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Paleoproterozoic evolution of Fennoscandia and Greenland

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The Paleoproterozoic evolution of Fennoscandia and Greenland can be divided into major rifting and orogenic stages. The Paleoproterozoic rifting of Fennoscandia started with 2.505–2.1 Ga, multi-phase, southwest-prograding, intraplate rifting. Both Fennoscandia and Greenland experienced 2.1–2.04 Ga drifting and separation of their Archean cratons by newly-formed oceans. The main Paleoproterozoic orogenic evolution of Fennoscandia resulted in the Lapland-Kola orogen (1.94–1.86 Ga) and the composite Svecofennian orogen (1.92–1.79 Ga). The Paleoproterozoic orogens in Greenland, from north to south, are the Inglefield mobile belt (1.95–1.92 Ga), the Rinkian fold belt/Nagssugtoqidian orogen (1.88–1.83 Ga) and the Ketilidian orogen (c. 1.8 Ga). The Lapland-Kola orogen, Inglefield mobile belt and the Rinkian fold belt/Nagssugtoqidian orogen are continent-continent collision zones with limited formation of new Paleoproterozoic crust, whereas the Ketilidian orogen displays a convergent plate-tectonic system, without subsequent collision. The composite Svecofennian orogen is responsible for the main Paleoproterozoic crustal growth of Fennoscandia.

Introduction

Fennoscandia is a geographical term used to describe the Scandinavian Peninsula, the Kola Peninsula, Russian Karelia, Finland and Denmark. The Fennoscandian crustal segment of the East European Craton (Gorbatshev and Bogdanova, 1993) is a geological term and consists of the Fennoscandian Shield, a southern continuation of the Shield with Precambrian rocks covered by platform sediments, and the reworked Precambrian crystalline rocks within the Caledonides in the west (Figure 1). In the Fennoscandian Shield, Archean crust and its Paleoproterozoic cover dominate in the east, Paleoproterozoic Svecofennian crust in the center, and Mesoproterozoic rocks are found in the southwest.

Greenland (Figure 2) consists of a cratonic Precambrian core, Phanerozoic orogens along the northern and eastern sides, and a rifted margin in the west which separates Greenland from the main part of Laurentia in Canada. The Archean core was extensively reworked during Paleoproterozoic time, but juvenile additions were minor. This is in contrast with Fennoscandia where the Paleoproterozoic was the most important crust-forming era.

Fennoscandia and parts of Greenland belong to the world's best known Precambrian regions. Most of Greenland is covered by inland ice, but the coastal regions show almost continuous exposure. Fennoscandia is mostly covered by glacial deposits and the outcrop density is low, but nevertheless voluminous lithological, petrological, geochronological, aeromagnetic, gravity, deep seismic reflection

Figure 1 Simplified geological map of Fennoscandia based on Koistinen et al. (2001). Upper inset: Major Paleoproterozoic orogens of the Fennoscandia. Lower inset: Hidden and exposed microcontinental nuclei and arcs older than 1.92 Ga in Fennoscandia (based on Lahtinen et al., 2005). Paleoproterozoic units in Kola peninsula: IA = Inari area; TT = Tersk terrane. Paleoproterozoic units in Finland: KA = Kittilä allochthon; SB = Savo belt; TB = Tampere belt; HB = Häme belt; UB = Uusimaa belt; Paleoproterozoic units in Sweden: SD = Skellefte district; BB = Bothnian basin; BA = Bergslagen area; OJB = Oskarshamn-Jönköping belt; TIB = Transscandinavian igneous belt. Localities: J = Jormua; K = Knaften.

and refraction, and geoelectric data are available and make it possible to understand its evolution in 3D. Our approach in this account is process-oriented and the focus is on major Paleoproterozoic events. We keep our usage of regional names to a minimum and only provide selected references.

Figure 2 Overview map of Greenland with Paleoproterozoic orogenic complexes.

Paleoproterozoic rifting

The main Paleoproterozoic rifting events of Archean Fennoscandia comprise two stages of intraplate, southwest-prograding rifting between 2.505–2.1 Ga, and c. 2.1 Ga drifting and separation of the cratonic components by newly-formed oceans. Figure 3 provides a general guide through the geotectonic, palaeogeographic and environmental developments related to each stage of rifting, whereas Figure 4 displays rocks recording the most important features of the Paleoproterozoic rift evolution.

Incipient rifting began in northeastern Fennoscandia and became widespread after the emplacement of 2.505–2.440 Ga, plume-related, layered gabbro-norite intrusions and dyke swarms (for references see Melezhik, 2006). Sedimentation occurred mainly in lacustrine basins with a short-term invasion of seawater (Figure 4A). The mantle plume-driven continental uplifts led to the emplacement of voluminous continental flood basalts that were subsequently uplifted, dissected by rifting, and affected by erosion and deep weathering, eventually resulting in the onset of the Huronian glaciation (Melezhik, 2006). At 2.44 Ga, tectonic inversion was followed by repeated rifting. Incipient intraplate rifting at c. 2.35 Ga (Figure 4B) was characterised by predominantly subaerial volcanism (Figure 4C) and thick accumulation of polymict conglomerates (Figure 4D) including diamictites. Peculiar terrestrial products of the c. 2.2–2.1 Ga advanced rifting were highly oxidised, alkaline lavas (Figure 4E) and associated hot-water spring travertines (Figure 4F). Marine environments were marked by deposition of abundant calcium sulphates (Figure 4G) and diverse stromatolites on extensive carbonate platforms (Figures 4H–I). Red beds (Figure 2J) formed thick formations in both subaerial and marine environments, providing firm evidence of an O₂-rich atmosphere. All carbonate deposits were anomalously enriched in ¹³C, signifying perturbation of the global carbon cycle during the Lomagundi-Jatuli Event (Melezhik et al., 2007).

Initial separation of the old, late Archean supercontinent, known as Kenorland, is roughly dated to 2.1 Ga (e.g., Daly et al., 2006). Formation of the Kola ocean and Svecofennian sea was marked by transition into marine conditions in most of the rifts fringing the continent (Figure 4K). Seafloor spreading was accompanied by voluminous submarine eruptions of MORB-like pillow basalts (Figure 4L), whereas thick turbiditic greywackes were deposited on the continental slopes (Figure 4M). The Lomagundi-Jatuli Event

Figure 3 Rifting-related paleotectonic and paleogeographic evolution of the Fennoscandian Shield during the early Paleoproterozoic.

Figure 4 Rocks recording the most important features of Paleoproterozoic evolution.

A. Wavy bedding in mud-sandstone deposited on a mixed tidal flat - indication of a short-term invasion of seawater into a rift-bound lacustrine system; Seidorechka Sedimentary Formation, Imandra/Varzuga Greenstone Belt, Russia. **B.** Sand-filled fractures in Archean pegmatite recording incipient extension in the Pasvik palaeorift, Norway; scale bar 10 cm. **C.** Columnar jointing in Kuetsjärvi alkaline basalt—a typical feature of subaerially erupted mafic lavas, Pechenga Greenstone Belt, Russia; hammer 60 cm long. **D.** Basal, polymict, poorly sorted conglomerate of the 250 m thick Neverskrukk Formation, Pasvik Greenstone Belt, Norway; scale bar 10 cm. **E.** Highly oxidised alkaline dacite—an indicator of probable oxidation processes in the upper mantle; Kuetsjärvi Volcanic Formation, Pechenga Greenstone Belt, Russia. **F.** Laminated, red (oxidised), ^{13}C -rich dolostone overlain by ^{13}C -depleted travertine, the oldest by far in the Earth's record; the dolostone represents the Lomagundi-Jatuli isotopic event, whereas the travertine is a product of a rift-bound, subaerial, hydrothermal system carrying mantle-derived carbon; Kuetsjärvi Sedimentary Formation, Pechenga Greenstone Belt, Russia. **G.** Dolomite-pseudomorph gypsum crystals in brown marine mudstone—indication of a sizeable seawater sulphate reservoir as a response to the oxic atmosphere; Tulomozero Formation, Onega basin, Russia. **H.** Pink (oxidised), ^{13}C -rich stromatolitic dolostone—an indication of perturbation of the global carbon cycle and oxic water; Tulomozero Formation, Onega basin, Russia. **I.** Exhumed stromatolite bioherm representing the subtidal facies of a carbonate platform; Peuranpalo, Tervola, Finland. **J.** Fresh drillcores of red beds representing a tidal carbonate flat and illustrating oxic depositional environments; Tulomozero Formation, Onega basin, Russia. **K.** Bimodal (herring-bone) cross-bedding in tidal channel sandstone; Kalix Greenstone Belt, Sweden; pencil 12 cm long. **L.** MORB-like, tholeiitic pillow basalt representing the 5 km-thick Pilgūjärvi Volcanic Formation deposited during a transition from rifting to drifting; Pechenga Greenstone Belt, Russia. **M.** Rhythmically bedded, sulphide-rich (pale yellow) greywacke deposited from turbidity currents on a continental slope; Pilgūjärvi Sedimentary Formation; Pechenga Greenstone Belt, Russia; **N.** Mudstone rich in organic carbon (55 wt%), representing an enhanced worldwide accumulation of organic matter (Shunga event); Zaonezhskaya Formation, Onega basin, Russia.

was terminated at 2.05 Ga, when previous 'red bed environments' were abruptly superseded by an unprecedented accumulation of formations rich in organic carbon (Figure 4N) representing the worldwide Shunga Event (Melezhik et al., 2005). The Jormua ophiolite (J in Figure 1), which either formed on a passive margin or in a continental rift setting at 1.95 Ga, is a unique example of Archaean sub-continental lithospheric mantle with a thin veneer of 1.95 Ga oceanic crust (Peltonen, 2005).

The last Paleoproterozoic rifting in Fennoscandia occurred at 1.65–1.54 Ga when the rapakivi granites formed. These granites both occur as large batholiths and smaller plutons; their genesis was

caused either by deep mantle plumes or distant convergent plate-tectonic processes (Rämö and Haapala, 2005).

The onset of Paleoproterozoic rifting and continental breakup of the Archean in Greenland is documented by emplacement of the prominent, c. 2.04 Ga Kangâmiut dyke swarm in the Nagssugtoqidian orogen (Figure 2; Mayborn and Leshner, 2006) as well as several other, variably deformed and mainly undated dyke swarms throughout most of West Greenland. The extensional thinning of the Archean crust also gave rise to several continental margin basins that were later caught up in the Paleoproterozoic orogens.

Arc magmatism and orogenic stages

Fennoscandia

No subduction-related magmatism aged between 2.5 and 2.1 Ga has yet been found in Fennoscandia. The 2.02 Ga volcanic rocks in the Kittilä allochthon (Figure 1; Hanski and Huhma, 2005) are the oldest Paleoproterozoic arc rocks of oceanic affinity. Other examples that are older than 1.9 Ga include the island arc magmatism at 1.98–1.96 Ga in the Kola region (Daly et al., 2006), 1.95 Ga supracrustal rocks and granitoids in Knaften (Wasström, 2005), and the 1.93–1.92 Ga island arc rocks in the Savo belt (Figure 1). However, petrological, geochemical and isotopic data from intrusive complexes surrounded by juvenile crust indicate the existence of older (2.1–2.0 Ga) lithosphere (mc in Figures 1 and 6; Lahtinen et al., 2005). The 1.90–1.79 Ga age span records the most prominent magmatic activity observed at the present erosion level of the Paleoproterozoic of Fennoscandia.

The Paleoproterozoic of Fennoscandia can be divided into the Lapland-Kola orogen (1.94–1.86 Ga; Daly et al., 2006), the controversial Gothian orogen (c. 1.6 Ga; Andersson et al., 2002) and the composite Svecofennian orogen (1.92–1.79 Ga; Lahtinen et al., 2005). The last of these is divided into the Lapland-Savo, Fennian, Svecobaltic and Nordic orogens (Figure 1). Whereas the Lapland-Kola orogen shows only limited formation of new crust, the composite Svecofennian orogen forms a large volume of Paleoproterozoic crust. It covers c. 1 million km² and extends southwards, under the Phanerozoic cover, beyond the Baltic countries, as far as the Tornquist zone in Poland. The crustal cross-section of Fennoscandia shown in Figure 5 images collisional structures and the traces of inferred subduction zone in the Svecofennian orogen.

The Lapland-Kola orogen (Figure 1) is the orogenic root of a mountain belt comprising reworked Archean crust with only a minimum amount of juvenile material. An island arc accretion stage at 1.945–1.92 Ga (Inari and Tersk in Figures 1 and 6a) preceded transpressional continent-continent collision between the Kola craton and the Karelian craton at 1.93–1.91 Ga (Daly et al., 2006). Similarly, the amount of juvenile material preserved in the northern segment (Kittilä allochthon) of the Lapland-Savo orogen is minor (see below). Both of these orogens record collision between Archean

continents, and only exhumed roots are preserved. However, Paleoproterozoic metasedimentary rocks (Figure 7A) are abundant in central Fennoscandia, and typically 60–70% of their detrital zircon populations are in the age range 2.1–1.92 Ga (Claesson et al., 1993; Lahtinen et al., 2002). The dominant 2.0–1.92 Ga zircons in the sediments could have their ultimate source in the former arcs in these orogens.

The Lapland-Savo orogen began to form in the north with collision of the Karelian craton with the Norbotten microcontinent at 1.92 Ga. Its main orogenic phase took place when the Keitele microcontinent collided with the Karelian craton at 1.92–1.91 Ga, followed by docking of the Bothnia microcontinent between 1.91–1.89 Ga (see Figures 6B–C). This latter docking terminated magmatism in the Knaften arc. The paleosubduction zone and associated accretionary wedge of the Knaften arc is seen as NE-dipping reflectors (C in Figure 5). The Karelian-Keitele continent-continent collision caused a change in the plate motions and led to a subduction reversal, with the onset of northward subduction at 1.90 Ga (TB in Figure 1). Subduction switchover occurred in the west, and N-directed subduction (B in Figure 5; Figure 6C) produced the Skellefte district volcanism (SD in Figure 1; Figure 7B) at 1.90–1.87 Ga (Allen et al., 2002). A doubly-plunging mantle reflector to the south of the Tampere belt, seen also further west in the BABEL 1 profile (B in Figure 5), indicates subduction also to the south under the Bergslagen microcontinent and attached island arc crust (HB in Figure 1).

The northward subduction under the Tampere belt itself ended at 1.89 Ga; this marked the beginning of the accretionary Fennian orogen (Figure 6C; Lahtinen et al., 2005). The N-directed subduction continued (reflector B in Figure 5, west of the proposed transform fault in Figure 6C), and produced 1.88–1.87 Ga back-arc volcanism in the Skellefte district. When the southern ocean had been consumed, the Bergslagen microcontinent was accreted to Keitele. This resulted in considerable shortening of the Tampere accretionary prism at 1.89–1.88 Ga (between TB and HB in Figure 1), and of the Häme and Uusimaa belts at 1.88–1.87 Ga (HB and UB in Figure 1). The Bergslagen area (BA in Figure 1) has a metamorphic peak at c. 1.87 Ga (Andersson et al., 2006). A second, younger peak at 1.86 Ga, dated from the southern part of the Bothnian Basin (BB in Figure 1) by Högdahl et al. (2007), prolongs the time window for the accretionary Fennian orogen to this date. Although it is widely

Figure 5 Simplified geological interpretation of the crustal structure of the Fennoscandian Shield along the BABEL and FIRE reflection profiles (after Korja and Heikkinen, in press). See Figure 1 for the geological map of Fennoscandia.

Figure 6 Selected stages of a schematic tectonic model for Fennoscandia between 1.93 Ga and 1.78 Ga (modified from Lahtinen et al., 2005).

Figure 7 Rocks from the composite Svecofennian orogen.

A. Metagraywacke in Bothnian Basin; arrows indicate four turbidite beds. White bands and lenses are epidote-bearing calc-silicates. South of Boliden, northern Sweden. Photo: R. Kumpulainen. B. Rhyolitic volcanic mass flow, c. 1.9 Ga; Glommersträsk, Skellefte District, northern Sweden. Photo: B. Kathol. C. Strongly sheared, transposed and retrogressed migmatitic structures in a WNW-ESE-trending, dextral ductile shear zone; Strängnäs, south-central Sweden. Photo: M.B. Stephens. D. Microcline megacryst-bearing granite, c. 1.8 Ga. Stordalsberget, central Sweden. Photo: T. Lundqvist.

accepted that crustal growth of central Fennoscandia (the Fennian orogen in this paper) took place by arc accretion, this model has also lately been strongly questioned (e.g., Skiöld and Rutland, 2006 and references therein).

Högdahl et al. (2007) proposed that N-directed subduction (Figure 6D) caused continental margin magmatism at 1.86–1.84 Ga between the Bothnian basin and the Bergslagen area (BB and BA in Figure 1), and linked this magmatism with an inferred retreating subduction zone (trace A in Figure 5). Subsequent crustal extension (Figure 6D) led to the formation of intra-orogenic clastic sedimentary basins (Bergman et al., 2007). Hermansson et al. (2007) proposed a different tectonic model for areas surrounding the Bergslagen area (BA in Figure 1). In their model, the tectonic domains north and south of Bergslagen itself were formed along a single active continental margin at 1.87–1.85 Ga, and were subsequently moved to their present positions.

The Svecobaltic orogen (1.83–1.80 Ga; Figure 6E) started as an oblique collision in the southeast affecting mainly the extended Fennian sequences. The transpressional tectonics can be observed as large-scale thrusting and margin-parallel E-W-trending shear zones (Figure 7C) in central Sweden and southern Finland. However, arc magmatism at 1.84–1.82 Ga (Mansfeld et al., 2005) is found in southern Sweden (OJB and BA in Figure 1). Thus, simultaneously with the collision in the southeast, a subduction regime was active at the southwestern margin (Figures 6E–F). A retreating Andean-type active margin at 1.85–1.80 Ga caused cyclic periods of subduction-type and marginal basin-type magmatism north (see above) and south of the Bergslagen area followed by terrane accretion (OJB in Figure 1) and final collision with an unknown microcontinent at 1.8 Ga (Figure 6F). One of the associated subduction zones is seen in BABEL B (A in Figure 5). The 1.81–1.77 Ga, NW-trending granitoids in southern Sweden, which belong to the older part of the Transscandinavian igneous belt (TIB1, Figures 1 and 5), are related to this stage.

The Nordic orogen includes the NNE-SSW-trending part of the 1.81–1.77 Ga TIB1 and related granitoids (Figure 1; Gorbatshev, 2004) and areas in central Sweden and Lapland affected by c. 1.8 Ga deformation. The granitoids resemble an Andean-type magmatic setting with an intruded batholith chain that presently occupies an area at least 800 km long and 100–200 km wide (Figure 7D). Lahtinen et al. (2005) proposed that the Nordic orogen resulted from a continent-continent collision between newly established Fennoscandia and Amazonia (Figure 6F). Another possibility is to interpret the Nordic orogen as an advancing Andean-type accretionary orogen, where the inboard deformation relates to retro-arc fold and thrust belts.

Younger TIB rocks occupy the western margin of Fennoscandia (TIB2 in Figure 1), where 1.71–1.67 Ga, N-S-trending batholiths and volcanic rocks discordantly truncate earlier WNW-ESE structural trends. Rocks of the Gothian orogen (c. 1.6 Ga) are present in the Sveconorwegian Province in parts of southwest Sweden and southern Norway. These rocks and parts of the TIB were heavily reworked during repeated orogenesis in the Meso- and Neoproterozoic (Andersson et al., 2002).

Greenland

Three roughly E-W-striking, Paleoproterozoic orogenies decreasing in age from north to south are found in Greenland (Figure 2). The deeply eroded Inglefield mobile belt, which straddles the north-western margin of Greenland and southeastern Ellesmere Island, is the northernmost and oldest of these orogens (Dawes, 2004). It is deeply eroded and mainly exposes intensely migmatized upper amphibolite to granulite facies paragneisses. A long-lived convergent orogenic stage accompanied by juvenile magmatism is documented by ≤ 2.0 Ga detrital zircon and syntectonic c. 1.95 Ga tonalites and diorites. Two late kinematic plutonic monzogranite and syenite complexes were emplaced at 1.92 Ga (Nutman et al., 2007).

The contiguous Rinkian fold belt and Nagssugtoqidian orogen in northern and central West Greenland, respectively (Figure 2), represent the northern and southern parts of a composite collisional orogen at ca. 1.88–1.83 Ga (Henderson and Pulvertaft, 1987; van Gool et al., 2002; Sidgren et al., 2006). The preceding arc magmatism is bracketed between 1.92 and 1.87 Ga. The northern part, the >600 km wide Rinkian fold belt, mainly comprises (meta)greywackes that were deposited unconformably on the Archean basement of the Rae craton (Figure 2; St-Onge et al., in press). The deformed unconformity can be used to trace large Paleoproterozoic structures in three dimensions: spectacular structures along steep fjord sides document first NE-, and then NW-directed tectonic transport by thrusting and recumbent isoclinal folding (Figures 8–9). The resulting flat-lying tectonic crustal structure favoured the development of huge late-stage, mid-crustal buckle folds during a final pulse of crustal shortening (Figure 10). The 1.87 Ga Prøven igneous complex in the central part of the Rinkian fold belt is a granitic laccolith with an Archean isotopic signature.

The 350 km wide Nagssugtoqidian orogen south of the Rinkian belt (Figure 2) consists of reworked Archean basement with some thrust and folded Paleoproterozoic rocks in its high-grade core. The c. 1.9 Ga Sisimiut and Arfersiorfik intrusive complexes have juvenile Paleoproterozoic isotopic signatures and were presumably fed from two parallel S-dipping subduction zones. The southern one is recognised as a suture in the core, and relics of a northern suture probably occur in the transitional area between the Nagssugtoqidian and Rinkian belts around Ilulissat (Figure 2; Connelly et al., 2006).

The youngest Ketilidian orogen (c. 1.8 Ga), at the southern tip of Greenland, displays a convergent plate-tectonic system without subsequent collision (Garde et al., 2002). It mainly comprises a juvenile, c. 1.85–1.80 Ga, c. 30,000 km² calc-alkaline arc (Julianehåb batholith, Figure 2). The arc formed at the southern, rifted margin of the Archean North Atlantic craton during long-lived sinistral transpression due to northward subduction. A small relic of the oceanic plate may now be preserved as obducted tholeiitic pillow lava in the northwestern border zone of the orogen. Southeast of the Julianehåb batholith a fore-arc basin is preserved. Detrital zircon ages show that the basin was filled very rapidly at around 1.79 Ga

Figure 8 Tectonically stacked quartzite lenses of the lower Karrat Group in the central Rinkian fold belt, West Greenland, indicating top-NE tectonic transport. View northwest, north side of Kangilleq fjord. Cliff side approximately 250 m high.

Figure 9 Granulite-grade metasedimentary schist of the Karrat Group at Red Head (75°N) in the northern Rinkian fold belt, West Greenland. Asymmetric quartz pods viewed north, derived from boudinaged quartz veins, indicate top-W tectonic transport.

with immature detritus derived from the rising batholith. Immediately thereafter, the basin was deformed and subjected to HT-LP metamorphism and partial melting; numerous synkinematic, ultramafic to intermediate dykes in the fore-arc may represent a deeper heat source related to the subduction system. Continued heating of the lower crust resulted in crustal inversion whereby granites and syenites were extracted from the middle or lower crust and emplaced as huge, tabular rapakivi-textured bodies at c. 1.75–1.72 Ga, which were subsequently gently folded into crustal-scale domes and narrow cusps.

Supercontinent stage

The global-scale network of Paleoproterozoic collisional orogens led to the formation of the Columbia/Hudsonia/Nuna supercontinent at c. 1.8 Ga (Zhao et al., 2002, 2004). Fennoscandia was located in the middle of this supercontinent and was surrounded by Laurentia in the northeast, Volgo-Uralia in the east, Sarmatia in the southeast, an unknown microcontinent in the southwest, and possibly Amazonia in the west (Lahtinen et al., 2005). Greenland is an essential part of Laurentia, and the Paleoproterozoic orogens in Greenland correlate well with Paleoproterozoic orogens in eastern Canada (St-Onge et al., in press). Correlations towards Fennoscandia are more problematic, not least due to the intervening Caledonian orogen. The Inglefield mobile belt and the Rinkian fold belt/Nagssugtoqidian orogen are both collision zones of Archean

Figure 11 Correlation of Paleoproterozoic orogens in Greenland and Fennoscandia at the c. 1.8 Ga supercontinent stage. Thin green lines indicate the orogenic grains of the Lapland-Savo, Fennia and Svecobaltic orogens (see Figure 1). See St-Onge et al. (in press) for correlation between northeastern Canada and Greenland. The reconstruction of continents at 1.83 Ga is based on Pesonen et al. (2003). Am = Amazonia; B = Baltica; L = Laurentia.

cratons, in which only a limited amount of Paleoproterozoic juvenile crust was formed and preserved. Similar collision zones in Fennoscandia are the Lapland-Kola orogen and the northern segment of the Lapland-Savo orogen. The collision age of the Inglefield mobile belt (c. 1.95–1.92 Ga) coincides with that of the Lapland-Kola orogen, and they are tentatively correlated with each other (Figure 11). The 1.88–1.83 Ga Rinkian fold belt/Nagssugtoqidian orogen is considerably younger than the Lapland-Kola orogen and partly overlaps in age with the Fennian and Svecobaltic orogens. However, the reworked Archean crust in the Rinkian fold belt/Nagssugtoqidian orogen, when compared with the major Paleoproterozoic accretionary crust in the two orogens in Fennoscandia does not favor correlation. The most obvious correlation is between the c. 1.8 Ga Ketilidian and Nordic orogens (Figure 11; Karlstrom et al., 2001), which both are characterized by linear Andean-type batholiths of similar age.

Figure 10 Dome-shaped anticlines and synclinal cusp of basement and cover ('Snepyrampen dome') in the central Rinkian fold belt, West Greenland: an example of crustal-scale, late-stage buckle folding of tectonically layered middle crust.

Conclusions

Fennoscandia and Greenland exemplify two major Paleoproterozoic events, the first of which is the prolonged rifting of Archean crust (part of a presumed supercontinent), which finally led to continental breakup at 2.1–2.04 Ga. The second event is formation of major new continental crust and reworking of older crust by accretionary and collisional orogens at 2.0–1.8 Ga, which finally led to formation of a new supercontinent at 1.8 Ga. The continental collision of Archean cratons in Greenland and northern Fennoscandia is characterised by large and wide orogenic roots which largely comprise reworked Archean crust and its cover, with only a minimum of juvenile material formed and preserved. Crustal growth in these cases is mainly seen in the derived sediment (e.g., Claesson et al., 1993; Kalsbeek et al., 1998; Lahtinen et al., 2002). The composite Svecofennian orogen forms a large Paleoproterozoic unit of new crust, covering c. 1 mill. km². It comprises 2.1–2.0 Ga microcontinents with unknown previous histories of evolution, juvenile arcs from >2.02 to ~1.8 Ga, and Andean-type vertical magmatic additions at 1.9–1.8 Ga, both in South Greenland and Fennoscandia. Subduction-related, folded rapakivi-textured granitoids at c. 1.75–1.72 Ga in South Greenland contrast with rifting-related rapakivi granites in Fennoscandia at 1.65–1.54 Ga.

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