GUIDE TO THE GEOLOGICAL MAP OF FINLAND – BEDROCK 1:1 000 000

by

Mikko Nironen

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This Guide is an explanation to the legend of the Geological Map of Finland – Bedrock 1:1 000 000. The background and approach to renewing the map are presented, and the reader is briefly introduced to the terminology of present and past geological units, and to the overall evolution of the Finnish bedrock. The fundamental concepts of tectonic province and province boundary are described, and each tectonic province, as well as the province boundaries, are briefly described. The legend of the map is based on tectonic province division. The core of the Guide is an evolution model from the Neoarchaean to the Cenozoic, linked to the map by legend titles.

Keywords: bedrock, tectonic, units, tectonics, accretion, plate collision, Svecofennian Orogeny, Archean, Proterozoic, Phanerozoic, Finland, Fennoscandian Shield

Mikko Nironen
Geological Survey of Finland, P.O. Box 96, FI-02151 Espoo, Finland

E-mail: mikko.nironen@gtk.fi / mikko.nironen@gmail.com

BACKGROUND AND APPROACH

Published maps of the bedrock of Finland at the scale 1:1 000 000, starting from early 1900s onwards, have included rock types and associations of various ages. The last bedrock map of this traditional type was published by Simonen (1980). The next one, the Bedrock map of Finland by Korsman et al. (1997), introduced a new approach: the legend reflected the tectonic model adopted by the compilers, and hence the map was a geological rather than a bedrock map. The name of the present map, the Geological Map of Finland – Bedrock (Nironen et al. 1016), was chosen because, like the map of Korsman et al. (1997), it includes a tectonic interpretation.

Twenty years have passed since the publication of the previous map, and several advances have taken place at GTK (Geological Survey of Finland): aerogeophysical mapping covered the whole of Finland by 2007; the 1:100 000 bedrock mapping programme was replaced by thematic mapping projects; and geological map data are now collected into the corporate GTK database. The first version of the harmonized map database was published in 2009. Another step forward included the implementation of stratigraphic classification based on geological units (Finstrati; see Luukas et al., this volume).

Since the time of the previous map compilation in 1997, a vast amount of new isotope data (U–Pb zircon and Sm–Nd data on magmatic rocks, and zircon data on detrital grains in sedimentary rocks) have been published. These data have enabled new ideas on the tectonic evolution from the build-up of an Archaean continent to its break-up and the development of a new Palaeoproterozoic continent during the Svecofennian orogeny, and many of these ideas have ended up in the tectonic model presented here.

The Geological Map of Finland – Bedrock is the printed version (and modification) of a bedrock database (DigiKP1M) at the scale 1:1 000 000. The ca. 200 rock types in the GTK database have been reduced to 119 rock associations for the present map. Similarly to the map of Korman et al. (1997), the rock associations are arranged according to a depositional (sedimentary and volcanic rocks) or intrusive (igneous rocks) setting in specific tectonic environments. A difference from the previous map compilation is that the legend is based on tectonic environments (geodynamic settings) rather than tectonic events; this is not a drastic change, because tectonic events and environments are closely linked. For most rock associations, the tectonic environment is specified, whereas some lack a detailed description, reflecting differences in understanding and available information.

The Archaean and Palaeoproterozoic rocks are variably metamorphosed; this is indicated in the map legend by the prefix 'meta' in the names of supracrustal rocks. However, in the case of igneous and subvolcanic rocks, the protolith name has been used. The Mesoproterozoic rocks are metamorphosed to a low grade (see also Hölttä and Heilimo, this volume).

Much effort has been put into structural interpretation, which is mainly based on aerogeophysical maps and to some extent on direct outcrop observations. The control of deformation zones on bedrock distribution has been emphasized. The interpretations, including the sense of displacement in deformation zones, tend to show the first, ductile deformation, not subsequent displacements that commonly take place in the same zones in brittle-ductile or brittle environments. The net displacements in oblique-slip faults may include e.g. early thrusting and subsequent horizontal displacement.

References are generally made to review articles in the book *Precambrian Geology of Finland – Key to the Evolution of the Fennoscandian Shield* (Lehtinen et al. 2005) instead of original papers published prior to 2005. References to selected later articles have been added to update to the present understanding of the bedrock in Finland and beyond.

GEOLOGICAL OUTLINE

Fennoscandia, used as a geographical term, consists of the countries Norway, Sweden, Finland and the northwestern part of Russia (Fig. 1a). Geologically, this area comprises the Archaean and Proterozoic rocks of the Fennoscandian Shield (also called the Baltic Shield; Gorbatschev & Bogdanova 1993), Neoproterozoic and Phanerozoic sedimentary rocks of the East European Platform, and allochthonous Palaeozoic rocks of the Caledonides (Fig. 1b). The bedrock of Finland mainly consists of crystalline rocks of the Fennoscandian Shield; there are only remnants of platform rocks in onshore Finland.

Fennoscandia is also the name of a crustal segment: the protocontinents Sarmatia and Volgo-Uralia assembled at ca. 2.0 Ga, and at 1.82–1.80 Ga the combined Volgo-Sarmatia collided with the protocontinent Fennoscandia to form the new continent Baltica, corresponding to the present East European Craton (Fig 1b; Gorbatschev & Bogdanova 1993, Bogdanova et al. 2008, Bogdanova et al. 2015).

The Archaean bedrock of Finland is dominated by TTG (tonalite–trondhjemite–granodiorite) group rocks with an age span of 3.5–2.73 Ga (Fig. 1c; Hölttä et al. 2012a and references therein). Other plutonic rocks include sanukitoids, quartz diorites and granites, with ages from 2.74 Ga to 2.66 Ga. Moreover, there are paragneisses and greenstone belts with variable lithologies: the Kuhmo

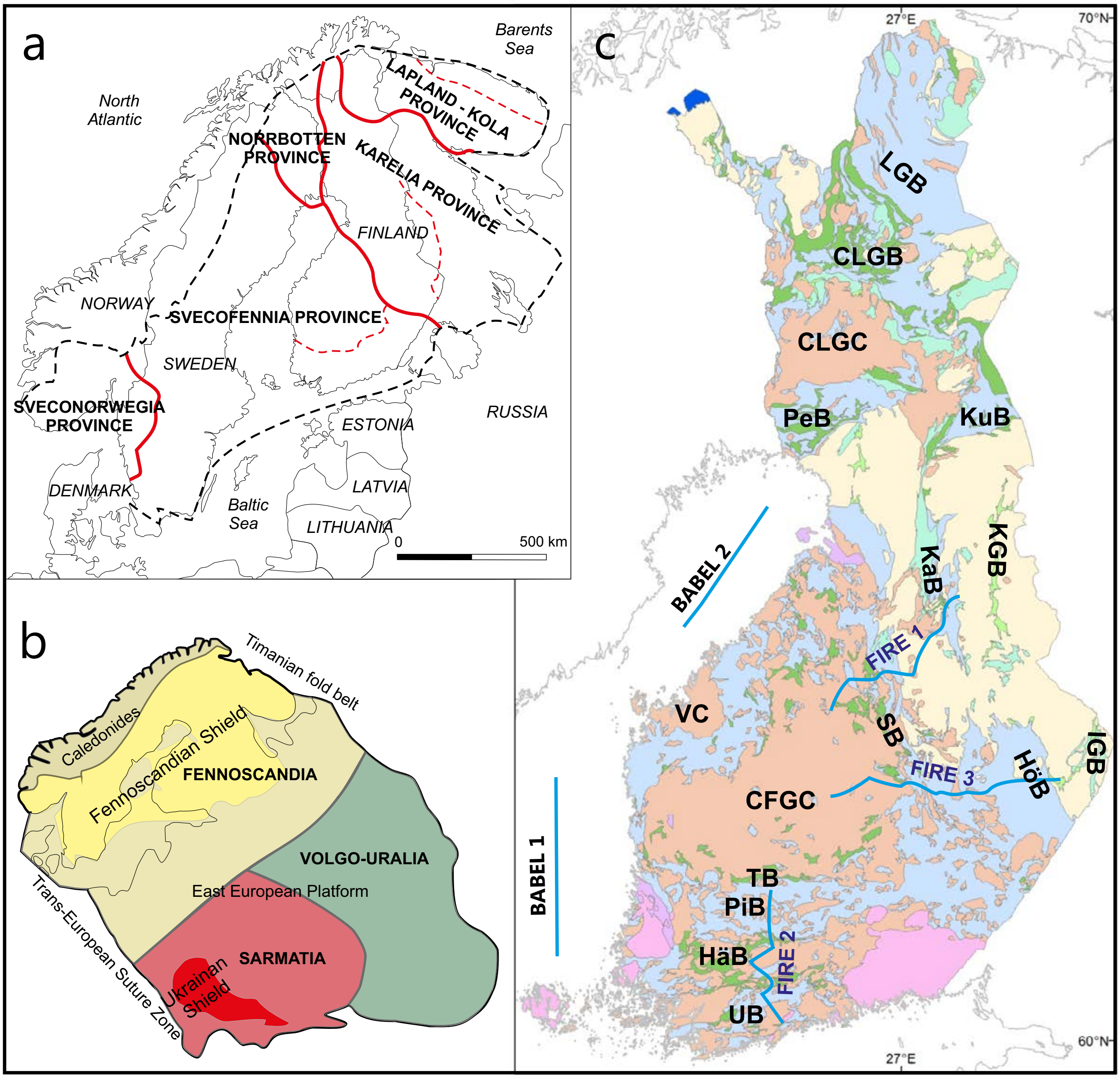


Fig. 1. a) Tectonic provinces in the Fennoscandian Shield (modified from Lahtinen et al. 2005, Daly et al. 2006, Bingen et al. 2008). Province boundaries are shown by solid red line, subprovince boundaries by broken red line. b) The East European Craton (Baltica) with crustal segments Fennoscandia, Sarmatia and Volgo-Uralia (modified from Gorbatschev and Bogdanova 1993). c) Generalized bedrock map of Finland. The locations of BABEL 1 and 2 deep seismic reflection profiles, and sections of FIRE 1, FIRE 2 and FIRE 3 profiles are shown by blue lines. LGB = Lapland granulite belt, CLGB = Central Lapland greenstone belt, CLGC = Central Lapland granitoid complex, PeB = Peräpohja belt, KuB = Kuusamo belt, KGB = Kuhmo greenstone belt, KaB = Kainuu belt, SB = Savo belt, IGB = Ilomantsi greestone belt, HöB = Höytiäinen belt, VC = Vaasa complex, CFGC = Central Finland granitoid complex, TB = Tampere belt, PiB = Pirkanmaa belt, HäB = Häme belt, UB = Uusimaa belt.

greenstone belt (Fig. 1c) consists of ca. 2.84–2.80 Ga komatiitic, basaltic rocks overlying felsic metavolcanic rocks, whereas the Ilomantsi greenstone belt is dominated by metasedimentary rocks, with 2.76–2.72 Ga felsic metavolcanic rocks and less abundant komatiitic rocks.

The Palaeoproterozoic metavolcanic–metasedimentary successions partially covering the Archaean basement in eastern Finland (and northwestern Russia) have traditionally been grouped together as Karelian formations. The deposition of these rocks onto the Karelia craton spans over 500 Ma, from 2.44 Ga to less than 1.92 Ga (Laajoki 2005, Lahtinen et al. 2010). The Karelian formations have been divided into the Sumi, Sariola, Jatuli and Kaleva tectofacies (Laajoki 2005) or chronostratigraphic systems (Hanski & Melezhik 2013). In the tectofacies concept, the tectofacies are separated by major unconformities, and they mark changes in the tectonic environment from intracontinental rifting to continental break-up: Sumi, Sariola and Jatuli represent continental–epicontinental deposits, and Kaleva represents a shift from a rift-marginal basin to a marine basin. The Sumi and Sariola rocks have a scattered occurrence in eastern and northern Finland, whereas the Jatuli rocks form the main part of the Proterozoic cover on the Archaean bedrock.

Layered mafic intrusions with ages around 2.44 Ga occur at the southern margin of the Peräpohja and Kuusamo belts, and in central Lapland (Hanski & Huhma 2005; Fig. 1c). These rocks, as well as 2.44 Ga felsic metavolcanic Sumi rocks, mark the onset of rifting of the Archaean crust in Finland. Mafic magmatism at 2.3 Ga, 2.2 Ga, 2.1 Ga and 2.05 Ga, in the form of intrusions, sills and dyke swarms, is further evidence of episodic extension of the crust (Hanski & Huhma 2005, Vuollo & Huhma 2005).

The depositional ages of the immature clastic metasedimentary and ultramafic–mafic metavolcanic Sariola rocks and the metavolcanic–metasedimentary Jatuli rocks range from 2.43 Ga to 2.3 Ga and from 2.3 Ga to 2.06 Ga, respectively (Laajoki 2005, Hanski & Melezhik 2013). The composition of the Jatuli rocks varies spatially: the remnant basin fills in the Höytiäinen belt (Fig. 1c) are quartzites, whereas arkosites and sedimentary carbonate rocks dominate further north; sedimentary carbonate rocks are missing in central Lapland.

The Kaleva rocks include turbiditic metagreywackes and minor metaconglomerates in the Höytiäinen, Savo, Kainuu and Peräpohja belts, and predominantly homogeneous deep marine metapsammites, occurring west of the Höytiäinen belt and in the Kainuu belt (Fig.1c). Remnants of 1.95 Ga ophiolite complexes occur as tectonic slivers in the metapsammites (Peltonen 2005, Peltonen et al. 2008). The Kaleva deposits have been variously divided (Lower/Upper, Eastern/Western), and they probably represent both early rift basin (ca. 2.1 Ga) and collision-related (ca. 1.9 Ga) foredeep to foreland deposits that together were thrust into complex allochthonous packages between 1.95 Ga and 1.92 Ga, during an initial stage of the Svecofennian orogeny (see below; Kohonen 1995, Lahtinen et al. 2010, Ranta et al. 2015).

The oldest rocks in the Finnish bedrock, not related to rifting of the Archaean crust, are the ca. 2.0 Ga oceanic mafic-ultramafic metavolcanic rocks and serpentinites in the Central Lapland greenstone belt (Fig. 1c; Hanski & Huhma 2005). They occur as slivers in an allochthonous–parautochthonous package of younger Palaeoproterozoic supracrustal rocks that was emplaced during an initial stage of the Svecofennian orogeny (Lahtinen et al. 2005).

In northernmost Finland, intrusive rocks with ages of 1.95–1.93 Ga (Meriläinen 1976), and the surrounding supracrustal rocks as well as the Lapland granulite belt (Fig. 1c), are parts of the Lapland–Kola orogen. The Lapland–Kola orogen developed at 1.95–1.91 Ga, initially as accretion and finally as continental collision (Daly et al. 2006). Noritic-enderbitic rocks intruded metasediments of the Lapland granulite belt at 1.92–1.91 Ga, the peak of granulite grade regional metamorphism was attained during 1.91–1.90 Ga, decompression and cooling (uplift stage) occurred during 1.90-1.88 Ga, and cooling was almost finished by 1.87 Ga (Tuisku & Huhma 2006).

A complex set of orogenic events, jointly named as the Svecofennian orogeny, occurred during 1.92–1.77 Ga (Korsman et al. 1999, Lahtinen et al. 2005) and caused the growth of new juvenile crust in central and southern Finland. The oldest Svecofennian rocks in this area are the 1.92 Ga primitive island arc rocks in the Savo belt (Fig. 1c; Kähkönen 2005). Nd isotopic evidence (Lahtinen & Huhma 1997, Rämö et al. 2001) suggest a ≥ 2.0 Ga nucleus for the Central Finland granitoid complex (Fig. 1c) and southernmost Finland. Island arc-type volcanic rocks and mafic-ultramafic intrusions, similar to rocks found in deeply eroded parts of modern oceanic and continental arcs, were emplaced into older, mainly turbiditic sequences (now biotite paragneisses and metagreywackes) at 1.90–1.87 Ga (Kähkönen 2005, Peltonen 2005), and together these assemblages form a curving (Z-type) belt that extends from the

Tampere belt to the northwest, encircles the Vaasa complex and continues to north-central Sweden (Skellefte district). Granitoids and minor gabbros (1.89–1.87 Ga) intruded the supracrustal rocks and formed the Central Finland granitoid complex (Nironen 2005). Together, these rocks comprise the Western Finland arc complex.

In southern Finland, the Häme belt is dominated by 1.89–88 Ga arc-type intermediate and arc rift-type mafic metavolcanic rocks (Fig. 1c; Kähkönen 2005). Synvolcanic mafic intrusions occur among the metavolcanic rocks (Peltonen 2005). In the Uusimaa belt, 1.90–88 Ga felsic to ultramafic metavolcanic rocks with arc-type to MORB-type geochemical characteristics, occurring among metagreywackes, quartz-feldspar gneisses and sedimentary carbonate rocks, are crosscut by 1.88–1.87 Ga granitoids (Kähkönen 2005). This part has been referred to as the Southern Finland arc complex.

Three regional metamorphic events overprinted the rocks in southern and central Finland during the Svecofennian orogeny. An early (ca. 1.91 Ga) metamorphic event preceded emplacement of pyroxene-bearing granitoids in the Savo belt (Fig. 1c; Lahtinen et al. 2015a). The main Svecofennian high-T/low-P type (1.88–1.87 Ga) event generally reached upper amphibolite grade, and can be detected throughout the Finnish Svecofennian (Korsman et al. 1999, Hölttä & Heilimo, this volume). The late Svecofennian high-T metamorphism (1.84–1.80 Ga) reached granulite grade in large areas in southernmost Finland and was associated with transpressional deformation and emplacement of 1.84–1.81 Ga granites. Two metamorphic events in western Lapland at 1.92–1.90 Ga and 1.86–1.85 Ga have recently been suggested (Lahtinen et al. 2015b). These may be associated with the Svecofennian orogeny, whereas a later (1.79–1.77 Ga) metamorphic event, recorded in large areas in central and western Finnish Lapland (Corfu & Evins 2002, Ahtonen et al. 2007, Niiranen et al. 2007), possibly has a different tectonic connotation.

After 1.76 Ga, orogenic activity ceased in the area of Finland and the continental crust was gradually stabilized. During 1.65–1.54 Ga, a crustal rifting episode caused emplacement of the rapakivi granites of southern Finland, as well as associated anorthosites and mafic dykes (Rämö & Haapala 2005). Rapakivi magmatism was the last major increment to the Precambrian bedrock of Finland.

TECTONIC PROVINCES

A major aim of the present Guide is to describe the evolution and relationships of the four fundamental, continent-scale lithospheric blocks, named as tectonic provinces (see inset map). The legend of the map is aimed to explain the stage of evolution *at 1.86 Ga, when four lithospheric blocks had been united to form a new Palaeoproterozoic continent, consisting of the tectonic provinces Karelia, Lapland–Kola, Norrbotten and Svecofennia*. The rocks coeval or older than the assembly are grouped under the title 'Tectonic provinces'. After the 1.86 Ga assembly, the tectonic provinces had a common tectonic history. The supracrustal or plutonic rocks that were deposited or intruded after the assembly, and some rocks that were emplaced in foreland environment before 1.86 Ga, are shown beyond the explanation of the tectonic provinces. In the following sections, the tectonic provinces and their boundaries are briefly described, with emphasis on the tectonic units within the provinces.

Tectonic province definition

In the past years, 'megaunits' in Fennoscandia have been referred to using several qualifying terms such as craton, orogen, belt, province and domain (e.g. Gáal & Gorbatschev 1987). Logical division of bedrock into major crustal blocks is conceptually difficult, and internationally established definitions such as geologic province (see Neuendorf et al. 2005) are rather vague. In this Guide, the geological province scheme of Geoscience Australia (http://www.ga.gov.au/applications/provexplorer/australian-geological-provinces) is taken as a basis. The tectonic provinces are groups of other types of provinces (sedimentary, igneous, structural, metamorphic or metallogenic) that are linked by a common tectonic history. Provinces may be small or large, and they are four-dimensional: in addition to an area and thickness, they represent a definable period of geological time. The definable period of geological time for the tectonic provinces under consideration starts from Palaeoproterozoic (1.86 Ga) amalgamation. In addition to the Geoscience Australia definition, tectonic provinces are here defined as lithospheric blocks that were separated from each other by oceanic lithosphere before the Palaeoproterozoic assembly. The conceptual difference between a tectonic province and an orogen is that a tectonic province is a component in accretion or collision, whereas orogen is a region that has been subjected to deformation during an orogenic cycle that results from accretion or collision; an orogen normally comprises parts of at least two provinces. As an example, in Finland, only the easternmost part of the Karelia Province is relatively unaffected by the Svecofennian orogeny, and consequently, the Svecofennian orogen covers much of the Karelia Province. The entire Fennoscandian Shield is divided according to the tectonic province scheme (Fig. 1a).

The **Karelia Province**, extending east to northwestern Russia and west to northern Sweden, is the largest province within the Fennoscandian Shield and consists of Archaean crust partly overlain by the Palaeoproterozoic cover sequence (the Karelian formations). The Finnish part of the Karelia Province has been divided into two subprovinces: the Western Karelia Subprovince, which consists of Mesoarchaean and Neoarchaean complexes and includes most of the Archaean bedrock of Finland, and the Central Karelia Subprovince, which barely extends to the easternmost part of Finland and is characterized by a short Neoarchaean crustal growth period (Fig. 1a; Hölttä et al. 2012b).

In the Finnish part of the Karelia Province, autochthonous Palaeoproterozoic cover rocks mainly occur in northern Finland. In addition to these, there are several allochthonous units mainly consisting of craton margin sequences: the Northern Karelia and Kainuu nappe complexes in eastern Finland, the Northern Ostrobothnia nappe system in central Finland and the Central Lapland nappe complex in Lapland (Fig. 2a; see also Luukas et al., this volume). The sequences in the Northern Savo nappe complex are of mixed origin.

The main part of the exposed **Lapland–Kola Province** is in the Kola Peninsula of Russia (Fig. 3). Daly et al. (2001) described the Lapland–Kola Province (orogen) as a collage of partially reworked late Archaean terranes and intervening belts of Palaeoproterozoic juvenile crust, welded together in the Palaeoproterozoic. Daly et al. (2001) considered the Archaean Belomorian belt part of the orogen. Although the Belomorian belt experienced metamorphic reworking during the Palaeoproterozoic that did not affect the 'Karelian protocraton' to the southwest (Bibikova et al. 2001), and early Palaeoproterozoic mafic–felsic volcanism occurred in the boundary area (Bogina et al. 2015), there is no evidence for separation of the Belomorian belt from the 'Karelian protocraton', with an oceanic basin between, during the Palaeoproterozoic. Therefore, according to the tectonic province definition given above, the Belomorian belt is included in the Karelia Province. Similarly, the area in the northern part, referred to as Murmansk craton, is here considered as a subprovince within the Lapland–Kola Province, because it remained assembled with other Archaean blocks throughout the Palaeoproterozoic Era (Daly et al. 2006). The predominant part of the Lapland–Kola orogen in Finland is the Palaeoproterozoic Northern Lapland nappe system (Fig. 2a). The Sodankylä nappe, located within the Karelia Province, is part of the Northern Lapland nappe system (see also Luukas et al., this volume).

The **Norrbotten Province** was recently introduced as a crustal block (Lahtinen et al. 2005). It is separated from the Karelia Province by a major shear zone (Lahtinen et al. 2015a), and consists of Archaean crust partly overlain by a Palaeoproterozoic sequences. The Ylläs nappe (Fig. 2a) in the eastern part of the Norrbotten Province consists of two clastic sedimentary rock sequences, with depositional ages of 1.92–1.90 Ga (Lahtinen et al. 2015a) and ≤1.88 Ga (Hanski & Huhma 2005).

An assemblage of Palaeozoic rocks within the Norrbotten Province, at the northwestern tip of Finland, are part of the fold and thrust belt that was displaced eastwards upon the Precambrian rocks of the Norrbotten Province during the 430–390 Ma Scandian event of the Caledonian orogeny (Roberts 2003). Since this subhorizontal lithologic package (Palaeozoic rocks on top of Precambrian rocks) does not have a common tectonic history, it is not considered to be a separate tectonic province.

The **Svecofennia Province** consists almost solely of Palaeoproterozoic rock assemblages. The Palaeoproterozoic juvenile crust of Finland is considered to have formed by accretion of the Western Finland and Southern Finland arc complexes to the Archaean continent during the Svecofennian orogeny (Nironen 1997, Lahtinen et al. 2005). These arc complexes are respectively denoted as the Western Finland Subprovince and Southern Finland Subprovince. The terms Central Svecofennian and Southern Svecofennian (arc complex) have also been used, but since these terms cover areas that extend beyond Finland and are not properly defined, the present subprovince division of Svecofennia is restricted to the area of Finland.

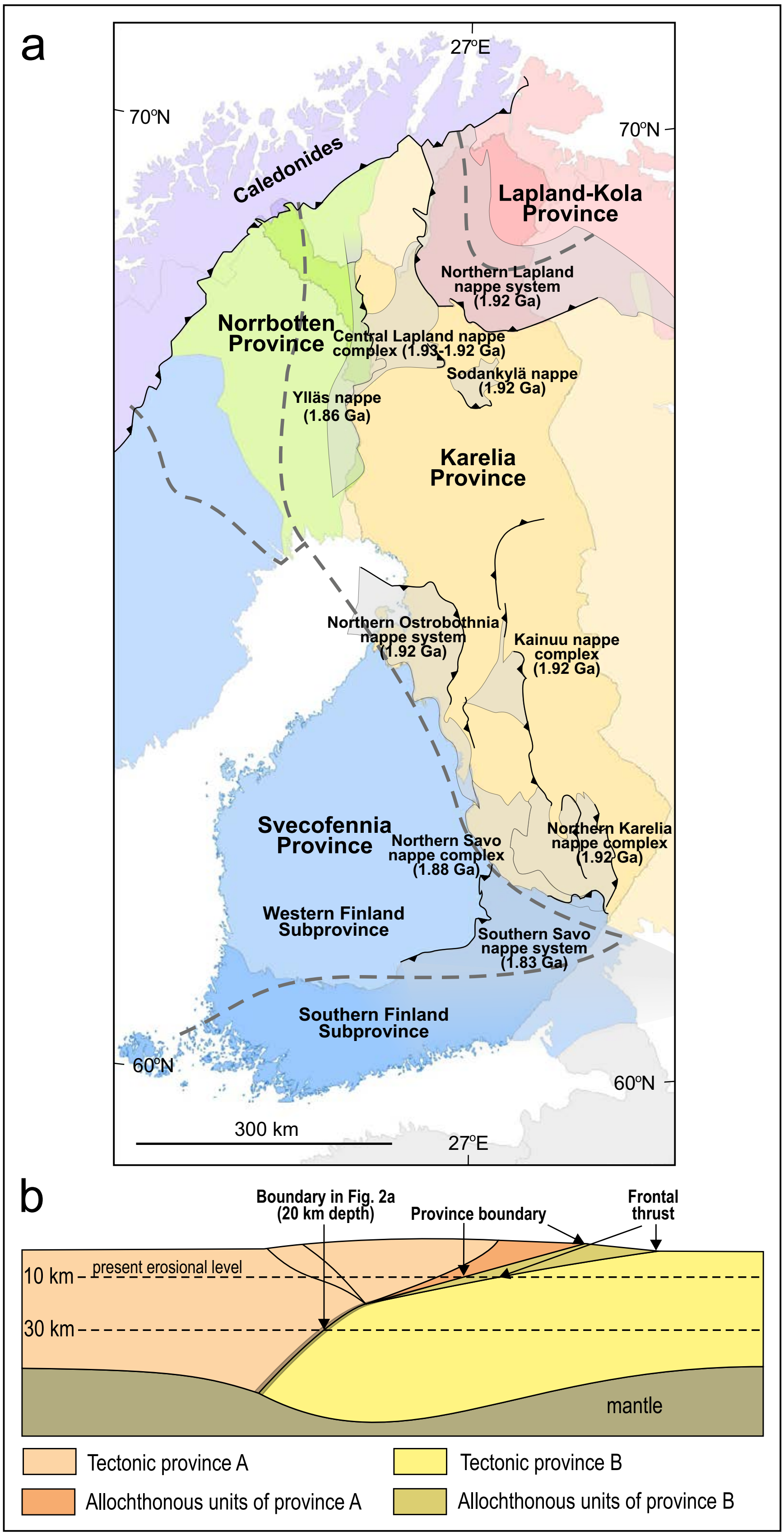


Fig. 2. a) Allochthonous units and tectonic boundaries in and around Finland. Allochthonous units are shown by grey shading. The ages given for allochthonous units denote the age of first thrusting; the units may have been involved in several thrusting events. Frontal thrusts and their dip directions are shown by thin black lines. Province boundaries are shown as they are from the surface down to the base of allochthonous units (= a few km depth). Sutures, interpreted at ca. 20 km depth, are shown by thick broken lines. b) Schematic presentation of the tectonic boundaries.

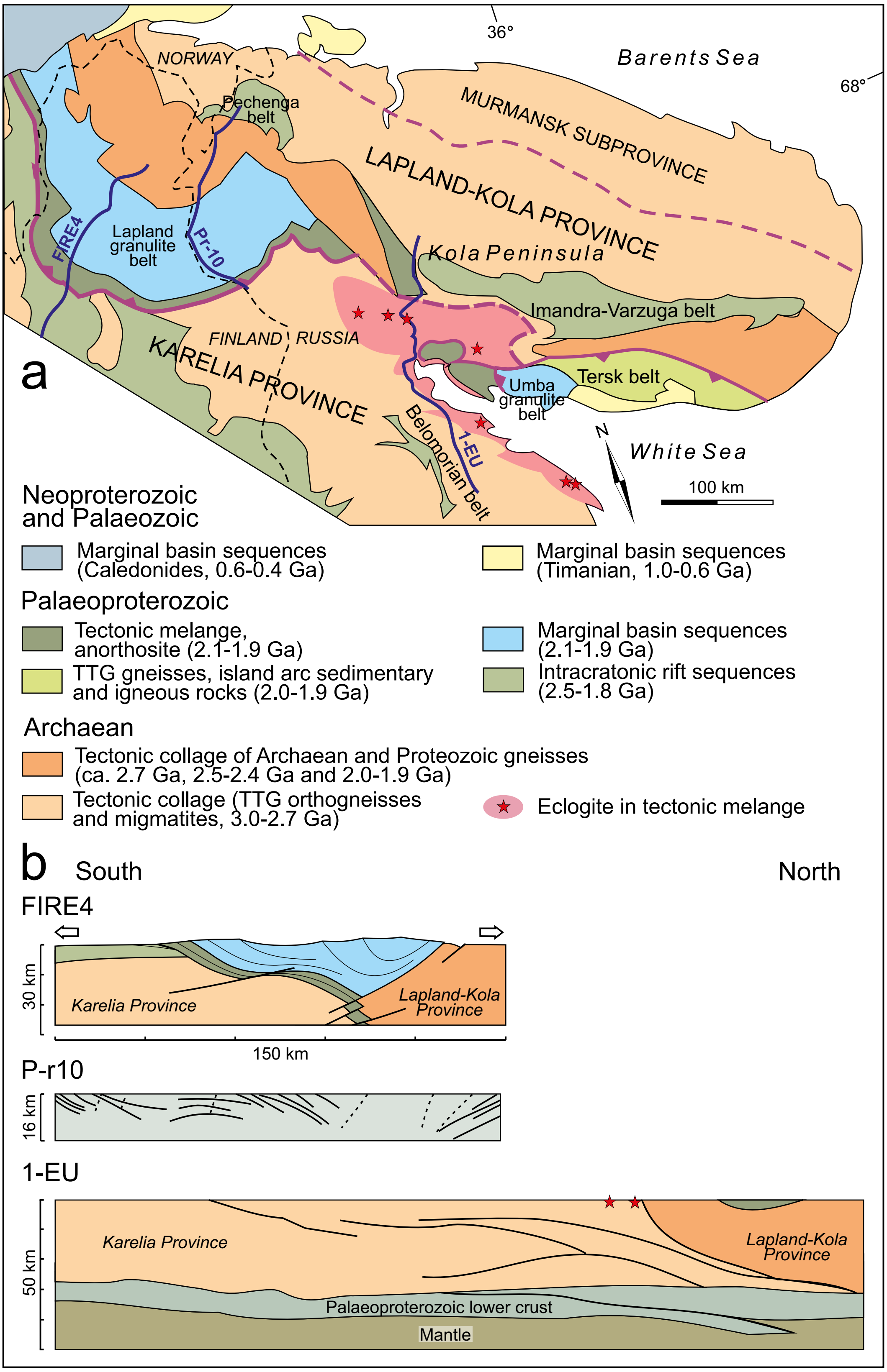


Fig. 3. a) Tectonic units in the Lapland–Kola Province (modified from Daly et al. 2006 and Mints et al. 2014). The tectonic province boundary (with inferred dip) is shown by a thick purple line (broken when poorly constrained). b) Interpretations of seismic profiles. The interpretation of profile FIRE4, modified from the present structure (Patison et al. 2006), is the assumed palaeosetting at 1.87 Ga. Thick black lines are thrust zones, thin lines indicate the general structural pattern. Interpretation of profile Pr–10 (without interpretation of crustal components) is a modification from Daly et al. (2006). Solid black lines are reflective layers, broken lines are fault zones. Profile 1–EU is a simplification of the one presented by Mints et al. (2014). Thick black lines are thrust zones. Stars denote eclogite occurrences.

Province boundaries

The definition of tectonic provinces – lithospheric blocks that were once separated from each other by oceanic lithosphere – implies that the province boundaries are marked by sutures. Sutures are cryptic zones that can usually be vaguely inferred due to orogenic reworking of the crust. In Figure 2a, tectonic province boundaries are interpreted to ca. 20 km crustal depth to give a three-dimensional impression. The interpretation to depth is based on reflection seismic data (Lapland – Kola – Karelia, Karelia – Svecofennia; Korja et al. 2006, Patison et al. 2006, Sorjonen-Ward 2006, Korja & Heikkinen 2008) and isotopic data (Karelia – Svecofennia; e.g. Huhma 1986; Patchett & Kouvo 1986, Lahtinen & Huhma 1997).

The foreland frontal thrusts are shown where they can be identified. Metamorphism in the Svecofennian orogen is characterized by fairly uniform 4–6 kbar pressures (Korsman et al. 1999, Hölttä & Heilimo, this volume), corresponding to 10–15 km crustal depth; thus, the Svecofennian frontal thrusts are comparable with deeper sections of similar thrusts in modern collisional orogens. The position of province boundary at various crustal depths is schematically presented in Figure 2b.

The Lapland – Kola – Karelia boundary is shown by a frontal thrust, dipping 10–20° NE. The shallow dip is well constrained by reflection seismic data (Fig. 3b). The province boundary at 20 km depth is at the intersection of seismic reflections: to the northeast, the reflections dip moderately SW.

The Norrbotten Province contains the Ylläs nappe, the extent of which is poorly constrained – hypothetical towards the north – but a frontal thrust can be defined for ca. 100 km length. The Norrbotten – Karelia boundary is inferred to dip to the west.

The Svecofennia – Karelia boundary, studied for decades, has been demonstrated to be complex. The boundary may be located fairly accurately to 10–20 km depth by reflection seismic data (Korja & Heikkinen 2008), and to Moho depth by refraction seismic data (Luosto et al. 1984, Korja et al. 1993). In the centre of the Savo belt, the suture is interpreted to be subvertical to 10–15 km depth, and to dip shallowly towards ENE at deeper levels (Fig. 4c). Further southeast, the Northern Karelia nappe complex overlies the Archaean crust to 10–15 km depth (Fig. 4d; Sorjonen-Ward 2006). Overall, the structure is interpreted to be a wedge of Archaean crust, underlain and partly overlain by Svecofennian crust (Korja & Heikkinen 2008).

It is difficult to assess the location of the boundary between the Western Finland and Southern Finland Subprovinces because of structural overprinting. In the east, the boundary is arbitrarily set to follow the frontal thrust of a late Svecofennian allochthonous unit (Southern Savo nappe system, Fig. 2a). However, rocks considered to be part of the Western Finland Subprovince are exposed as a tectonic window within the Southern Savo nappe system, implying that the boundary is actually further south. Continuation to the west is also ambiguous: in the centre, the boundary is defined by shear zones, and in the west the subprovince boundary was interpreted on a lithological basis. Reflection seismic data (Korja & Heikkinen 2008) suggest that in the central part, the boundary is subvertical (cf. Figs. 2a and 4e).

TECTONIC EVOLUTION FROM THE NEOARCHAEAN TO THE CENOZOIC

The following model is aimed to give an overall tectonic framework for the Geological Map. Nevertheless, it necessarily extends beyond the national boundaries of Finland and incorporates tectonic evolution of the present Fennoscandian Shield. Two periods fundamental to the evolution are emphasized: 1) the break-up of the Neoarchaean continent (Karelia craton) at 2.1–2.05 Ga; and 2) the assembly of four lithospheric blocks (Karelia, Lapland–Kola, Norrbotten and Svecofennia) at 1.92–1.86 Ga to form a new Palaeoproterozoic continent (protocontinent Fennoscandia).

The model (Figs. 5–6, 8–10) is presented so that the Karelia Province is the ‘fixed block’, and the other crustal components move with respect to it. The coordinates in the text refer to present directions. The numbers in parentheses [ ] refer to numbered rock associations in the map legend. The palaeogeographic reconstructions are outlined.

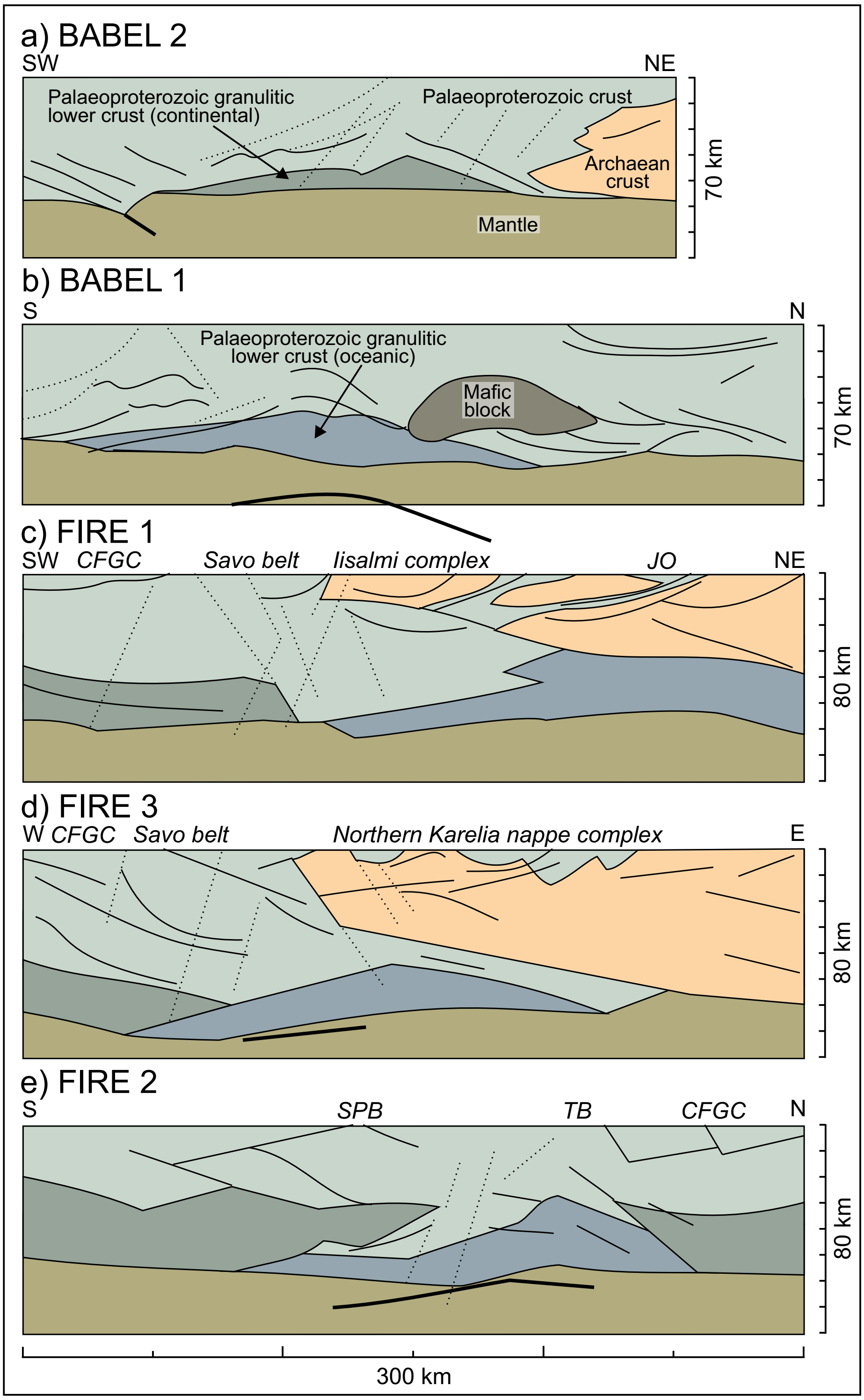


Fig. 4. Selected sections along BABEL and FIRE deep seismic reflection profiles (simplified from Lahtinen et al. 2009a). No vertical exaggeration. Crustal reflections are shown as thin black lines, mantle reflections as thick black lines, and faults/shear zones as broken lines. CFGC = Central Finland granitoid complex, JO = Jormua ophiolite, TB = Tampere belt, SPB = Subprovince boundary. For the profile locations, see Figure 1.

Assembly of the Neoarchaean continent (3.1–2.68 Ga)

The Archaean bedrock of the Karelia Province has been extensively studied during the past 15 years (e.g. Slabunov et al. 2006, Hölttä et al. 2012a, Mints et al. 2014). Previously scattered information (mainly in Russian) on the Archaean terranes in the Lapland–Kola Province has recently been compiled (Daly et al. 2001, Daly et al. 2006), whereas little is known of the Archaean bedrock in the Norrbotten Province.

The komatiitic and basaltic rocks of the Kuhmo greenstone belt resemble geochemically modern oceanic plateau basalts, and Hölttä et al. (2012b) suggested that the komatiites were erupted onto a rhyolitic-dacitic-tholeiitic oceanic plateau. In contrast, the metavolcanic rocks of the Ilomantsi greenstone belt probably represent arc magmatism within an attenuated continental margin. Overall, the geochemical data from Finland suggest the evolution of Archaean rocks in various geodynamic settings. The oldest Archaean terranes within the Fennoscandian Shield are Mesoarchaean in age (the 3.5–3.1 Ga Vodlozero microcontinent; Slabunov et al. 2006). Two core areas of the Neoarchaean assembly were the Karelia and Murmansk cratons, and the Belomorian belt was an orogenic belt between these cratons. Eclogitic bodies in Neoarchaean mélanges evidence Mesoarchaean (2.87 Ga; Mints et al. 2010) and Neoarchaean (2.79–2.73 Ga; Dokukina et al. 2014) eclogite facies metamorphic events in rocks of oceanic origin, and the metamorphic events as well as regional-scale nappe complexes have been associated to subduction and collision (Slabunov et al. 2006, Mints et al. 2009, Mints et al. 2014). Terrane accretion initiated at 3.1 Ga (Slabunov et al. 2006), but in Finland the exotic terranes probably only accreted at ca. 2.83–2.75 Ga. Collisional crustal stacking that resulted in the formation of a new Neoarchaean continent occurred at 2.73–2.68 Ga (Hölttä et al. 2012a). The continent included the blocks represented by the tectonic provinces Karelia and Lapland–Kola. The palaeogeographic setting of the Norrbotten Province at this stage is unknown.

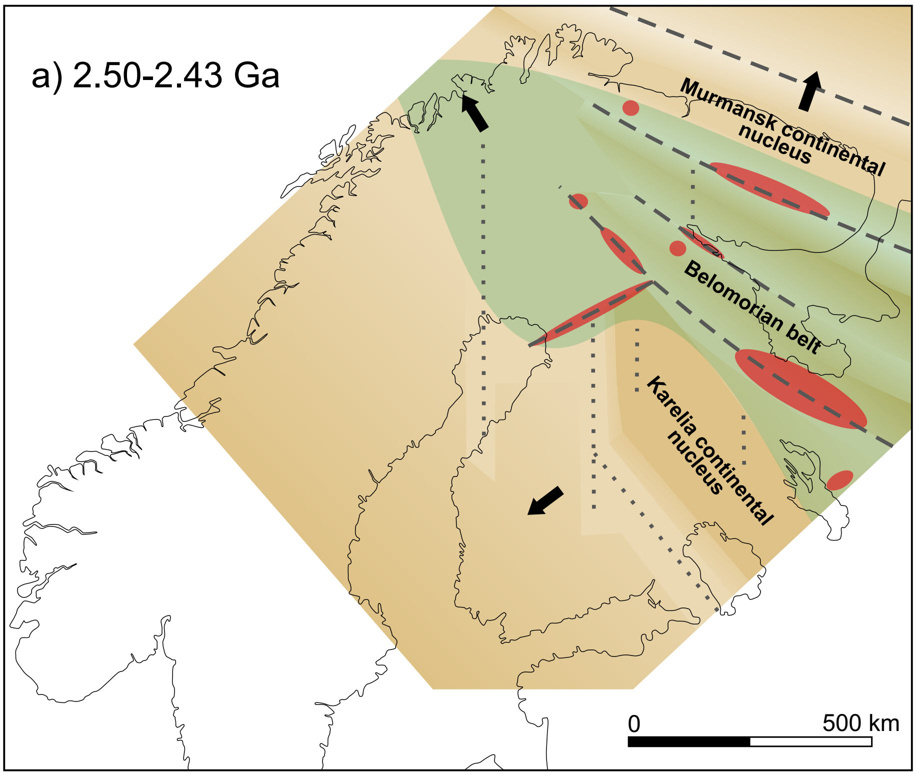
In palaeogeographic reconstructions the 2.7 Ga accretions have been assigned to assembly of the supercontinent Kenorland (Williams et al. 1991). Bleeker (2003) proposed the existence of separate supercratons Superia, Sclavia and Vaalbara instead of a single supercontinent. Based on palaeomagnetic evidence, at 2.7–2.6 Ga Karelia (i.e. Archaean rocks of the Karelia Province) was located at high latitudes (80–60°; Mertanen & Korhonen 2011).

Cratonic stage with intermittent extensional episodes (2.68–2.1 Ga)

After collisional orogeny, the new continent stabilized to form a craton, referred to as the Karelia (or Karelian) craton. The shear zones in at least N–S and ENE–WSW directions in the Karelia Province have a long history: they possibly developed during amalgamation of the Archaean terranes and were reactivated during subsequent extensional events. An indication of such a long history is the 2.61 Ga Siilinjärvi carbonatite [83] in east-central Finland (O'Brien et al. 2005): it occurs as a N–S-trending vertical tabular body, was probably emplaced in a Neoarchaean rift, and is presently delimited by Palaeoproterozoic shear zones.

The depositional age of quartzites and arkosites in Central Lapland [87] and quartz-feldspar paragneisses in the Kainuu belt [92] have been interpreted as Neoarchaean (Lahtinen et al. 2015a). If this interpretation is correct, the first extensional rift basin(s) developed in the Neoarchaean continent already during 2.6–2.5 Ga. The first unequivocal examples of Palaeoproterozoic extension are the 2.50 Ga layered mafic intrusions in the Kola Peninsula, associated with crustal uplift and erosion (Fig. 5a; e.g. Amelin et al. 1995, Bayanova et al. 2009).

A major extensional event at 2.44–2.43 Ga resulted in emplacement of layered mafic intrusions, felsic intrusions and intermediate–felsic volcanic rocks of the Sumi system [82-80] in intracontinental rift basins in northern Finland (Hanski & Huhma 2005, Hanski & Melezhik 2013) and basalts in Russia (Bogina et al. 2015). The layered intrusions and metavolcanic rocks occur in NW-SE trending zones, and in an almost orthogonal zone at the Archaean–Proterozoic boundary in the Peräpohja and Kuusamo belts (Fig. 1; Iljina & Hanski 2005, Hanski & Huhma 2005). The assumption here is that these intrusions were originally emplaced in an ENE–WSW-trending extensional zone (Fig. 5a). These two directions suggest large-scale extension in the present northern Fennoscandia, probably as the result of a mantle plume upwelling (Hanski & Huhma 2005, Melezhik & Hanski 2013). Basalts (Kulikov et al. 2010) and basaltic dykes and sills in the Archaean basement (Vuollo & Huhma 2005,



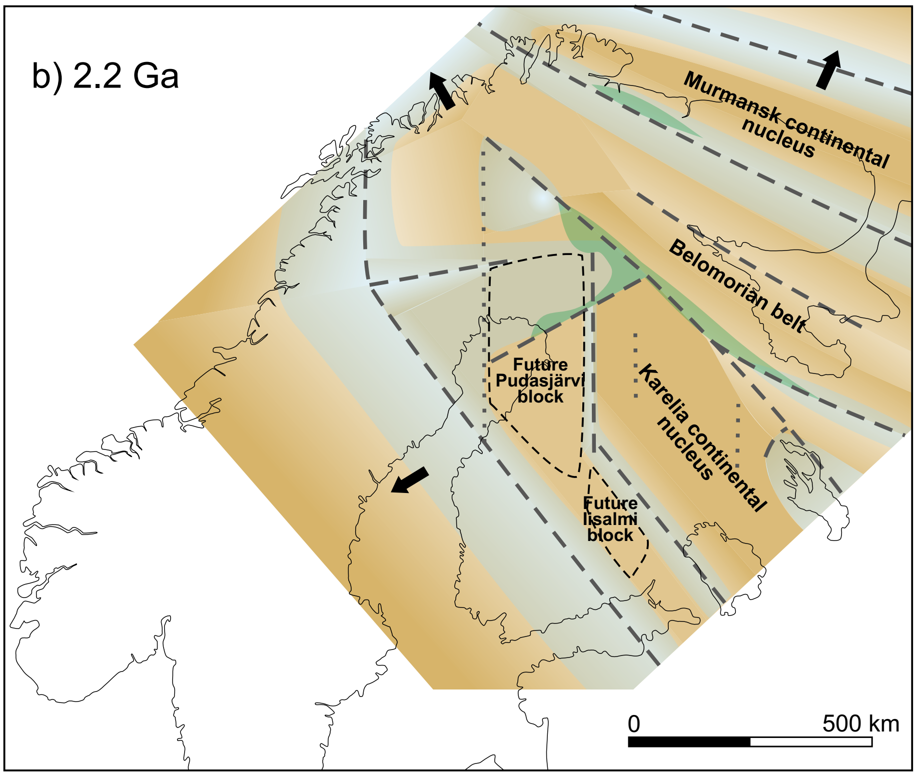


Fig. 5. a) Archaean continental crust at 2.50–2.43 Ga (modified from Melezhik and Hanski 2012). The present location of 2.50–2.43 Ga mafic plutonism is shown schematically by red areas. The inferred extent of continental flood basalt magmatism is shown by green shading, old deformation zones are shown by dotted lines and active deformation zones by broken lines. Note that the active deformation zones are drawn according to the magmatic centres in their present settings, not in their palaeosettings during 2.50–2.43 Ga. Directions of tectonic transport are shown by black arrows. b) Archaean continental crust at 2.2 Ga. Variation in crustal thickness is shown by a variable intensity of colouring (darker = thicker crust, lighter = thinner crust). Mafic volcanism is shown by green shading, sedimentary basins by blue shading.

Stepanova & Stepanov 2010) suggest the formation of a large flood-basalt province (Fig. 5a; cf. Melezhik & Hanski 2013).

The deposition of Sumi system rocks was followed by a period of weathering in semi-arid conditions, and the removal of much of the deposits during rift inversion (Laajoki 2005, Karinen 2010). Immature sedimentary rocks [78], including glaciogenic rocks, represent the Sariola system and are found at the margins of the Höytiäinen, Kainuu and Peräpohja belts (Fig. 1). Ultramafic and mafic rocks, occurring in the Kuusamo belt and in central Lapland [77], were deposited upon the sedimentary Sariola rocks or directly on Archaean gneisses. The Sariola rocks were probably deposited in reactivated rift basins during 2.43– 2.35 Ga.

The occurrences of a palaeosol and a major unconformity imply intense weathering and deep erosion in a warm and humid climate after the deposition of the Sariola rocks (Laajoki 2005). The subaerial palaeosol may have covered large areas of the continent, with a low topographic relief during the deposition of the Jatuli system sediments.

The beginning of Jatuli has been set at 2.3 Ga (Hanski & Melezhik 2013). Quartzites and arkosites of the Jatuli system [76] are voluminous in the Peräpohja belt, Kuusamo belt, and Central Lapland greenstone belt; Laajoki (2005) proposed a relative rise in the epicontinental sea level in the Karelia craton and the deposition of transgressive fluvial and deltaic sediments. The quartzites and arkosites are intruded by 2.2 Ga sills and dykes that evidence an extensional event within the Karelia craton (Vuollo & Huhma 2005), although the volume of mafic magmatism was probably less than during 2.44–2.43 Ga (Fig. 5b).

Palaeomagnetic studies indicate the drift of Karelia to near-equatorial palaeolatitudes by ca. 2.50–2.45 Ga (Mertanen & Korhonen 2011 and references therein, Pesonen et al. 2012). Based on palaeomagnetic evidence from dyke swarms, Bleeker and Ernst (2006) suggested that Karelia was attached to the Superior craton at 2.45 Ga, and that both were part of the supercraton Superia. However, a more recent palaeomagnetic study of dykes in eastern Finland (Salminen et al. 2014) does not support such an assembly.

Attenuation and break-up of the craton (2.1–2.05 Ga)

The quartzites [74], deposited after 2.2 Ga, and an increasing amount of metasedimentary carbonate rocks [73] record a shift from rift-controlled, fluvial-dominated system to open sea conditions, with development of carbonate platforms (Fig. 5b; Laajoki 2005, Melezhik & Hanski 2013). Graphite-sulphide paraschists [72-71] in the Lapland greenstone belt suggest the development of a separate euxinic basin.

The final break-up of the Karelia craton initiated at 2.1 Ga (Fig. 6a; see Lahtinen et al. 2010). Extension is evidenced as 2.1 Ga granites in northern Finland [62] (Ahtonen et al. 2007, Lauri et al. 2012), and especially as ca. 2.1 Ga tholeiitic dyke swarms that extend from eastern Finland to Russia (Vuollo & Huhma 2005, Stepanova & Stepanov 2010). Mafic intrusions in northern and central Finland [61] and ultramafic-mafic volcanic rocks in central Lapland [70] are of 2.06–2.05 Ga age. Emplacement of the 2.1 Ga dykes and 2.06–2.05 Ga magmatic rocks has been associated with mantle plume activity (Hanski & Huhma 2005, Lahtinen et al. 2015b). Moreover, the cratonic break-up is coeval with end of the 'Lomagundi-Jatuli isotopic excursion', an increase in $\delta^{13}C$ values of Jatulian carbonate rocks during 2.2–2.1 Ga, and drop at 2.06 Ga (Karhu 2005, Hanski & Melezhik 2013). The relationship between these two events is unclear.

A divergent triple point is modelled on top of a plume head, with continental break-up and development of a failed rift in northern Finland (Fig. 6a). An unknown amount of continental mass drifted away, continental ribbons and smaller extensional blocks were separated from the attenuated continental margin, and a volcanic margin developed in the southwest. In Figure 6a, a flood-basalt province is outlined in the continental nucleus. Part of the Kaleva system rocks [67] possibly represent sedimentation in the newly formed marginal basins.

Daly et al. (2006) modelled the opening of a narrow oceanic basin between the Karelia and Murmansk continental blocks as the result of 2.5–2.1 Ga rifting and final break-up. Palaeoproterozoic (1.98–1.94 Ga) juvenile crust, found as magmatic rocks of the Lapland granulite, Umba and Tersk belts (Fig. 3) was formed, presumably in an island-arc setting. The volcanic-sedimentary rocks of the Pechenga belt represent a continental rift environment, with attenuation of the crust between 2.06 Ga and 2.02 Ga (Fig. 6a; Melezhik et al. 2007, Hanski et al. 2014).

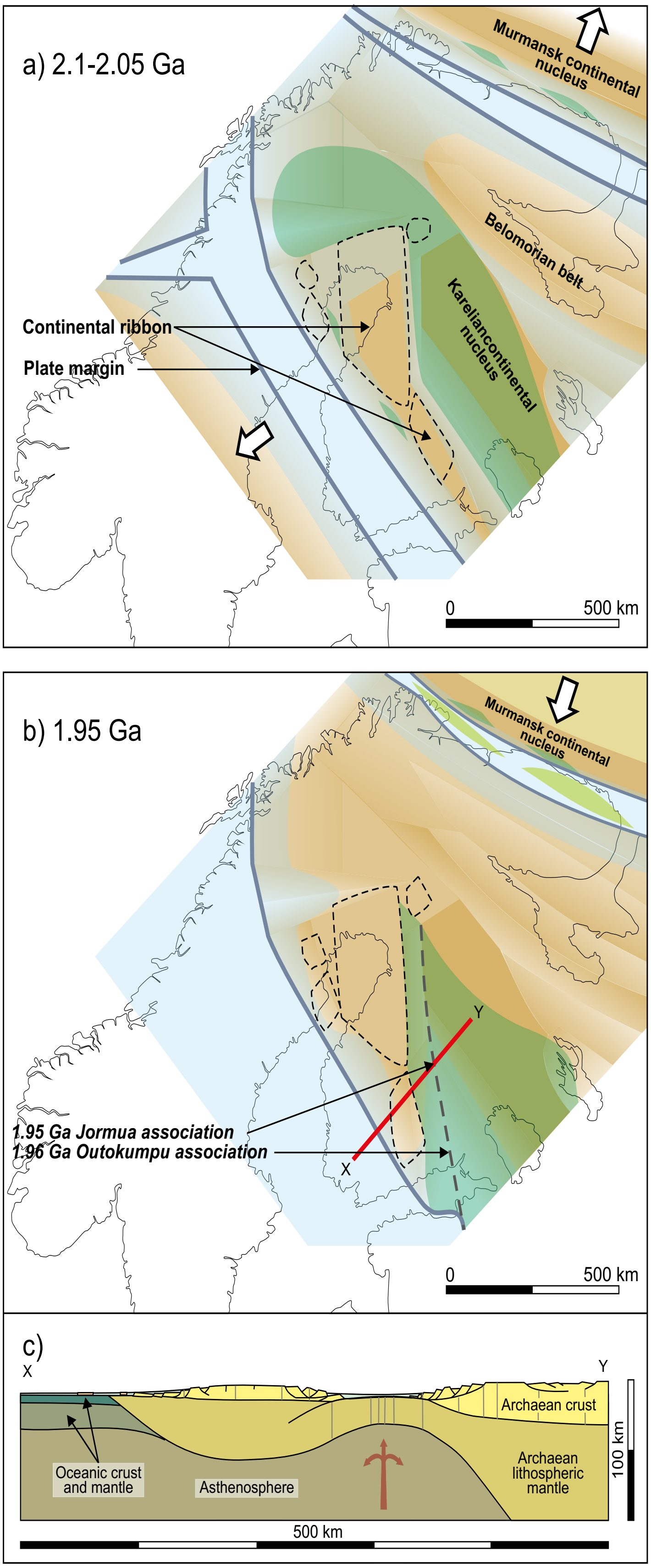


Fig. 6. a) Break-up of Archaean continental crust at 2.1–2.05 Ga. Newly separated continental ribbons and smaller extensional blocks are shown. b) Margins of the remnant Archaean continent at 1.95 Ga. Two ophiolite complexes (Outokumpu, Jormua) have developed in a V-shaped basin at the southwestern margin, and accretion begins at the northeastern margin. Rift-related volcanism is marked by darker green, arc volcanism by lighter green shading . c) Cross-section (red line in Fig. 6b) including the site of the Jormua complex upon exhumed continental mantle. No vertical exaggeration. Modified from Peltonen (2005).

A global rise in atmospheric oxygen would be expected in association with the break-up of a supercontinent (due to an increase in shelf areas). Bleeker and Ernst (2006) suggested the break-up of the supercontinent Superia at around 2.0 Ga, whereas according to Bradley (2008), the break-up occurred earlier, at around 2.3 Ga. The reasons for the global rise in atmospheric oxygen at 2.3–2.1 Ga, and the 2.3–2.06 Ga Lomagundi-Jatuli isotopic excursion remain unresolved.

Change from divergent to convergent tectonics (2.05–1.92 Ga)

Dyke swarms and sills with ages around 1.98–1.95 Ga (Vuollo & Huhma 2005, Lubnina et al. 2016), as well as carbonate rocks and metabasalts ('Ludicovian system'; Hanski & Melezhik 2013) indicate that rifting, sedimentation and magmatism continued in the remnant continent with the development of a flood-basalt province (Fig. 6b). At 1.95 Ga, protoliths of the future ophiolite complexes Jormua and Outokumpu [55] were formed. In addition to lava rocks, sheeted dykes and gabbros, the main part of the Jormua complex consists of subcontinental lithospheric mantle. Peltonen et al. (2003) concluded that Jormua represents a seafloor that developed in an initial stage of continental breakup; a mantle diapir caused extreme thinning of the crust along detachment zones and uplift of the Archaean subcontinental mantle at the surface (Fig. 6c). The Outokumpu complex, consisting of numerous mantle tectonite massifs and associated polymetallic sulphide deposits, is not as complete as Jormua, but Peltonen (2005) suggested a common origin to the Jormua complex, with a more oceanic setting; later, Peltonen et al. (2008) presented an analogy to modern sulphide deposits in slow-spreading oceanic ridges. The present basin set-up is adopted from Kohonen (1995): extreme thinning of the continental crust led to the development of a northward-propagating V-shaped basin and a continental ribbon (Iisalmi block; Fig. 6b), similar to present continental ribbons and smaller fragments at the margins of the Atlantic Ocean (Péron-Pinvidic & Manatschal 2010). In the present model, the Jormua association represents the incipiently opened northern part of the basin, whereas the Outokumpu association developed closer to the continental margin.

Apart from the 1.95 Ga ophiolite complexes in eastern Finland, and the 1.95 Ga granitoid rocks and associated extrusive rocks in northern Sweden (Knaften; Wasström 2005), there are no reliably dated rocks in the Palaeoproterozoic bedrock of Fennoscandia in the age range 1.95–1.93 Ga. Rather poorly constrained age data come from northern Finland: 1.95–1.93 Ga ages have been obtained from intrusive rocks in northernmost Finland (Meriläinen 1976), and the youngest detrital zircons in the metasediments of the Lapland granulite complex have ages around 1.94 Ga (Tuisku & Huhma 2006). Daly et al. (2006) proposed the onset of accretion at 1.95 Ga between the Lapland-Kola and Karelia continental blocks. Initial accretion, coeval with extension in southwestern margin of the remnant Archaean continent, is shown in Figure 6b. Docking of the continent Laurentia with the Murmansk continental nucleus is modelled as the reason for accretion. The possible connection between contractional and extensional tectonics in opposite parts of the Karelia continental nucleus is discussed by Lahtinen et al. (2010).

Main collisional stage (1.92–1.87 Ga)

In contrast to the scarcity of properly dated rocks in the age range 1.95–1.93 Ga, considerably more age data on 1.92 rocks exists (Lahtinen & Huhma 1997, Lahtinen et al. 2002), and 1.92 Ga presumably marks the prelude to the main Palaeoproterozoic collisional period that led to the development of a new lithospheric block – a predecessor of the Svecofennia Province.

Plate tectonic models for the western part of the Lapland–Kola orogen involve 1) subduction of an oceanic crust northwards and closure of the ocean with thrusting of the Lapland–Kola orogen upon the Karelia Province (Barbey et al. 1984, Daly et al. 2006), and 2) subduction southwards, with the metasediments and magmatic rocks of the Lapland granulite belt representing a back-arc basin fill (Berthelsen & Marker 1986, Lahtinen et al. 2005). The Lapland granulite belt, in the orogenic core, is here interpreted as a crustal pop-up structure with opposing structural vergences (Fig. 3b), although other interpretations exist (e.g. Perchuk et al. 2000). The structure along the seismic profile 1-EU suggests a Palaeoproterozoic overprint on Neoarchaean structures, with Palaeoproterozoic subduction towards the northeast. It is easier to envisage burial of the metasediments in the Lapland granulite belt to relatively high pressures (5–9 kbar; Tuisku et al. 2006) and back to the surface in a subduction zone than in a back-arc environment. Therefore, the model favoured here is with subduction towards the northeast. In this model, the Pechenga–Varzuga belt represents a rift basin in the overriding plate.

Two main tectonic models have been presented for the Svecofennian orogeny. In the simplest model, sediments were accumulated and volcanic rocks were erupted in a widespread basin inboard an active continental margin (Rutland et al. 2001, 2004, Williams et al. 2008). In more detailed versions of a single active continental margin, several island arcs were accreted during continuous subduction, with retreat of the subduction zone to the southwest (Gáal & Gorbatschev 1987, Gorbatchev & Bogdanova 1993, Bogdanova et al. 2015), or with longer periods of subduction retreat and shorter periods of advance, leading to extensional and compressional cycles in the overriding plate (Hermansson et al. 2008, Saalmann et al. 2009). The second model includes the collision of arc complexes (microcontinents) with the Archaean continent (Nironen 1997). A refined model (Lahtinen et al. 2005, Lahtinen et al. 2009a) consists of several orogenic stages and

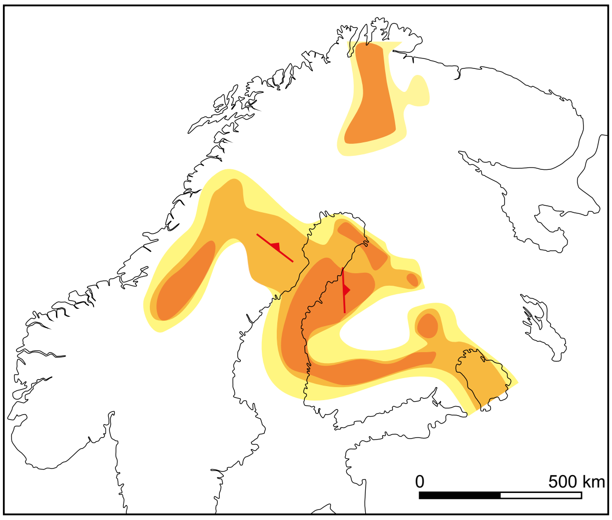
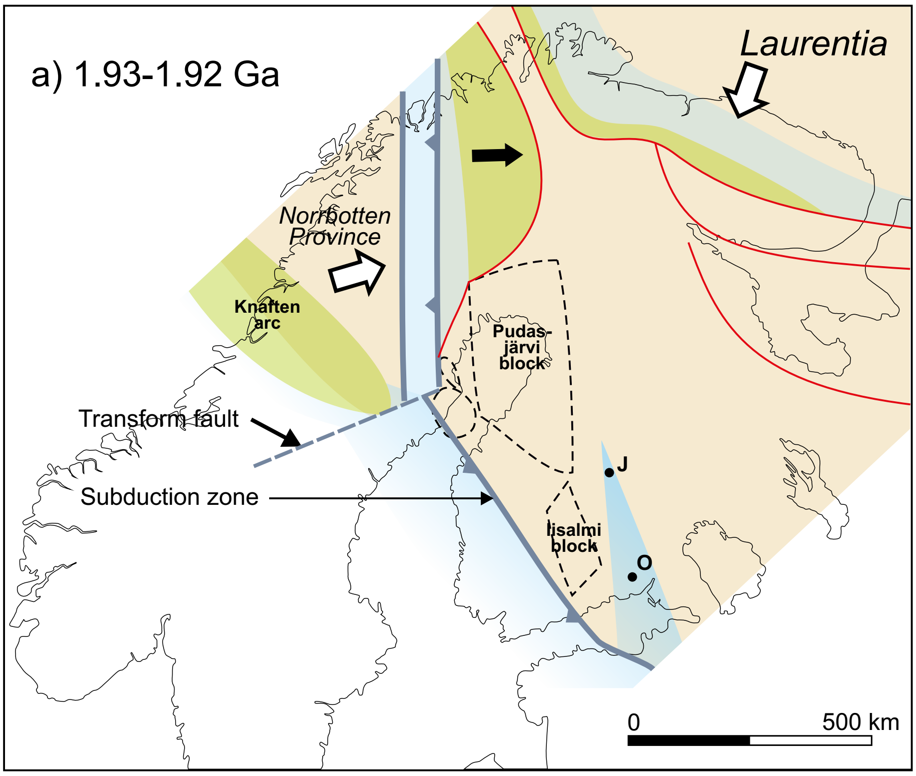


Fig. 7. Fennoscandian crustal conductance pattern at 0–60 km depth (simplified from Korja et al. 2002). Orange ≥4 $\log_{Siemens}$, light orange 3.5–4 $\log_{Siemens}$, yellow 2.5–3.5 $\log_{Siemens}$. Dip directions from Korja et al. (1993).

orogenies, including microcontinent accretion, continental extension, continental collision, and finally orogenic collapse before crustal stabilization. The existence of microcontinents comes from Nd and Hf isotopic evidence, suggesting a ≥2.0 Ga crustal component in central and southernmost Finland and in central Sweden (Lahtinen & Huhma 1997, Rämö et al. 2001, Andersson et al. 2011). In subsequent modifications of the second model, the configuration of the microcontinents was reinterepreted (Lahtinen et al. 2009b), and in order to explain the curving structural and metamorphic zoning pattern in supracrustal rocks and in crustal conductance (Fig. 7), Lahtinen et al. (2014) presented a model in which an originally linear orogen along the Archaean continental boundary was buckled into a coupled orocline. The model presented here is a refinement of the model involving several collisional events during the Svecofennian orogeny (Nironen 1997, Lahtinen et al. 2005).

At 1.93–1.92 Ga, blocks of Archaean crust and an oceanic basin (containing protolith sediments of the Lapland and Umba granulite belts and magmatic rocks of the Tersk belt; Fig. 3) were being assembled between the Karelia and Murmansk continental blocks (Fig. 8a). Lahtinen et al. (2010) proposed a scenario in which the growing mountain belt provided detritus that was transported south, into the passive margin, to form the main part of the Kaleva sediments. Partly simultaneously with



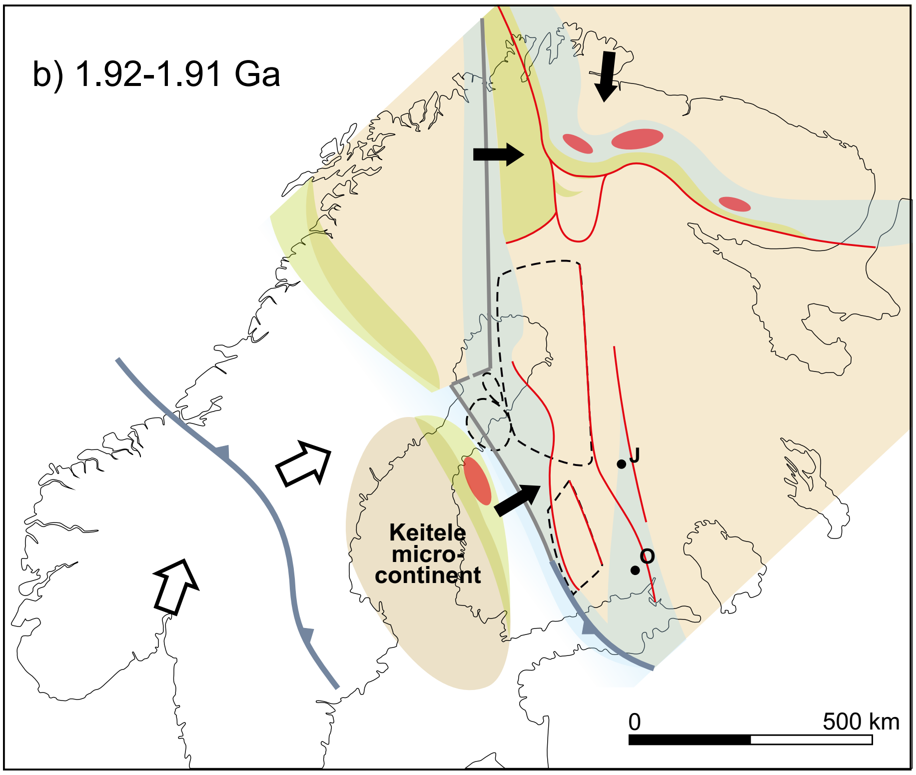


Fig. 8. a) Initiation of the main collisional stage of the Svecofennian orogeny (including Lapland–Kola area) at 1.93–1.92 Ga. Archaean crustal units are shown by a light-brown colour, Palaeoproterozoic microcontinent and areas dominated by arc-type volcanism are shown in darker brown and light green, respectively. Sedimentary belts mainly consisting of passive margin sequences are shown in blue; deposits related to continental rift are not shown. Relative plate motion is shown by a white arrow, the direction of tectonic transport by black arrow. Red lines are active deformation zones. J = Jormua association, O = Outokumpu association. b) Collision at 1.92–1.91 Ga, with jump of the subduction zone outboard of the Keitele microcontinent and subduction flip. Magmatism is shown by a red colour.

the Lapland – Kola – Karelia collision, a NNW–SSE-trending active continental margin closed at 1.92 Ga, and the Svecofennian orogeny initiated by collision of the continental blocks Karelia and Norrbotten; a volcanic arc (Knaften arc) had been attached to the Norrbotten block before continental collision (cf. Lahtinen et al. 2005). The reason for subduction towards the southwest is that there is little evidence of subduction-related magmatism in the Karelia continental margin. Remnants of oceanic crust that had existed between the two Archaean continental blocks [58–57] were sliced between passive margin sequences, and this package was thrust eastwards during collision (Kittilä allochthon–parautochthon, presently a component of the Central Lapland nappe complex; Fig. 2a). The orogenic belt developed diachronously so that at 1.92 Ga, foreland sediments of the Karelia – Norrbotten collision [25] were deposited simultaneously with accretion of a microcontinent (Keitele microcontinent) and attached volcanic arc to the Karelia continental margin (Fig. 8b). Accretion of the Keitele microcontinent caused slab break-off, subduction zone jump outboard of the accreted microcontinent, and reversal of subduction polarity along the margin of the Norrbotten block and attached magmatic arc (Knaften arc; Fig. 8b). Accretion caused a metamorphic event that culminated at ca.1.91 Ga. Kaleva sediments and tectonically enclosed fragments of ophiolitic bodies [56–55] were thrust upon the continental margin, leading to development of the Kainuu nappe complex (Figs. 3a and 8b). Sediments from the southwestern plate [67, 64] were mixed with the passive margin and early foredeep sediments to form allochthonous packages of the Northern Karelia nappe complex and Northern Ostrobothnia nappe system. Continental blocks of varying size were attached to the attenuated continental margin. Whether the blocks were continental ribbons (Pudasjärvi and Iisalmi blocks) or smaller blocks, their mantle components were detached from the crust during attachment (Fig. 4).

To explain the development of the 1.89–1.88 Ga Tampere belt (Fig. 1), with a rather simple E–W-trending basin structure (Kalliomäki et al. 2014 and references therein), Nironen (1997) and Lahtinen et al. (2005) modelled a change in plate movement direction (Figs. 8b, 9a). Another microcontinent (Bothnia microcontinent) approached from the south, and a transform fault developed between the two microcontinental blocks. Subduction in opposite directions caused gradual subduction of the oceanic plate in the east (Fig. 9b). A third microcontinent (Bergslagen microcontinent) approached from the south, leading to orthogonal collision of the Berslagen and Keitele microcontinents (Fig. 10a). In the west, a change in subduction polarity towards north is modelled to explain the north-dipping mantle reflector in the BABEL 1 profile (Fig. 4b; cf. Korja & Heikkinen 2005).

Age data indicate magmatism and deformation already occurring at 1.89 Ga, but the peak of early Svecofennian deformation, metamorphism and magmatism occurred at 1.88–1.87 Ga, when the two microcontinents Bergslagen and Keitele collided (Fig. 10a). As the result of this collision, the Pudasjärvi and Iisalmi blocks were thrust northwards, and foreland sequences [24] were deposited in front of the thrust zones. As modelled by Pajunen et al. (2008), the Keitele microcontinent started to rotate dextrally, coevally with mafic-felsic magmatism [43–39]. The Bothnia microcontinent docked at 1.87 Ga, finishing subduction in central Sweden (Fig. 10b). The oroclinal bend developed as the result of rotation of the Keitele microcontinent and docking of the Bothnia microcontinent.

Assembly of the lithospheric blocks to form the protocontinent Fennoscandia (1.87–1.86 Ga)

At 1.87 Ga, the continental and microcontinental blocks and remnants of oceanic crust as well as intervening arc and passive margin sequences were united to form a coherent continental block – the protocontinent Fennoscandia. The term protocontinent Fennoscandia means here the continental block before attachment to Volgo-Sarmatia.

Figure 10b shows the configuration of the microcontinental blocks at 1.87 Ga. The northern suture explains the curving structural-metamorphic pattern in supracrustal rocks and in crustal conductance (cf. Fig. 7). The blocks are not everywhere in contact with each other, but remnants of oceanic crust and volcanic-sedimentary arc material are squeezed between and overlying the continental blocks, following the interpretation of seismic profiles (Fig. 4). The Keitele microcontinent extends northwest as a thin layer at the bottom of the crust and is almost in contact with the Bothnia microcontinent (Fig. 4a). The boundary between the Western and Southern Finland Subprovinces roughly follows the suture between the Keitele and Bergslagen microcontinents.

Svecofennian collisional orogenic evolution during 1.87–1.86 Ga was diachronic. Peak metamorphic

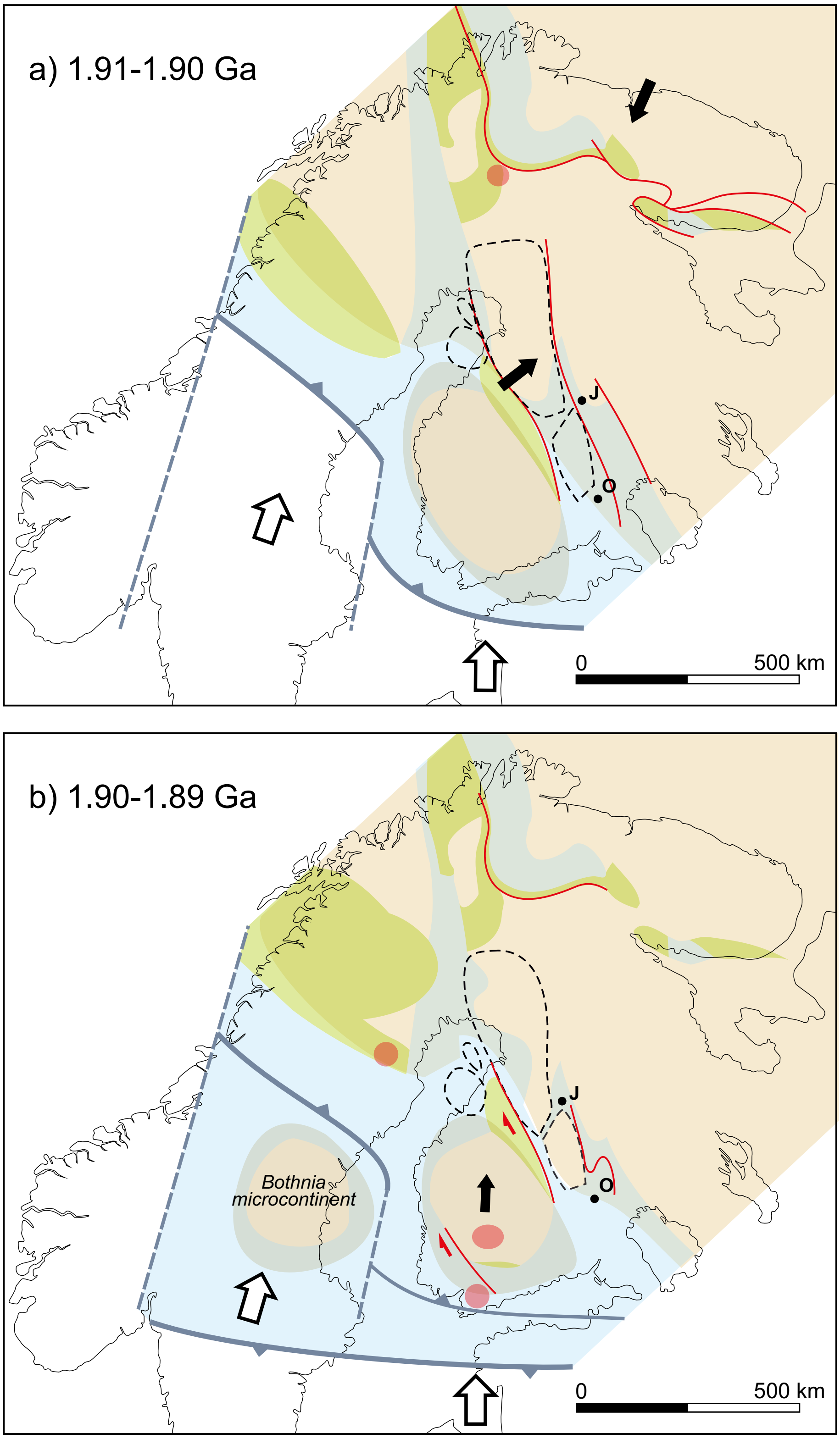


Fig. 9. Early Svecofennian accretion. a) 1.91–1.90 Ga. b) 1.89 Ga. Thick and thinner blue-grey lines denote an active subduction zone and inactive zone (suture), respectively. For colours and other symbols, see Figure 8.

conditions were attained in central Sweden and Finland at 1.88–1.87 Ga, with the development of the metamorphic-intrusive Vaasa complex [37–36] (Figs. 1 and 10b; Högdahl et al. 2012, Mäkitie et al. 2012, Kotilainen et al. 2016). Contractional deformation in central Finland changed into transtensional at ca. 1.87 Ga, with lateral spreading and thinning of the middle crust (Korja et al. 2009) and emplacement of an A-type granitoid suite [38] along transtensional shear zones (Nironen 2005),

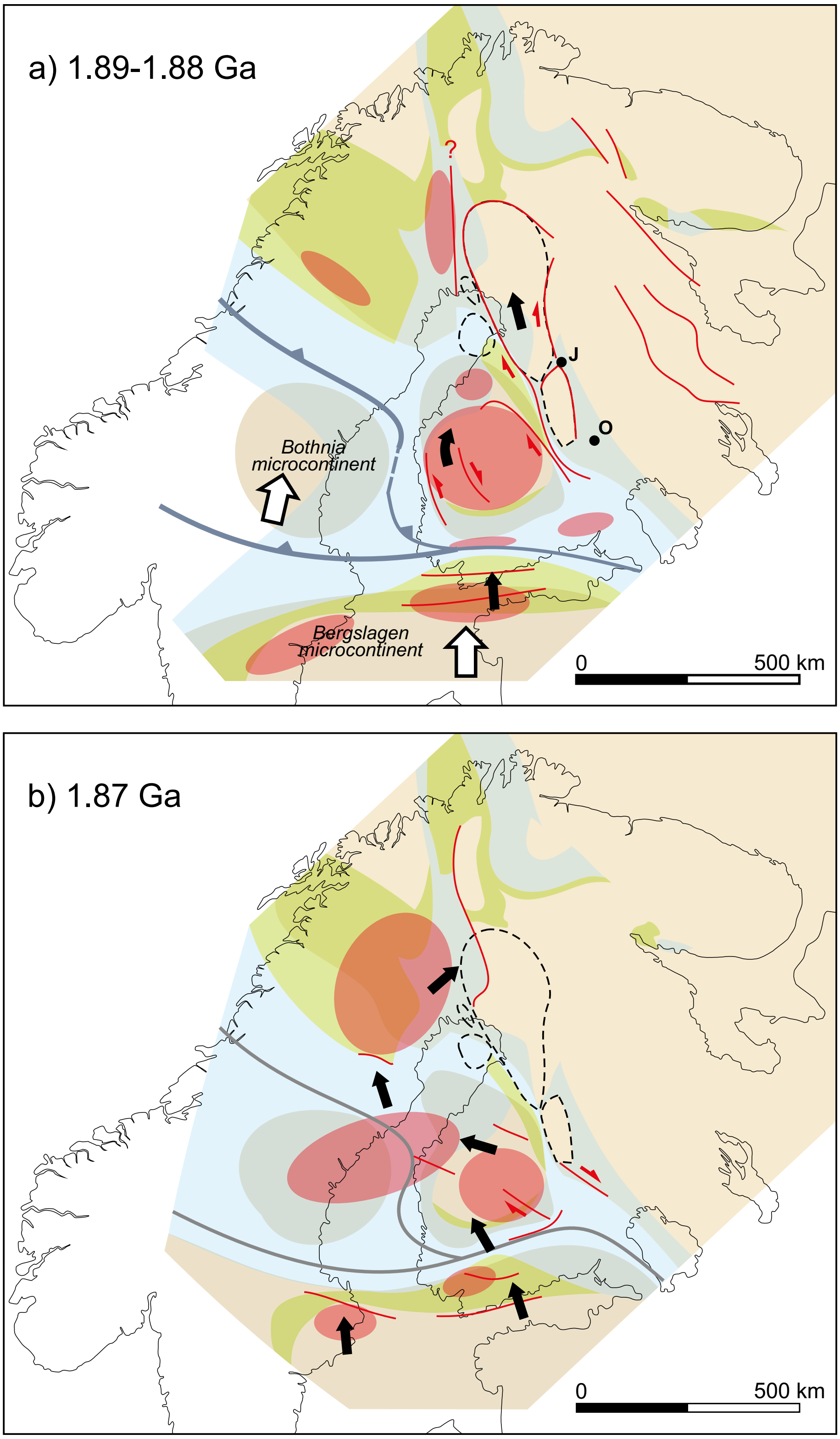


Fig. 10. Early Svecofennian accretion a) 1.89–1.88 Ga. b) 1.87 Ga. For colours and symbols, see Figure 8.

while contractional deformation still prevailed in southern Finland (Väisänen et al. 2002, Skyttä et al. 2006). In south-central Sweden, the metamorphic peak was only attained at 1.86 Ga (Stephens & Andersson 2015).

Fennoscandia attained its maximum crustal thickness during the 1.92–1.87 Ga collisions. The thickest crust developed in the area extending from the Savo belt to the Bothian Sea between Finland and Sweden, e.g. in the area where the microcontinents accreted (Fig. 10), and the crustal thickness in this area is still > 50 km (Lahtinen et al. 2009a, Bagherbandi et al. 2015).

In a global context, the main collisional period has been assigned to the amalgamation of the supercontinent Columbia (also referred to as Nuna or Hudsonland) at 1.9–1.8 Ga (e.g. Zhao et al. 2004, Rogers & Santosh 2009, Evans & Mitchell 2011). A recent study of dykes in central Finland (Klein et al. 2016) suggests that at 1.87 Ga, Fennoscandia (Baltica in the palaeomagnetic literature) was at low latitudes (ca. 20°) and had not yet been assembled with Laurentia to form Columbia.

Intra-orogenic stabilization and late Svecofennian crustal extension, continued accretion in the south (1.86–1.83 Ga)

Metasandstones, including orthoquartzites [20], have a scattered occurrence in southern Finland, and similar rocks are found in south-central Sweden (Fig. 11a). Their depositional age may be constrained by detrital zircon data between 1.87 Ga and 1.83 Ga (Bergman et al. 2008). The occurrence of ultramature quarzite (Lahtinen & Nironen 2010) implies upper crustal stabilization and deposition of quartz sands in a continental environment.

The period referred to as late Svecofennian initiated at 1.84 in southern Finland by crustal extension and associated mafic to felsic magmatism (Nironen 2005, Pajunen et al. 2008). The Late Svecofennian period is characterized by high-T metamorphism and associated granite magmatism [16]. The high-grade metamorphism and magmatism are thought to be the result of crustal extension, leading to high heat flow from the mantle (Korsman et al. 1999, Lahtinen et al. 2005), or of the combined effect of crustal thickening and radioactive heat production (Kukkonen & Lauri 2009).

Again, evolution was diachronic: at 1.83 Ga when crustal extension prevailed in southern Finland and quartz sands were redeposited in intracontinetal rift valleys (Fig. 11a; Skyttä & Mänttäri 2008, Nironen & Mänttäri 2011), a volcanic arc (the Oskarshamn–Jönköping belt in southern Sweden) was developing at a continental margin (Fig. 11b; Mansfeld et al. 1995). This volcanic arc probably has a continuation in Lithuania (Skridlaite & Motuza 2001, Siljauskas & Skridlaite 2016). Following the general idea of Gorbatschev and Bogdanova (1993), Lahtinen et al. (2005) proposed retreating of an Andean-type active margin to explain this southwestward crustal growth.

Little is known of tectono-metamorphic evolution in northern Finland during 1.86–1.83 Ga. A granite in the Central Lapland granitoid complex (Fig. 1) yielded a 1.85 Ga age (Ahtonen et al. 2007), and overgrowths in detrital zircons indicate a metamorphic event at 1.86–1.85 Ga in western Finnish Lapland (Lahtinen et al. 2015b). Both crustal contraction and extension can explain these observations.

Late collisions, orogenic collapse and formation of the continent Baltica (1.82–1.76 Ga)

Late Svecofennian deformation, characterized by NW–SE trending shear zones (Fig. 11b), has been attributed to transpression with bulk compression towards the NW (Fig. 11b; Väisänen & Skyttä 2007 and references therein). Deformation was most intensive in southeastern Finland, where an allochthonous unit developed (Southern Savo nappe system, Fig. 2a). A broad, dextral shear zone developed in southern Fennoscandia, extending from Estonia to the archipelago of SW Finland and possibly to central Sweden (Fig. 11; Torvela et al. 2008, Högdahl et al. 2009). The amount of displacement along this zone is unknown.

Lahtinen et al. (2005) modelled continental collision at 1.84–1.80 Ga (Svecobaltic orogeny) to explain the late Svecofennian deformation and metamorphism in southern Finland. Oblique collision between two protocontinents Fennoscandia and Volgo-Sarmatia caused transpression in southern Finland; transpressional deformation initiated earlier but collision occurred at 1.82–1.80 Ga (Fig. 11b). Together, Fennoscandia and Volgo-Sarmatia formed the continent Baltica (or East European Craton; Gorbatschev & Bogdanova 1993, Bogdanova et al. 2015). Hereafter, Fennoscandia refers to part of the continent Baltica. Southward–southeastward accretionary growth of Fennoscandia continued during and after the collision.

The late Svecofennian metamorphism and deformation partly overlap in age with voluminous magmatic activity that produced the N–S trending Transscandinavian igneous belt in Sweden (TIB;

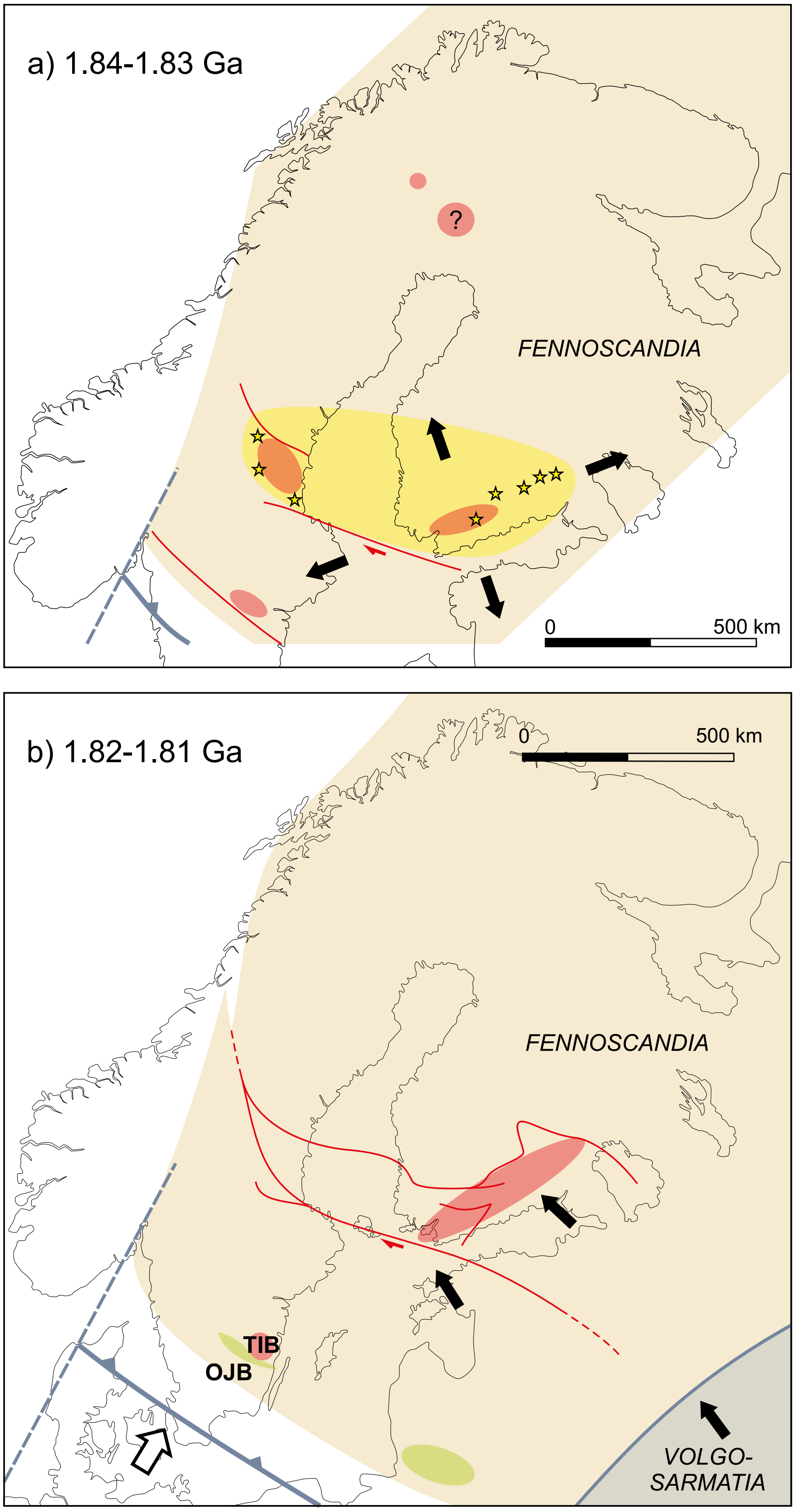


Fig. 11. a) Late Svecofennian extension (1.84–1.83 Ga). b) Late Svecofennian collision (1.82–1.81 Ga). Quartzite occurrences are shown by stars, inferred distribution of mature sands by yellow, (mainly granitic) magmatism by orange (from Åhäll and Larsson 2000, Pajunen et al. 2008, Lahtinen and Nironen 2010). Active deformation zones are shown as red lines (partly modified from Torvela et al. 2008, Högdahl et al. 2009). TIB = Transscandinavian igneous belt magmatism, OJ = Oskarshamn-Jönköping belt.

Gorbatschov and Bogdanova 1993). The TIB has been considered to represent magmatism in a convergent continental margin environment in three cycles: 1.81–1.77 Ga, 1.72–1.69 Ga and 1.68–1.66 Ga (Åhäll & Larson 2000). Lahtinen et al. (2005) proposed collision (Nordic orogeny) at 1.82–1.79 Ga between Baltica and Amazonia to explain TIB magmatism. Another possibility, presented by Lahtinen et al. (2009) and adopted here, is that the western margin of Fennoscandia was the convergent margin of an advancing Andean-type accretionary orogen (Fig. 12a). Models of subduction during TIB magmatism vary from a single curving zone (Andersson et al. 2004) to two zones (Lahtinen et al. 2005, this study); overall, the modelled plate kinematic arrangement at the northwestern margin of Fennoscandia (Figs. 11 and 12) is speculative because of the subsequent orogenic overprint.

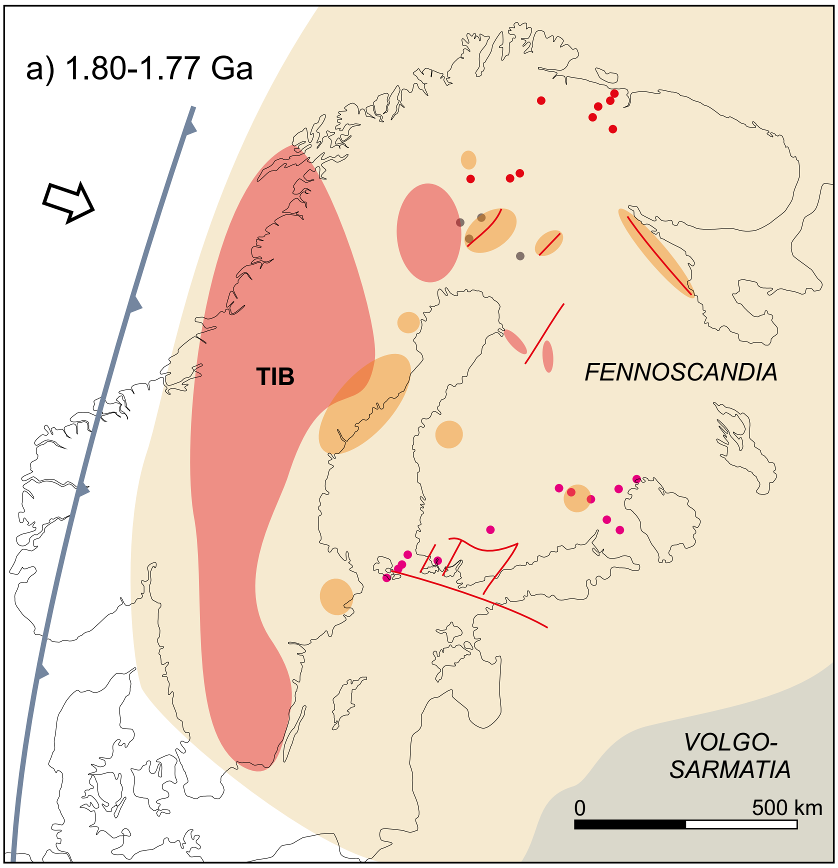
In the model of Lahtinen et al. (2005), gravitational collapse that followed the two (proto)continental collisions, Fennoscandia–Volgo-Sarmatia and Fennoscandia–Amazonia, caused large-scale crustal extension, metamorphism and magmatism in the northern part of Fennoscandia. In northern Finland, a suite of mafic rocks [18] and another one of granites [17], and in southern Finland a suite of bimodal shoshonitic rocks [16], all in the age range 1.81–1.77 Ga, represent post-collisional magmatism (Fig. 12a; Nironen 2005, Ahtonen et al. 2007, Heilimo et al. 2009). Ages of 1.80–1.76 Ga have been obtained from pegmatites and migmatite leucosomes in Sweden, Finland and Russia (Lahtinen et al. 2005, Ahtonen et al. 2007), indicating late metamorphism that affected Fennoscandia. These late ages may be attributed to crustal extension: the FIRE4 reflection seismic profile reveals a horizontal, strongly reflective zone at the base of the crust in central Lapland (Patison et al. 2006), which is probably the result of lower crustal extension.

The palaeogeographic reconstructions for the supercontinent Columbia include an initial assembly of Laurentia, Baltica and Amazonia by 1.8 Ga (e.g. Johansson 2009). Here, the name Baltica is used to mean the terrane that was assembled during the Palaeoproterozoic (East European Craton), and was a part of various continental assemblages or a separate continental unit from the Palaeoproterozoic to the present (cf. Bogdanova et al. 2008).

Crustal stabilization (1.76–1.65 Ga)

During 1.72–1.66 Ga, accretionary orogenic activity continued at the southwestern margin of Fennoscandia and produced TIB magmatism (Andersson et al. 2004). Little is known of the interior of Baltica during the period 1.76–1.65 Ga, but an indication of crustal stabilization comes from the Suursaari island in the Gulf of Finland (Fig. 13), where an ultramature (quartz arenitic) conglomerate, containing quartz arenitic boulders, lies nonconformably upon Svecofennian basement (Pokki et al. 2013a). Correlative occurrences are found in southern Finland and central Sweden, and together these suggest deep erosion of the Svecofennian crust (migmatitic basement) and development of a large peneplain (quartz arenite) in the interior of Baltica before the emplacement of the oldest rapakivi granites at 1.65 Ga (see below). The depositional age of the Suursaari conglomerate can be bracketed between 1.65 Ga and 1.63 Ga. Deposition of the conglomerate marks a shift from tectonically stable conditions to incipient intracontinental rifting.

A problem still unresolved is that the Svecofennian metasedimentary rocks generally yield 4–6 kbar pressures, corresponding to 10–15 km depth, and yet there is little evidence as to when and how these rocks were exhumed. Exhumation occurred between 1.80 Ga and 1.65 Ga, and preferably in the earlier part of this period considering the tectonically stable conditions in the later part. Was the mechanism large-scale normal faulting and associated erosion during orogenic collapse? Where is the eroded material?



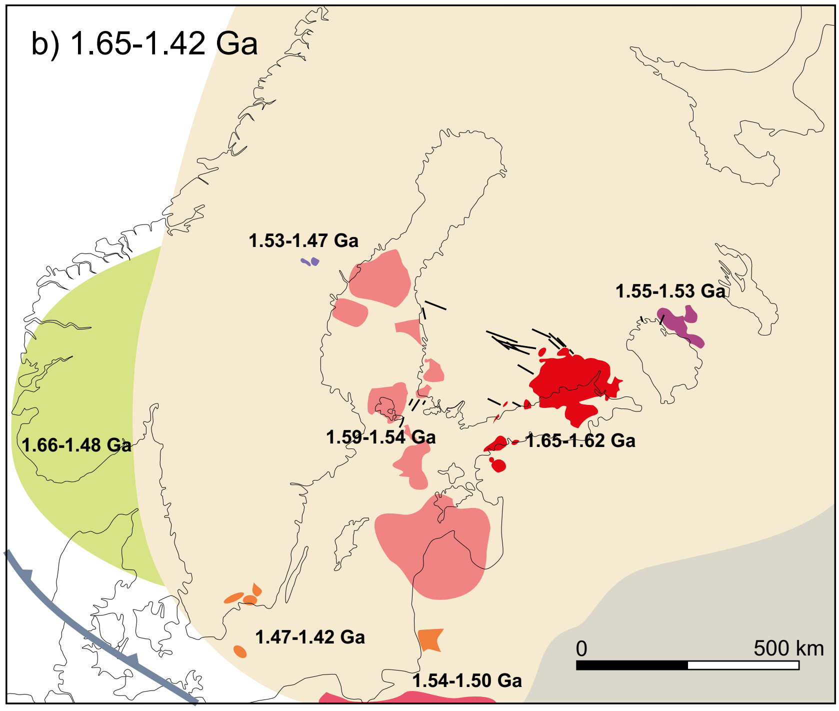


Fig. 12. a) Magmatism in Fennoscandia during 1.80–1.77 Ga (post-collisional magmatism in Finland). Plutonic bodies shown by points (magenta = bimodal, brown = mafic, red = granitic; Haapala et al. 1987, Eklund et al. 1998, Eklund and Shebanov 2005, Ahtonen et al. 2007, Heilimo et al. 2009). Red areas show granitic magmatism (Åhäll and Larsson 2000), orange areas sites of metamorphism and intrusion of pegmatites (Romer and Smeds 1994, 1997, Alviola et al. 2001, Bibikova et al. 2001, Ahtonen et al. 2007). TIB = Transscandinavian igneous belt. Active deformation zones are shown by red lines. b) A-type magmatism in Fennoscandia during 1.65–1.42 Ga (Laitakari et al. 1996, Koistinen et al. 2001, Bogdanova et al. 2008, Brander and Söderlund 2009). Plutons with red and bluish colours are rapakivi granites, plutons with an orange colour are related to Hallandian–Danopolonian orogeny. Age ranges for each colour are given. The green area represents Gothian and Telemarkian accreted units (Bingen et al. 2008) before the Sveconorwegian orogeny.

Intracontinental rifting and magmatism, orogenic activity at the continental margin (1.65–0.9 Ga)

There is evidence of rifting and magmatism in the interior of Baltica during 1.65–1.1 Ga, and orogenic activity in the southwestern margin during 1.66–1.42 Ga. A bimodal association of rapakivi granites [13–12], gabbro-anorthosites [14] and dykes is found in onshore and offshore southern Finland and central Sweden (Fig. 13). The age of rapakivi magmatism spans 1.65–1.47 Ga, with four age groups (Fig. 12b; Rämö & Haapala 2005, Rämö & Mänttäri 2015). The rapakivi granites are part of a more extensive anorthosite-mangerite-charnockite-granite (AMCG) suite of that includes the 1.54–1.50 Ga bimodal Mazury complex in Lithuania and northeastern Poland. The mafic dyke swarms in southern Finland have two main orientations: the WNW-ESE-oriented dykes are coeval with the oldest (1.65–1.62 Ga) granites, whereas the NNE-SSW-oriented ones are associated with younger granites. The established concept is that the rapakivi magmas were derived in an extensional setting from the lower crust, whereas the mafic magmas had a variably crustal-contaminated mantle origin. However, the ultimate cause of the bimodal magmatism is controversial (Rämö & Haapala 2005).

Sandstone [11], found in the western coast of Finland but mainly occurring as fillings in offshore basins and Lake Ladoga, flank or overlie the magmatic rocks (Fig. 13). Based on the BABEL 1 seismic profile, Korja et al. (2001) concluded that the sandstone reaches 3–4 km thickness in the Bothnian Bay area (see northern part of Fig. 4b), and suggested that the basin initiated as a cauldron subsidence on top of the rapakivi intrusions. Rifting, sedimentation and magmatism probably occurred intermittently (Pokki et al. 2013b).

Mesoproterozoic clastic sedimentary rocks [9] are preserved in a fault-bounded basin in the western coast of Finland (Fig. 13), and continue into the Bothnian Bay area (Kohonen & Rämö 2005, Paulamäki & Kuivamäki 2006). Drilling at the 75 Ma Lappajärvi meteorite impact site [1] (Fig. 13) revealed a 20-m-thick pile of Mesoproterozoic sedimentary rocks [9] below and around the impact rocks, suggesting that Mesoproterozoic sediments covered much wider areas than the preserved occurrences. Late Mesoproterozoic evolution involved short extensional periods, as exemplified by the few 1.3–1.1 Ga mafic dykes found in northern Finland, as well as the 1.27 Ga dolerite sills and dykes in southwestern Finland [10].

Partly coeval with rapakivi magmatism, several allochthonous terranes, presently comprising the Sveconorwegia Province (Fig. 1), were accreted to the western margin as two events, first at 1.66–1.52 Ga (the Gothian event; Åhäll & Connelly 2008), and at 1.52–1.48 Ga (the Telemarkian event, Bingen et al. 2008). A third event, referred to as the Hallandian-Danopolonian orogeny (Bogdanova et al. 2008), extended from southern Sweden to Lithuania and Poland and resulted in AMCG magmatism during

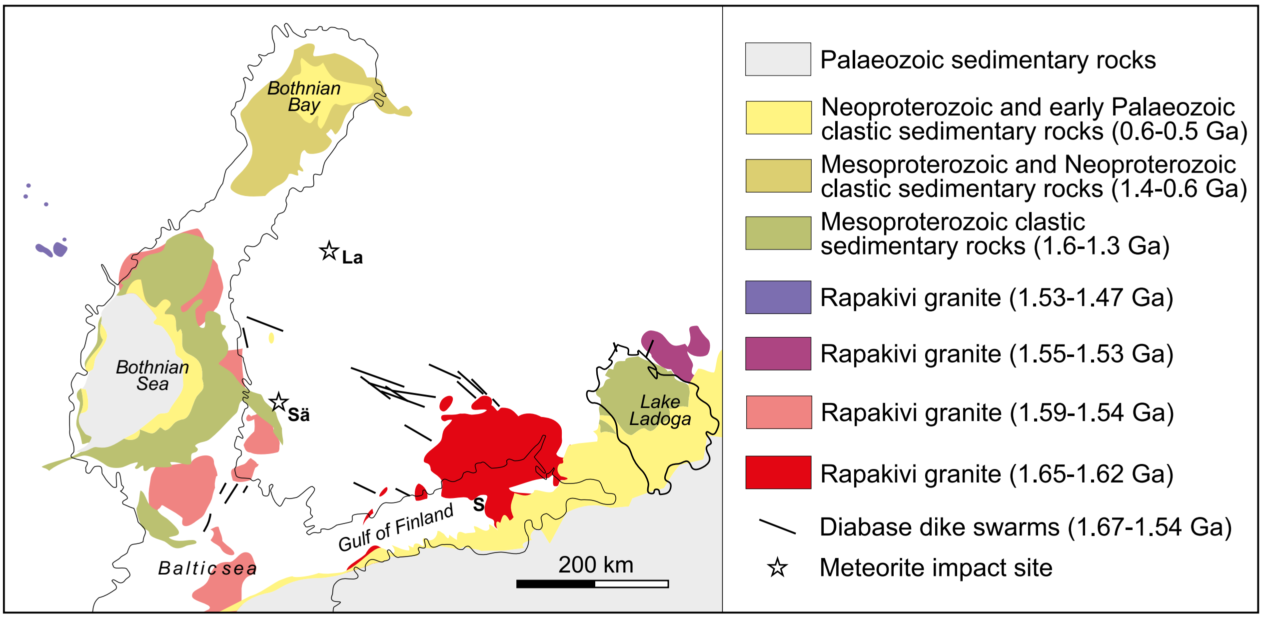


Fig. 13. Mesoproterozoic and younger lithological units in the eastern part of the Fennoscandian Shield (generalized from Koistinen et al. 2001). Dyke swarms associated with rapakivi magmatism are shown crudely. S = Suursaari island, La = Lappajärvi impact site, Sä = Sääksjärvi impact site.

1.47–1.42 Ga. In Figure 12b, a subduction zone is delineated to explain the Gothian, Telemarkian and Hallandian–Danopolonian events. The orientation of subduction would match with development of the two dyke swarms associated with rapakivi magmatism (older parallel, younger orthogonal). The episodic activity and different orientations of subduction in southwestern Fennoscandia could explain inboard rapakivi magmatism and the non-linear age distribution, as suggested by Åhäll et al. (2000).

As postulated above, a shallow basin probably developed in central Fennoscandia during the rapakivi magmatic event. Emplacement of the 1.27–1.1 mafic dikes and sills preceded the 1.14–09 Ga Sveconorwegian (–Grenvillian) orogeny that was prominent in the area of southern Norway and southwestern Sweden (Bingen et al. 2008, Roberts & Slagstad 2015). The dykes, sills and normal faults bounding the present occurrences of Mesoproterozoic cover rocks are probably expressions of foreland extension during the initial stage of the Sveconorwegian orogeny. During the orogeny, the basin deepened into a foreland basin that received huge masses of detritus from the orogenic front area (Fig. 14a; Kohonen & Rämö 2005). Extension of the basin to the east is speculative.

In some palaeogeographic reconstructions (e.g. Zhao et al. 2004), AMCG magmatism in Baltica (1.65–1.45 Ga) was part of the initial fragmentation of the supercontinent Columbia, whereas in other ones the maximum assembly of Columbia was achieved as late as 1.5 Ga (Salminen et al. 2017). The final break-up probably occurred at 1.2 Ga. Plate organization during the break-up of Columbia and re-assembly into the supercontinent Rodinia at 1.1–1.0 Ga, with Baltica as one component, appears to include a complex set of continental extensions, accretions and collisions, including the Sveconorwegian orogeny (cf. Bogdanova et al. 2008, Johansson 2009, Pisarevsky et al. 2014, Roberts & Slagstad 2015). The degrees of freedom in the palaeomagnetic method are evident in the variable reconstructions of the Columbia – Rodinia reorganization (Meert 2014).

Neoproterozoic uplift and denudation (900–640 Ma)

Early- and Mid-Neoproterozoic (Tonian–Cryogenian, 1000–640 Ma) rocks are missing in onshore Finland, but exist in the Bothnian Bay area (Fig. 13). In Estonia, late Neoproterozoic (Ediacaran or 'Vendian') or Cambrian sedimentary rocks occur directly upon the crystalline Svecofennian basement (Puura et al. 1996). Based on deep drilling, a kaolinitic palaeosol, with a thickness of 0.5–2 m, covers the Svecofennian basement, indicating an ancient peneplain. In addition to Estonia, evidence for the peneplain, referred to as Pre-Vendian, Sub-Vendian or Sub-Cambrian, has been detected in southern Norway, southern Sweden, southern Finland, and in the Kola Peninsula area (Riis 1996, Kohonen & Rämö 2005, Lidmar-Bergström et al. 2013, Hall 2015).

The peneplain indicates crustal uplift, removal of most of the foreland sediments that had been deposited during the Sveconorwegian orogeny, and exposure of the crystalline basement. Gradual uplift and denudation occurred during most of the Neoproterozoic and resulted in a large, virtually flat shield area at the end of the Cryogenian Period (Fig. 14b). The northwestern margin of Baltica was a rift margin, and shallow rift basins (including the Bothnian Bay basin) formed or were reactivated in northwestern Baltica, whereas Mesoproterozoic aulacogens in the east (present East European Platform area) were filled with near-shore and shallow-marine sediments (Grachev & Nikolaev 2006, Pease et al. 2008). The northeastern margin of Baltica was a passive margin throughout the Neoproterozoic, filled with a km-scale pile of sediments.

The position and orientation of Baltica in the supercontinent Rodinia in the Neoproterozoic is controversial (e.g. Meert & Torsvik 2003, Cawood & Pisarevsky 2006). In the palaeogeographic reconstruction of Li et al. (2008), Baltica was within moderate southern latitudes during the break-up of the supercontinent Rodinia at 750–600 Ma.

Late Neoproterozoic and early Palaeozoic subsidence and platform sedimentation (640–430 Ma)

Late Neoproterozoic (Ediacaran) and Palaeozoic rocks are virtually missing in onshore Finland: the Lauhavuori sandstone [8] in western Finland is possibly Early Cambrian (Paulamäki & Kuivamäki 2006), and Cambrian sedimentary rocks [4] overly the Archaean basement in the northwestern tip of Finland. In contrast, an Ediacaran to Devonian sequence covers Precambrian basement rocks in the

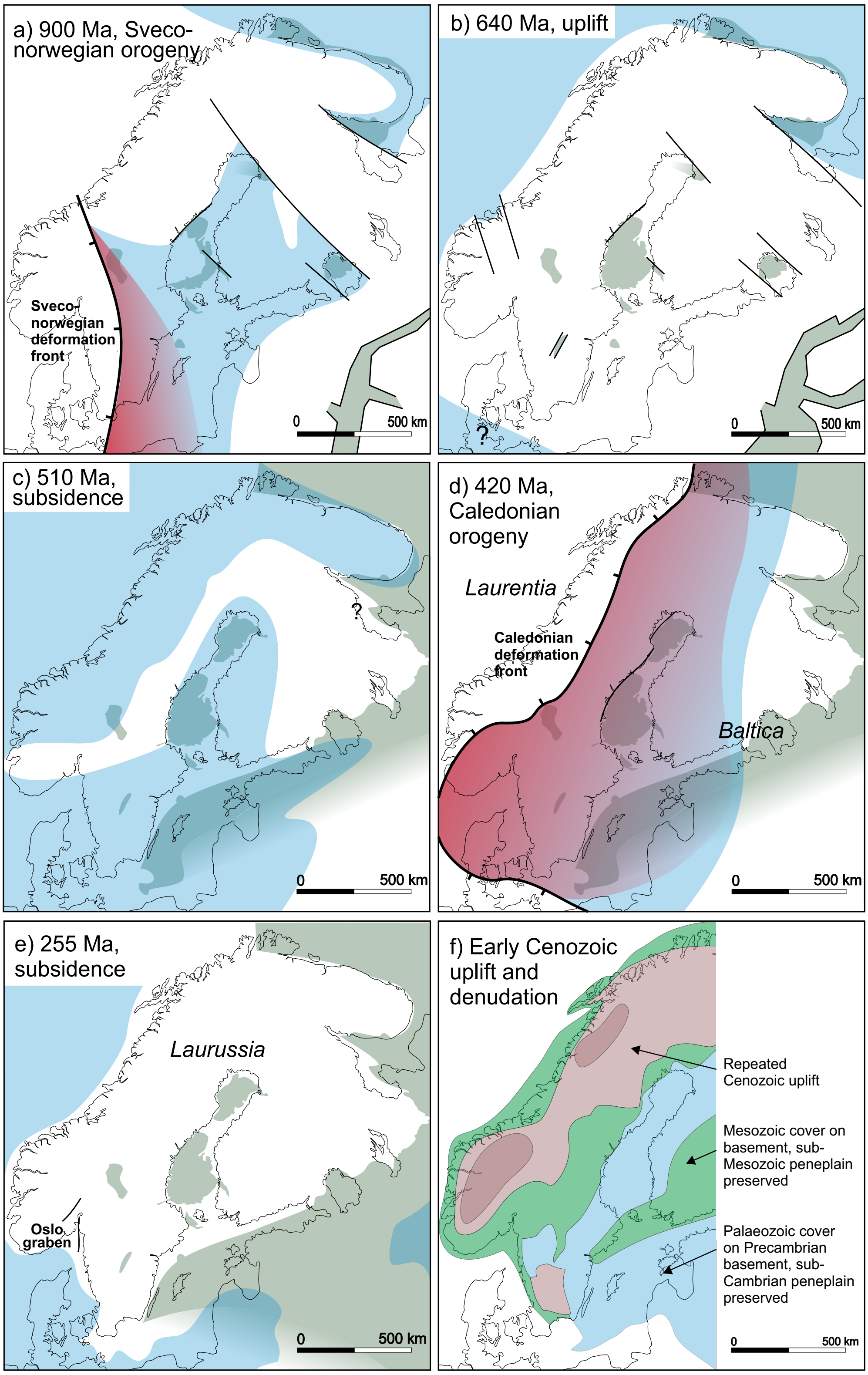


Fig. 14. Model of a few stages in the evolution of the platform cover in Fennoscandia. Light greenish areas indicate the present distribution of cover rocks that had been deposited during each palaeogeographic snapshot, transparent blue areas show the inferred extent of the epicontinental sea or oceanic shelf, red areas represent orogenic depositories. Black lines denote inferred faults. a) 900 Ma, end of Sveconorwegian orogeny. Modified from Kohonen and Rämö (2005). b) 640 Ma, final stage of crustal uplift in Fennoscandia. Modified from Vidal and Moczydlowska (1995) and Bogdanova et al. (2008). c) 510 Ma, maximum flooding stage. Modified from Nielsen and Schvovsbo (2011). d) 420 Ma, Caledonian orogeny. Modified from Kohonen and Rämö (2005). Note that the Caledonian frontal thrust is drawn where it is now; it was probably located elsewhere at 420 Ma. e) 255 Ma, Permian subsidence. Modified from Ruebsam et al. (2017). f) Early Cenozoic uplift and denudation. Modified from Lidmar-Bergström et al. (2013).

Baltic countries: the thickness of the sequence increases southwards from northern Estonia up to ca. 2000 m in Latvia, where some 800 m of Devonian rocks form the uppermost part (Puura et al. 1996). In the Bothnian Bay area (Fig. 13), the uppermost sandstone is probably Ediacaran in age (Paulamäki & Kuivamäki 2006). Cambrian sandstone and Ordovician limestone and shale are found in the Bothnian Sea area.

The Ediacaran and early Palaeozoic sedimentation has been assigned to the opening of the Iapetus Ocean and development of a passive continental margin in northwestern Baltica (Kohonen & Rämö 2005). During the Ediacaran, central Baltica slowly submerged under an epicontinental sea, the Palaeobaltic Basin, and Ediacaran sedimentation extended to (present-day) Estonia. Ediacaran sedimentation in the area of Finland is controversial: the sequences in northern Estonia suggest that sedimentation also occurred in southern Finland, whereas Ediacaran and Early Cambrian sediments in Estonia were supplied by detritus from the Precambrian basement and rapakivi granites, suggesting that most of Finland was land area during the Ediacaran and beginning of the Cambrian (Isozaki et al. 2014). During the Early Cambrian, transgression in Fennoscandia and regression in central Baltica included a series of subsidence and uplift events of the order of tens to a maximum of hundreds of metres (Nielsen & Schovsbo 2011); Figure 14c shows the interpreted Early–Middle Cambrian maximum flooding stage.

During the Ordovician and Silurian a carbonate platform extended from southern Norway to the St. Petersburg area, with a deeper shelf that opened southwest to the surrounding ocean. The Cambrian and Ordovician rocks in the Bothnian Sea area are tectonically preserved remnants of an originally extensive Cambrian–Ordovician sequence that covered much of Finland. It is speculative whether the carbonate platform extended to the area of Finland during the Silurian (cf. Baarli et al. 2003, Kohonen & Rämö 2005).

The early Palaeozoic subsidence has been assigned to the isolation of Baltica from Laurentia and opening of the Iapetus Ocean at 600–550 Ma (Pease et al. 2008, Li et al. 2008). Palaeomagnetic and geological data suggest that Baltica was at high southern palaeolatitudes during the early Ediacaran (Li et al. 2008), and drifted northwards to low palaeolatitudes during the Cambrian and Ordovician; during the mid-Silurian (425 Ma), Baltica was within tropical latitudes (Cocks & Torsvik 2005). From 450 Ma onwards, continental collisions occurred at the western margin of Baltica (see below).

Caledonian and Variscan orogenies and development of the Permian basins (430–250 Ma)

Late Palaeozoic rocks have mainly been recognized in southwestern Fennoscandia: Silurian to Devonian sedimentary rocks occur in Latvia, and Permian rocks are found in the Oslo graben in Norway. However, fission track studies (e.g. Cederbom et al. 2000) suggest that late Palaeozoic deposits covered large areas of Fennoscandia, presumably as a result of the Caledonian orogeny. The Caledonian orogeny initiated in the early Ordovician (490–470 Ma), but the main, Scandian phase lasted from the Silurian to the Devonian (430–390 Ma; Roberts 2003, Gee et al. 2008). The foreland fold and thrust belt sequence, with eastward lateral transport of allochthonous units for several hundreds of kilometres, dominates the geology of Norway, but barely extends to the northwestern part of Finland (Fig. 2). The allochthonous nappe units [7–5] build up the Finnish part of the Caledonides (Kohonen & Rämö 2005).

As the result of the Caledonian orogeny, a foreland basin formed and received detritus from the growing mountain belt (Fig. 14d). Detritus probably buried the foreland basin, including most of Finland, with an estimated thickness of >2.5 km in western Sweden, thinning to 1.5–0.5 km in Finland (Larson et al. 1999).

The Variscan (or Hercynian) orogeny formed a continuation with the Caledonian orogeny, with initiation set at 400 Ma and ending at 300 Ma (e.g. Kroner & Romer 2013). The orogeny severely modified the crust in the present central and western Europe south of the Trans-European Suture Zone (Fig. 1b), whereas the effects in Baltica were mainly restricted to the southwestern margin as transpressional deformation in the Trans-European Suture Zone. Fission-track analysis suggests a late Carboniferous (314–307 Ma) uplift and exhumation event in southern Sweden (Jaspen et al. 2016). Post-Variscan extension caused development of the intracontinental Permian Basins that encompassed the present Denmark, northern Germany and Poland areas and reached Latvia at the end of the Palaeozoic Era, but most of Baltica was presumably land area (Fig. 14e; Ruebsam et al. 2017).

Baltica remained an independent continental block until the Iapetus Ocean started to close: it was docked firstly with Avalonia at the end of the Ordovician, and with Laurentia to form Laurussia during the Silurian (Cocks & Torsvik 2005, Torsvik et al. 2012). Docking with Laurentia caused the Caledonian orogeny, and plate convergence continued during the Variscan orogeny. Fennoscandia was foreland to both orogenies. These orogenies were parts in the assembly of the supercontinent Pangea.

Mesozoic and Cenozoic uplift and subsidence events, and development of the Fennoscandian Shield (250–0 Ga)

The virtual lack of Phanerozoic sedimentary rocks in Finland indicates considerable uplift and erosion, but keys to the timing are few. According to Kohonen and Rämö (2005), the occurrence of clastic dykes of Cambrian sandstone in southwestern Finland and the preservation of the 560 Ma Sääksärvi impact structure (Fig. 13) indicate that the present erosional level in southern Finland is close to the Sub-Vendian (Sub-Cambrian) peneplain. As mentioned above, a pile of Mesoproterozoic sedimentary rocks are found below and around the 75 Ma Lappajärvi meteorite impact site [1] (Fig. 13). The preservation of them is explained by downfaulting of the precursory sediments during the impact in the late Cretaceous (Kohonen & Vaarma 2001). Since the thickness of the sedimentary rock package is only 20 m, most of the pre-existing sediments had been eroded by the late Cretaceous.

Geological correlation and landscape analysis (Riis 1996, Lidmar-Bergström et al. 2013) suggest uplift in the early Mesozoic, during which most of Scandinavia and central Finland was exposed to denudation and the development of a peneplain in the Jurassic. Uplift was followed by transgression, and during late Cretaceous, almost entire Fennoscandia was below sea level. Early Cenozoic opening of the North Atlantic Ocean, or more specifically the Norwegian Sea, caused major but differential late Cenozoic uplift in Fennoscandia, during which the present Fennoscandian Shield was exhumed (Fig. 14f). The estimated Cenozoic uplift decreased from 1000 m in Norway to less than 200 m in southern Finland (Riis 1996).

Amalgamation of the supercontinent Pangea was complete by 250 Ma. Opening of the North Atlantic Ocean initiated at ca. 60 Ma as part of the break-up of Pangea.

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