Accepted Manuscript

Geological Society, London, Special Publications

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DOI: https://doi.org/10.1144/SP542-2023-37

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Received 15 February 2023
Revised 6 June 2023
Accepted 6 June 2023

Reconstructing the Paleoproterozoic heart of Nuna, from Fennoscandia to Northeastern Laurentia

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Abstract

Accurate paleogeographic reconstructions of past supercontinents are necessary to test models of supercontinent cyclicity, secular variation in plate tectonics, and mantle geodynamics. However, numerous factors limit our ability to reconstruct past supercontinents as is evident in the ongoing debates regarding the construction phase and geometry of Earth's most recent supercontinent, late Paleozoic to early Mesozoic Pangea. An important factor in these debates, and a focus of our study is how best to palinspastically restore, in section and in plan-view, the orogenic belts along which supercontinents were stitched together. We utilize the orogenic belts spanning Baffin Island, Greenland and Fennoscandia that are inferred to record the assembly of the Paleoproterozoic supercontinent Columbia to test the accepted reconstruction of the Nuna core of Columbia. We show that as reconstructed in the Nuna model, each of Baffin Island, Greenland (with some complications) and Fennoscandia are characterized by an older cratonic backstop that gives way to younger accretionary complexes toward the inferred oceanic domain that lay to the south and which is inferred to have closed by subduction beneath northeastern Laurentia / Fennoscandia during Columbia assembly. This southward transition from cratonic backstop to accretionary orogen is a hallmark of upper plates in modern convergent plate margins and is consistent with the construction of the Columbia supercontinent, including its Nuna core, through plate tectonic processes, and provides a broad validation of the Nuna reconstruction and hence for Columbia as a whole. Map view curvature of the cratonic backstops is restricted to long wavelength, open bends consistent with the Archean crust having been characterized by significant lithospheric strength. The more accretionary southern portions of Nuna are, however, characterized by sinuous

orogens that developed by oroclinal bending of formerly more linear belts, significantly complicating their palinspastic restoration and rendering detailed correlation of juvenile orogenic belts across Nuna problematic.

1.0

Brendan Murphy, together with his long-time collaborator Damian Nance, was amongst the first to recognize that the deep-time, long-wavelength 'cyclicity' recognized in magmatic suites and orogen-forming events could be explained as a result of the cyclic formation and breakup of supercontinents (Murphy and Nance, 1990). This insight, and subsequent efforts to understand the interplay between lithosphere and mantle that drove the cyclicity (Murphy and Nance, 2003; 2008; 2013; Murphy et al, 2009) led to a massive ongoing surge in interest in, and research on, the supercontinents. The result is an ever-expanding appreciation of the role of supercontinents in the evolution of life on Earth, climate change, the mineral and energy endowments of continents, and the evolving structure and behaviour of Earth's deep mantle and lithosphere through deep time.

Testing supercontinent models and explaining their cyclicity is difficult. There is consensus regarding the paleogeographic distribution of the main continental components of Pangea, Earth's most recent supercontinent (Veevers, 2004). However, the timing of Pangea formation and break-up, the stability and duration of its reign, and the tectonic processes and orogenic events that led to its formation, all remain topics of considerable debate (Irving, 1977; Domeier et al., 2012; Murphy & Nance, 2008; Shaw and Johnston, 2016). Stepping backwards through time to earlier supercontinents compounds these uncertainties. It remains uncertain if the proposed late Neoproterozoic supercontinental predecessor to Pangea, Pannotia, ever existed (Murphy et al., 2021). The assembly of a supercontinent (Rodinia) is accepted as a hallmark of the end of the Paleoproterozoic and the beginning of the Neoproterozoic. However, reconstructing Rodinia remains akin to a jigsaw puzzle with little agreement regarding the shape and composition of the various continental 'pieces' (Torsvik, 2003).

There are numerous factors that limit our ability to accurately reconstruct supercontinents that existed hundreds of millions of years ago, including the fragmentary nature of the geological record, and our limited ability (mostly through paleomagnetic studies) to constrain the latitude, longitude and orientation (relative to the geographic poles) of the continental fragments included within past supercontinents. Additional complications, and the focus of this study, concerns debate over 1) how best to palinspastically restore the orogenic belts along which the supercontinents were stitched together, and 2) the role of ribbon continents and juvenile arc terranes, neither of which are likely to retain their shape during accretion and collision. Using our studies of the orogenic belts along which the 'Nuna' core (Meert, 2012) of the Paleoproterozoic supercontinent 'Columbia' (Fig. 1) was constructed, we document differing methods for palinspastic restoration of these sinuous orogens and discuss the implications for continental reconstructions.

We first review the evidence for there having been a Paleoproterozoic Columbian supercontinent and describe the regional geology of Baffin Island, Greenland and Fennoscandia. We then look at the Paleoproterozoic orogenic belts in each of these three regions, documenting the map-view geometry and vergence directions of the belts, and summarizing the consensus tectonic interpretations for each region. We end by considering the contrasting interpretations of these orogenic belts and discuss whether the differing interpretations are due to geologically meaningful differences between each of the orogenic belts. Changes in the orientation and vergence directions of orogenic belts across Baffin Island and Greenland are assumed to reflect primary paleogeographic features and are commonly interpreted within a collisional 'Indentation tectonics' framework. In contrast, the sinuous character of the Baltican orogens are interpreted to result from oroclinal bending of a previously more linear accretionary orogen. Reconciling these contrasting interpretations requires significantly more data, especially from the Laurentian orogens.

2.0 GEOLOGICAL BACKGROUND - COLUMBIA

The arguments for the existence of a Paleoproterozoic supercontinent (Columbia), as first proposed by Rogers and Santosh (2002), include the widespread development of 1.7 to 2.1 Ga orogenic belts after a >400 Myr interval of tectonic quiescence. The ubiquity of orogenic belts points to the interaction of most all of the continents existing at that time, and is consistent with the growth of a significant landmass. The subsequent Mesoproterozoic development of widely distributed, major sedimentary basins likely represents rifting heralding the subsequent break-up of Columbia. Paleomagnetic data are consistent with the presence of a coherent supercontinent which peaked in size at about 1.5 Ga, immediately prior to or even during the onset of its break-up (Meert and Santosh, 2017).

Paleomagnetic data and geological connections have been used to argue that Laurentia and Baltica came together during the formation of Columbia, forming a coherent continental block which is referred to as Nuna (Fig. 1) and which outlasted Columbia (Meert, 2012). Our focus is on the boundary region along which Laurentia and Baltica were thought to be joined, including Baffin Island and Greenland of Laurentian provenance and the exposed northern portion of Baltica, primarily in Finland. In both Laurentia and Baltica, pre-existing Archean cratons were assembled into larger continental masses along Paleoproterozoic orogenic belts (Fig. 1).

In Laurentia, a southerly (present-day coordinates) Superior craton is inferred to have been attached along a north-facing passive margin to oceanic lithosphere to the north that underlay what is referred to as the Manikewan Ocean (Hoffman, 1988). Subduction of the Manikewan oceanic lithosphere to the north resulted in ocean closure and brought the Superior craton into collision with an upper plate Churchill continent to the north giving rise to the Trans-Hudson Orogen (THO, Figs 2, 3). Complicating this picture are smaller cratonic 'terranes' that are presumed to have resided within the Manikewan ocean and which collected along the southern convergent margin of the Churchill continental domain forming an accretionary complex prior to the terminal arrival of the Superior craton (Hoffman, 1988, 1990; Corrigan et al, 2009, 2021; St Onge et al, 2006, 2009). In this model, Baffin Island and Greenland (Fig. 1) lay within the upper 'Churchill' plate, were characterized by south-facing convergent southern margins, and to varying extents, were characterized by accretionary orogen development prior to terminal collision.

Baltica is (based on geological and paleomagnetic constraints) inferred to have lain east of Greenland but rotated counter-clockwise such that northern Baltica lay adjacent to southeastern Greenland. The accretion of Archean cratonic blocks (Kola and Karelia) and development of an accretionary orogen characterized by juvenile arc terranes is consistent with Baltica occupying an upper plate tectonic setting similar to the more westerly Greenland and Baffin Island regions (Fig. 1). However, Baltica lay east of the easternmost extent of the Superior craton and likely avoided participation in the terminal continental collision with this lower plate continent (Lahtinen et al, 2009).

3.0 PALEOPROTEROZOIC OROGENS

3.1 Baffin Island

NNW-SSE trending Baffin Island is narrow (250 - 500 km wide) and long (>1500 km) and is oriented at a high angle to the E-W strike of Archean - Paleoproterozoic geological belts that characterize the easternmost Canadian portion of the Laurentian craton. Hence, the island provides an ideal transect of the Paleoproterozoic structures and assemblages that developed during and which record the assembly of Nuna. However the island is remote and sparsely populated. Despite its size (slightly larger than Spain), the island's population is only ~13,000, 60% of whom live in the southern Baffin town of Iqaluit. As a result, like the rest of Canada's High Arctic, there are virtually no roads and negligible infrastructure throughout Baffin island. Despite significant recent campaigns aimed at mapping southern (in the early 2000s) and northern (2017-19) Baffin Island by the Geological Survey of Canada and the Canada-Nunavut Geoscience Office, little of the island is mapped in any detail (figures 1-2 of St-Onge et al., 2020, provide the spatial distribution of most recent mapping efforts), and there remains a paucity of geochronological constraints on the timing of deformation, metamorphism and magmatism. The lack of topical studies and detailed mapping is a significant impediment to the development and testing of models of Nuna development.

Broadly, the island is divisible into a northern cratonic Rae domain and a southern accretionary complex (Fig. 2). We first describe the southern accretionary complex and then the northerly Rae domain.

3.1.1) South Baffin Accretionary Complex

The southern Baffin accretionary complex is divisible into a southern Dorset fold belt, a central Meta Incognita 'microcontinent' or continental terrane, and a northern Foxe fold belt. The Meta Incognita continental terrane is assumed to underlie much of southern Baffin Island. Meta Incognita continental basement rocks are exposed in the Meta Incognita, Hall and Foxe inliers (Fig. 2) and consist of Neoarchean (~2.7 Ga) crust and early Paleoproterozoic intrusions (St-Onge et al., 2006). The Meta Incognita continental terrane and to a lesser extent the fringing fold belts are overprinted by voluminous magmatic suites including the 1.90-1.89 Ga Qikiqtarjuaq plutonic suite and the younger (1.87 -1.84 Ga) Cumberland Batholith. The Dorset and Foxe fold belts record south- and north-verging crustal shortening, respectively.

The Dorset fold belt (Jackson and Taylor, 1972) is characterized by imbricate thrust stacks, asymmetric tight to isoclinal folds, and a penetrative foliation formed during granulite facies metamorphism. Crustal repetition records significant crustal thickening. South to southwest vergence during crustal thickening is indicated by thrust faults that are characterized by a flat-ramp-flat geometry and which cut structurally up-section towards the southwest. Coeval folding gave rise to tight to isoclinal south to southwest-verging folds and a penetrative axial planar foliation (Scott, 1997).

The basal thrust of the Dorset fold belt is interpreted as a suture referred to as the Soper River suture (St-Onge et al, 2009). The footwall to the north-dipping Soper River consists of the 1.86 - 1.82 Ga Narsajuaq Arc sequence. Gneissic basement rocks as old as 2.98 Ga and younger (1.91 - 1.90 Ga) meta-sedimentary (Lake Harbour Group) and intrusive (Lona Bay and Schooner Harbour) sequences comprise the Meta Incognita terrane hangingwall to the suture. Granitoid rocks of the 1.87 - 1.84 Ga Cumberland batholith intrude the Meta Incognita hangingwall sequence. Ocean closure and suture formation is younger than the youngest Narsajuaq Arc footwall rocks (1842 +5/-3 Ga) and is inferred to have immediately post-date the youngest 1.84 Ga monzogranite phase of the hangingwall Cumberland batholith.

The Dorset fold belt is interpreted to have developed during and provide an upper plate record of north-dipping subduction of oceanic lithosphere beneath the Meta Incognita continental terrane leading to the closure of the Manikewan ocean. In this model, the Narsajuaq Arc sequence is inferred to have been embedded within the Manikewan oceanic lithosphere. Entry of the buoyant Narsajuaq lithosphere into the Soper River trench resulted in collision, crustal thickening and Dorset fold belt formation (St-Onge et al., 2009). The voluminous Cumberland batholith is one of a series of Paleoproterozoic plutonic complexes and belts that, based on geochemistry, age and assumed upper plate location, have been interpreted as Andean-type continental arcs that at least in part record subduction of the Manikewan oceanic lithosphere (Theriault et al., 2001; Whalen et al, 2010).

The Foxe fold belt, like the Dorset fold belt, is characterized by imbricate thrust stacks, asymmetric tight to isoclinal folds, and a penetrative foliation. Structural thickening within the fold belt is indicated by repeated structural duplication of the basement-cover contact by folding and along numerous, low-angle thrust faults. North to NNE vergence during crustal thickening is indicated by a strongly developed N-S trending lineation at shallow structural levels. Low-angle thrust faults cut shallowly up-section to the north and are characterized by E-W trending branch lines (Corrigan et al., 2001), implying top-to-the-north vergence. Although evidence for early S-verging kinematics is locally documented within Paleoproteroic strata near the southern margin of the belt (Tippett, 1984; Henderson et al., 1989), these structures are fairly minor and have been largely overprinted by N-verging structures. Strong ductile deformation of cover sequences in imbricate stacks of basement-cover thrust sheets implies thrust fault developed after significant ductile crustal thickening. Metamorphic grade increases to the south, culminating in a migmatite zone within the highly thickened hinterland of the fold belt.

The uppermost thrust of the Foxe fold belt is interpreted as a suture referred to as the Baffin suture (St-Onge et al., 2009). Hence, the fold belt, which is underlain by continental margin, siliciclastic-carbonate strata of the 2.16 - 1.90 Ga Piling Group (e.g., Partin et al., 2014; Wodicka et al., 2014 and references therein) and its depositional Rae craton basement, forms the footwall to a south-dipping suture. Rocks of the Meta Incognita continental terrane, including its cratonic basement and the 1.91 - 1.90 Lake Harbour Group sedimentary cover sequence, form the hangingwall to the Baffin suture. Granitoid rocks of the 1.87 - 1.84 Cumberland batholith and the older 1.90 - 1.89 Ga Qikiatarjuaq plutonic suite crop out across, obscure and post-date formation of the Baffin suture. Hence, ocean closure and suture formation are tightly constrained to ~1.90 Ga as they must post-date Piling Group deposition (2.16 - 1.90 Ga) but predate Qikiatarjuaq plutonic suite magmatism (1.90 - 1.89 Ga).

The Foxe fold belt is interpreted to have developed during and provide a lower-plate record of south-dipping subduction of oceanic lithosphere beneath the north margin of the Meta Incognita continental terrane. Subduction is inferred to have led to closure of an oceanic domain that lay south of the Rae craton but north of the Meta Incognita continental terrane. Hence, in this model, the Meta Incognita continental terrane lay within the Manikewan ocean and was subsequently accreted to the south margin of the Rae craton by south-dipping subduction of the intervening oceanic lithosphere beneath the continental terrane (St-Onge et al., 2009). The Foxe fold belt is inferred to provide a lower plate record of entry of the buoyant Rae craton into the Baffin trench and coeval overthrusting by the accreting upper plate Meta Incognita continental terrane.

In summary, the Baffin Accretionary complex consists of a central accreted Meta Incognita continental terrane that is bound by fringing fold belts. The north-verging Foxe fold belt developed along the north margin of the Meta Incognita continental terrane and records its accretion to the south margin of the Rae craton at 1.90 Ga. Accretion was followed by the initiation of north-dipping subduction beneath the south margin of the accreted Meta Incognita continental terrane. Subduction closed the remaining oceanic domain by ~1.83 Ga leading to entry of the buoyant Narsajuaq Arc terrane into the Soper River trench, giving rise to the collisional Dorset fold belt along the south margin of the Meta Incognita continental terrane. The sum result is an accretionary complex characterized by a 1.83 Ga south-verging Dorset fold belt along its southern margin, a central, accreted Meta Incognita continental terrane, and a northern, north-verging 1.90 Ga Foxe fold belt that forms the boundary between the Meta Incognita continental terrane and the south margin of the Rae craton (Fig. 2).

3.1.2) Northern Cratonic Domain

Work carried out northwest of the Foxe Fold Belt, within basement gneiss and subordinate granite-greenstone belts of the Rae Craton (2.90-2.71 Ga; Skipton et al., 2019), suggests that Archean basement was affected by Paleoproterozoic tectonic events. Interpretations of the framework and dynamics of northern Baffin Island with respect to the THO hinge on the origin of the Isortoq Shear Zone (ISZ; Jackson, 2000; Jackson and Berman, 2000; Bethune and Scammell, 2003; Saumur et al., 2018). The latter is expressed as a prominent NE-SW trending linear aeromagnetic anomaly extending ~200 km from the SW coast of Baffin Island at Grant-Suttie Bay to the centre of the island NW of Barnes Ice Cap, where its aeromagnetic response becomes more diffuse. The ISZ has been interpreted as an extensional structure that accommodated significant crustal unroofing during gravitational collapse of over-thickened crust (Jackson and Berman, 2000). Regionally, the southeast-dipping shear zone's footwall consists of orthopyroxene-bearing felsic plutonic rocks and granulite facies assemblages (Dexterity granulite belt), metamorphosed at 1.82 Ga (Bethune and Scammell, 1997; Jackson and Berman, 2000), whereas hangingwall rocks to the southeast consist of lower grade, amphibolite–facies rocks. However, more recent mapping (Skipton et al., 2020a, b) shows that the extent of the footwall granulite belt is not coincident with the ISZ, and that the shear zone is thin (unobservable to <10m-scale), locally cryptic, likely consists of a series of discontinuous shears, and juxtaposes similar gneissic felsic units (Saumur et al., 2018). The shear zone dips moderately SE, and shear sense indicators indicate dextral-normal sense shearing (Bethune and Scammell, 2003; Saumur et al., 2018). In the Grant-Suttie Bay area (Saumur et al., 2022) the ISZ occurs within the overturned lower limb of a south-verging Isotorq Nappe (Saumur et al., 2022) and accommodated thinning of the inverted limb during nappe emplacement.

South-verging structures have been previously documented in the vicinity of Steensby Inlet near the western extent of the ISZ (Fig. 3; "Steensby Domain" of Jackson, 2000; Jackson and Berman, 2000). A 20 km scale asymmetric antiformal fold, the Steensby antiform, with a long southeastern limb that parallels and is kinematically linked to the shear zone, involves panels of Piling Group strata folded along with Archean felsic gneiss and granitoids that sit structurally below older Mary River Group strata within the lower, overturned limb of the Isortoq Nappe (Fig. 3; Morgan, 1982; Skipton et al., 2020a). Steeply southeast-dipping upright strata cropping out 30 km southeast of the ISZ form the upper limb of the Isotorq Nappe. Although previously attributed entirely to the Mary River Group, geochronological studies (Folkesson et al., 2023) show that Piling group strata overlie the Mary River Group within the upright nappe limb (Saumur et al., 2022; Folkesson et al., 2023). These relationships limit nappe emplacement to the Paleoproterozoic, at the earliest. The high-grade, high-tonnage Mary River Fe-ore deposit (e.g., Jackson, 2000; Harrison et al., 2022), located ~ 200 km northwest of the ISZ (see “MR” in Figure 2), provides a similar record of Paleoproterozoic deformation, including i a metamorphic thermal event associated with late-stage orogenic collapse linked to the THO (MacLeod, 2012). Although further work is required, the footprint of the THO extends much further northward into the Rae Craton than previously acknowledged.

Along the northeastern coast of Baffin Island, Jackson (2000) and Jackson and Berman (2000) recognized a 1.825-1.81 Ga, SW-vergent Northeast Baffin Orogen, which appears to overprint previous regional tectonic fabrics, retrogressively metamorphosed rocks of the Dexterity granulite belt, and affects Piling Group strata. Little is otherwise known on this event and how it may relate to other events documented on Baffin Island and Greenland.

3.2 Greenland

Greenland is the largest island on Earth with an area of over two million square kilometers. It is up to 1300 km from east to west and 2700 km from north to south, extending from 60º to 84º Lat North, with its northern end representing the northernmost land area in the world. Crystalline rocks comprising the continental crust of Greenland formed during a series of Archaean and Paleoproterozoic orogenic events (Fig 4). Archaean rocks (3200–2600 Ma, with local units up to 3800 Ma) that were almost unaffected by Paleoproterozoic orogenic activity belong to the North Atlantic craton. Juvenile Paleoproterozoic rocks (1950–1750 Ma) were emplaced during at least four Paleoproterozoic orogenic events that were responsible for reworking of the Rae and North Atlantic cratons (Thrane, 2021). Roughly E-W trending orogenic belts (Inglefield, Nagssugtoqidian, Ketilidian) young slightly from north to south while the roughly N-S trending Rinkian orogen, part of the THO, is geographically bracketed between the Inglefield orogen to the north and the Nagssugtoqidian to the south (Fig. 4).

3.2.1) Inglefield Orogen

The Precambrian basement of southeastern Ellesmere Island, eastern Devon Island and northern West Greenland (Fig. 4) is represented by polydeformed granulite facies, metasedimentary and metaigneous rocks of the Ellesmere-Devon Terrane and Inglefield mobile belt (Frisch and Dawes, 1982; Frisch, 1988; Dawes et al., 1988). Metapelitic and quartz-feldspathic paragneiss and schist of the Etah Group, also comprising marble and calc-silicate, amphibolite, and ultramafic units, have been bracketed between 1980 Ma and 1950 Ma. The supracrustal rocks were intruded by intermediate to felsic orthogneiss, megacrystic monzogranite, quartz diorite, syenite, and subordinate metagabbro of the Etah metaigneous complex between 1949±13 and 1915±19 Ma (Nutman et al. 2008a).

3.2.2) Rinkian Orogen

Supracrustal rocks in central West Greenland belong to the Karrat Group that initiated after c. 2000 Ma in an intra-cratonic rift basin with basal quartzites overlaying Archean gneiss of the Rae Craton (Guarnieri et al., 2022a; Guarnieri et al., 2023). The rift basin evolved to a back-arc basin with associated sub-alkalline volcanism (Guarnieri et al., 2022b; DeWolfe et al., 2023; DeWolfe and Sørensen, in review), concomitant with the emplacement of arc-related granitoids of the Prøven Igneous Complex between 1900 and 1850 Ma) (Thrane et al., 2005; Guarnieri et al., 2022c; Kokfelt et al., 2023). The arc-continent collision initiates the east-ward structuring of the Rinkian Orogen as back-arc fold and thrust system antithetic to north-eastward oceanic subduction. The high-temperature Rinkian metamorphism at c. 1830–1800 Ma increases in metamorphic grade from greenschist facies in the south (Guarnieri and Baker, 2022) to granulite facies in the north, where it is marked by migmatization and the emplacement of S-type leucogranites. The Rinkian Orogen is interpreted as part of the THO (Figs 1, 4) (e.g., Hoffman, 1988; Connelly et al., 2006; Guarnieri et al., submitted).

3.2.3) Nagssugtoqidian Orogen

In the Nagssugtoqidian Orogen of West Greenland (Fig. 4), Palaeoproterozoic psammitic and pelitic schist and gneiss, with banded iron formation, metagreywacke, fine-grained metavolcanic rocks, and marble belong to the Nunatarsuaq supracrustal rocks (Garde & Steenfelt 1999), Naternaq supracrustal belt (Østergaard et al. 2002; Thrane & Connelly 2006) and Nordre Strømfjord supracrustal suite (Van Gool et al. 1999). The supracrustal rocks are dominated by 2200–1950 Ma detrital zircon grains with little to no Archaean input. The calc-alkaline Sisimiut and Arfersiorfik plutonic suites emplaced from ca. 1920 Ma to c. 1870 Ma are interpreted as subduction-related arc magmatism (Kalsbeek et al., 1987; Kalsbeek and Nutman, 1996; Whitehouse et al., 1998; Connelly et al., 2000). High-temperature metamorphism during collision is dated at c. 1850 Ma and continues until c. 1825 Ma (Connelly et al., 2000). The Nagssugtoqidian Orogen of South-East Greenland consists of Meso- to Neoarchaean orthogneiss, amphibolite and ultramafic rocks of the North Atlantic and Rae cratons, which formed the hinterland and foreland of the north-verging orogen, respectively (Kolb, 2014). Up to three kilometers of Paleoproterozoic sedimentary and volcanic rocks belonging to the Síportôq Supracrustal Association (Hall et al., 1989a, 1989b), overlie the Archean rocks. High-temperature metamorphism is associated with compression at c. 1860–1825 Ma (van Gool et al., 2002) post-dating the emplacement of arc-related intermediate intrusions of the Ammassalik Intrusive Complex at c. 1880 Ma (Hansen and Kalsbeek, 1989; Nutman et al., 2008b).

3.2.4) Ketilidian Orogen

The Ketilidian Orogen (Fig. 4) defines the southern margin of the North Atlantic Craton where the Archean basement is overlain by shallow marine metasedimentary rocks and deep marine metagraywackes of the Paleoproterozoic Vallen Group Along a basal unconformity that is progressively obscured by ductile deformation (Chadwick and Garde 1996; Garde et al., 2002). The Vallen Group is tectonically overlain by metabasaltic pillow lavas, pillow breccias and sills, with minor intercalations of mudstone and calcareous rocks of the Sortis Group (Bondesen 1970). Sinistral transpression and arc-related granite emplacement followed between 1850–1800 Ma with the calc-alkaline Julianehåb batholith, related to northward subduction, largely emplaced between 1854–1795 Ma. Fore-arc sedimentation, pervasive deformation together with high-temperature metamorphism and anatexis occurred between 1795–1785 Ma (Garde et al., 2002).

3.3 Fennoscandia

Fennoscandia is in the northern part of (proto) Baltica or the East European Craton (EEC; Fig. 5a, b; Gorbatschev and Bogdanova 1993; Bogdanova et al. 2008). The Fennoscandian Shield comprises Archean crust with Paleoproterozoic cover in the east, Paleoproterozoic crust in the central part, and progressively younger Meso- to Neoproterozoic rocks in the southwest (Fig. 5a). From 2.5 to 2.1 Ga, the Archean craton/cratons were rifted, and disturbed, and small oceans were opened. Eventually a wide passive margin was developed along the present edges of the former cratons (Lahtinen et al., 2005; Bogdanova et al., 2008). The main Paleoproterozoic orogenic evolution occurred between 1.92 Ga and 1.77 Ga and orogenic areas have been divided into the composite Svecofennian (Lahtinen et al., 2005; 2009) and Lapland-Kola Orogens (Daly et al., 2006; Lahtinen and Huhma, 2019), which are the focus of this study. The composite Svecofennian Orogen forms a large volume of Paleoproterozoic crust, whereas the Lapland-Kola Orogen comprises mainly of reworked Archean with only limited preservation of new juvenile crust. The partly Paleoproterozoic Gothian orogeny (1.7–1.5 Ga), which was overprinted during the Sveconorwegian orogeny, is not discussed. Similarly, relics of the Svecofennian rocks present in the Blekinge–Bornholm, Sveconorwegian (eastern part) and Caledonide orogens (Fig. 5a; e.g., Stephens 2020), are left out of this discussion.

3.3.1) Lapland-Kola orogen

The Lapland-Kola Orogen (Fig 5a, c) is a deeply exhumed continent-continent collision zone (Daly et al., 2006) and comprises juvenile arc rocks (IA; 1.96–1.92 Ga; Daly et al. 2006; Lahtinen and Huhma 2019), granulite belts (Meriläinen, 1976; Berthelsen and Marker, 1986a; Barbey and Raith, 1990; Glebovitsky et al., 2001; Daly et al., 2006; Tuisku et al., 2006; Lahtinen and Huhma 2019), collisional mélanges (Daly et al., 2006), and reworked Archean (e.g., Belomorian in Fig. 5a; e.g., Bibikova et al. 2001). A cryptic suture (Fig. 5a) is predicted to separate a northeast-verging Kola continent (lower plate) to the northeast (Daly et al. 2006, Mints et al. 2009; Mudruk et al., 2022 and references therein) and a southwest-verging Karelian continent (retro-arc) and arc-rocks (upper plate) to the southwest (Lahtinen and Huhma 2019).

Eclogites characterize the Lapland-Kola orogen in Belomorian and north of it along the suture (Fig. 5a), but the age of these eclogites is debated. Some consider them as products of Archean metamorphism overprinted in the Paleoproterozoic (e.g., Dokukina et al. 2021), while others indicate a Paleoproterozoic age (e.g. Yu et al. 2019 and references therein). Archean and Paleoproterozoic eclogitic events have also been proposed (Balagansky et al. 2015; Volodichev et al. 2021). Li et al. (2023) concluded that at least the Salma eclogites record cold subduction (peak conditions of 2.2–2.3 GPa, 650–670 °C) prior to 1.93 Ga.

Based on the tectonic model by Lahtinen and Huhma (2019) a collisional lower plate foreland in the NE, comprises the Kola continent (Archean), Paleoproterozoic cover rocks and allochthonous upper plate rocks. The NW–SE oriented cryptic suture is characterized by eclogites. The upper plate comprises the Karelia continent (Archean), Paleoproterozoic cover rocks and the juvenile Inari arc (IA in Fig. 5a). The arc's hinterland is composed of granulite grade retro-arc basin strata. During collision (D1) at 1915–1910 Ma large thrust nappes formed on the foreland and attempted subduction of lower plate Archean crust down to depths of ≥60 km led to the eclogite facies metamorphism. Collisional shortening in the retro-arc basin is seen as recumbent folding and shearing. Renewed NE-SW shortening (D2), due to far-field effects in the SW at 1.88–1.87 Ga, caused a large-scale crustal duplexing (cf. Tuisku et al. 2006) and eclogite exhumation.

A switch in the stress field from NE–SW (D2) to NNW–SSE led to orogen-parallel contraction and buckling that resulted in reactivation of the suture zone as a crustal-scale dextral strike-slip fault (Fig. 5a). Buckling is seen in the bending of existing fabrics about a vertical axis, the widespread development of gentle-open to close folds orthogonal to the D1–D2 fabrics, and growth of abundant radial fractures. The end result is a mega-scale multi-layer vertically-plunging parallel fold, the Inari secondary orocline (Fig. 5a). The orocline involves the Inari arc, retro-arc basin and possibly also the heated retro-arc foreland. The age of orocline formation is not well defined but is younger than 1.87–1.86 Ga. We tentatively correlate it with shield-wide tectonic event D4 of Lahtinen et al. (2022, submitted) at 1.83–1.82 Ga.

Different interpretations have been proposed. Daly et al (2006) invoked opposing subduction systems whereas others (Barbey et al. 1984; Tuisku et al. 2006; Nironen,2017) employed NE-dipping subduction beneath the Inari craton (Fig. 5a) . The curvature of the Inari belt has been modelled as a progressive orocline that formed during thrusting (D1; e.g., Gaál et al., 1989; Daly et al., 2006). As discussed in Lahtinen and Huhma (2019), in the absence of paleomagnetic data, stretching lineations (partly D1) show varying transport directions around the orocline's limbs and D1 and D2 structures follow the oroclinal bending. Although some primary curvature or progressive bending is possible, a secondary orocline best explains the geometry of the D1 and D2 structures.

3.3.2) Svecofennian orogen

The Paleoproterozoic Svecofennian (terminology Svecokarelian used in Sweden; e.g., Stephens, 2020) orogeny (1.91–1.78 Ga) formed the composite Svecofennian orogen, one of the largest collages of Paleoproterozoic orogens globally (Lahtinen et al., 2005). Within the orogen, the Archean Karelia continent is separated from Paleoproterozoic Svecofennia by a 1.91 Ga cryptic suture zone (Koistinen, 1981) (Fig. 5a). The Archean–Proterozoic (AP) boundary region is characterized by anomalously thick crust (> 56 km; Korja et al., 1993) and lithospheric mantle (O'Brien and Lehtonen, 2012) favoring a crocodile jaw configuration (Peltonen and Brügmann 2006; Lahtinen et al. 2016).

The suture continues north where it separates lower plate Karelia from the upper plate Norrbotten continent (Lahtinen et al. 2005, 2015, submitted; Bingen et al. 2015). Ocean closure at ca. 1.91 Ga resulted in obduction of 2.02 Ga supra-subduction ophiolites and boninites (fore-arc of an oceanic arc) over the lower plate (in the CLB in Fig. 5a; Hanski 1997; Hanski and Huhma 2005) and formation of the >600-km-long and in part >300 km-wide E-vergent Lapland foreland fold and thrust belt in the lower Karelia plate (CLB, PB and KB in Fig. 5a; Lahtinen and Köykkä 2020). Intraplate and dominantly strike-slip interpretations for the N–S-trending boundary zone have been proposed by Berthelsen and Marker (1986b), Kärki et al. (1993), Nironen (1997), and Bergman and Weihed (2020).

The 1.91 Ga collision zone was defined as the Lapland-Savo orogen of the composite Svecofennian orogen by Lahtinen et al. (2005). The orogenic belt, mainly east of the sutures (Fig. 5a), and the Lapland-Kola orogen (Fig. 5c) both show linearity, which is lacking from the central Svecofennian orogen. The equidimensional 1.90–1.88 Ga Paleoproterozoic crustal domain forming the center of the Svecofennian orogen (Figs. 5a, c), was defined as the Fennian orogen (Lahtinen et al., 2005). This orogen is characterized by two continuous large arcuate structures; a southerly convex to the west bend that connects through to a more northerly convex to the east bend. These structures coincide with a crustal conductance anomaly (Korja et al., 2002) and have been interpreted as a pair of coupled Bothnian oroclines (Fig. 5a; Lahtinen et al., 2014). These oroclines continue south into a third, convex to the east Saimaa orocline (Fig. 5a; Lahtinen et al., 2022).

The center line of the Bothnian and Saimaa oroclines consists of 1.90–1.88 Ga supracrustal arc rocks and arc-related plutonic rocks (SD–PoB–TB–HB/UB in Fig. 5a). Vectors defined by structural vergence, decreasing metamorphic grade and direction toward oceanward lithologies vary consistently as a function of structural trend around the oroclines. Thus, an originally linear hot orogen buckled to form the present equidimensional crustal domain. The Saimaa orocline has been severely overprinted by post-orocline deformation and its continuation towards the SW is under investigation. The $D_{orocline}$ is rather well dated at 1875–1865 Ma (Lahtinen et al. 2022) whereas the proposed Bergslagen orocline (Beunk and Kuipers, 2012) has been dated at 1.83–1.82 Ga. Thus, the Bergslagen orocline is younger and possibly coeval with the Inari orocline (Fig. 5a). The similar age (1.90–1.80 Ga) arc volcanism, lithologies and ore types in the Uusimaa belt (UB) and in the Bergslagen area (BA; Fig. 5a) favor correlation of these belts and suggest that they formed parts of the same microcontinent (Allen et al. 1996; Nironen, 1997; Lahtinen et al. 2005).

Deformation in the Norrbotten (Bergman and Weihed 2020) and Bothnia–Skellefteå (Skyttä et al. 2020) lithotectonic domains (Fig. 5a) started at 1.88 Ga or earlier as part of a shield-wide D2 tectonic event (Lahtinen et al. 2022 and references therein). In contrast, the first deformation event in the Ljusdal (Högdahl and Bergman, 2020) and Bergslagen (Stephens and Jansson 2020) lithotectonic domains (Fig. 5a) started at 1.87 Ga, and at 1.86–1.85 Ga in the Småland lithotectonic domain (Wahlgren and Stephens 2020). We tentatively correlate these 1.87–1.85 Ga deformation events as part of a shield-wide D3 tectonic event (Lahtinen et al. 2022 and references therein).

A N–S shortening event at 1.83–1.82 Ga in Bergslagen caused intense buckling and oroclinal bending (Bergslagen orocline in Fig. 5a) and folding along variably but generally steeply plunging folds (Beunk and Kuipers 2012). A shield-wide D4 tectonic event (Lahtinen et al. 2022 and references therein) caused NNW–SSE shortening in southern Finland and northern Fennoscandia and is related to far-field collisional effects. N–S to NE–SW shortening in central and southern Sweden (Beunk and Kuipers 2012; Högdahl and Bergman 2020; Stephens and Jansson 2020; Wahlgren and Stephens 2020) is linked to accretionary growth (Stephens 2020b and references therein) of crust to the northeast that now lies covered in the Baltic countries (Bogdanova et al. 2015).

Formation of a juvenile 1.83–1.82 Ga volcanic arc (Mansfeld et al. 2005) in the Småland lithotectonic domain (Fig. 5a) was followed by voluminous bimodal 1.81–1.76 Ga Transscandinavian Igneous Belt magmatism (TIB-1 in Högdahl et al., 2004). The igneous belt comprises an elongated array of batholiths extending c. 1400 km from southeasternmost Sweden to northwestern Norway (WTBC in Fig. 5a) that post-dates D4 and lies nearly orthogonal to 1.86 – 1.81 Ga belts (Bogdanova et al., 2015) and D4 structural trends. The D5 1780–1760 Ma NE–SW shortening event (Lahtinen et al. 2022) is the first major deformation event recorded in the West Troms Basement Complex (WTBC in Fig. 5a, e.g., Bergh et al. 2010; Paulsen et al. 2021) and in Kiruna (K in Fig. 5a; Andersson et al. 2021).

4.0) DISCUSSION

4.1) Issues with Data Availability and Correlations

Our review of the Nuna heart of the supercontinent Columbia points to a number of factors that complicate reconstructions of past continents and place limits on palinspastic restorations of ancient orogens. Perhaps the most salient feature of our review is the vast superiority of the Fennoscandian geological database (Fig. 5) relative to eastern Laurentia of Baffin Island and Greenland (Figs. 2, 4). Geological mapping is restricted to Greenland's coastal regions and hence, provides a geographically limited perspective on its geological evolution. However, the significant topographic relief available in Greenland's coastal fjords provides for superb 3-dimensional mapping, particularly through combining 3D-photogeology with field structural data (Guarnieri and Baker, 2022). Geophysical surveys allow us to place constraints on the nature of the vast ice-covered core of Greenland, but without any way of ground-truthing the geophysical data, geological correlations across Greenland's enormous expanse are conjectural. Similarly, Baffin Island's almost total lack of infrastructure has limited mapping of the island to geological survey campaigns (St-Onge et al., 2020), at scales of 1:100,000 at best. The result is a paucity of detailed maps and topical studies. In contrast, Fennoscandia, despite a lack of significant topographic relief across much of its glacially peneplaned interior, has been studied in significant detail. The result is a much richer and more mature understanding of the geological evolution of Fennoscandia than is available for eastern Laurentia. But relating the D1 through D5 tectonic events characterizing Fennoscandian orogenic belts (e.g., Lahtinen et al., 2022 and references therein) with the much more sparsely constrained orogens of Greenland and Baffin Island represents a significant challenge.

A related data constraint concerns geochronological data. High-quality U-Pb zircon age determinations are available for most of the orogenic belts spanning eastern Laurentia and Fennoscandia. Using this valuable data set to test correlations of orogenic belts and geological provinces is, however, complicated. Significant age ranges are reported from most of the orogenic belts, in part because samples include pre-existing basement, and pre-, syn- and post-orogenic magmatic belts. A good example is provided by the Nagssugtoqidian and Lapland-Kola orogens of eastern Greenland and Fennoscandia, respectively. Nagssugtoqidian orogen crystallization age determinations range over ~100 Ma, stretching from 1.92 to 1.82 Ga. Lapland-Kola orogen crystallization ages range over ~200 Ma, from 1.98 to 1.77 Ga. These age determinations are permissive of correlation of the two orogens, but fail to provide a robust test of the correlation. Geologists can never acquire enough geochronological data, but testing of correlations across Nuna requires better constraints on specific orogenic features (e.g. syn-collisional stretching lineations).

Ultimately, robust testing of the paleogeographic and tectonic reconstructions will require the development of detailed 'Correlation charts' in which the age, evolution, vergence and character of magmatic suites, metamorphic belts, and deformed domains that span Nuna are plotted. However, at this time data availability, especially from Arctic Canada and Greenland, remain problematically weak.

4.2) Non-unique tectonic interpretations

An ongoing challenge in tectonic analyses concerns the significance of specific orogenic features, including magmatic, fold and thrust, and sinuous orogenic belts. Classification of magmatic belts through the use of geochemistry-based tectonic discrimination diagrams provides only the weakest of tests of correlations. Modern analogues suggest that such geochemical designations can be non-unique. Modern lavas erupting in Mexico today include alkaline intraplate lavas mixed in and erupted with typical calc-alkaline ones, all in the same arc 'province'. More useful would be the development of magmatic vectors including younging directions and contoured geochemical maps. Such ‘vector maps’ provide a far more robust test of proposed correlations than absolute ages or specific geochemical characteristics.

Fold and thrust belts provide one of the most robust tectonic vectors within orogenic belts as they are normally characterized by folds and thrust faults that verge overwhelmingly in a single direction. Beyond vergence direction, however, there are numerous differing interpretations of the tectonic setting indicated by such thrust belts. Modern analogues (the Alpine - Himalayan system, St-Onge et al., 2006) imply that fold and thrust belts develop adjacent to collisional sutures, verge toward the incoming lower plate, and reflect the geometry of the subduction zone/suture along which collision occurred. This interpretation is employed for both the Foxe and Dorset fold belts of Baffin Island (Fig. 2). However, the Foxe fold belt is inferred to imbricate and thicken lower plate crust and to lie in the footwall of a collisional ‘Baffin’ suture. In contrast, the Dorset fold belt is inferred to imbricate and thicken upper plate crust, and to form the hangingwall to the Soper River collisional suture. These contrasting interpretations are not supported by any significant differences between the two fold belts, this despite the fact that lower plate orogens are commonly distinguishable (passive margin sedimentary strata, thin-skinned deformation) from upper plate orogens (sediment poor, thick-skinned deformation). Instead, the two fold belts are so similar they have been referred to as mirror images of one another (Jackson & Berman, 2000).

Fold belts are also thought to develop within the retroarc region of collisional orogens, where they are characterized by a geometry and vergence direction opposite that of the collisional suture zone. A retro-arc setting for the north-verging Foxe fold belt would imply that the fold belt is not proximal to any suture zone (which might explain why the inferred Baffin suture is ‘cryptic’), that it lies north of and is separated from a more southerly north-dipping suture zone by an upper plate magmatic arc. Such an interpretation would be consistent with development of the Dorset fold belt immediately adjacent to and above the relevant suture, coeval with the development of the oppositely verging Foxe fold belt in a retroarc setting. However, available geochronological data suggest that the Dorset fold belt is younger than and post-dates the Foxe fold belt perhaps by 50 my. Models that include a north-dipping, south-verging Baffin suture proximal to and immediately south of the south-dipping, north-verging Foxe fold belt could only work if the Foxe fold belt was younger than and cut the Baffin Suture, a relationship which has not been documented.

4.3) Reconstructing Nuna

Despite the challenges outlined above, our review provides confidence in the overall interpretation of northeastern Laurentia and Fennoscandia as having occupied an upper plate position on a plate tectonic Earth during the formation of Columbia, throughout ocean closure and subsequent collision. This includes oceanward younging of the upper plate, and the development of accretionary complexes adjacent to the inferred suture (Figs. 2, 4, 5). The distribution, geometry and kinematics of structural and lithological elements are analogous to Phanerozoic orogenic systems. From a uniformitarian perspective, this suggests that collisional geodynamic processes have not changed greatly since the Paleoproterozoic - although some differences, as discussed below, could be attributable to the involvement of cold, strong cratons.

The processes responsible for the development of sinuous or bent orogenic belts are much debated. Large-scale curvature of the Paleoproterozoic orogens of Nuna have been attributed to indentation tectonics (Hoffman, 1988; Corrigan et al., 2021). In these models, convex and concave curvature of the orogenic belts are inferred to reflect original paleogeographic promontories and recesses that characterized the margins of colliding continents. Alternatively, sinuous orogenic belts of Fennoscandia have been interpreted as the result of bending of a previously more linear orogen (Lahtinen et al, 2014, 2022). The Nuna reconstruction may, however, help reconcile these contrasting interpretations.

Each of Baffin Island (Fig. 2), Greenland (Fig. 4, with some complications) and Fennoscandia (Fig. 5) are characterized by an older cratonic backstop that gives way to younger accretionary complexes toward the inferred Manikewan Oceanic domain that lay to the south and which is inferred to have closed by subduction beneath northeastern Laurentia / Fennoscandia. These Archean cratonic backstops include the Rae Craton of northern Baffin Island and northern Greenland, and the reworked Karelian - Kola cratons of northeastern Fennoscandia. Orogenic curvature is limited within these cratonic backstops to open, long wavelength gentle bends of the reworked Karelian/Kola cratonic crust of Fennoscandia. Younger accretionary complexes consisting of fold and thrust belts and accreted continental terranes lie south of the Rae craton on Baffin Island and Greenland and SW of the Karelian/Kola backstop in Fennoscandia. On Baffin Island, the Foxe and Dorset fold belts enclose the accreted Meta Incognita terrane. The accretionary central and southern Greenland features the Rinkian, Nagssugtoqidian and Ketilidian fold and thrust belts that developed during and record accretion of the Aasiaat domain and the North Atlantic craton. Fennoscandia southwest of the Karlia includes early Paleoproterozoic continental crust and juvenile accreted arc terranes. Orogenic curvature within the accretionary complexes, most notably in Fennoscandia, includes the development of multiple linked convex and concave bends (Fig. 5). Structural vergence, metamorphic grade, crustal structure and stratigraphy vary as a function of trend around these bends indicating that the bends are oroclines that formed by bending previously linear or nearly linear belts (Johnston et al., 2013).

These observations suggest that cratonic Archean crustal provinces are likely too strong to deform by orocline formation and may therefore preserve original continental paleogeographic features including promontories and recesses. In contrast, younger accretionary crust appears to have been weak enough to have deformed by bending and folding about vertical axes of rotation giving rise to multiple coupled oroclines. Applying this 'oroclinal' model to the accretionary portions of Greenland and Baffin Island suggests that the N-S trending Rinkian thrust belt of western Greenland may be an eastward continuation of a Baffin Island fold belt that links through a 90-degree oroclinal bend. Similarly the arcuate, concave to the north curvature of the Nagssugtoqidian may reflect subsequent oroclinal bending. And the Dorset and Foxe fold belts of Baffin Island are commonly portrayed as intersecting west of Baffin Island. Neither of these fold belts can be followed west of their intersection point suggesting that they may be continuously traced into one another around a concave to the west orocline. Oroclinal bending of accreting ribbon continents and juvenile arc terranes during the construction of the Columbia supercontinent has previously been documented in the Australian continental domain (Betts et al., 2016). Australia is thought to have lain along the opposite (western) margin of Laurentia within Columbia suggesting that the bending and deformation of formerly linear terranes and ribbon continents was a globally significant process during Paleoproterozoic supercontinent construction (Sayab et al., 2021).

Finally, accretionary complexes characterized by voluminous pre-, syn- and post-tectonic magmatic suites that young oceanward are a hallmark of modern convergent plate margins. Crust added to the upper plate from the subducting lower plate is accommodated by trench retreat and related advance of the arc magmatic over the newly accreted crust (Oldow et al., 1990). Syn- and post-tectonic magmatism records the cessation of subduction and collision-related crustal thickening and slab break-off (Davies and von Blanckenburg, 1995). The north to south transition of Nuna from Archean cratonic backstops into Paleoproterozoic accretionary orogens characterized by voluminous magmatic suites and oroclinal deformation are consistent with construction of the Columbia supercontinent, including its Nuna core, through plate tectonic processes fundamentally similar to those that continue to shape the Earth today.

5.0) CONCLUSION

The Nuna reconstruction is consistent with an upper plate continent that was subject to long term subduction beneath its southern margin. Arc magmatism and accretionary orogenesis, presumably facilitated by trench retreat, resulted in southward growth of the continent. The resulting continental architecture exhibits a north to south transition from a strong cratonic backstop that maintained its lithospheric integrity throughout supercontinental assembly to a weak accretionary domain within which palinspastic reconstructions and orogenic correlations are inherently difficult. This cratonic to accretionary transition provides a robust tectonic vector that constitutes a first-order test of paleogeographic reconstructions of supercontinents.

ACKNOWLEDGMENTS

We thank Brendan Murphy for his scientific insights but more importantly for the passion, enthusiasm and camaraderie that he has brought to our scientific community. STJ and BMS acknowledge the Geological Survey of Canada (in particular Marc St Onge and Garth Jackson), Baffinland Iron Mines Corporation and the National Science and Engineering Research Council of Canada (grants RGPIN-2019-04275 and RGPIN-2021-03306, respectively) for supporting their research on northern Baffin Island. PG thanks the GEUS, Denmark for supporting his Greenland mapping campaigns, and RL thanks the GTK, Finland for support of his Fennoscandian research programs. Cecilio Quesada, Volume Editor, and two reviewers, Gabriel Gutiérrez-Alonso and Mohammad Sayab, are thanked for their comments, suggested revisions and advice, all of which significantly improved the manuscript.

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Figure Captions

Figure 1. Proposed paleogeography of Columbia (after Zhao et al., 2002, 2003). Northeastern Laurentia and Fennoscandia, which constitute part of the NUNA core of Columbia, are highlighted. Abbreviations of selected cratons: A-Australian Cratons; B-Baltica; EA- East Antarctica; IN-India; NA-North America; NC-North China; S-Siberia; SA-South America; SAF-South Africa; SC-South China; WAF-West Africa.

Figure 2. Summary of the main lithostructural elements of Baffin Island. Dominant vergence directions are indicated with red arrows. The Accretionary complex includes the Foxe (FFB) the Dorset (DFB) fold belts, an accreted oceanic arc (N: Narsajuaq Arc, in orange) and a series of quartzo-feldspathic gneissic domains (H: Hall; MI: Meta Incognita; F: Foxe, all shown in lilac) that together are thought to constitute a coherent 'Meta Incognita' continental terrane. C-Q: Igneous rocks of the Cumberland Batholith and the Qikiqtarjuaq Plutonic Suite (pink). Mesoproterozoic and younger sedimentary rocks are shown in grey. Localities discussed in the text: SI – Steensby Inlet, MR: Mary River, ISZ: Isortoq Shear Zone. See text for relevant references.

Figure 3. Simplified Geology of the southeastern Steensby Inlet area ("domain 2C" of Jackson, 2000), location indicated in Fig. 2. Folded panels of Piling group and gneissic units of the Rae Craton define an asymmetric NE-plunging fold, named here the Steensby Antiform, which is kinematically linked to the dextral-normal ISZ. Geology after Jackson et al. (1975), Morgan (1982), Bethune and Scammell (1997), Skipton et al. (2020) and Saumur et al. (2022).

Figure 4. Composite geological map of Greenland showing the main cratons and orogenic belts. Dominant vergence directions are shown with red arrows (as in Fig. 2). See text for discussion and relevant references.

Figure 5. a) Simplified geological map of the Fennoscandian Shield based on Koistinen et al. (2001). Oroclines from Beunk and Kuipers (2012), Lahtinen et al. (2014, 2022), and Lahtinen and Huhma (2019). WTBC = West Troms Basement Complex. IA = Inari arc; CLB = Central Lapland Belt; CLGC = Central Lapland Granitoid Complex; KB = Kuusamo belt; PB = Peräpohja belt; K = Kiruna; SB = Savo belt; CLGC = Central Lapland granitoid complex; SD = Skellefteå district; PoB = Pohjanmaa belt; TB = Tampere belt; PB = Pirkanmaa belt; HB = Häme belt; UB = Uusimaa belt; BA = Bergslagen area. Lithotectonic domains in Sweden (Stephens, 2020): No = Norrbotten; Ök = Överkalix; B-S = Bothnia–Skellefteå; Lj = Ljusdal, Be = Bergslagen; Sm = Småland. b) Inset: Baltica/East European Craton and TESZ = Trans-European Suture Zone. TIB = Transscandinavian Igneous Belt. c) Inset: Lapland-Kola and Svecofennian orogens.

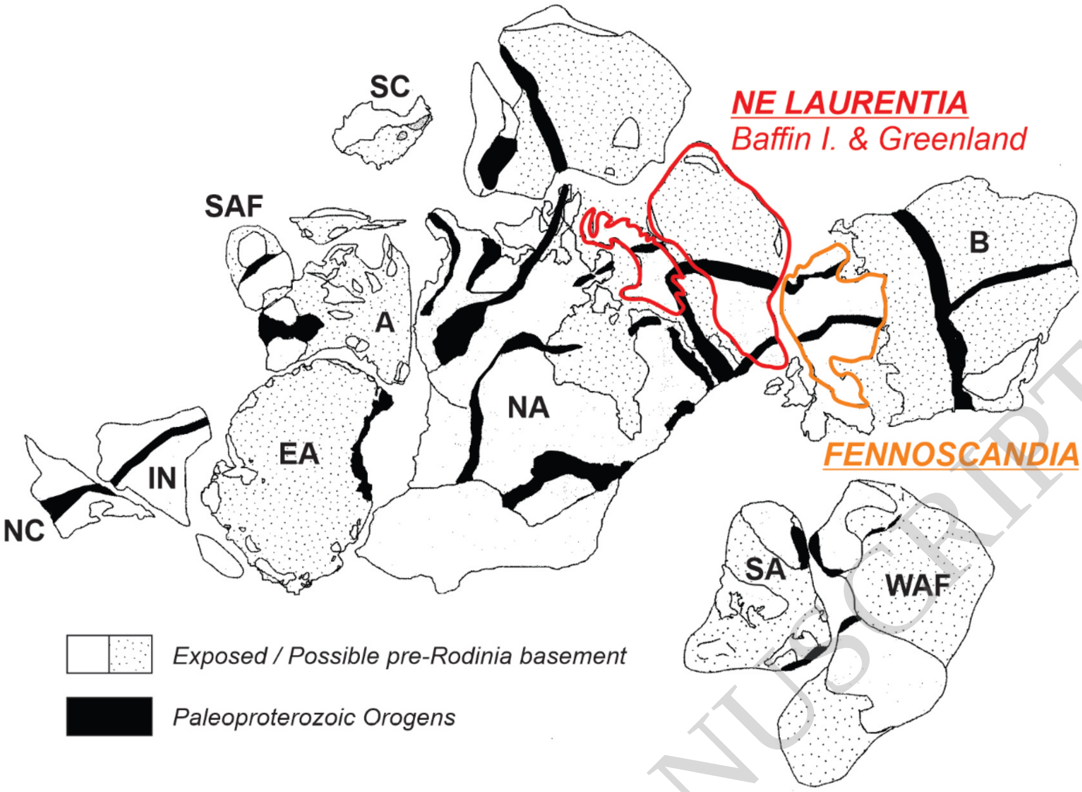
SC
NE LAURENTIA
Baffin I. & Greenland
SAF
B
A
NA
EA
IN
NC
FENNOSCANDIA
SA
WAF
Exposed / Possible pre-Rodinia basement
Paleoproterozoic Orogens

Figure 1

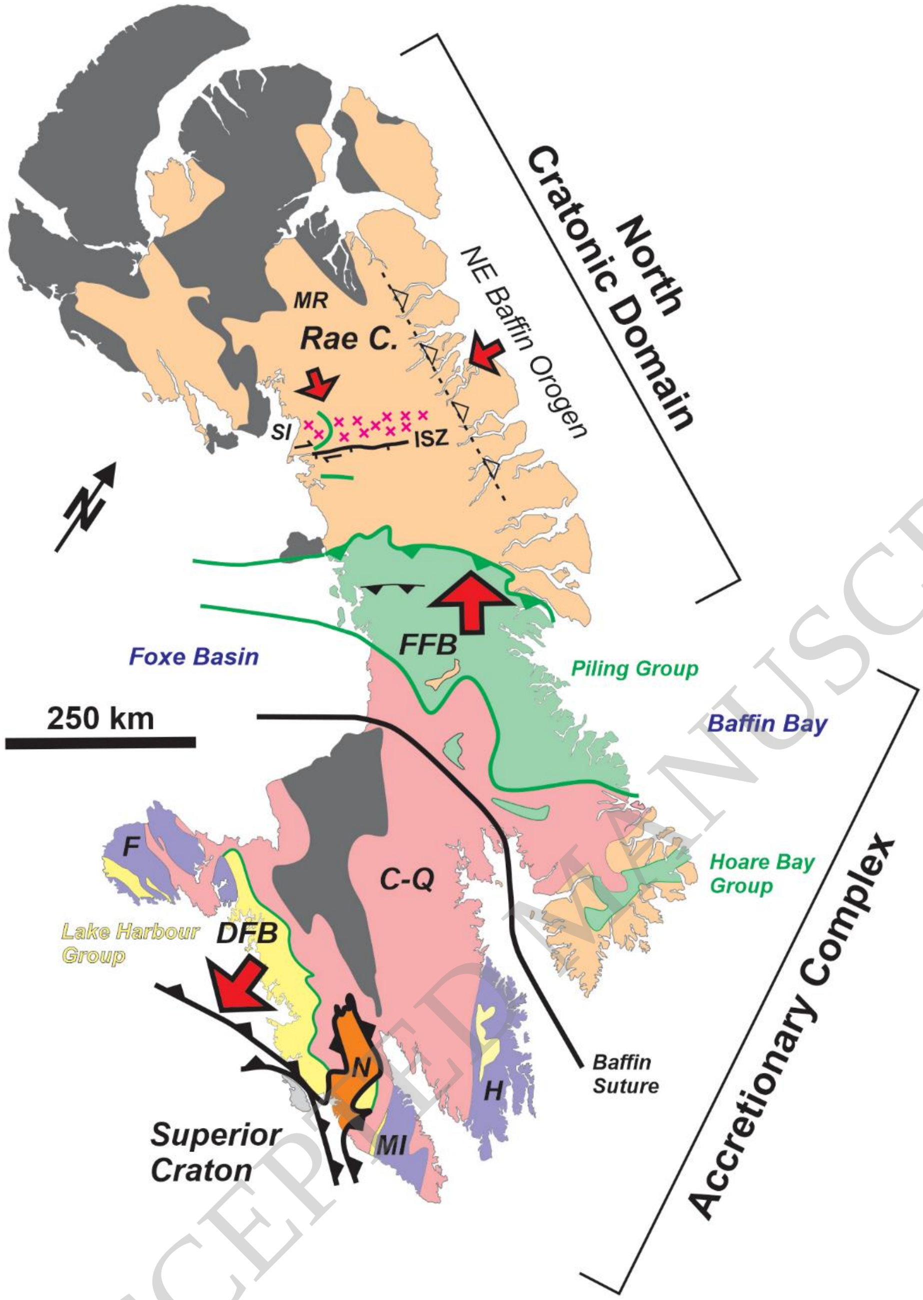
North Cratonic Domain
NE Baffin Orogen
MR
Rae C.
SI
ISZ
N
Foxe Basin
FFB
Piling Group
250 km
Baffin Bay
F
C-Q
Hoare Bay Group
Lake Harbour Group
DFB
N
H
Baffin Suture
Superior Craton
MI
Accretionary Complex

Figure 2

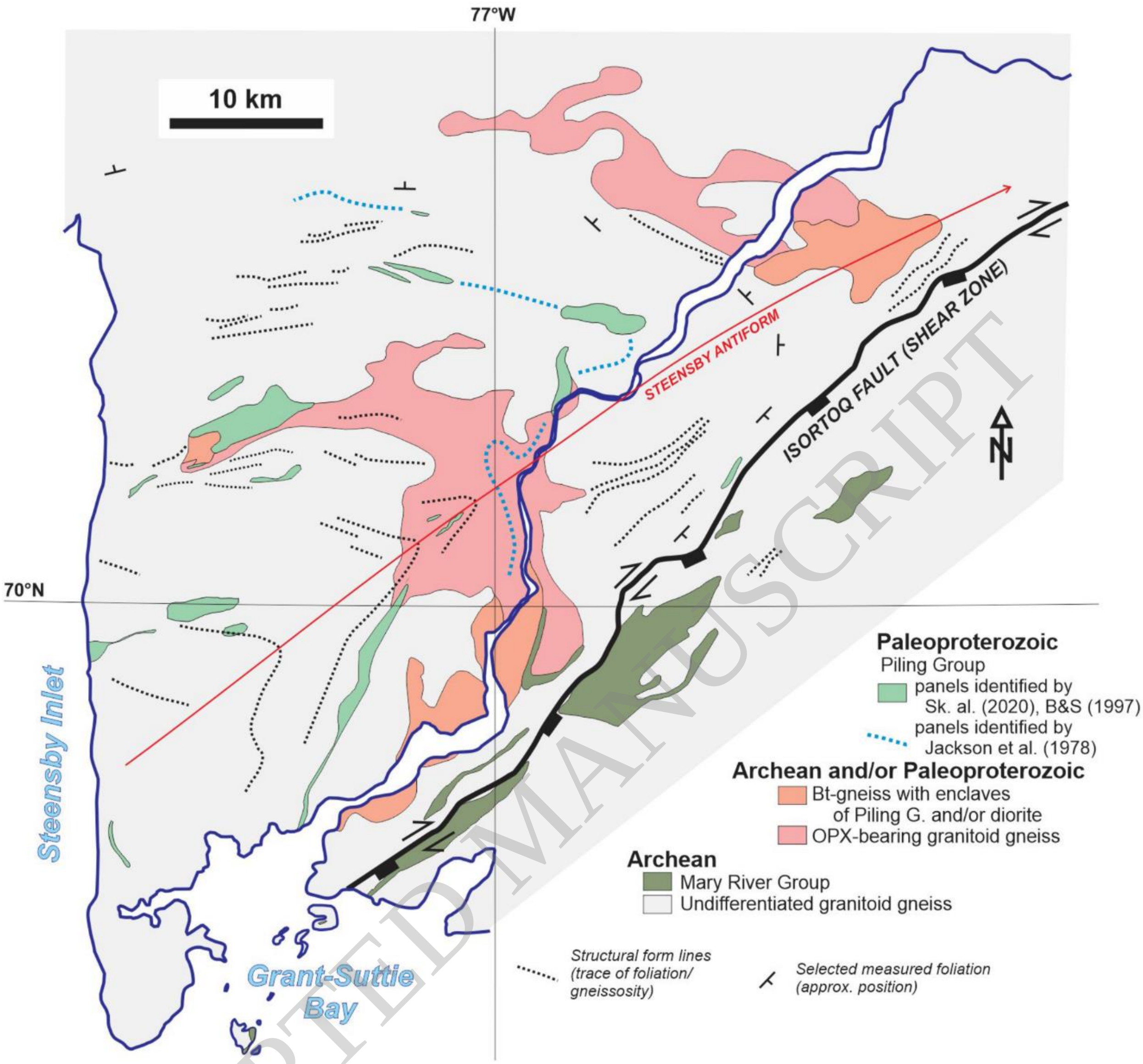
77°W
10 km
70°N
STEENSBY ANTIFORM
ISORTOQ FAULT (SHEAR ZONE)
N
Steensby Inlet
Grant-Suttie Bay
Paleoproterozoic
Piling Group
panels identified by Sk. al. (2020), B&S (1997)
panels identified by Jackson et al. (1978)
Archean and/or Paleoproterozoic
Bt-gneiss with enclaves of Piling G. and/or diorite
OPX-bearing granitoid gneiss
Archean
Mary River Group
Undifferentiated granitoid gneiss
Structural form lines (trace of foliation/ gneissosity)
Selected measured foliation (approx. position)

Figure 3

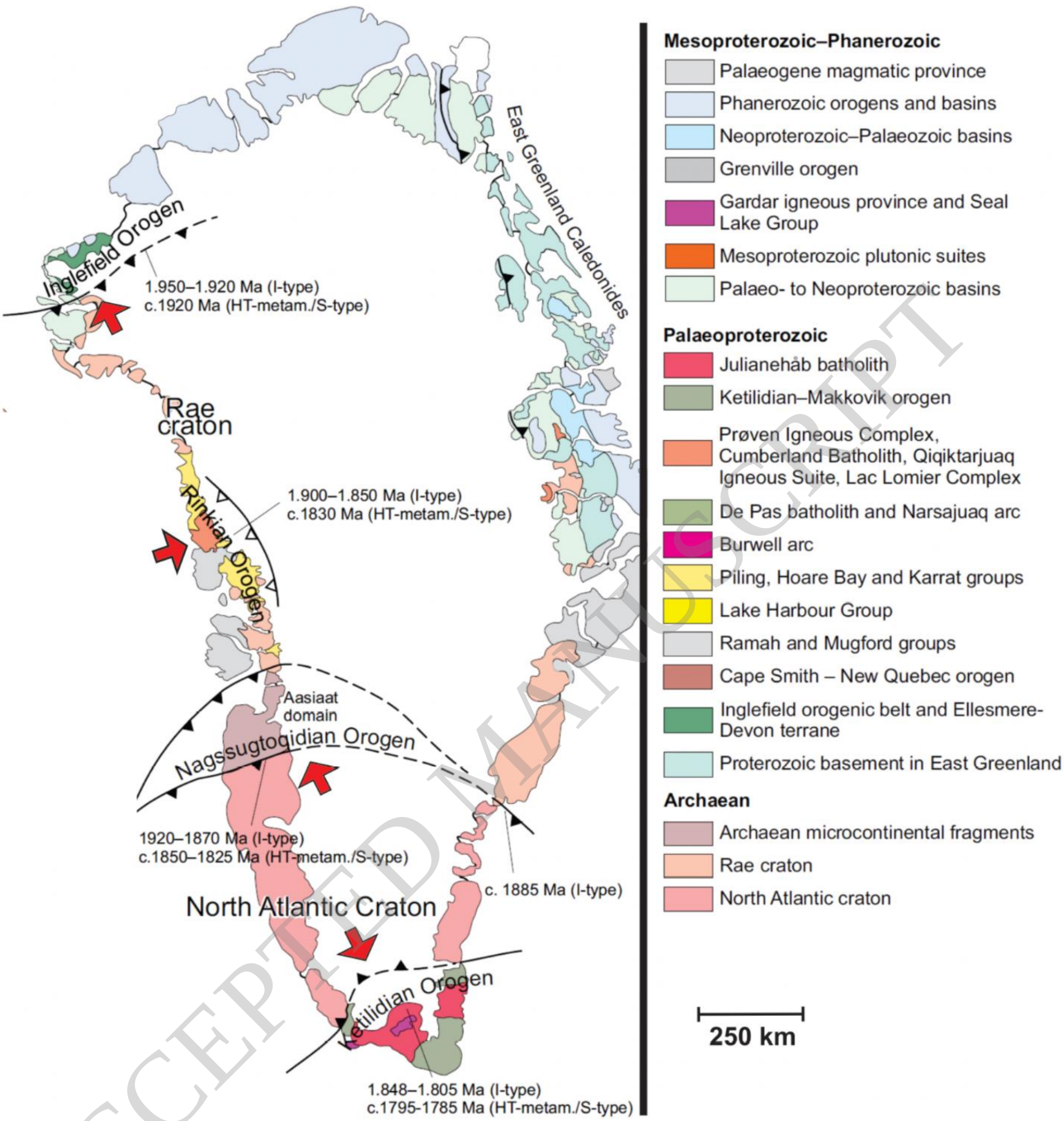
East Greenland Caledonides
Inglefield Orogen
1.950–1.920 Ma (I-type)
c.1920 Ma (HT-metam./S-type)
Rae
craton
Rinkian Orogen
1.900–1.850 Ma (I-type)
c.1830 Ma (HT-metam./S-type)
Aasiaat
domain
Nagssugtoqidian Orogen
1920–1870 Ma (I-type)
c.1850–1825 Ma (HT-metam./S-type)
c. 1885 Ma (I-type)
North Atlantic Craton
Ketilidian Orogen
1.848–1.805 Ma (I-type)
c.1795-1785 Ma (HT-metam./S-type)
Mesoproterozoic–Phanerozoic
Palaeogene magmatic province
Phanerozoic orogens and basins
Neoproterozoic–Palaeozoic basins
Grenville orogen
Gardar igneous province and Seal Lake Group
Mesoproterozoic plutonic suites
Palaeo- to Neoproterozoic basins
Palaeoproterozoic
Julianehåb batholith
Ketilidian–Makkovik orogen
Prøven Igneous Complex, Cumberland Batholith, Qiqiktarjuaq Igneous Suite, Lac Lomier Complex
De Pas batholith and Narsajuaq arc
Burwell arc
Piling, Hoare Bay and Karrat groups
Lake Harbour Group
Ramah and Mugford groups
Cape Smith – New Quebec orogen
Inglefield orogenic belt and Ellesmere-Devon terrane
Proterozoic basement in East Greenland
Archaean
Archaean microcontinental fragments
Rae craton
North Atlantic craton
250 km

Figure 4

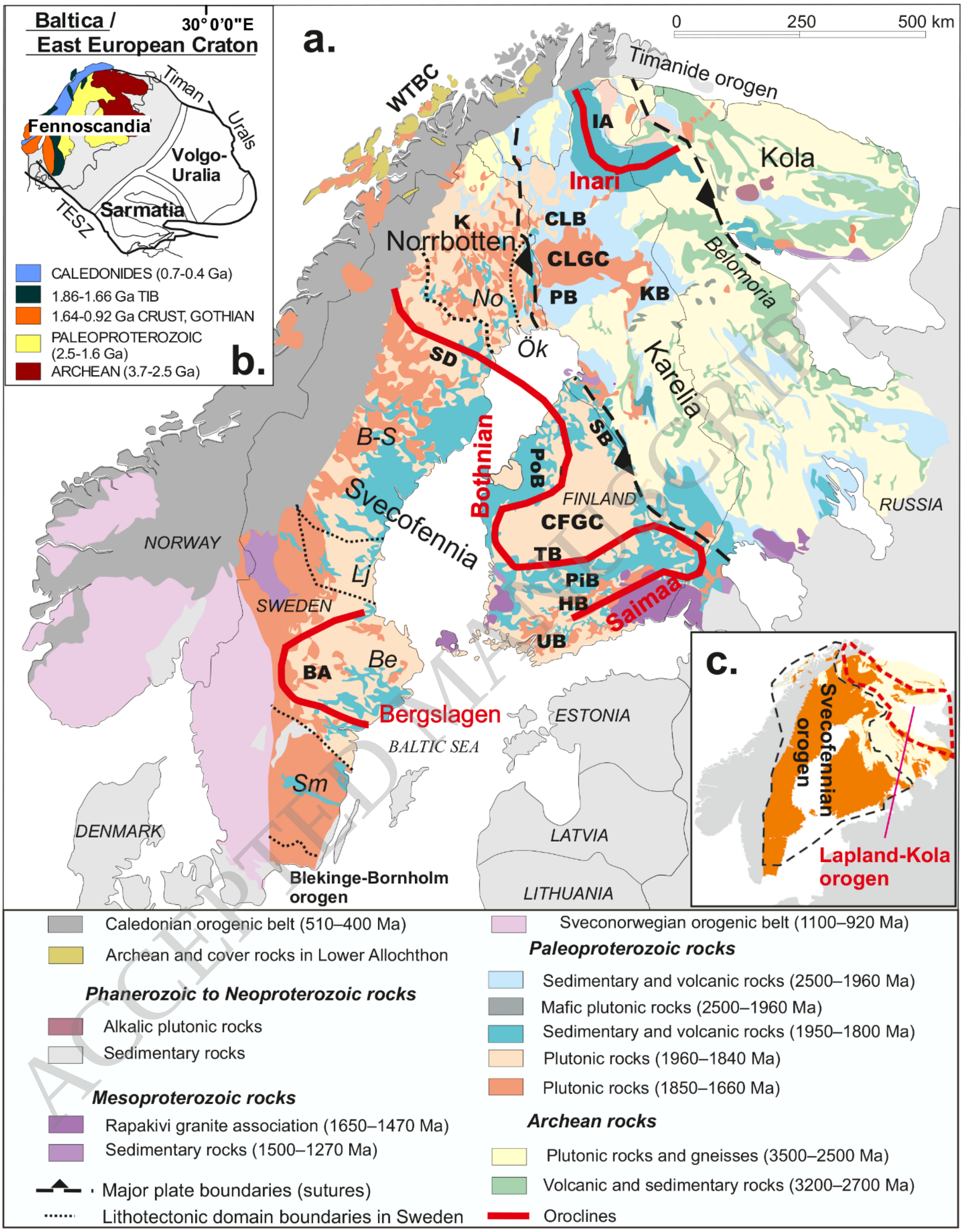
Baltica /
East European Craton
30° 0'0"E
Fennoscandia
Timan
Urals
Volgo-
Uralia
Sarmatia
TESZ
CALEDONIDES (0.7-0.4 Ga)
1.86-1.66 Ga TIB
1.64-0.92 Ga CRUST, GOTHIAN
PALEOPROTEROZOIC
(2.5-1.6 Ga)
ARCHEAN (3.7-2.5 Ga)
b.
a.
0
250
500 km
WTBC
Timanide orogen
IA
Inari
Kola
CLB
K
Norrbotten
CLGC
PB
KB
No
Ök
Belomoria
SD
Karelia
SB
B-S
PoB
Bothnian
Svecofennia
FINLAND
CFGC
RUSSIA
NORWAY
TB
Lj
PiB
HB
Saimaa
SWEDEN
UB
BA
Be
Bergslagen
ESTONIA
BALTIC SEA
Sm
DENMARK
LATVIA
Blekinge-Bornholm
orogen
LITHUANIA
c.
Svecofennian
orogen
Lapland-Kola
orogen
Caledonian orogenic belt (510–400 Ma)
Archean and cover rocks in Lower Allochthon
Phanerozoic to Neoproterozoic rocks
Alkalic plutonic rocks
Sedimentary rocks
Mesoproterozoic rocks
Rapakivi granite association (1650–1470 Ma)
Sedimentary rocks (1500–1270 Ma)
Major plate boundaries (sutures)
Lithotectonic domain boundaries in Sweden
Sveconorwegian orogenic belt (1100–920 Ma)
Paleoproterozoic rocks
Sedimentary and volcanic rocks (2500–1960 Ma)
Mafic plutonic rocks (2500–1960 Ma)
Sedimentary and volcanic rocks (1950–1800 Ma)
Plutonic rocks (1960–1840 Ma)
Plutonic rocks (1850–1660 Ma)
Archean rocks
Plutonic rocks and gneisses (3500–2500 Ma)
Volcanic and sedimentary rocks (3200–2700 Ma)
Oroclines

Figure 5