides. Up to 4 km of sediments ("Old Red Sandstone") is interpreted to be deposited on the Caledonian foreland basin east of the orogen. The extensional collapse of the orogeny caused the uplift of the foreland basin, and the subsequent erosion of the basin fill started during the mid-Devonian. The Devonian, post-Caledonian tectonics are characterized, especially in western and southern Norway, by major extensional detachments and extensional reactivation of major shear zones.

During the Late Palaeozoic, the continents were gradually assembled into the supercontinent Pangaea. The final completion of Pangaea took place at Late Carboniferous when the northern margin of Gondwana collided with the combination of Laurentia and Baltica, known as Laurussia, resulting in Variscan orogeny in Europe. During the late Variscan orogeny, rifting and volcanism occurred throughout north-western Europe. The tectonics related to the late Variscan orogeny resulted in a dextral transpressional and/or transtensional strike-slip faulting along the NW-SE trending Tornquist Zone in the south-western margin of the Fennoscandian Shield. The faulting was associated with volcanic activity and intrusion of dyke swarms, the ages of which range from 300 to 240 Ma.

The Permian rifting and volcanism is demonstrated by ca. 400 km long Oslo Rift where mafic to silicic extrusive and intrusive rocks formed in several phases during the time period of 305 - 240 Ma. The Oslo Rift is characterized by major N-S and NNE-SSW to NE-SW trending faults, which have been formed in several phases under both ductile and brittle tectonic regimes.

At the beginning of the Mesozoic, all the continents were assembled into a supercontinent Pangaea, formed during the Palaeozoic. Major tectonic activity was concentrated to the break-up of Pangaea, which begun in the Triassic. During the Jurassic, the Tethys-Central

Atlantic-Gulf of Mexico rift system developed into a principal fracture system along which the Central Atlantic opened by sea-floor spreading initiating the Laurasia-Gondwana separation during Middle Jurassic to Early Cretaceous. Although rifting occurred in the Norwegian-Greenland Sea rift throughout the Jurassic and Cretaceous, the final crustal separation between Greenland and Europe took place only during the Early Tertiary. There was hardly any deformation within the Fennoscandian Shield during the Mesozoic, however, the south-western margin of the shield, the Tornquist Zone, was repeatedly reactivated during the Mesozoic rifting phases. From Early Triassic to Early Cretaceous the stress regime in the Tornquist Zone was dominantly transtensional.

In the Tertiary, the evolution of north-western Europe was dominated by the opening of the North Atlantic and the initiation of sea-floor spreading in the Norwegian-Greenland Sea and by onset of Alpine continent-continent collision. The latest phase of deformation, mainly affecting the margins of the Fennoscandian Shield, was the uplift of western Scandinavia. The first phase of uplift, amounting close to 1500 m in northern Sweden occurred in the Palaeogene, in connection with the opening of the North. The second major episode of uplift occurred in Neogene starting from late Oligocene and had its centre in southern Norway. During Late Pliocene and Pleistocene, the tectonic uplift was amplified by isostatic rebound in response to the glaciation. In southern Scandinavian Caledonides, a Neogene uplift of about 1000 metres has been inferred.

The orientation of maximum horizontal compression present within the Fennoscandian Shield today is in good accordance with the direction of ridge push forces from the Mid-Atlantic Ridge, which, accordingly, seems to be the major stress-generating mechanism. However, the current stress field in Fennoscandia is likely a combination of plate boundary forces with local sources (e.g. glacial rebound and local geology). The stresses in the Fennoscandian Shield are mainly considered to be due to three mechanism: 1) residual stress left from the isostatic rebound after deglaciation, 2) concentration of the stress due to creep, and 3) stress caused by the spreading of the Mid-Atlantic Ridge.

Neotectonic movements (postglacial and recent crustal movements) are clear evidences of young bedrock movements and offer therefore good possibilities for studying the behaviour of bedrock. Large postglacial faults were discovered in 1960's in northern Finland and later all over the northern Fennoscandia. Small post-glacial faults located in ice polished bedrock outcrops and with a scarp height 0–20 cm have been found in southern Finland, but so far larger post-glacial faults have not been recognised. Investigations on postglacial bedrock movements have revealed that the postglacial faults studied so far are situated in old, reactivated fracture zones. Estimations of earthquake magnitudes connected with postglacial faults in the Finnish Lapland have varied from 5.3 to 7.5.

REFERENCES

Abels, A., Bergman, L., Lehtinen, M. & Pesonen, L. J. 2000. Structural constraints and interpretations on the formation of the Söderfjärden and Lumparn impact structures, Finland. *In:* Plado, J. & Pesonen, L.J. (eds.) Meteorite impacts in Precambrian shields. 4^{th} ESF Workshop, Lappajärvi-Karikkoselkä-Sääksjärvi, Finland, May 24-28, 2000. Espoo: Helsinki: Geological Survey of Finland : University of Helsinki. Programme and abstracts, p. 26.

Åhäll, K.-I. 1995. Crustal units and role of the Mylonite Zone system in the Varberga-Horred region, SW Sweden. GFF 117, pp. 185-198.

Åhäll, K.-I. & Gower, C.F. 1997, The Gothian and Labradorian orogens: variations in accretionary tectonism along a late Paleoproterozoic Laurentia-Baltica margin. GFF 119, p. 181-191.

Åhäll, K.-I. & Connelly, J. 1996. Proterozoic plate geometry in the North Atlantic region: constraints from persistent 1.51-1.15 Ga anorogenic magmatism in Scandinavia. GFF 118 Jubilee Issue, pp. A5-A6.

Åhäll, K-I & Connelly, J. 1998. Intermittent 1.53-1.13 Ga magmatism in western Baltica age constraints and correlations within a postulated supercontinent. Precambrian Research 92 (1), pp. 1-20.

Åhäll, K-I & Schöberg, H. 1999. The 963 Ma Vinga intrusion and post-compressional deformation in the Sveconorwegian Orogen, SW Sweden. GFF 121 (2), pp. 101-106.

Åhäll, K.-I., Daly, J.S. & Schöberg, H. 1990. Geochronological constraints on Mid-Proterozoic magmatism in the Östfold-Märstrand Belt: Implications for crustal evolution in SW Sweden. *In:* Gower, G., Rivers, T. & Ryan, B. (eds.) Mid-Proterozoic Laurentia-Baltica. Geological Association of Canada Special Paper 38, pp. 471-483.

Ahlberg, P., Clarkson, E.N.K. & Taylor, C.M. 1996. Trilobites and faunal dynamics in the Upper Cambrian of Sweden. GFF 118 Jubilee Issue, pp. A57-58.

Ahlin, S. 1980. Description of the solid rocks map of Borås, Sweden (in Swedish with an English summary). Sveriges Geologiska Undersökning, Serie Af 130, 114 p.

Ahlin, S. 1987. Phanerozoic faults in the Västergotland basin area, SW Sweden. Geologiska Föreningens i Stockholm Förhandlingar 109 (3), pp. 221-227.

Ainsaar, L., Kirsimäe, K. & Meidla, T. 1996. Regression in Caradoc: evidence from south-western Estonia (Ristiküla core). Geological Survey of Denmark and Greenland Report 98, pp. 5-12.
Alviola, R. 1988. Search of Ordovician limestones from the Valkiameri basin during summer 1984 (in Finnish). Geological Survey of Finland, unpublished report M 9/1131/84, 10 p.

Amantov, A. 2001. Geological structure of the eastern (Baltic-White Sea) subaquaeous Fennoscandian margin. Geonorus, journal of geology online (http://www.conset.net/aaman63/fenmarg.htm).

Amantov, A. & Muller, J. 2001. Geological history and bedrock geology of the St. Petersburg area. *In:* Geological basis for water planning in the St. Petersburg area. Scientific cooperation on sustainable water management in agricultural areas around St. Petersburg (SCOPE). Reports (http://www.scope.ruc.dk/bedrockgeo.htm).

Amantov, A., Laitakari, I. & Poroshin, Ye. 1996. Jotnian and Postjotnian: sandstones and diabases in the surroundings of the Gulf of Finland. In: Koistinen, T. J. (ed.) Explanation to the map of Precambrian basement of the Gulf of Finland and surrounding area 1 : 1 mill.. Espoo: Geological Survey of Finland. Special Paper 21, pp. 99-113.

Andersen, T.B. 1998. Extensional tectonics in southern Norway: An overview. Tectonophysics 285, pp. 333-351.

Andersen, T.B., Torsvik, T.H., Eide, E.A., Osmundsen, P.E. & Faleide, J.I. 1999. Permian and Mesozoic extensional faulting within the Caledonides of central south Norway. Journal of the Geological Society of London 156, pp. 1073-1080.

Andersson, A., Dahlman, B., Gee, D. G. & Snall, S. 1985. The Scandinavian alum shales. Sveriges Geologiska Undersökning, Serie Ca N:o 56, 50 p.

Andersson, J. 2001. Sveconorwegian orogenesis in the south-western Baltic Shield - Zircon geochronology and tectonothermal setting of orthogneisse in SW Sweden. Lund University, Faculty of Science, doctoral dissertation, 155 p.

Andersson, J., Söderlund, U., Cornell, D., Johansson, L. & Möller, C. 1999. Sveconorwegian (-Grenvillian) deformation, metamorphism and leucosome formation in SW Sweden, SW Baltic Shield; constraints from a Mesoproterozoic granite intrusion. Precambrian Research 98 (1-2), pp. 151-171.

Andersson, J., Söderlund, U., Möller, C. & Johansson, L. 2001. Sveconorwegian (-Grenvillian) orogenesis in the SW Baltic Shield – timing and tectonic framework in light of complex zircon geochronology. The Geological Society of America Annual Meeting abstracts, November 5-8, 2001, Session 179.

Andréasson, P.G. 1994. The Baltoscandian margin in the Neoporterozoic-early Palaeozoic times, some constraints on terrane derivation and accretion in the Arctic Scandinavian Caledonides. Tectonophysics 231, pp. 1-32.

Andréasson, P. G. & Rodhe, A. 1990, Geology of the Protogine Zone south of Lake Vättern: a reinterpretation: Geologiska Föreningens i Stockholm Förhandlingar 112, pp. 107-125.

Andréasson, P.G. & Albrecht, L. 1995. Derivation of 500-Ma eclogites from the passive margin of Baltica and a note on the tectonometamorphic herogeneity of eclogite-bearing crust. Geological Magazine 132, pp. 729-738.

Andréasson, P-G., Svenningsen, O. M. & Albrecht, L. 1998. Dawn of Phanerozoic orogeny in the North Atlantic tract; Evidence from the Seve-Kalak Superterrane, Scandinavian Caledonides. GFF 120 (2), pp. 159-172.

Arkonsuo, A. 2000. Impaktimetamorfoosi ja Karikkoselän kraateri Petäjävedellä. Impact metamorphism and Karikkoselkä crater at Petäjävesi (in Finnish). University of Helsinki, Msc. Thesis, 123 p.

Artyushkov, E.A., Artyushkov, Y. A., Lindström, M. & Popov, L. E. 2000. Relative sea-level changes in Baltoscandia in the Cambrian and Early Ordovician; the predominance of tectonic factors and the absence of large scale eustatic fluctuations. Tectonophysics 320 (3-4), pp. 375-407.

Asklund & Kulling, O. 1926. Nya data till Ålands geologi II. Den ny-upptäcta östersjönkalken i Lumparnfjärden (in Swedish). Geologiska Föreningens i Stockholm Förhandlingar 48, pp. 503-509.

Axberg, S., 1980. Seismic stratigraphy and bedrock geology of the Bothnian sea, Northern Baltic. Stockholms Contributions in Geology 36, pp. 153 - 213.

BABEL WG 1993. Integrated seismic studies of the Baltic Shield using data in the Gulf of Bothnia region. Geophysical Journal International 112 (3), pp. 305-324.

van Balen, R.T. & Heeremans, M. 1998. Middle Proterozoic-early Palaeozoic evolution of central Baltoscandian intracratonic basins; evidence for asthenospheric diapirs. Tectonophysics 300 (1-4), pp. 131-142.

Bergman, L. 1982. Paleozoic sediments in the rapakivi area of the Åland Islands. Geological Survey of Finland. Bulletin 317, pp. 7-27.

Bergman, L. & Lindberg, B. 1979. Phanerozoic veins of galena in the Åland rapakivi area, south-western Finland. Bulletin of the Geological Society of Finland 51 (1-2), pp. 55-62.

Bergström, J. & Gee, D.G. 1985. The Cambrian in Scandinavia. *In:* Gee, D. G. & & Sturt, B.A (eds.) The Caledonide Orogen; Scandinavia and related areas; Vol. 1. Chichester;, United Kingdom: John Wiley & Sons, pp. 247-271.

Bergström, J. & Kornfält, K-A. 1998. Outline of the geology of Scania. *In:* Ahlberg, P. (ed.) Guide to excursions in Scania and Västergötland, southern Sweden. Lund Publications in Geology 141, pp. 17-19.

Bingen, B., Demaiffe, D. & van Breemen, O. 1998. The 616 Ma old Egersund basaltic dike swarm, SW Norway, and late Neoproterozoic opening of the Iapetus Ocean. Journal of Geology 106 (5), pp. 565-574.

Bruton, D. L., Lindström, M., Owen, A. W. 1985. The Ordovician of Scandinavia. *In:* Gee, D. G. & Sturt, B.A. (eds.) The Caledonide Orogen; Scandinavia and related areas; Vol. 1. Chichester, United Kingdom : John Wiley & Sons, pp. 273-282.

Buchan, K. L.; Mertanen, S.; Park, R. G.; Pesonen, L. J.; Elming, S. A.; Abrahamsen, N.; Bylund, G. 2000. Comparing the drift of Laurentia and Baltica in the Proterozoic; the importance of key palaeomagnetic poles. Tectonophysics 319 (3), pp. 167-198.

Bugge, T., Prestvik, T. & Rokoengen, K. 1980. Lower Tertiary volcanic rocks off Kristiansund, mid-Norway. Marine Geology 35, pp. 277-286.

Cederbom, C. 1997. Fission track thermochronology applied to Phanerozoic thermotectonic events in central and southern Sweden. Göteborg University, Licenciate thesis, 47 p.

Cederbom, C. 2001. Phanerozoic, pre-Cretaceous thermotectonic events in southern Sweden revealed by fission track thermochronology. Earth and Planetary Science Letters 188 (1-2), pp. 199-209.

Cederbom, C., Larson, S.Å., Tullborg, E.-L. & Stiberg, J.-P. 2000. Fission track thermochronology applied to Phanerozoic thermotectonic events in central and southern Sweden. Tectonophysics 316 (1-2), pp. 153-168.

Chen, R. 1991. On the horizontal crustal deformations in Finland. Publications of the Finnish Geodetic Institute 91 (1), 98 p.

Chen, R.& Kakkuri, J. 1994. Feasibility study and technical proposal for long-term observations of bedrock stability with GPS. Report YJT-94-02. 33 p.

Chen, R.& Kakkuri, J. 1998. GPS operations at Olkiluoto, Kivetty and Romuvaara for 1997. Posiva Working Report 98-08e. 44 p., 222 app. pages.

Clauss, B., Marquart, G. & Fucks, K., 1989. Stress orientations in the North Sea and Fennoscandia, a comparison to the Central European stress field. *In:* Gregersen, S. & Basham, P.W. (ed.) Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound. NATO ASI Series, Series C: Mathematical and Physical Sciences - Vol. 266. Dordrecht/Boston/London: Kluwer Academic Publishers, pp. 277-287.

Connelly, J. N., Berglund, J. & Larson, S. Å. 1996. Thermotectonic evolution of the Eastern Segment of south-western Sweden; tectonic constraints from U-Pb geochronology. *In:* Brewer, T. S. (ed.) Precambrian crustal evolution in the North Atlantic region. Geological Society of London Special Publications 112, pp. 297-313.

Connelly, J.N, Åhäll, K.-I. & Brewer, T. 2001. Accretionary growth of the Sveconorwegian Province of the Baltic Shield between 1.7-1.5 Ga and links to intracontinental magmatism. Geological Society of America, 2001, annual meeting. Geological Society of America, Abstracts with Programs 33 (6), p. 30.

Cornell, D.H., Larson, S.Å., Berglund, J., Connelly, J.N., Armstrong, R., Nesbitt, B. & Milton, A. 1996. Genesis and U-Pb dating of zirkon rims in migmatite. 22nd Nordic Geological Winter Meeting, Turku 1996, Abstracts, p. 32.

Dalziel, I.W.D. 1997. Neoproterozoic-Paleozoic geography and tectonics: Review, hypothesis, environmental speculation. Geological Society of America Bulletin 109, pp. 16-42.

Dalziel, I.W.D., Dalla Salda, L.H. & Gahagan, L.M. 1994. Palaeozoic Laurentia-Gondwana interaction and the origin of the Appalachian-Andean mountain system. Geological Society of America Bulletin 106, pp. 243-252.

Deeks, N.R. & Thomas, S.A. 1994. Basin inversion in a strike-slip regime; the Tornquist Zone, southern Baltic Sea. Norsk Geologisk Tidsskrift 74 (3), p. 174.

Deeks, N.R. & Thomas, S.A. 1995. Basin inversion in a strike-slip regime; the Tornquist Zone, southern Baltic Sea. *In:* Buchanan, J.G., Buchanan, P.G. (eds.) Basin inversion. Geological Society of Londonn Special Publications 88, pp. 319-338.

Donner, J. 1996. On the origin and glacial transport of erratics of Jotnian sandstone in south-western Finland. Bulletin of the Geological Survey of Finland 68 (2), pp. 72-83.

Ekman, M. 1996. A consistent map of the postglacial uplift in Fennoscandia. Terra Nova 8, pp. 158-165.

Edelman, N. 1949. Some morphological details of the roches moutonnees in the archipelago of SW Finland. Bulletin de la Commission géologique de Finlande 144, 129–137.

Edelman, N. 1951. Glacial abrasion and ice movement in the area of Rosala-Nötö, SW-Finland. Bulletin de la Commission géologique de Finlande 154, pp. 157-169.

Eliasson, T. & Schöberg, H. 1991. U-Pb dating of the post-compressional Sveconorwegian Bohus granite, SW Sweden: Evidence of restitic zircon. Precambrian Research 51, pp. 337-350.

Eide, E.A., Torsvik, T.H. & Andersen, T.B. 1997. Absolute dating of brittle fault movements: Late Permian and late Jurassic extensional fault breccias in western Norway. Terra Nova 9, pp. 135-139.

Elming, S-Å & Matsson, H. 2001. Post Jotnian basic intrusions in the Fennoscandian Shield, and the break up of Baltica from Laurentia; a palaeomagnetic and AMS study. Precambrian Research 108 (3-4), pp. 215-236.

Elo, S. 1982. On the gravimetric studies of the bedrock in Satakunta (in Finnish). Espoo: Geological Survey of Finland. Unpublished report Q 20/21/1982/1, 17 p.

Elo, S., Mattsson, A. & Kurimo, M. 1993a. The additional geophysical investigations of the Kauttua-Virttaankangas tunnel route in 1992 (in Finnish). Espoo: Geological Survey of Finland. Unpublished report.

Elo, S.. Kuivasaari, T., Lehtinen, M., Sarapää, O. & Uutela, A. 1993. Iso-Naakkima, a circular structure filled with Neoproterozoic sediments, Pieksämäki, south-eastern Finland. Bulletin of the Geological Society of Finland 65 (1), pp. 3-30.

Elvhage, C. & Lidmar-Bergström, K. 1987. Some working hypothesis on the geomorphology of Sweden in the light of a new relief map. Geogr. Ann. A 69, pp. 343-358.

Eriksson, B., Grönlund, T. & Uutela, A. 1999. Biostratigraphy of Eemian sediments at Mertuanoja, Pohjanmaa (Ostrobothnia), western Finland. Boreas 28 (2), pp. 274-291.

Erlström, M., Thomas, S. A., Deeks, N. & Sivhed, U. 1997. Structure and tectonic evolution of the Tornquist Zone and adjacent sedimentary basins in Scania and the southern Baltic Sea area. Tectonophysics 271 (3-4), pp. 191-215.

Fenner, J. 1988. Occurrences of pre-Quaternary diatoms in Scandinavia reconsidered. Meyniana 40, pp. 133-141.

Fjelskaar, W. & Cathles, L. 1991. The present rate of uplift in Fennoscandia implies a low-viscosity astenosphere. Terra Nova 3, pp. 393-400.

Flodén T., 1980**.** Seismic stratigraphy and bedrock geology of the central Baltic. Stockholm Contributions in Geology 35, 240 p.

Fossen, H. & Rykkelid, E. 1992. Postcollisional extension of the Caledonide orogen in Scandinavia: Structural expressions and tectonic significance. Geology 20, pp. 737-740.

Fossen, H. & Dunlap, W.J. 1998. Timing and kinematics of Caledonian thrusting and extensional collapse, southern Norway: evidence from $^{40}Ar/^{39}Ar$ thermochronology. Journal of Structural Geology 20, pp. 765-781.

Franke, W. 1996. Evolution of the central European Variscides. GFF 118 Jubilee Issue, pp. A31-A32.

Gee, D.G. 1985. A tectonic model for the central part of the Scandinavian Caledonides. American Journal of Science 275, pp. 468-515.

Gorbatschev, R. 1967. Petrology of Jotnian rocks in the Gävle area, east central Sweden. Sveriges Geologiska Undersökning, Serie C 621, 50 p.

Gorbatschev, R., Lindh, A., Solyom, Z., Laitakari, I., Aro, K., Lobach-Zhuchenko, S. B., Markov, M. S., Ivliev, A. I., Ivliyev, A. I. & Brynhi, I. 1987. Mafic dyke swarms of the Baltic Shield. *In:* Halls, Henry C. & Fahrig, W.F. (eds.) Mafic dyke swarms; a collection of papers based on the proceedings of an international conference. Geological Association of Canada Special Paper 34, pp. 361-372

Gregersen, S., Korhonen, H. & Huesebye, E.S. 1991. Fennoscandian dynamics: Present-day earthquake activity. Tectonophysics 189, pp. 333-344.

Grönlund, T. 1977. The occurrence of Tertiary diatoms in Lapland (in Finnish with an English abstract). Espoo: Geological Survey of Finland. Report of Investigations 17, pp. 19-30.

Hagenfeldt, S. E. 1988. Acritarch assemblages of Early and Middle Cambrian age in the Baltic Depression and south-central Sweden. *In:* Winterhalter, B. (ed.) The Baltic Sea : papers prepared for a colloquium on Baltic Sea marine geology, Parainen, Finland, 27-29, May 1987. Espoo: Geological Survey of Finland. Special Paper 6, pp. 151-161.

Hagenfeldt, S.E. 1989a. Lower and Middle Cambrian acritarchs from the Baltic Depression and south-central Sweden, taxonomy, stratigraphy and palaeogeographic reconstruction. University of Stockholm, Department of Geology. Doctoral thesis, 32 p.

Hagenfeldt, S.E. 1989b. Lower and Middle Cambrian acritarchs from the Baltic Depression and south-central Sweden, taxonomy and biostratigraphy. Stockholm Contributions in Geology 41, pp. 1-250.

Hagenfeldt, S.E. 1994. The Cambrian File Haidar and Borgholm Formations in the Central Baltic depression and south central Sweden. Stockholm Contributions in Geology 43, pp. 69-110.

Hagenfeldt, S.E. 1995. Erratics and Proterozoic-Lower Palaeozoic submarine sequences between Åland and mainland Sweden. Geological Survey of Sweden, Research papers, SGU series Ca 84, 35 p.

Hagenfeldt, S.E. 1997. The development of the Cambrian in Baltoscandia. *In:* Cato, I. & Klingberg, F. (eds.) Proceedings of the Fourth Marine Geologcal Conference – the Baltic, Uppsala 1995. Geological Survey of Sweden, Research papers, SGU series Ca 86, pp. 61-66.

Hagenfeldt, S.E. & Söderberg, P. 1994. Lower Cambrian sandstone erratics and geophysical indications of sedimentary rock in the Stockholm area, Sweden. GFF 116, pp. 185-190.

Hansen, B. T., & Lindh, A. 1991, U-Pb zircon age of the Görbjörnap syenite in Skåne, southern Sweden. Geologiska Föreningens i Stockholm Förhandlingar 113, pp. 335-337.

Hints, L. 1998. Oandu Stage (Caradoc) in central Estonia. Proceedings of Estonian Academy of Sciences, Geology 47, pp. 158-172.

Hints, L., Meidla, T., Nõlvak, J. & Sarv, L. 1989. Some specific features on the Late Ordovician evolution in the Baltic Basin. Proceedings of the Academy of Sciences of the Estonian SSR, Geology 38, pp. 83-87.

Hirvas H. & Tynni R., 1976. Tertiary clay deposit at Savukoski, Finnish Lapland, and observations of tertiary microfossils, preliminary report (In Finnish). Geologi 28, pp. 33 - 40.

Hirvas, H., Kujansuu, R. & Tynni, R. 1976. Till stratigraphy in northern Finland. *In:* Easterbrook, D. J. & Sibrava, V. (eds.) Quaternary glaciations in the northern hemisphere: report no. 3 on the session in Bellingham, Washington, USA, September 1975. Prague: Geological Survey, pp. 256-273.

Hokkanen, K. 2002. Occurrence of unmetamorphosed limestones in Åland (in Finnish). Espoo: Geological Survey of Finland. Unpublished report P 31.4.031, 19 p.

Holmer, L.E. & Popov, L.E. 1990. The acrotretacean brachiopod Ceratreta tanneri (Metzer) from the Upper Cambrian of Baltoscandia. Geologiska Föreningens i Stockholm Förhandlingar 112 (3), pp. 249-263.

Huff, W.G., Bergström, S.M. & Kolata, D.R. 1992. Gigantic Ordovician ash fall in North America and Europe; biological, tectonomagmatic, and event-staritigraphic significance. Geology 20, pp. 875-878.

Hurich, C. A. 1996. Kinematic evolution of the lower plate during intracontinental subduction: An example from the Scandinavian Caledonides. Tectonics 15(6):1248-1263.

Jaeger, H. 1984. Einige Aspekte der geologischen Entwicklung Südskandinaviens im Altpaläozoicum (in German with an English summary). Zeitschrift für angewandte Geologie 30 (1), pp. 17-32.

Jaanusson, V. 1963. Classification of the Harjuan (upper Ordovician) rocks of the mainland of Sweden. Geologiska Föreningens i Stockholm Förhandlingar 85 (1), pp. 110-144.

Jaanusson, V. 1973. Aspects of carbonate sedimentation in the Ordovician of Baltoscandia. Lethaia 6 (1), pp. 11-34.

Jaanusson, V. 1976. Faunal dynamics in the Middle Ordovician (Viruan) of Balto-Scandia. *In:* Basset, M.G. (ed.) The Ordovician System: Proceedings of a Palaentological Association Symposium, Birmingham, Sebtember 1974, pp. 301-326.

Johansson, Å. 1990. Age of the Onnestad Syenite and some gneissic granites along the southern part of the Protogine Zone, southern Sweden. *In:* Gower, C. F., Rivers, T., & Ryan, B. (eds.) Mid-Proterozoic Laurentia-Baltica. Geological Association of Canada Special Paper 38, pp. 131-148.

Johansson, L. & Johansson, Å. 1990, Isotope geochemistry and age relationships of mafic intrusions along the Protogine Zone, southern Sweden: Precambrian Research 48, pp. 395-414.

Johansson, L., A. Lindh, and C. Möller, 1991, Late Sveconorwegian (Grenville) high-pressure granulite facies metamorphism in southwest Sweden. Journal of Metamorphic Geology 9, p. 283-292.

Jensen, L.N. & Schmied, B.J. 1992. Late Tertiary uplift and erosion in the Skagerrak area; magnitude and consequences. Norsk Geologisk Tidsskrift 72 (3), pp. 275-279.

Jensen, L. N., Riis, F., Boyd, R. (eds.) 1992. Post-Cretaceous uplift and sedimentation along the western Fennoscandian shield. Norsk Geologisk Tidsskrift 72 (3).

Kakkuri, J. 1985. Die Landhebung in Fennoskandien im Lichte der heutigen Wissenscaft. Zeitschrift für Vermessungswesen 110 (2), pp. 51-59.

Kakkuri, J. & Chen, R. 1992. On horizontal crustal strain in Finland. Bulletin Géodésique 66 (1), pp. 12-20.
Kalla, J. 1960. Deep drillling in the Muhos formation at Liminka Tupos (in Finnish). Vuoriteollisuus 1, pp. 53-54.

Katzung, G., Giese, U., Walter, R. & Von Winterfeld, C. 1993. The Rügen Caledonides, northeast Germany. Geological Magazine 130, pp. 725-730.

Kesola, Reino 1981. Clastic dykes belonging to the the Muhos formation in basement rocks (in Finnish with an English abstract). Geologi 33 (9-10), pp. 133-135.

Kleesment, A. 1997. Devonian sedimentation basin. *In:* Raukas, A. & Teedumae, A. (eds.) Geology and mineral resources of Estonia. Tallinn: Estonian Academy Publishers, pp. 205-208.

Klingspor, I. 1976. Radiometric age-determination of basalts, dolerites and related syenite in Skåne, southern Sweden. Geologiska Föreningens i Stockholm Förhandlingar 98, pp. 195-216.

Knoll, A.H. & Walther, M.R. 1992. Latest Proterozoic stratigraphy and Earth history. Nature 356, pp. 673-678.

Knudsen, T.-L. & Andersen, T. 1999. Petrology and geochemistry of the Tromøy Gneiss Complex, South Norway, an Alleged Example of Proterozoic Depleted Lower Continental Crust. Journal of Petrology 40 (6), pp. 909-933.

Kohonen, J. & Vaarma, M. 2001. Sedimentary rocks in Finnish impact structures: pre-impact or post-impact? (in Finnish with an English summary). Geologi 53 (7), pp. 111-118.

Kohonen, J., Pihlaja, P., Kujala, H. & Marmo, J. 1993. Sedimentation of the Jotnian Satakunta sandstone, western Finland. Espoo: Geological Survey of Finland. Bulletin 369, 35 p.

Koistinen, T., Stephens, M. B., Bogatchev, V., Nordgulen, Ø., Wennerström, M., Korhonen, J. (comp.) 2001. Geological map of the Fennoscandian Shield, scale 1:2 000 000. Espoo : Trondheim : Uppsala : Moscow: Geological Survey of Finland : Geological Survey of Norway : Geological Survey of Sweden : Ministry of Natural Resources of Russia.

Korja, A. & Heikkinen, P. J. 1995. Proterozoic extensional tectonics of the central Fennoscandian Shield: results from the Baltic and Bothnian Echoes from the Lithosphere experiment. Tectonics 14 (2), pp. 504-517.

Korja, A., Heikkinen, P. & Aaro, S., 2001. Crustal structure of the northern Baltic Sea palaeorift. Tectonophysics 331 (4), pp. 341-358.

Kotilainen, A. 2001. A preliminary structural interpretation of the acoustic seismic data from the Lake Pyhäjärvi, SE-Finland. Unpublished report in Finnish. Geological Survey of Finland.

Kousa, J. & Lundqvist, T. 2000. Meso- and Neoproterozoic cover rocks of the Svecofennian Domain. *In:* Lundqvist, Th. & Autio, S. (eds.) Description to the bedrock map of central Fennoscandia (Mid-Norden). Espoo: Geological Survey of Finland. Special Paper 28, 75-76.

Kouvo, O. 1976. On the chronostratigaphy of the Finnish bedrock (in Finnish). Stratigraphy symposium 8.9.1976. Helsinki: Geological Society of Finland and Geological Union. Educational manifold N:o 2, pp. 1 - 13.

Kramm, U., Kogarko, L. N., Kononova, V. A. & Vartiainen, H. 1993. The Kola alkaline province of the CIS and Finland: precise Rb-Sr ages define 380-360 Ma age range for all magmatism. Lithos 30 (1), 33-44.

Kresten, P., Printzlau, I., Rex, D., Vartiainen, H. & Woolley, A. 1977. New ages of carbonatitic and alkaline ultramafic rocks from Sweden and Finland. Geologiska Föreningens i Stockholm Förhandlingar 99, pp. 62-65.

Kuivamäki, A. & Vuorela , P. 2001. Role of bedrock in disposal safety. Pp. 18-34 in: Rasilainen, K. (editor) 2001. Nuclear Waste Management in Finland. Final Report of the Public Sector's Research Programme JYT2001 (1997-2001). Ministry of Trade and Industry Finlad. Studies and Reports 15/2002.

Kuivamäki, A. Vuorela, P. & Paananen, M. 1998. Indications of postglacial and recent bedrock movements in Finland and Russian karelia. Geological Survey of Finland, Nuclear Waste Disposal Research. Report YST- 99.

Kujansuu, R. 1964. Nuorista siirroksista Lapissa. Summary: Recent faults in Lapland. Geologi 6, 30–36.

Kujansuu, R. & Uutela, A. 1997. Palaeozoic acritarchs in till-covered sand deposits at Kauhajoki, western Finland. *In:* Autio, S. (ed.) Geological Survey of Finland, Current Research 1995-1996. Espoo: Geological Survey of Finland. Special Paper 23, pp. 93-98.

Kukkonen, I. & Kuivamäki, A. 1985. Geologisia ja geofysikaalisia havaintoja Pasmajärven ja Suasseljän postglasiaalisista siirroksista. Abstract: Geological and geophysical observations of the Pasmajärvi and Suasselkä postglacial faults. Geological Survey of Finland, Nuclear Waste Disposal Research, Report YST-46. 14 p., 18 app. pages.

Kumpulainen, R. & Nystuen, J. P. 1985. Late Proterozoic basin evolution and sedimentation in the westernmost part of Baltoscandia. *In:* Gee, D. G. & Sturt, B.A. (eds.) The Caledonide Orogen; Scandinavia and related areas; Vol. 1. Chichester, United Kingdom: John Wiley & Sons, pp. 213-232.

Lagerbäck, R. 1979. Neotectonic structures in Northern Sweden. Geologiska Föreningens i Stockholm Förhandlingar 100, 263–269.

Laitakari, I. 1983. The Jotnian (upper Proterozoic) sandstone of Satakunta. *In:* Laajoki, K. & Paakkola, J. (eds.) Exogenic processes and related metallogeny in the Svecokarelian geosynclinal complex. Espoo; Geological Survey of Finland. Guide 11, pp. 135-139.

Larson, S. Å., 1996. The Gothian and Sveconorwegian terranes of SW Sweden. GFF 118 Jubilee Issue, p. A17.

Larson, S. Å., J. Berglund, J. Stigh, and E.-L. Tullborg, 1990, The Protogine Zone, soutwest Sweden: a new model-an old issue. *In:* Gower, C. F., Rivers, T. & Ryan, B. (eds.) Mid-Proterozoic Laurentia-Baltica. Geological Association of Canada Special Paper, pp. 317-333.

Larson, S.Å. & Tullborg, E-L. 1994. Tectonic regimes in the Baltic Shield during the last 1200 Ma - a review. Stockholm: Swedish Nuclear Fuel and Waste Management Co. Technical Report TR-94-05, 77 p.

Larson, S.Å. & Tullborg, E-L. 1998. Why Baltic Shield zircons yield late Paleozoic, lower-intercept ages on U-Pb concordia? Geology (Boulder) 26 (10), pp. 919-922.

Larson, S.Å., J. Stigh & E-L.Tullborg, 1986. The deformation history of the eastern part of the southwest Swedish gneiss belt. Precambrian Research 31, pp. 237-257.

Larson, S.Å., Tullborg, E.-L., Cederbom, C. & Stiberg, J.-P. 1999. Sveconorwegian and Caledonian foreland basins in the Baltic Shield revealed by fission-track thermochronology. Terra Nova 11 (5), pp. 210-215.

Lauerma, R. 1987. The diabase dykes in Salla (in Finnish with an English abstract). *In:* Aro, K. & I. Laitakari (eds.) Diabases and other mafic dyke rocks in Finland. Espoo: Geological Survey of Finland. Report of investigation 76, pp. 185-187.

Lehmuskoski, P. 1996. Active fault line search in southern and central Finland with precise levellings. Reports of the Finnish Geodetic Institute 96:5. 16 p.

Lehtovaara, J.J. 1982a. Palaeozoic sedimentary rocks in Finland. Annales Academiae Scientiarum Fennicae. Series A. III. Geologica - Geographica 133, 35 p.

Lehtovaara, J. J. 1982b. Stratigraphical section through Lower Cambrian at Söderfjärden, Vaasa, western Finland. Bulletin of the Geological Society of Finland 54 (1-2), pp. 35-43.

Lehtovaara, J. J. 1984. Söderfjärden, Vaasa, western Finland : a crater covered with sedimentary material since its formation in the Cambrian (in Finnish with an English summary). Terra 96 (1), pp. 23-33.

Lehtovaara, J. J. 1988. The palaeosedimentology of the autochthon of the Finnish Caledonides. *In:* Laajoki, K. & Paakkola, J. (eds.) Sedimentology of the Precambrian formations in eastern and northern Finland: proceedings of IGCP 160 Symposium at Oulu, January 21-22, 1986. Espoo: Geological Survey of Finland. Special Paper 5, pp. 255-264.

Lehtovaara, J. J. 1992. Söderfjärden : a Cambrian impact crater in western Finland. *In:* Pesonen, L. J. & Henkel, H. (eds.) Terrestrial impact craters and craterform structures with a special focus on Fennoscandia: papers from a symposium held in Espoo and Lappajärvi, Finland, May 29-31, 1990. Tectonophysics 216 (1-2), pp. 157-161.

Lehtovaara, J. J. 1995. Pre-Quaternary rocks of the Kilpisjärvi and Halti map-sheet areas (in Finnish with an English summary).Geological map of Finland 1:100 000. Explanation to the maps of Pre-Quaternary rocks. Sheets 1823 ja 1842. Espoo: Geological Survey of Finland, 64 p.

Lehtovaara, J. J. & Tynni, R. 1983. Fossiliferous boulders of Lower Cambrian phosphoritic sandstone in south-western Finland. Bulletin of the Geological Society of Finland 55 (2), pp. 85-99.

Lindh, A. & Persson, P.-O. 1990. Proterozoic granitoid rocks of the Baltic Shield - trends of development. *In:* Gower, C.F., Rivers, T. & Ryann A.B. (eds.) Mid-proterozoic Laurentia-Baltica. Geological Association of Canada Special Paper 38, pp. 23-40.

Lidmar-Bergström, K. 1991. Phanerozoic tectonics in southern Sweden. Z. Geomorph. N.F. 82, pp. 1-16.

Lidmar-Bergström, K. 1996. Long term morphotectonic evolution in Sweden. Geomorphology 16, pp. 33-59.

Lidmar-Bergström, K. 1999. Uplift histories revealed by landforms of the Scandinavian domes. *In:* Smith, B.J., Whalley, W. B. & Warke, P.A. (eds.) Uplift, erosion and stability : perspectives on long-term landscape development. Geological Society of London Special publication 162, pp. 85-91.

Lidmar-Bergström, K. & Näslund, J.O. 2000. Landforms and uplift in Scandinavia. European Geophysical Union, 25th General Assemply, Nice, France, 2000. Programme and abstracts (http://www.copernicus.org/EGS/egsga/nice00/programme/abstracts/aac8130.pdf)

Lukashov, A. D. 1995. Paleoseismotectonics in the northern part of Lake Onega (Zaonezhskij Peninsula, Russsian Karelia). Geological Survey of Finland. Nuclear Waste Disposal Research, Report YST-90, 36 p.

Lundqvist, T., Bøe, R., Kousa, J., Lukkarinen, H., Lutro, O., Roberts, D., Solli, A., Stephens, M. & Weihed, P. 1996. Bedrock map of Central Fennoscandia. Scale 1 : 1 000 000. Espoo : Trondheim : Uppsala: Geological Survey of Finland : Geological Survey of Norway : Geological Survey of Sweden.

Löfgren, A. 1985. Early Ordovician conodont biozonation at Finngrundet, south Bothnian Bay, Sweden. Bulletin of the Geological Institutions of the University of Uppsala, New Series 10, pp. 135-142.

Männil, R. 1966. Evolution of the Baltic basin during Ordovician (in Russian with English summary). Tallin, Estonia: NSV Teaduste Akadeemia Geoloogia Instituut, 199 p.

Männil, R. & Meidla, T. 1994. The Ordovician System of the East European Platform (Estonia, Latvia, Lithuania, Byelorussia, parts of Russia, the Ukraine and Moldova). *In:* Webby, B.D., Ross, R..J. Jr. & Zhen, Y.Y. (eds.) The Ordovician system of the European Platform and Tuva (south-eastern Russia); correlation charts and explanatory notes. International Union of Geological Sciences Publication 28, pp. 1-52.

Månsson, A.G.M., 1996. Brittle reactivation of ductile basement structures; a tectonic model for the Lake Vättern basin, southern Sweden. GFF 118 Jubilee Issue, p. A19.

Mänttäri, I. & Koivisto, M. 2001. Ion microprobe uranium-lead dating of zircons from the Lappajärvi impact crater, western Finland. Meteoritics & Planetary Science 36 (8), pp. 1087-1095.

Marek, R. 2000. Palaeozoic structures at the margin of the Baltic Shield revealed by new and reprocessed marine reflection seismic data from Kattegat, south-west Scandinavia. Tectonophysics 327, pp. 293-309.

Martinsson, A. 1968. Cambrian palaentology of Fennoscandian basement fissures. Lethaia 1, pp- 137-155.

Martinsson, A. 1974. The Cambrian in Norden. *In:* Holland, C. H. (ed.) 1974. Cambrian of the British Isles, Norden and Spitsbergen. London: John Wiley & Sons, pp. 185-284.

Marttila, E. 1969. On the sedimentation of the Satakunta sandstone. Helsinki: University of Helsinki , Department of geology. Academic dissertation, 157 p.

Matisto, A. 1964. Onko Tyrvään Vaunujoella hiekkakiveä? Is there sandstone at Vaunujoki in Tyrvää? (in Finnish). Geologi 16 (10), pp. 153-154.

Meert, J.G., Torsvik, T.H., Eide, E.A. & Dahlgren, S. 1998. Tectonic significance of the Fen Province, S. Norway: Constraints from Geochronologogy and Palaeomagnetism. Journal of Geology 106, pp. 553-564.

Mens, K. A. & Pirrus, E. A. 1986. Stratigraphical characteristics and development of Vendian-Cambrian boundary beds on the East European Platform. Geological Magazine 123, pp. 357-360.

Mens, K., Bergström, J. & Lendzion 1987. The Cambrian System on the East European Platform. Correlation Chart and Explanotory Notes. Tallinn: Valgus, 119 p.

Merrill, G. K. 1980. Ordovician conodonts from the Åland islands, Finland. Geologiska Föreningens i Stockholm Förhandlingar 101 (4), pp. 329-341.

Mertanen, S. & Pesonen, L. J. 1997. Paleomagnetic evidence for the drift of the Fennoscandian Shield. *In:* Pesonen, L. J. (ed.) The lithosphere in Finland - a geophysical perspective. Geophysica 33 (1), pp. 81-98.

Mertanen, S., Pesonen, L. J., Elming, S.-Å. & Buchan, K. L. 1996. Palaeomagnetic evidence of drift of the Fennoscandian (Baltic) Shield and reconstructions of Fennoscandia and Laurentia during the Proterozoic. GFF 118 Jubilee Issue, A20-A21.

Mikkola, E. 1932. On the physiography and late-glacial deposits in northern Lapland. Bulletin de la Commission géologique de Finlande 96, 88 p.

Milne, G. A., Davis, J.L., Mitrovica, J.X., Scherneck, H.-G., Johansson, J.M., Vermeer, M., Koivula, H., 2001. Space-geodetic constraints on glacial isostatic adjustement in Fennoscandia. Science, Vol. 291, 23 March 2001, pp 2381–2385.

Milnes, A. G., Wennberg, O. P., Skår, O. & Köstler, A. G. (1997). Contraction, extension and timing in the South Norwegian Caledonides: The Sognefjord transect. *In:* Burg, J.-P. & Ford, M. (eds.) Orogeny Through Time. Geological Society of London Special Publication 121, pp. 123-148.

Milnes, A.G., Gee, D.G. & Lund, C.-E., 1998. Crustal structure and regional tectonics of SE Sweden and the Baltic Sea. Svensk Kärnbränslehantering AB, Technical Report SKB TR 98-21, 63 p.

Mogensen, T.E. 1994. Palaeozoic structural development along the Tornquist Zone, Kattegat area, Denmark. Tectonophysics 240 (1-4), pp. 191-214.

Mogensen, T.E. 1995. Triassic and Jurassic structural development along the Tornquist Zone, Denmark. Tectonophysics 252 (1-4), 197-220.

Möller, C. & Söderlund, U. 1997, Age constraints on the regional deformation within the Eastern Segment, S. Sweden: Late Sveconorwegian granite dyke intrusion and metamorphic-deformational relations: GFF 119, pp. 1-12.

Muir Wood, R. 1989. Extraordinary deglaciation reverse faulting in northern Fennoscandia. In: S. Gregersen and P.W. Basham (eds) 1989. Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound. Proceedings of the NATO Advanced Research Workshop on Causes and Effects of Earthquakes at Passive Margins and in Areas of Postglacial Rebound on Both Sides of the Atlantic, Vordingborg, Denmark , May 9–13, 1988. Dordrecht: NATO ASI Series C, Mathematical and Physical Sciences 266, 141–173.

Muir Wood, R. 1995. Reconstructing the tectonic history of Fennoscandia from its margins: The past 100 million years. Svensk Kärnbränslehantering AB, Techn. Rep. SKB TR 95-36, 85 p.

Murrel, G.R. & Andriessen, P.A.M. 2000. Longterm morphotectonic evolution in Finland; constraints on quantification and timing of vertical motions from fission track thermochronology. Fission track 2000; 9th international conference on fission track dating and thermochronology. Abstracts - Geological Society of Australia 58, pp. 233-234.

Mörner, N. A. 2001. The Boda Cave and its surroundings. The 9663 BP paleoseismic event [online]. [cited 31.8.2002]. Portable Document Format. Available from:< htpp://www.pog.su.se/04public/public.htm

Nenonen, J. & Huhta, P. 1993. Quaternary glacial history of the Hattu schist belt and adjacent parts of the Ilomantsi district, eastern Finland. In: Nurmi, P. & Sorjonen-Ward, P. (eds.) 1993. Geological development, gold mineralization and exploration methods in the late Archean Hattu schist belt, Ilomantsi, eastern Finland. Geological Survey of Finland, Special Paper 17, 185–191.

Nestor, H. & Einasto, R. 1997. Ordovician and Silurian carbonate sedimentation basin. *In:* Raukas, Anto & Teedumae, A. (eds.) Geology and mineral resources of Estonia. Tallinn: Estonian Academy Publishers, pp. 192-204.

Neumann, E. R., Olsen, K. H., Baldridge, W. S. & Sundvoll, B. 1992. The Oslo Rift; a review. Tectonophysics 208 (1-3), pp. 1-18.

Niemelä, J., Ikonen, L. & Kinnunen, K. 1985. Pebbles of Cambro-Ordovician sedimentary rocks in potholes of Kattiluoto, SW-Finland (in Finnish with an English abstract). Geologi 37 (9-10), pp. 169-175.

Nielsen, A.T. 1992. Intercontinental correlation of the Arenigian (Early Ordovician) based on sequence and ecostartigraphy. *In:* Webby, B.D. & Laurie, J.R. (eds.) Global

Perspectives on Ordovician Geology. Proceedings of the Sixth Internation Symposium on the Ordovician System. Rotterdam: Balkema, pp. 367-379.

Nikishin, A.M., Ziegler, P.A., Stephenson, R.A., Cloetingh, S.A.P.L., Furne, A.V., Fokin, A., Ershov, A.V., Bolotov, S.N., Korotaev, M.V., Alekseev, A.S., Gorbachev, V.I., Shipilov, E.V., Lankreijer, A., Bembinova, E.Y. & Shalimov, I.V. 1996. Late Precambrian to Triassic history of the East European Craton: dynamics of sedimentary basin evolution. Tectonophysics 268, pp. 23-63.

Nőlvak, J. 1997. The Ordovician in Estonia. *In:* Raukas, A. & Teedumae, A. (eds.) Geology and mineral resources of Estonia. Tallin: Estonian Academy Publishers, pp. 54-55.

Nõlvak, J., Meidla, T. & Uutela, A. 1995. Microfossils in the Ordovician erratic boulders from south-western Finland. Bulletin of the Geological Society of Finland 67 (2), pp. 3-26.

Norling, E. & Bergström, J. 1987. Mesozoic and Cenozoic tectonic evolution of Scania, southern Sweden. Tectonophysics 137 (1-4), pp. 7-19.

Norton, M.G. 1986. Late Caledonide extension in western Norway: A response to extreme crustal thickening. Tectonics 5, pp. 195-204.

Nyström, J.O. 1983. Pumpellyite-bearing rocks in cenral Sweden and extent of host rock alterations as a control of pumpellyite composition. Contributions to Mineralogy and Petrology 83, pp. 159-163.

Nyström, J.O. & Levi, B. 1980. Pumpellyite-bearing Precambrian rocks and post-Svecokarelian regional metamorphism in central Sweden. Geologiska Föreningens i Stockholm Förhandlingar 102, pp. 37-39.

Obst, K. & Solyom Z. 2000. A Permo-Carboniferous mafic dyke swarm and lithospheric extension in southernmost Sweden. Abstracts for oral and poster presentations for the "Geodynamis controls on Permo-Carboniferous rifting and magmatism in NW Europe" symposium during Geoscience2000 conference in Manchester University, Great Britain, 17-20 April 2000 (http://earth.leeds.ac.uk/whatsnew/Geoscience2000%abstracts.htm).

Ödman, O.H. 1957. Description to Map of the Pre-Cambrian Rocks of the Norbotten county, ecl. the Caledonian Mountain Range (in Swedish with an English summary). Sveriges Geologiska Undersökning Ca 41, 151 p.

Öhman, T., Pesonen, Lauri J., Raitala, J., Uutela, A. & Tuisku, P. 2000. The Saarijarvi Crater; older and larger than assumed? *In:* Plado, J. & Pesonen, L.J. (eds.) Meteorite impacts in Precambrian shields. 4th ESF Workshop, Lappajärvi-Karikkoselkä-Sääksjärvi, Finland, May 24-28, 2000. Espoo : Helsinki: Geological Survey of Finland : University of Helsinki. Programme and abstracts, p. 82.

Olesen, O. 1984. The Masi fault, evidence of neotectonics in the Precambrian of Finnmark, northern Norway. In: Binzer, K., Marcussen , I. & Konradi, P. (eds.) 1988. 18. Nordiske Geologiske Vintermøde, København 1988. Abstracts. Copenhagen: Danmarks Geologiske Undersøgelse, 332–333.

Olesen, O., Dehls, J., Bungum, H., Riis, F., Hicks, E., Lindholm, C., Blikra, L. H., Fjelskaar, W., Olsen, L., Longva, O., Faleide, J. I., Bockmann, L., Rise, L., Roberts, D., Braathen, A. and Brekke, H. 2000. Neotectonics in Norway, Final Report. NGU Report 2000.002.

Öpik, A.A. 1956. Cambrian (Lower Cambrian) of Estonia. El sistema Cámbricio, su paleogeografía y el problema de su base. Symbosium I, Part I: Europa, Africa, Asia. XX Congreso Geólogico Internacional, XX Sesión, México, pp. 97-126.

Paananen, M. 1987. Venejärven, Ruostejärven, Suasseljän ja Pasmajärven postglasiaalisten siirrosten geofysikaalinen tutkimus. Abstract: Geophysical studies of the Venejärvi, Ruostejärvi, Suasselkä and Pasmajärvi postglacial faults in northern Finland. Geological Survey of Finland, Nuclear Waste Disposal Research, Report YST-59. 97 p., 45 app. pages.

Paananen, M. 1989. Sähköiset luotaukset, geofysikaaliset reikämittaukset ja hydrauliset testit Pasmajärven postglasiaalisella siirroksella. Abstract: Resistivity soundings, geophysical borehole measurements and hydraulic tests at Pasmajärvi postglacial fault. Geological Survey of Finland. Nuclear Waste Disposal Research, Report YST-69. 25 p., 28 app. pages.

Page, L., Möller, C. & Johansson, L. 1996. Ar/Ar geochronology across the Mylonite Zone and the South-western Granulite Province in the Sveconorwegian Orogen of S. Sweden. Precambrian Research 79, pp. 239-259.

Park, R. G., Åhäll, K.-I. & Boland, M.P. 1991, The Sveconorwegian shear-zone network of SW Sweden in relation ot Mid-Proterozoic plate movements. Precambrian Research 49, pp. 245-260.

Paulamäki, S., Paananen, M, & Elo, S. 2002. Structure and geological evolution of the bedrock of southern Satakunta, SW Finland. Helsinki: Posiva Oy. Report Posiva 2002-04**,** 125 p.

Pegrum, R.M. 1984. The extension of the Tornquist Zone in the Norwegian North Sea. Norsk Geologisk Tidsskrift 64 (1), pp. 39-68.

Pekkala, Y. & Sarapää, O. 1989. Kaolin exploration in Finland. *In:* Autio, S. (ed.) Geological Survey of Finland. Current Research 1988. Espoo: Geological Survey of Finland. Special Paper 10, pp. 113-118.

Peltonen, P., Huhma, H., Tyni, M. & Shimizu, N. 1999. Garnet peridotite xenoliths from kimberlites of Finland: Nature of the continental mantle at an Archaean Craton – Proterozoic mobile belt transition. *In*: Guerney, J.J., Guerney, J.L., Pascoe, M.D. & Richardson, S.H. (eds.) Proceedings of the 7th International Kimberlite Conference, Vol. 2, pp. 664-676.

Pesonen, L. J., Abels, A., Lehtinen, M. & Plado, J. 2000. Meteorite impact cratering - implications for the Fennoscandian lithosphere. *In:* Pesonen, L. J., Korja, A. & Hjelt, S.-E. (eds.) Lithosphere 2000 : a symposium on the structure, composition and evolution of the lithosphere in Finland, Espoo, Otaniemi, October 4-5, 2000: programme and extended abstracts. Institute of Seismology. University of Helsinki. Report S-41, pp. 113-119.

Pesonen, L. J., Mertanen, S. & Elming, S.-Å. 2001. Reconstructions of continents during the Proterozoic - a way towards Rodinia. *In:* Sircombe, K. N. & Li, Z. X. (eds.) From basins to mountains: Rodinia at the turn of the century : Chris Powell memorial symposium, Perth, Australia, 30 September - 2 October 2001. Geological Society of Australia. Abstracts 65, pp. 82-84.

Peulvast, J.P. 1985. Postorogenic morphotectonic of the Scandinavian Caledonides during the Mesozoic and Cenozoic *In:* Gee, D. G. & Sturt, B.A. (eds.) The Caledonide Orogen; Scandinavia and related areas; Vol. 1. Chichester, United Kingdom: John Wiley & Sons, pp. 979-995.

Pipping, F. 2000. Impact craters. *In:* Lundqvist, Th. & Autio, S. (eds.) Description to the bedrock map of central Fennoscandia (Mid-Norden). Espoo: Geological Survey of Finland. Special Paper 28, pp. 108-114.

Plink-Björklund, P. & Björklund 1999. Sedimentary response in the Baltic Devonian Basin to post-collisional events in the Scandinavian Caledonides. GFF 121, pp. 79-80.

Priem, H. N. A., Mulder, F. G., Boelrijk, N. A. I. M., Hebeda, E. H., Verschure, R. H. & Verdurmen, E. A. Th. 1968. Geochronological and palaeomagnetic reconnaissance survey in parts of central and southern Sweden. Physics of the Earth and Planetary Interiors 1 (6), pp. 373-380.

Poprawa, P, Sliaupa, S., Stephenson, R. & Lazauskiene, J. 1999. Late Vendian-early Palaeozoic tectonic evolution of the Baltic Basin; regional tectonic implications from subsidence analysis. Tectonophysics 314 (1-3), pp. 219-239.

Puura, V., Amantov, A., Tikhomirov, S., Laitakari, I., 1996. Latest events affecting the Precambrian basement, Gulf of Finland and surrounding areas. *In:* Koistinen, T. (ed.) Explanation to the Map of Precambrian Basement of the Gulf of Finland and Surrounding Areas, 1:1 mill. Espoo: Geological Survey of Finland. Special Paper 21, pp. 115-125.

Puura, V., Vaher, R. & Miidel, A. 1997. Tectonics. *In:* Raukas, A. & Teedumae, A. (eds.) Geology and mineral resources of Estonia. Tallinn: Estonian Academy Publishers, pp. 163-180.

Puura, V., Vaher, R. & Tuuling 1999. Pre-Devonian landscape of the Baltic Oil-Shale Basin, NW of the Russian Platform. *In:* Smith, B.J., Whalley, W. B. & Warke, P.A. (eds.) Uplift, erosion and stability: perspectives on long-term landscape development. Geological Society of London Special Publication 162, pp 75-83.

Ramberg, I.B. & Larsen, B.T. 1978. Tectonomagmatic evolution. *In:* Dons, J. A., Larsen, B. T. (eds.) The Oslo Paleorift; a review and guide to excursions. Norges Geologiske Undersökelse 45, Nr. 337, pp. 55-73.

Riis, F. 1996. Quantification of Cenozoic vertical movements of Scandinavia by correlation of morphological surfaces with offshore data. Global and Planetary Change 12, pp. 331-357.

Risku-Norja, H. 1992. Geochemistry of the dolerite dykes in Södermanland, eastern central Sweden. Geologiska Föreningens i Stockholm Förhandlingar 114 (1), pp. 67-91.

Ro, H. E., Larsson, F. R., Kinck, J. J. & Husebye, E. S. 1990. The Oslo Rift; its evolution on the basis of geological and geophysical observations. Tectonophysics 178 (1), pp. 11-28.

Rohde, A. 1987. Depositional environments and lithostratigraphy of the Middle Proterozoic Almesåkra Group, southern Sweden. Sveriges Geologiska Undersökning Ser. Ca, NR 69, 80 p.

Rohrman, M. & van der Beek, P.A. 1996. Cenozoic post-rift domal uplift of North Atlantic margins; An asthenospheric diapirism model. Geology (Boulder) 24, pp. 901-904.

Romer, R.L. 1996. Contiguous Laurentia and Baltica before the Grenvillian-Sveconorwegian orogeny? Terra Nova 8, pp. 173-181.

Romer, R.L. & Smeds, S.-A. 1996. U-Pb columbite ages of pegmatites from Sveconorwegian terranes in south-western Sweden. Precambrian Research 76, pp. 15-30.

Rõõmusoks, A. 1960. Stratigraphy and paleogeography of the Ordovician in Estonia. International Geological Congress, Report of the Twenty-First Session Norden, Part VII, Proceedings of Section 7: Ordovician and Silurian stratigraphy and correlations, pp. 58.69.

Saadre, T. 1992. Distribution pattern of the Ordovician discontinuity surfaces, East Baltic region. Bulletin of the Geological Survey of Estonia 2 (1), pp. 16-26.

Saari, J. 1992. A review of the seismotectonics of Finland. Helsinki: Nuclear Waste Commission of Finnish Power Companies. Report YJT-92-29, 76 p.

Samulsson, J. & Middleton, M. 1998. The Caledonian foreland basin in Scandinavia: constrained by thermal maturation of the Alum Shale. GFF 120, pp. 307-314.

Sarapää, O. 1996. Proterozoic primary kaolin deposits at Virtasalmi, south-eastern Finland. Espoo: Geological Survey of Finland, 152 p.

Sauramo, M. 1916. Über das Vorkommen von Sandstein in Karstula, Finland (in German). Fennia 13 (7), 13 p.

Séranne, M. 1992. Late Palaeozoic kinematics of the Møre-Trøndelag Fault Zone and adjacent areas, central Norway. Norsk Geologisk Tidskrift 72, pp. 141-158.

Simonen, Ahti 1960. Pre-Quarternary rocks in Finland. Bulletin de la Commission géologique de Finlande 191, pp. 1-49.

Simonen, A. & Kouvo, O. 1955. Sandstones in Finland. Comptes Rendues de la Societe géologique de Finlande 28, pp. 57-87; also Bulletin de la Commission géologique de Finlande 168.

Skordas, E., Mayer, K. Olsson, R. & Kulhanek, K. 1991. Causality between interplate (North Atlantic) and intraplate (Fennoscandia) seismicities. Tectonophysics 185 (3-4), pp. 295-307.

Slunga, R.S. 1989. Focal mechanism and crust stresses in the Baltic Shield. NATO ASI Series.Series C: Mathematical and Physical Sciences 266.

Söderberg, P. 1993. Seismic stratigraphy, tectonics and gas migration in the Åland Sea, northern Baltic proper. Stockholm Contributions in Geology 43 (1), pp. 1-67.

Söderberg, P. & Hagenfeldt, S.E. 1995. Upper Proterozoic and Ordovician submarine outliers in the archipelago northeast of Stockholm, Sweden. GFF 117, pp. 153-161.

Söderlund, U. 1999. Geochronology of Tectonothermal Events in the Paraautochthonous Eastern segment of the Sveconorwegian (Grenvillian) Orogen, South-western Sweden. Lund University, Faculty of Science. Doctoral dissertation, 142 p.

Solyom, Z., Lindqvist, J. E. & Johansson, I. 1992. The geochemistry, genesis, and geotectonic setting of Proterozoic mafic dyke swarms in southern and central Sweden. Geologiska Föreningen i Stockholm Förhandlingar 114 (1), pp. 47-65.

Starmer, I.C. 1991. The Proterozoic evolution of the Bamble-sector shear belt, southern Norway: correlations across southern Scandinavia and the Grenville controversy. Precambrian Research 49, pp. 107-139.

Starmer, I. C. 1993, The Sveconorwegian Orogeny in southern Norway, relative to deep crustal structures and events in the North Atlantic Proterozoic Supercontinent. Norsk Geologisk Tidsskrift 73, pp. 109-132.

Stephansson, O. 1988. Ridge push and glacial rebound as rock stress generators in Fennoscandia. Bull. Geol. Inst. Univ. Uppsala N.S. 14, pp. 39-48.

Stephansson, O., Dahlström, L.-O., Bergström, K., Särkkä, P., Väätäinen, A., Myrvang, A., Fjeld, O. K. & Hanssen, T. H. 1987. Fennoscandian rock stress data base - FRSDB. Luleå University Research Report TULEA 1987 06, 35 p.

Stephens, M.B. 1988. The Scandinavian Caledonides: a complexity of collisions. Geology Today, Jan-Feb 1988, pp. 20-25.

Stephens, M.B. & Gee, D.G. 1989. Terranes and polyphase accretionary history in the Scandinavian Caledonides. *In:* Dallmeyer, R. D. (ed.) Terranes in the Circum-Atlantic Paleozoic orogens. Geological Society of America Special Paper 230, pp. 17-30.

Stephens, M. B., C.-H. Wahlgren, R. Weijermars, and A. R. Cruden, 1996, Left-lateral transpressive deformation and its tectonic implications, Sveconorwegian orogen, Baltic Shield, south-western Sweden. Precambrian Research 79, pp. 261-279.

Stuevold, L-M. & Eldholm, O. 1996. Cenozoic uplift of Fennoscandia inferred from a study of the mid-Norwegian margin. Global and Planetary Change 12 (1-4), pp. 359-386.

Sturt, B.A. & Ramsay, D.M. 1999. Early Ordovician terrane-linkages between oceanic and continental terranes in central Caledonides. Terra Nova 11, pp. 79-85.

Sundvoll, B., Neumann, E. R., Larsen, B. T., Tuen, E. 1990. Age relations among Oslo Rift magmatic rocks; implications for tectonic and magmatic modelling. Tectonophysics 178 (1), pp. 67-87.

Suominen, V. 1991. The chronostratigraphy of south-western Finland with special reference to Postjotnian and Subjotnian diabases. Espoo: Geological Survey of Finland. Bulletin 356, 100 p.

Svenningsen, O.M. 1994. The Baltica-Iapetus passive margin dyke complex in the Sarektjåkkå Nappe, northern Swedish Caledonides. Geological Journal 29, pp. 323-354 .

Svenningsen, O.M. 1995. Extensional deformation along the Late Precambrian-Cambrian Baltoscandian passive margin: the Sarektjåkkå Nappe, Swedish Caledonides. Geologische Rundschau 84, pp. 649-664.

Svenningsen, O.M. 2001. Onset of seafloor preading in the Iapetus Ocean at 608 Ma: precise age of the Sarek Dyke swarm, northern Swedish Caledonides. Precambrian Research 110, pp. 241-254.

Swensson, E. 1990. Cataclastic rocks along the Nesodden Fault, Oslo region, Norway; a reactivated Precambrian shear zone. Tectonophysics 178 (1), pp. 51-65.

Tait, J., Schaetz, M., Bachtadse, V. & Soffel, H. 2000. Palaeomagnetism and Palaeozoic palaeogeography of Gondwana and European terranes. *In:* Franke, W., Haak, V., Oncken, O. & Tanner, D. Orogenic processes; quantification and modelling in the Variscan Belt. London: Geological Society Special Publications 179, pp. 21-34.

Talbot, C. & Slunga, R. 1989. Patterns of active shear in Fennoscandia. Dondrecht/Boston/London: Kluwer Academic Publishers, pp. 441-446.

Talvitie, Jouko 1979. Remote sensing and geobotanical prospecting in Finland. Bulletin of the Geological Society of Finland 51 (1-2), 63-73.

Tanner, V. 1911. Über eine Gangformation von fossilführenden Sandstein auf der Halbinsel Långbergsödaöjoen im Kirchspiel Saltvik, Åland Inseln (in German). Bulletin de la Commission géologique de Finlande 25, 13 p.

Thorslund, P. 1960. Cambro-Silurian. *In:* Magnusson, Nils H., Brotzen, F., Kulling, O. & Thorslund, P. (eds.) Description to accompany the map of the pre-Quaternary rocks of Sweden. Sveriges Geologiska Undersökning, Serie Ba 16, pp. 69-110.

Thorslund, P. & Axberg, S. 1979. Geology of the southern Bothnian Sea; Part I. Bulletin of the Geological Institutions of the University of Uppsala, New Series 8, pp. 35-62.

Tirén, S.A. & Beckholmen, M. 1992. Rock block map analysis of southern Sweden. Geologiska Föreningens i Stockholm Förhandlingar 114 (3), pp. 253-269.

Tjernvik, T.E. & Johansson, J.V. 1979. Description of the upper portion of the drill-core from Finngrundet in the South Bothnian bay. Bulletin of the Geological Institutions of the University of Uppsala, New Series 8, pp. 173-204.

Torsvik, T.H. 1998. Palaeozoic palaeogeography: A North Atlantic viewpoint. GFF 120, pp. 109-118.

Torsvik, T.H. & Rehnström, E.F. 2001. Cambrian palaeomagnetic data from Baltica; implications for true polar wander and Cambrian palaeogeography. Journal of the Geological Society of London 158 (2), pp. 321-329.

Torsvik, T.H., Trench, A., Svensson, I. & Walderhaug, H.J. 1993. Palaeogeographic significance of mid-Silurian palaeomagnetic results from southern Britain - major revision of the apparent polar wander path for eastern Avalonia. Geophysical Journal International 113, pp. 651-668.

Torsvik, T.H., Lohmann, K.C. & Sturt, B.A. 1995. Vendian glaciations and their relation to the dispersal of Rodinia; paleomagnetic constraints. Geology (Boulder) 23 (8), pp. 727-730.

Torsvik, T.H., Smethurst, M. A., Meert, J. G., Van der Voo, R., McKerrow, W. S., Brasier, M..D., Sturt, B.A. & Walderhaug, H.J. 1996. Continental break-up and collision in the Neoproterozoic and Palaeozoic; a tale of Baltica and Laurentia. Earth-Science Reviews 40 (3-4), pp. 229-258.

Tullborg, E-L. 1997. Recognition of low-temperature processes in the Fennoscandian Shield. Publ. -Earth Sciences Centre A17, 33 p.

Tullborg, E.-L., Larsson, S. Å., Björklund, L., Samuelsson, L. & Stigh, J. 1995 Thermal evidence of Caledonide foreland, molasse sedimentation in Fennoscandia. Stockholm: Swedish Nuclear Fuel and Waste Management Co. Technical Report TR 95-18, 38 p.

Tullborg, E-L., Larson S.Å. & Stiberg, J-P. 1996. Subsidence and uplift of the present surface in the south-eastern part of the Fennoscandian Shield. GFF 118, pp. 126-128.

Tynni, R. 1965. Lateglacial faults in the bedrock of Otaniemi (in Finnish). Geologi 7, pp. 97-98.

Tynni, R., 1966. Über spät- und postglaziale Uferverschiebung in der Gegend von Askola, Südfinnland. Bulletin de la Commission géologique de Finlande 223. 97 p.

Tynni, R. 1974. Microfossils in a specimen of Cambrian (?) sandstone from Karstula, Central Finland. Bulletin of the Geological Society of Finland 46 (1), pp. 9-13.

Tynni, R. 1975. Ordovician hystrichospheres and chitinozoans in limestone from the Bothnian Sea. Espoo: Geological Survey of Finland. Bulletin 279, 59 p.

Tynni, R. 1978a. Microfossils of the Muhos formation (in Finnish with an English abstract). Espoo: Geological Survey of Finland. Report of Investigation 30, 18 p.

Tynni, R. 1978b. Lower Cambrian fossils and acritarchs in the sedimentary rocks of Söderfjärden, western Finland. *In:* On the geology and the Cambrian sediments of the circular depression at Söderfjärden, western Finland. Espoo: Geological Survey of Finland. Bulletin 297, pp. 39-63.

Tynni, R. 1982a. New results of studies on the fossils in the Lower Cambrian sediment deposits of the Söderfjärden basin. Bulletin of the Geological Society of Finland 54 (1-2), pp. 57-68.

Tynni, R. 1982b. On Paleozoic microfossils in clastic dykes in the Åland Islands and in the core samples of Lumparn. Espoo: Geological Survey of Finland. Bulletin 317, pp. 35-114.

Tynni R., 1982c. The reflection of geological evolution in Tertiary and interglacial diatoms and silicoflagellates in Finnish Lapland. Geological Survey of Finland. Bulletin 320, 40 p.

Tynni, R. & Donner, J. 1980. A microfossil and sedimentation study of the Late Precambrian formation of Hailuoto, Finland. Espoo: Geological Survey of Finland. Bulletin 311, 27 p.

Tynni, R. & Hokkanen, K. 1982. Traces of crawling by annelids in Lauhanvuori sandstone (in Finnish with an English summary). Geologi 34 (7), pp. 129-134.

Tynni, R. & Uutela, A. 1984. Microfossils from the Precambrian Muhos formation in Western Finland. Espoo: Geological Survey of Finland. Bulletin 330, 38 p.

Tynni, R. & Uutela, A. 1985. Late Precambrian shale formation of Taivalkoski in northern Finland (in Finnish with an English summary). Geologi 37 (4-5), pp. 61-65.

Uutela, A. 1986. A microfossil study of the sandstone of the Precambrian Ryydys (in Finnish with an English summary). Geologi 38 (8), pp. 206-210.

Uutela, A. 1989. Age and dispersal of sedimentary erratics on the coast of south-western Finland. Espoo: Geological Survey of Finland. Bulletin 349, 100 p.

Uutela, A. 1990. Proterozoic microfossils from the sedimentary rocks of the Lappajärvi impact crater. Bulletin of the Geological Society of Finland 62 (2), pp. 115-120.

Uutela, A. 1998. Extent of the northern Baltic Sea during the Early Palaeozoic era - new evidence from Ostrobothnia, western Finland. Bulletin of the Geological Society of Finland 70 (1-2), pp. 51-68.

Uutela, A. 2001. Proterozoic and early Palaeozoic microfossils in the Karikkoselkä impact crater, central Finland. Bulletin of the Geological Society of Finland 73 (1-2), pp. 75-85.

Uutela, A. & Sarjeant, W. A. S. 2000. The Ordovician acritarch genera Tranvikium and Ampullula: their relationship and taxonomy. Review of Palaeobotany and Palynology 112 (1-3), pp. 23-38.

Vaarma, M. & Pipping, F. 1997. Pre-Quaternary rocks of the Alajärvi and Evijärvi map-sheet areas (in Finnish with an English summary). Geological maps of Finland 1:100 000. Explanation to the maps of Pre-Quaternary rocks. Sheets 2313 ja 2314. Espoo: Geological Survey of Finland, 83 p.

Vaasjoki, M. (1996) . Pb isotope composition of galena in fracture infilling at Palmottu (in Finnish). Espoo: Geological Survey of Finland. Unpublished report 3.1.1996, 3 p.

Veltheim, V. 1969. On the pre-Quaternary geology of the Bothnian Bay area in the Baltic Sea. Bulletin Commission géologique de Finlande 239, 56 p.

Veriö, A. 1992. Geodetic observations connected with an earthquake in the Fennoscandian Shield. Surveying Science in Finland 10 (1), 70-83.

Veriö, A., Kuivamäki, A. & Vuorela, P. 1993. Kallioperän murroslinjojen nykyliikunnoista. Maanmittauslaitoksen murroslinjavaaitukset 1974–1992. Summary: Recent displacements within reactivated fracture systems in Finland; results and analysis of levelling measurements carried out by the National Land Survey of Finland between 1974–1992.Geological Survey of Finland, Nuclear Waste Disposal Research, Report YST-84. 189 p.

Veräjämäki, A. 1998. Pre-Quaternary rocks of the Kokemäki map-sheet area (in Finnish with an English summary). Geological maps of Finland 1:100 000. Explanation to the maps of Pre-Quaternary rocks. Sheet 1134. Espoo: Geological Survey of Finland, 51 p.

Vejbæk, O. V. (1990). The Horn Graben, and its relationship to the Oslo Graben and the Danish Basin. Tectonophysics 178, pp 29-49.

Vidal, G. 1982. The pre-Palaeozoic bedrock. *In:* Wikman, H., Bruun, Å., Dahlman, B & Vidal, G. (eds.) Description to the bedrock map Hjo NO (in Swedish with an English summary). Sveriges Geologiska Undersökning Af 120, 112 p.

Vidal, G. & Moczydłowska, M., 1995. The Neoproterozoic of Baltica - stratigraphy, palaeobiology and general geological evolution. Precambrian Research 73, pp. 197-216

Vorma, A. & Niemelä, R. 1994. Geological map of Finland 1:100 000: Pre-Quaternary rocks, sheet 1133 Yläne. Espoo: Geological Survey of Finland.

Vuorela, P., 1990. Neotektoniska rörelser. Sammanfattande rapport från NKA-projekt KAV 330. (Report in Swedish) 18 p.

Wahlgren, C.-H., Stephens, M.B., Weijermars, R. & Cruden, A.R. 1993.The Mylonite Zone north of lake Vänern - a left-lateral transpressive deformation zone in the Sveconorwegian orogen, south-western Sweden. *In:* Stephens, M.P. & Wahlgren, C.-H. (eds.) Ductile shear zones in the Swedish segment of the Baltic Shield. Sveriges Geologiska Undersökning, Rapporter och meddelanden nr 76, p. 21.

Wahlgren, C.-H., Cruden, A. R. & Stephens, M. B. 1994. Kinematics of a major fan-like structure in the eastern part of the Sveconorwegian orogen, Baltic Shield, south-central Sweden: Precambrian Research 70, pp. 67-91.

Wahlström, R., & Assinovskaya, B. A. 1998. Seismotectonics and Lithospheric Stresses in the Northern Fennoscandian Shield. Geophysica, 34 (1-2), p. 51–61.

Wang, X., Page, L.M. & Lindh, A. 1996. $^{40}Ar/^{39}Ar$ geochronologic constraints from the southersternmost part of the Eastern Segment of the Sveconorwegian orogen: implications for timing of granulite facies metamorphism. GFF 118, pp. 1-8.

Wannäs, K. 1989. Seismic stratigraphy and tectonic development of the Upper Proterozoic to Lower Paleozoic of the Bothnian Bay, Baltic Sea. Stockholm Contribution in Geology 40 (3), pp. 83-168.

Wannäs, K.O. & Flodén, T., 1994. Tectonic framework of the Hanö Bay area, southern Baltic Sea. Svensk Kärnbränslehantering AB, Technical Report SKB TR 94-09, 40 p.

Winterhalter, B. 1967. The Sylen and Solovjeva shoals as observed by a diving geologist. Geologiska Föreningens i Stockholm Förhandlingar 89, pp. 205-217.

Winterhalter, B. 1972. On the geology of the Bothnian Sea : an epeiric sea that has undergone Pleistocene glaciation. Espoo: Geological Survey of Finland. Bulletin 258, 66 p.

Winterhalter, B. 1982. The bedrock geology of Lumparn Bay, Åland. Espoo: Geological Survey of Finland. Bulletin 317, pp. 115-130.

Winterhalter B., Flodén T., Ignatius H., Axberg S. & Niemistö L., 1981. Geology of the Baltic Sea. *In:* Voipio A.(ed.) The Baltic Sea . Amsterdam: Elsevier, pp. 1-121.

Zeck, H. P., Andriessen, P. A. M., Hansen, K., Jensen, P. K. & Rasmussen, B. L. 1988. Paleozoic paleo-cover of the southern part of the Fennoscandian Shield; fission track constraints. Tectonophysics 149 (1-2), 61-66.

Zheng, F. 1996. Tectonic development of the Bohus Granite (SW Sweden) and its adjoining areas. Stockholm Contributions in Geology 44, pp. 1-208.

Ziegler, P.A. 1988. Evolution of the Arctic-North Atlantic and the western Tethys. The American Association of Petroleum Geologists Memoir 43.

Ziegler, P.A. 1989. Evolution of Laurussia: a Study in Late Palaeozoic Plate Tectonics. , Dordrecht/Boston/London: Kluwer Academic Publishers, 102 p.

Ziegler, P. A. 1990. Geological Atlas of Western and Central Europe. Shell Internationale Maatschappij B. V., Geol. Soc. London, Elsevier, Amsterdam, 2nd edition.