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Paleoproterozoic Supercontinent: Origin and Evolution of Accretionary and Collisional Orogens Exemplified in Northern Cratons

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Abstract—The evolution of the North American, East European, and Siberian cratons is considered. The Paleoproterozoic juvenile associations concentrate largely within mobile belts of two types: (1) volcanic–sedimentary and volcanic–plutonic belts composed of low-grade metamorphic rocks of greenschist to low-temperature amphibolite facies and (2) granulite–gneiss belts with a predominance of high-grade metamorphic rocks of high-temperature amphibolite to ultrahigh-temperature granulite facies. The first kind of mobile belt includes paleosutures made up of not only oceanic and island-arc rock associations formed in the process of evolution of relatively short-lived oceans of the Red Sea type but also peripheral accretionary orogens consisting of oceanic, island-arc, and backarc terranes accreted to continental margins. The formation of the second kind of mobile belt was related to the activity of plumes expressed in vigorous heating of the continental crust; intraplate magmatism; formation of rift depressions filled with sediments, juvenile lavas, and deposits of pyroclastic flows; and metamorphism of lower and middle crustal complexes under conditions of granulite and high-temperature amphibolite facies that, in addition, spreads over the fill of rift depressions. The evolution of mobile belts pertaining to both types ended with thrusting in a collisional setting. Five periods are recognized in Paleoproterozoic history: (1) origin and development of a superplume in the mantle that underlay the Neoarchean supercontinent; this process resulted in separation and displacement of the Fennoscandian fragment of the supercontinent (2.51–2.44 Ga); (2) a period of relatively quiet intraplate evolution complicated by locally developed plume- and plate-tectonic processes (2.44–2.0 (2.11) Ga); (3) the origin of a new superplume in the subcontinental mantle (2.0–1.95 Ga); (4) the complex combination of intense global plume- and plate-tectonic processes that led to the partial breakup of the supercontinent, its subsequent renascence and the accompanying formation of collisional orogens in the inner domains of the renewed Paleoproterozoic supercontinent, and the emergence of accretionary orogens along some of its margins (1.95–1.75 (1.71) Ga); and (5) postorogenic and anorogenic magmatism and metamorphism (<1.75 Ga).

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INTRODUCTION

The Early Precambrian geological evolution of our planet, the models of formation of the continental crust, and the origin and evolution of continents and supercontinents in the Early Precambrian have been topics of active discussions in recent decades. In these discussions, the Paleoproterozoic time often has been regarded as a transitional stage, a peculiar bridge between the preplate-tectonic evolution in the Archean and the present-day plate tectonics. Furthermore, the sequence of origination and breakdown of supercontinents in the geological history of the Earth is a subject of discussion. The onset of Paleoproterozoic evolution is marked by rifting within the oldest large continent (the Pangea-0 supercontinent [39]). I previously adverted to the specific Paleoproterozoic evolution mainly in reference to the Fennoscandian Shield and, to a lesser extent, to the North American Craton [33, 106]. On the basis of key (in my opinion) geochronological, petrologic, geochemical, and lithologic data on the East European, North American, and Siberian cratons, a comprehensive substantiation of the Paleoproterozoic evolution of the oldest supercontinent was presented in [109]. This publication, which unfortunately turned out to be virtually inaccessible to Russian specialists, has been used as a basis for this paper and has been supplemented with new data and illustrations.

The paleogeographic reconstructions for the Early Precambrian meet difficulties because of many gaps in the paleomagnetic database [59]. Nevertheless, it has been established that the paths of apparent wandering of the paleomagnetic pole over most of the Paleoproterozoic Eon virtually coincide for the main shields, providing evidence for a large supercontinent that existed from ~2.9 to 1.1 Ga ago [116]. The currently discussed reconstructions of the former supercontinents are mainly based on the correlation of geological units and events established in their present-day fragments. The basically different models of the origin and evolution of Proterozoic supercontinents known to date

[66, 75, 122] demonstrate substantial disagreements in the understanding of the nature of the Early Precambrian tectonic units.

The Paleoproterozoic evolution was predated by a number of occurrences of the marginal continental type of magmatic activity at the end of the Archean, 2.7 Ga ago, that were recorded in an expressive frequency peak of age estimates [3, 66]. This specific feature may be interpreted as evidence for the collision of large continental masses and the origin of the first supercontinent or a few large composite continents [66]. The subsequent Paleoproterozoic evolution that began ~2.5 Ga ago is regarded a process of breakup (at least partial) of the Archean supercontinent [39, 67, 75]. The recovery of the supercontinent's integrity was preceded by a mass increment of the juvenile crust that began about 2.0–1.9 Ga ago and ended with the fast accretion of island-arc systems and the formation of accretionary orogens along the margins of the reorganized supercontinent 1.88–1.84 Ga ago.

The Paleoproterozoic juvenile associations largely participated in the formation of two types of tectonic units: (1) volcanic–sedimentary and volcanic–plutonic belts composed of low-grade metamorphic rocks of greenschist to low-temperature amphibolite facies and (2) granulite–gneiss belts with a predominance of high-grade metamorphic rocks of high-temperature amphibolite to ultrahigh-temperature granulite facies. The mobile belts of the first type are commonly interpreted as sutures of collisional orogens or continental rifts that completed their evolution in a setting of general compression. In addition, volcanic–plutonic associations localized along margins of the ancient continent are considered accretionary orogens [89, 153]. The interpretation of the tectonic and geodynamic role of the granulite–gneiss (granulite for the sake of brevity) belts is not so obvious. As follows from structural evidence, most large Paleoproterozoic granulite complexes are related to collisional events in some way. However, the geochemical signatures of intrusive charnockites and enderbites from these belts are close to the respective parameters of calc-alkaline and tonalite–trondhjemite–granodiorite series. Because of this circumstance, the suprasubduction (island-arc or marginal continental) origin is commonly ascribed to these rocks [92, 148], and granulite belts are interpreted as paleosutures (collisional orogens) together with the volcanic–sedimentary belts of the first type [30, 67, 69, 89, 123]. However, several attributes of granulite belts (see below) are inconsistent with such an interpretation.

The factual information in this paper is presented as a review of reference geochronological, petrologic, geochemical, and lithologic data on the evolution of the East European, North American, and Siberian cratons. The age estimates, as a rule, are based on U–Pb determinations; the use of Sm–Nd data is always specified; and the references to the sources of information are given in the text. The periodization of the main events and the nomenclature of the stages in the evolution of the lithosphere that were proposed in [109] are accepted.

The main objectives of the paper are to reveal evidence for the existence of Paleoproterozoic supercontinent and to develop a model of its geological evolution, including periodization and the specific features of this process as well as comparison of the Paleoproterozoic evolution with both the preceding Archean events and the subsequent stages of geological history. Special attention will be paid to the interaction of events related to plume and plate tectonics in the process of Paleoproterozoic crust formation. Another aim of this paper is to reappraise the existing genetic models of granulite belts and the spatial and temporal relationships between granulite and volcanic–sedimentary belts on the basis of available data.

The manifestations of magmatic and thermal activity in the early Paleoproterozoic are largely concentrated in the ancient continent called Laurentia. According to [64], this continent embraced the North American and Fennoscandian cratons. The term Lauroscandia is used below, as recommended by V.E. Khain, since most researchers apply this term only to the North American Craton.

In the structural grain of the present-day lithosphere, Fennoscandia (Fennoscandian crustal segment, after [55]) is situated in the northwestern portion of the East European Craton, which also includes the Volga–Ural and Sarmatian segments (see Fig. 6). Sarmatia existed as an independent continent until at least 2.05–2.0 Ga ago, and its amalgamation with Fennoscandia ended about 1.7 Ga ago [55]. In contrast to the early Paleoproterozoic, the middle and late Paleoproterozoic rock associations are widespread in almost all continents. This study is focused on the structure and evolution of the main Paleoproterozoic tectonic units of Lauroscandia and the Siberian Carton. Owing to relatively weak post-Paleoproterozoic reworking, these units may be accepted as typical objects. The temporal relations between the main Paleoproterozoic rock associations in tectonic belts of Lauroscandia and Siberia are shown in Fig. 1.

GEOLOGICAL EVOLUTION OF NORTHERN CONTINENTS IN THE PALEOPROTEROZOIC: A REVIEW OF KEY EVENTS

Geological Evolution of Lauroscandia

The Inner Domain of the Continent (Intracontinental Collisional Oorogens)

Superplume and initial rifting of the Archean continent (2.51–2.44 Ga). The onset of the Paleoproterozoic evolution of Lauroscandia (Figs. 1–3, see Fig. 1, 6) about 2.5 Ga ago is marked by the emplacement of mantle-derived magma and the formation of mafic dikes and mafic–ultramafic intrusions in the crust. The age of the layered peridotite–gabbro–norite intrusions localized in the upper crust is estimated at 2.51–2.44 and ~2.4 Ga

[42, 45, 111]. The gabbroanorthosite bodies in the lower crust were formed at virtually the same time (2.49–2.43 Ga) [24, 38, 98, 110]. Their emplacement was accompanied by the metamorphism of high-temperature amphibolite and granulite facies. The formation of the Kolvitsa gabbroanorthosite pluton in the Kola Peninsula 2.46 Ga ago was followed by numerous mafic injections into tension cracks. Virtually all of the dikes contain xenoliths of foliated and metamorphosed gabbroanorthosite [48]. The *PT* conditions of metamorphism (700–900°C and 10–12 kbar) that correspond to a depth of 35–43 km have been obtained for gabbroanorthosite, country mafic granulites, and small gabbro and diorite intrusive bodies at a short distance from the Kolvitsa pluton [48, 53, 80].

From 2.5 to 2.1 Ga, the western (in present-day coordinates) margin of the Kola–Karelian continent was gradually overlapped by terrigenous sediments, dolomites, black shales, and tholeiitic lava flows (Lapponian and Kalevian groups), i.e., by the rock association characteristic of passive margins complicated by rifting.

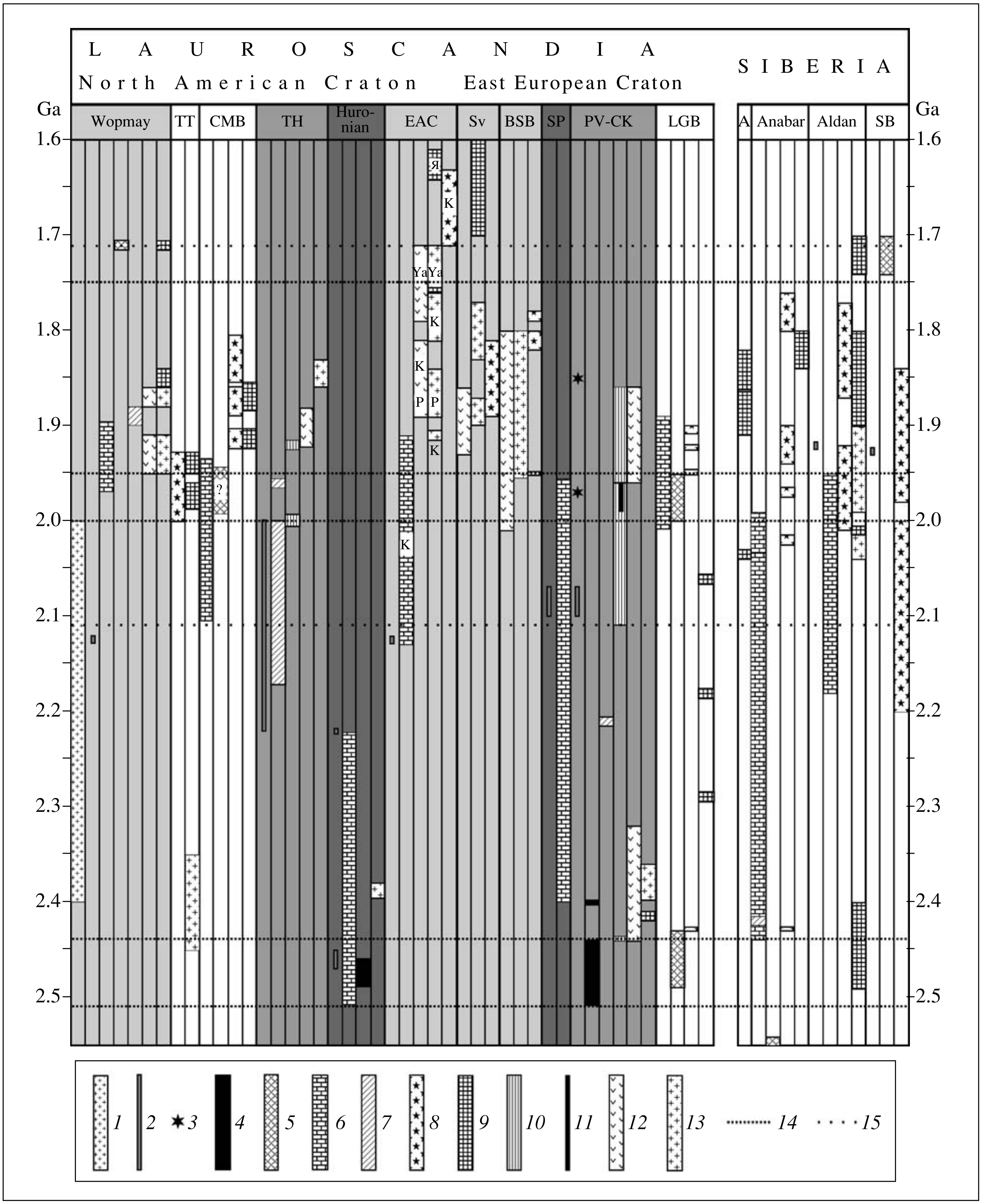
The initial rifting in the framework in the Superior Craton of the Canadian Shield is marked by the formation of the Huronian Supergroup that unconformably overlies the Archean basement. The lowermost portion of the section comprising plateau basalts, felsic lavas, and arkosic metasediments, is cut through by gabbroanorthosite, mafic–ultramafics, and granitoid intrusions 2.49–2.46 Ga in age as well as by mafic dikes of the Matachewan swarm, which is dated at 2.47–2.45 Ga. The upper part of the section composed of tillite, shale, quartzite, and carbonate rocks and crowned by cross-bedded sandstone, is, in turn, cut by metadolerite dikes 2.22 Ga in age [68, 89].

A certain similarity of the early Paleoproterozoic magmatism in the North American and Karelian cratons stimulated the development of the models that suggested the participation of these cratons in the structure of the common Archean continent of Lauroscandia. The similarity of mafic and ultramafic intrusions that formed 2.51–2.44 Ga ago in the aforementioned territories [149], as well as the similarity of the Huronian and Lapponian shelf sediments (2.49–2.22 and 2.5–2.3 Ga, respectively), is evident. The time of the suggested breakup of Lauroscandia by plume-initiated rifting is estimated differently. Heaman [88] considered this event a result of the opening of the Matachewan ocean about 2.45 Ga ago and the transformation of the Huronian sedimentary basin into a passive margin. In the opinion of Ernst and Blicker [40], the correlation of the dike-swarm orientation in the Karelian and Superior cratons indicates that such a division occurred 2.1–2.0 Ga ago. The paleomagnetic data show that these cratons occupied distinct spatial positions already about 2.45 Ga ago [60, 105]. However, the breakup of Lauroscandia could have immediately followed the initial stage of rifting 2.51–2.50 Ga ago, i.e., 50–70 Ma before the time recorded in the paleomagnetic data. The structural image of the reconstructed continent consisting of a number of oval (in plan view) and concentrically arranged tectonic belts [108] (see Fig. 6), provides an argument in favor of this suggestion.

In a broader view, the character of the magmatism, high-grade metamorphism, and sedimentation from 2.51 to ~2.44 Ga indicates extension conditions and an influx of mantle heat to the inner domain of a rather vast continent. As will be shown below, similar processes were inherent at that time to other continental domains. The synchronism of these processes in distant areas shows that they probably belonged to a common continent (probably a supercontinent). The emplacement of mantle-derived magma, in combination with high-grade metamorphism, allows the suggestion that these processes were related to the plume type of mantle phenomena. The wide occurrence of these processes is consistent with the idea of superplumes.

Quiescent within-plate development (2.44–2.0 (2.11) Ga). The period from ~2.44 to 2.0 Ga was characterized by moderate tectonic and magmatic activity both in Lauroscandia and elsewhere [66]. The volcanic–sedimentary sequences with abundant mafic volcanics dated at 2.44–2.40 Ga correspond to the onset of this period. They are predominant among the oldest rocks of the Pechenga–Imandra–Varzuga (Pechenga–Varzuga for the sake of brevity) and Circum-Karelian belts [21, 104, 119, 131]. According to the seismic data, these belts are through-crustal over- and underthrust assemblies [17–19, 21, 134]. The low-K tholeiite at the base of the sequence is associated with shallow-water sandstone and overlain by lacustrine quartz sandstone and low-Ti basaltic andesite that are prevalent in volume. Low-Ti and high-Mg basalts (basaltic komatiites) are less abundant, and andesite and dacite occur in subordinate amounts. The section is completed with dacite and rhyolite 2.44–2.42 Ga in age and subaerial andesite 2.32 Ga in age [36, 45]. Basaltic komatiite and basaltic andesite that formed 2.41 Ga ago dominate in the Vetreny Belt (the southwestern segment of the Circum-Karelian Belt) [119]. Low-Ti basaltic andesite is dominant in the lower portion of the section of the Shomba and Lehta structural units of the Circum-Karelian Belt, and rhyolitic ignimbrite dated at 2.44 Ga is abundant [10, 11]. The geochemistry of mafic volcanic rocks indicates their relations to the plume or T-MORB types. Some features may be a result of the crustal contamination of suprasubduction melts [11, 35].

The upper part of the section that pertains to this period is composed of alkali basalt and trachybasalt in association with alkali picrite and rhyolite that date to ~2.2 Ga and intercalate with epicontinental red beds, stromatolite limestone and dolomite, and phosphate- and manganese-bearing interlayers. In the Superior Craton, similar sequences are known at the passive margin complicated by rifting. The fragments of this



sequence have been retained in the New Québec and Trans-Hudson belts of the North American Craton [89]. From 2.3 to 2.1 Ga ago, most of the area of the Karelian Craton was a passive margin covered with shelf sediments. The beginning of a new stage of rifting is recorded in the emplacement of metadolerite sills that occurred 2.2–1.97 Ga ago [150]. Similar events in the North American Craton are recorded in dike swarms that formed 2.22–1.99 Ga ago [73 and references therein].

Fig. 1. Correlation of the main events in the evolution of the Paleoproterozoic orogens of Lauroscandia and Siberia, modified after [109]. (*1*) Poorly studied continental crust of presumably subduction-related origin; (*2*) mafic dikes; (*3*) alkaline mafic–ultramafic intrusions; (*4*) layered mafic–ultramafic intrusions; (*5*) gabbroanorthosite; (*6*) sedimentary sequences with platform and rift-related volcanics; (*7*) rift-related volcanic–sedimentary complexes; (*8*) granulite–gneiss complexes; (*9*) intraplate granite; (*10*) MORB-type volcanics and ophiolitic complexes; (*11*) OIB-type mafic and ultramafic rocks; (*12*) island-arc volcanics; (*13*) island-arc granitoids; (*14, 15*) age boundaries: (*14*) major and (*15*) subordinate. Types of tectonic units (colors of columns): granulite–gneiss belts (white), accretionary orogens (light gray), volcanic–sedimentary belts affected by low-grade metamorphism (gray), and passive margins (dark gray). Orogenic belts: (TT) Taltson–Thelon, (CMB) belts related to the Cumberland batholith, (TH) Trans-Hudson, (EAC) east of the North American Craton, (K) Ketilidian, (P) Penokean, (Ya) Yavapai–Mazatzal, (Sv) Svecofennian Accretionary Orogen, (BSB) Belarus–South Peribaltic Orogen, (SP) Svecofennian passive margin, (PV-CK) Pechenga–Varzuga and Circum-Karelian belts, (LGB) Lapland Granulite Belt, (A) Akitkan Belt, (SB) Stanovoi Belt.

Superplume and onset of tectonic activity ((2.11) 2.0–1.95 Ga). The indications of the second superplume are known from many regions. The activity of this superplume was realized in rifting that locally passed to oceanic spreading in the region of future collisional orogens. These events are recorded in the fragmented ophiolitic sections of Purtuniq (2.0 Ga) [129] and Jormua (1.95 Ga) [100] (the Cape Smith Belt, the northern segment of the Circum-Superior Belt, and the Kainuu Belt in the western Karelian Craton, respectively) as well as in the Kittilä ophiolite-like complex in northern Finland (2.0 Ga) [84]. The T-MORB-type tholeiitic pillow lavas in the Pechenga Synform (2.11–1.96 Ga) are close in age. Tholeiites are combined with picritic lava flows and felsic ash flows; in geochemistry, these rocks resemble the products of volcanic eruptions on oceanic islands 1.99–1.96 Ga ago. Subvolcanic gabbrowehrlite related to the activity of these volcanoes intruded the lower portion of the tholeiitic sequence but has a much greater abundance in the form of lenticular tectonic boudins in the imbricate complex of sulfidized terrigenous rocks [21, 104, 131]. This complex, known as a productive sequence of the Pechenga Cu–Ni ore field, is interpreted as a fragment of accretionary prism [21]. The formation of oceanic structural units was controlled by longitudinal and transform faults. The role of intraoceanic transform faults in the evolution of magmatism within the future Pechenga Synform is illustrated in [21]. The transform faults that penetrated into the adjacent continent have been retained to a certain extent in the moved up continental plates. The continental continuation of transform faults in the northeastern and eastern framework of the Pechenga Synform are known as the Näsäkkä dike swarm. In geochemical signatures, the dikes are close to picrite lavas and Ni-bearing gabbrowehrlite of the Pechenga Synform. The intrusive bodies of the Karikjärvi peridotite–gabbronorite complex dated at 2.04–1.94 Ga are located at the extension of the dike swarm [1, 36]. Alkaline mafic and ultramafic rocks and carbonatites of the Gremyakha-Vyrmes pluton that was retained in the subducted continental plate were formed 1.97–1.95 Ga ago [8, 34]. The drowned southeastern margin of the Karelian Craton is overlain by plateau basalts dated at 1.98 Ga [118].

The most important event of this period was the formation of a new generation of gabbroanorthosite plutons 2.0–1.95 Ga in age [14, 27, 51, 107] at the same depth in the crust where gabbroanorthosite bodies that were emplaced 2.49–2.43 Ga ago were residing.

Interaction of plate and plume tectonics and formation of collisional orogens (1.95–1.75 (1.71) Ga). *The Lapland–Peribaltic Orogen.* The geochronology of sedimentary and volcanic protoliths that were subsequently transformed into the Lapland Granulite Belt (in a broad sense, i.e., including the Kandalaksha–Kolvitsa fragment) shows that a vast basin was formed about 2.0 Ga ago between the Pechenga–Varzuga Belt and the northern branch of the Circum-Karelian Belt. The age of detrital zircon grains from khondalite (metagraywacke) covers an interval of 2.71 to 1.88 Ga. Together with Sm–Nd data, this estimate indicates that the sedimentation was completed about 1.9 Ga ago [2, 58, 69, 91, 135]. Since second-generation gabbroanorthosite intruded the lower portion of the thick sedimentary–volcanic sequence and was jointly affected by high-grade metamorphism, it should be stated that deposition of the sedimentary sequence started no earlier than 2.0 Ga ago. The morphology of detrital zircon grains shows that a granite–granodiorite provenance is the most probable. The calc-alkaline volcanic rocks erupted about 1.96 Ga ago and subsequently transformed into gneisses that occur in the southern framework of the Imandra–Varzuga segment [69]. At the present-day erosion level, these rocks are exposed in the hanging wall of over- and underthrust assemblies. Therefore, the commonly postulated relations of graywacke metasediments to subduction remain equivocal.

Independent occurrences of high-grade metamorphism were documented after the emplacement of gabbroanorthosites of both generations. The same metamorphic events (M0 at 2.46–2.43 Ga and M1 at 2.0–1.95 Ga [13, 14, 27]) characterized by similar *PT* conditions (860–960°C and 10.3–14.0 kbar, see Fig. 4) were established at the base of the nappe assembly of the Lapland Belt. These events were virtually coeval with the emplacement of gabbroanorthosite. The *PT* parameters of event M2 decreased from 860 to 800°C and from 12.4 to 5.8 kbar. The always recorded event M3 proceeded with the *PT* parameters decreasing from 770°C and 10.7 kbar to 640°C and 4.8 kbar. The high-temperature metamorphisms M2 and M3 affected both igneous and sedimentary rocks ~1.95–1.92 and 1.92–1.90 Ga ago, respectively [6, 13, 14, 98, 107]. The collisional events that included thrusting and exhumation of deep

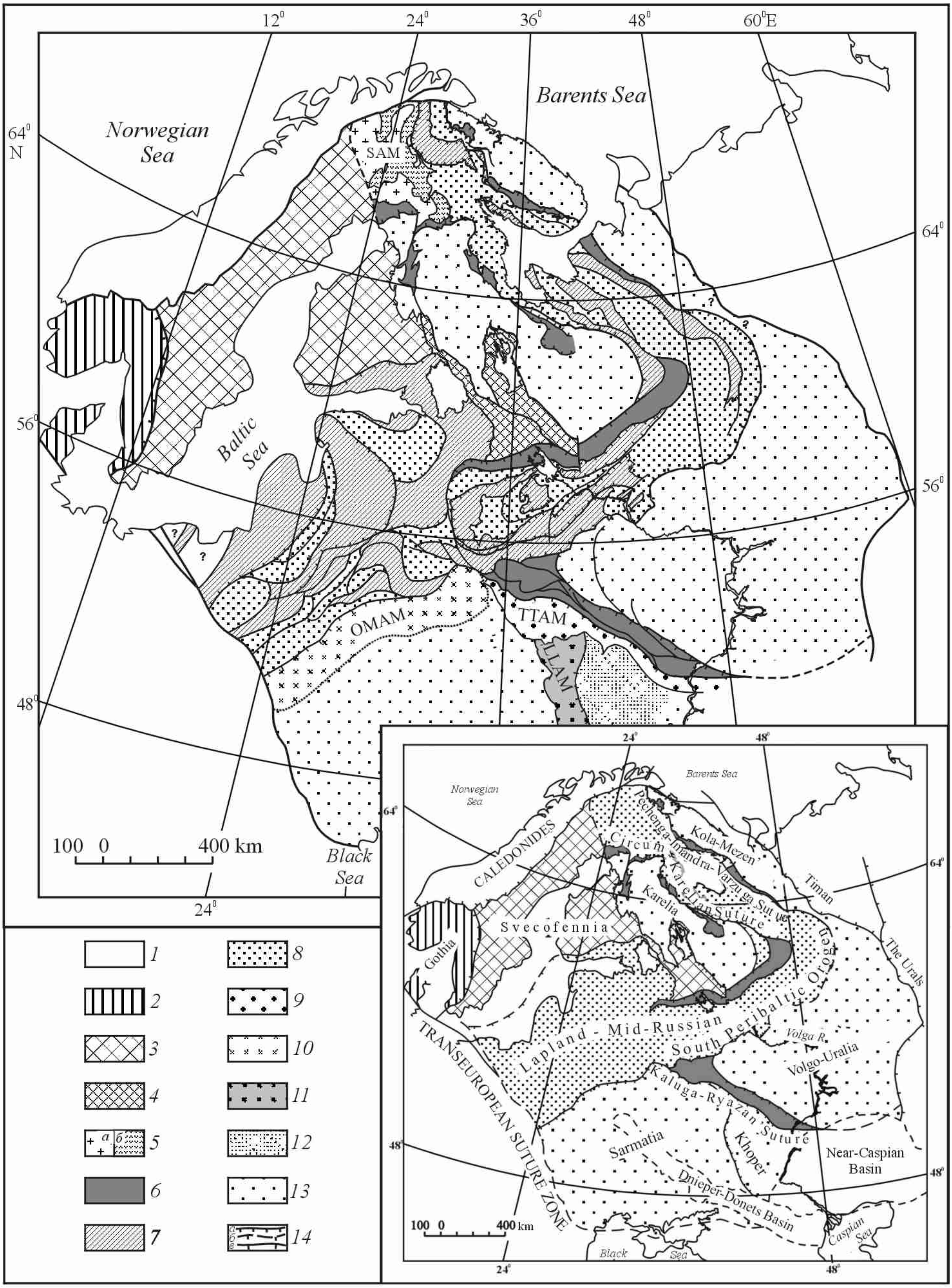


Fig. 2 -corrected

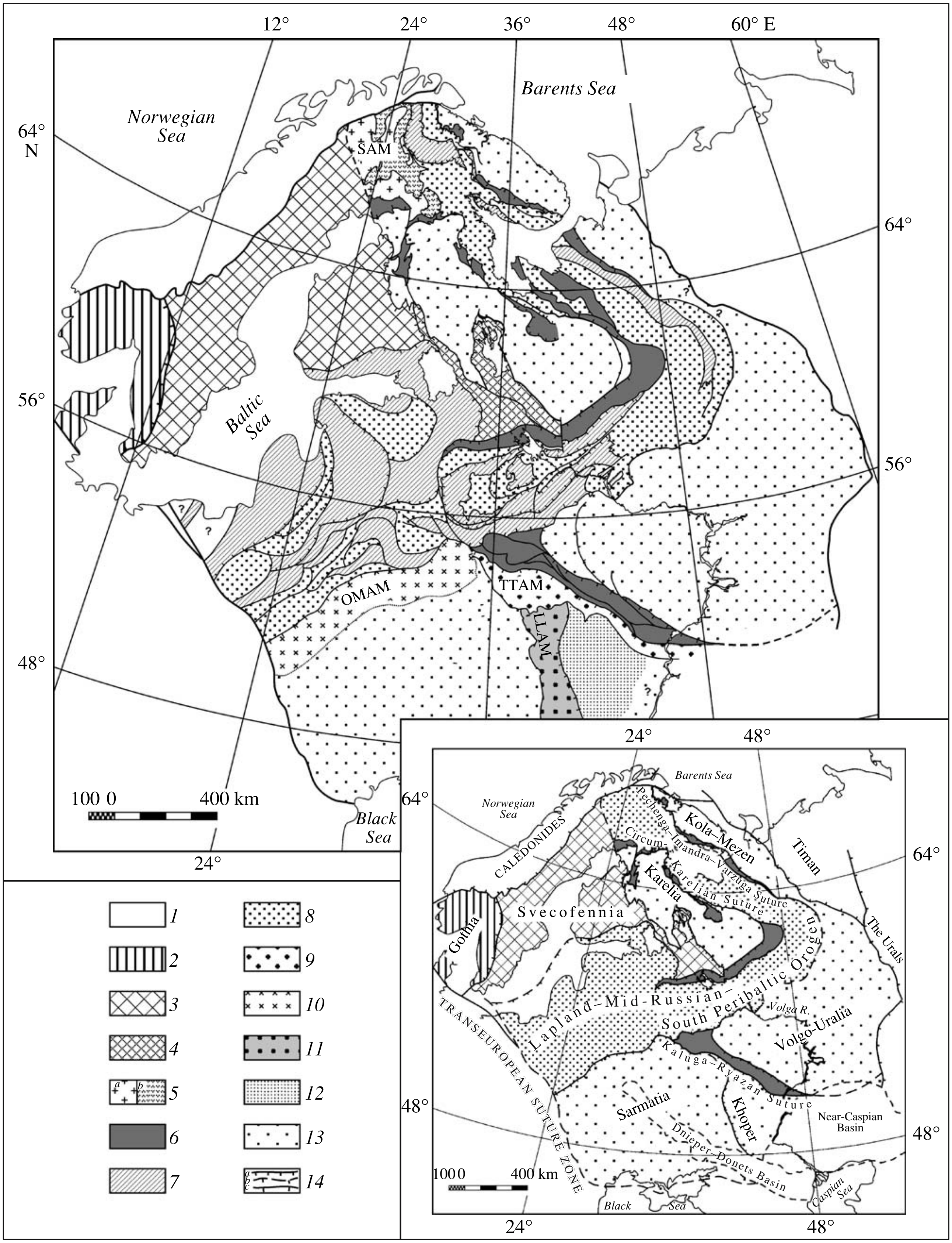

12°
24°
36°
48°
60° E
Barents Sea
Norwegian Sea
64° N
SAM
64°
56°
Baltic Sea
56°
OMAM
TTAM
LLAM
48°
100 0 400 km
Black Sea
24°
1
2
3
4
5
6
7
8
9
10
11
12
13
14
24°
48°
Barents Sea
Norwegian Sea
64°
CALEDONIDES
Pechenga–Imandra–Varzuga Suture
Kola–Mezen
Timan
Circum-Karelian Suture
Karelia
64°
Svecofennia
Gothia
The Urals
Lapland–Mid-Russian–South Peribaltic Orogen
Volga R.
Volgo-Uralia
TRANSEUROPEAN SUTURE ZONE
Kaluga–Ryazan Suture
48°
Sarmatia
Khoper
Near-Caspian Basin
48°
Dnieper–Donets Basin
1000 400 km
Black Sea
Caspian Sea
24°
48°

Fig. 2. Paleoproterozoic orogens and tectonic belts of the East European Craton, modified after [22]. The inset is a scheme of tectonic demarcation of the Early Precambrian crust in the East European Craton, Archean crustal segments, and Paleoproterozoic orogens, modified after [52]. (*1*) Phanerozoic framework of the East European Craton; (*2*) Gothian Belt (~1 Ga) of Paleoproterozoic complexes reworked in the course of the Sveco-Norwegian collision; (*3–12*) *Paleoproterozoic tectonic units*: (*3*) Svecofennian Accretionary Orogen, 1.90–1.75 Ga; (*4*) rocks complexes of the passive margin thrust over the Karelian Craton during formation of the accretionary orogen; (*5*) Svecofennian active margin (SAM): (*a*) volcanic and (*b*) intrusive complexes, 1.90–1.75 Ga; (*6*) sedimentary–volcanic belts affected by low- and moderate-grade metamorphism (sutures and deformed rifts), 2.50–1.85 Ga; (*7*) granulite–gneiss belts with protoliths dated at 2.45–1.90 Ga and granulite metamorphism dated at 1.95–1.85 Ga; (*8*) amphibolite–gneiss and migmatite belts with Neoarchean and Paleoproterozoic protoliths that metamorphosed 1.95–1.85 Ga ago; (*9–11*) volcanic–plutonic complexes of active margins that arose during formation of Sarmatia: (*9*) Tula–Tambov (TTAM), 1.95–1.75 Ga; (*10*) Osnitsa–Mikashevichi (OMAM), 2.0–1.95 Ga; (*11*) Lipetsk–Losevo (LLAM), ~2.1 Ga; (*12*) terrigenous complexes of the Vorontsovka Belt, ~2.1 Ga; (*13*) Archean granite–greenstone and granite–gneiss domains; (*14*) boundaries of the main tectonic units: (*a*) proved, (*b*) inferred, and (*c*) boundary of the Early Precambrian crust of the East European Craton. Tectonic units (*2–7*) pertain to the continent of Fennoscandia.

crustal levels are marked by inverted metamorphic zoning in para-autochthonous rocks. The transport of hot tectonic sheets composed of granulite complexes to the erosion level 1.89–1.87 Ga ago [6, 43, 69, 95, 147] was the immediate cause of this metamorphism (event M4, 650–550°C and 8.4–4.5 kbar). The deepest granulites were not involved in thrusting; they were preserved in the lower crust. Fragments of this crust are available for study as xenoliths entrapped by Devonian lamprophyre dikes and pipes on the coast of Kandalaksha Bay. The xenoliths commonly consist of garnet granulite of the norite–gabbroanorthosite composition that is similar to the rocks of the Lapland granulite complex [7, 26, 96, 97]. The systematic variation of the parameters of metamorphism through the composite crustal section that comprises granulites from the Lapland Belt and xenoliths corresponds to the total crustal thickness of approximately 70 km. This crust was affected by granulite-facies metamorphism within a depth interval from 70 to ~25 km. The tectonic sheet of the Lapland Belt corresponds to a crustal section from 45–50 to ~25 km in depth for event M2 (Fig. 4) [107].

The youngest (1.87–1.86 Ga) rock complexes of the Pechenga–Varzuga Belt that were formed after exhumation of the deep crustal levels that formed Lapland Granulite Belt included rhyolite, dacite, andesite, picrite, and high-Mg basalt of the suprasubduction type; N-MORB, in association with volcaniclastic rocks and black shale, occur in subordinate amounts [104, 131]. These deposits likely formed in the process of closure of residual basins. The formation of the Tiksheozero alkaline mafic–ultramafic complex with carbonatites in northern Karelia 1.85 Ga ago testifies to the development of plume-related processes [5].

The geophysical fields make it possible to trace the Lapland Belt to the East European Craton rather reliably. The geophysical data, combined with the results of deep drilling, were used as the basis for mapping of the basement structure and allowed recognition of arcuate granulite and gneiss–amphibolite belts about 2000 km in total extent. The system of belts that surrounds the Karelian Craton in the east and the south (Fig. 2) makes up the Lapland–Mid-Russian–South Peribaltic intracontinental collisional orogen (Lapland–Peribaltic Orogen [23] for the sake of brevity). In the central portion of the platform crossed by reference geotraverse 1-EB, the gently dipping granulite complexes compose the Nelidovo Synform that is oval in plan view; its centroclinal closure faces westward [22, 23]. The rock complexes that fill this synform remain poorly studied. Detailed geochronological and petrologic data are available only for rocks of the Moscow Granulite Belt. Here, leuconorite and enderbite were emplaced 1.98 Ga ago; granulite-facies metamorphism (~1000°C and 10–12 kbar [54]) took place afterward. Both the parameters and age of metamorphism coincide within error limits with the parameters and age of metamorphism M1 at the base of the nappe assembly of the Lapland Belt.

The basement of the East European Platform in the Baltic region and a considerable portion of Belarus may be regarded as a part of the Svecofennian accretionary orogen [82]. However, in contrast to the Svecofennides in the central part of the Fennoscandian Shield, where moderately metamorphosed volcanic–sedimentary and plutonic complexes are predominant, this domain is made up of a system of arcuate belts consisting of the rocks pertaining to greenschist, epidote-amphibolite, and high amphibolite–granulite facies of metamorphism. Unlike the Nelidovo Synform, the arcuate belts face eastward (Fig. 2).

The easternmost granulite belt in the Belarus–Baltic segment extends along the northern coast of the Gulf of Finland through the northern Ladoga region farther southward to Lake Ilmen and the town of Staraya Russa (Staraya Russa–South Finland Granulite Belt). This belt is a tectonic nappe that is overthrust in the northeastern direction. In the north Ladoga region, the peak parameters of metamorphism dated at 1.88–1.85 Ga decrease from 950–840°C and 6.4 kbar at the base of nappe to 860–780°C and 4.8 kbar upsection. The subsequent metamorphic event that occurred 1.80–1.78 Ga ago is characterized by a *PT* range of 680–600°C at 5.5 kbar to 560–620°C at 3.2 kbar. As a result, inverted metamorphic zoning arose in para-autochthonous rocks: 550°C and 3 kbar immediately beneath the granulite belt to 450°C and 1 kbar at a distance from the granulite belt (my interpretation based on the data reported in [4, 29]).

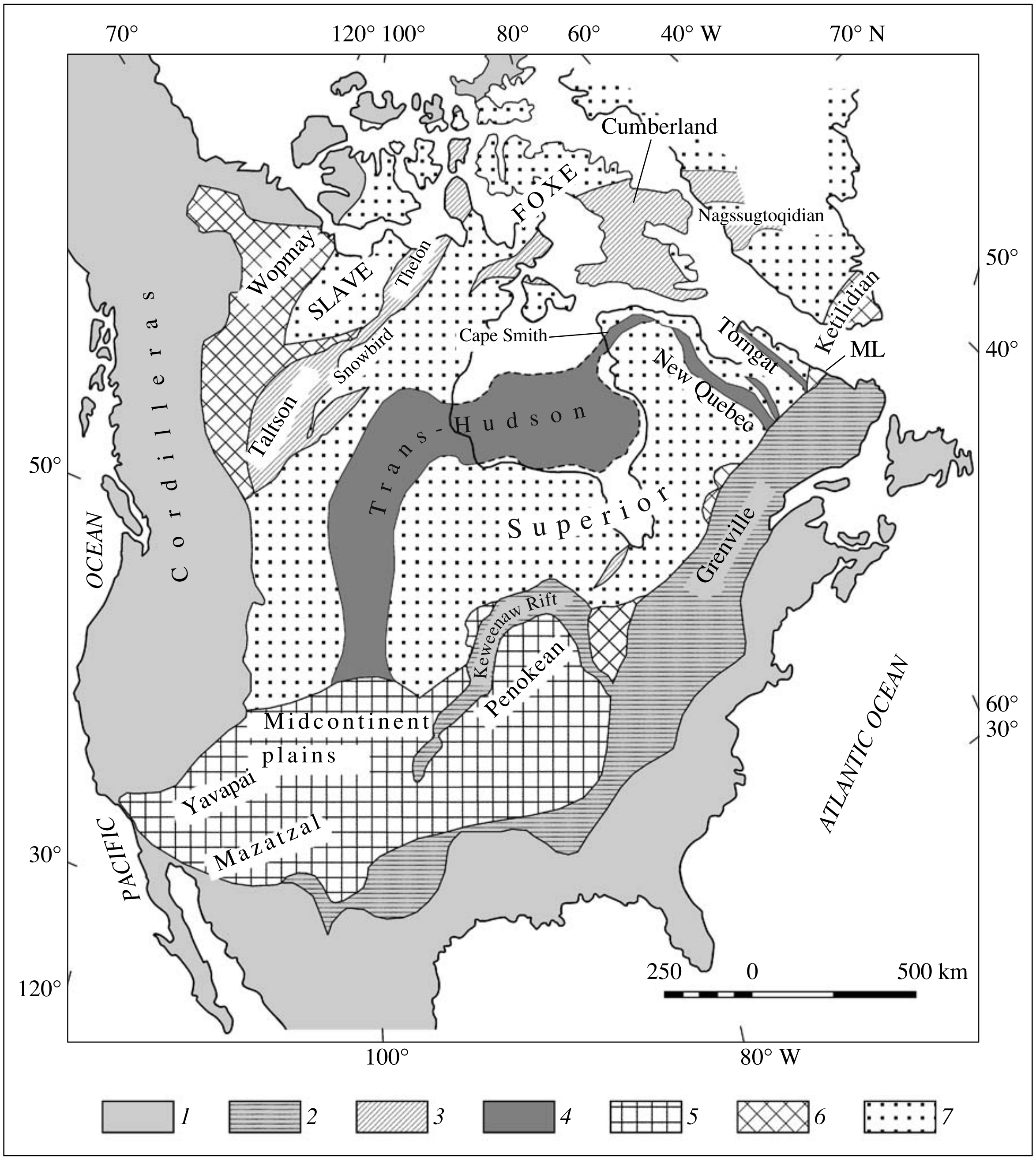


Fig. 3. Paleoproterozoic orogens and tectonic belts of the North American Craton, modified after [89]. (*1*) Phanerozoic; (*2*) Neoproterozoic; (*3–6*) Paleoproterozoic: (*3*) granulite–gneiss belt, 2.0–1.7 Ga, composed largely of juvenile Paleoproterozoic protoliths; (*4*) sedimentary–volcanic belts affected by low-grade metamorphism, 2.0–1.8 Ga; (*5, 6*) accretionary orogens: (*5*) 1.9–1.6 Ga and (*6*) 1.9–1.8 Ga; (*7*) Archean granite–greenstone and granite–gneiss domains. ML is the Makkovik–Labrador Orogen.

The island-arc volcanism started in the Belarus–Baltic domain somewhat earlier than in the Svecofennian Orogen proper. The volcanic and intrusive complexes were formed here from 2.1 to 1.8 Ga, becoming younger westward. Granulite-facies metamorphism at a temperature up to 900°C and a pressure of 8–10 kbar [130, 140] was related to two events: 1.82–1.80 and 1.79–1.78 Ga. Granulite-facies metamorphism likewise reveals a tendency to rejuvenate in the western direction. The youngest metamorphic event (1.63–1.61 Ga) likely accompanied anorogenic magmatism. Terrigenous protoliths of granulite associations contain detrital zircon dated at 2.45–1.98 Ga; an insignificant admixture of Archean populations is noted [63]. It is noteworthy that some of the detrital zircons were derived from an older Paleoproterozoic source unknown within this region. It is suggested that the high-grade metamorphism was related to the backarc and/or postcollision extension. Moreover, the record of thermal events includes indications of heating 1.55–1.45 Ga ago in connection with a pulse of anorthosite–rapakivi granite magmatism [56]. The signs of attachment of the juvenile Belarus–Baltic Terrane to the main body of the Svecofennian Orogen reveal an age somewhat earlier than 1.96 Ga [63].

The characteristic feature of the deep structure of the northern Lapland–Belomorian branch of the orogen is its localization above the counter-plunging tectonic sheets of the Pechenga–Varzuga and Circum-Karelian belts [17]. The crust of this orogen was extruded upward, a phenomenon that was probably one of the main causes of the exhumation of deep-seated granulite complexes. Underthrusting of the Karelian Craton beneath the Lapland–Baltic Orogen was established in the southern sector of the orogen as well [23]. However, the opposite margin of the orogen, in contrast, has been underthrust in the southern direction beneath Sarmatia and Volgo-Uralia [22].

The Taltson–Thelon Orogen that extends in the near-meridional direction across the North American Craton (Fig. 3) is close in many respects to the Lapland Belt. The Taltson Magmatic Zone is largely composed of I- and S-type granitoids (1.99–1.96 and 1.95–1.93 Ga, respectively) in association with metasedimentary granulites. (The *PT* parameters of metamorphism are 6–8 kbar and >900°C.) It was suggested that the origin of the Taltson Magmatic Zone was initiated by subduction of the oceanic crust beneath the margin of the Archean Churchill Province (continent) 1.99–1.96 Ga ago and completed with the collision between this continent and the Buffalo Head Terrane, whose crust was 2.4–2.0 Ga in age [89, 103, 124]. However, the subsequent geochemical study has shown that granitoids of both types are products of intracrustal melting [70]. These data, along with high-grade metamorphism, indicate that the rocks of the Taltson Magmatic Zone were formed under intraplate conditions [61, 74] and probably with participation of plume-type heat sources. The Thelon Orogen was regarded by Hoffman [89] as a result of oblique right-lateral collision between the Archean cartons of the Slave Province (foreland) and Rae Province (hinterland). In the Thelon Orogen, granodioritic plutons are hosted in the rocks that underwent high-temperature metamorphism [89, 144]. The northern extension of the Thelon Orogen (the Boothia Peninsula–Somerset Island and Devon–Ellesmere Island terranes) comprises complexes of orthogneisses, graphite-bearing paragneisses, and marbles with lenses of mafic and ultramafic granulites similar in metamorphic grade. The parameters of metamorphism dated at 1.9 Ga are estimated at 740–900°C and 6–8 kbar [99].

The origin of the branching Paleoproterozoic belts (Nagssugtoqidian, Rinkian, Foxe, and Torngat [89]) and orthogneiss complexes in structurally interrelated tectonic units (the Lake Harbour Group, the Narsajuaq Arc, and the Ramsay River [137]) may be interpreted in the same way. Granulite gneisses that metamorphosed at a temperature reaching 950°C and at a pressure of ~4 to ~12 kbar dominate in these belts. The granulite's protoliths in the lower parts of the sections mainly consist of metasedimentary rocks of platform and rift types with the participation of evaporites, mafic and ultramafic volcanic rocks, anorthosite sills and other bodies. The age of detrital zircons indicates that juvenile Paleoproterozoic rocks of unknown origin dated at 2.4–1.93 Ga were the sources of the sediments. A significant contribution of older, Archean rocks is noted as well. Sedimentation started about 2.0 Ga ago and ended 1.95–1.93 Ga ago. The oldest manifestation of granulite-facies metamorphism is recorded in charnockite of the Sisimiut Complex (1.92–1.90 Ga), i.e., at the end of sedimentation or immediately after its cessation. The Cumberland charnockite batholith was emplaced into the central part of the system 1.87–1.85 Ga ago. The age of the main phase of the granulite metamorphism is estimated at 1.85–1.80 Ga or somewhat older. The age of thrusting and exhumation of deep-seated metamorphic complexes varies from 1.85 to 1.74 Ga [92–94, 112, 128, 141, 148].

It is suggested that, as in the northern branch of the Lapland–Peribaltic Orogen, the crust of the Taltson–Thelon Orogen was located above a subducted slab [47, 124].

The Trans-Hudson Orogen reaches 500 km in width and embraces a considerable portion of the Circum-Superior Belt. The oceanic opening is recorded in the fragmented Purtuniq ophiolite section 2.0 Ga in age (the Cape Smith Belt and the northern segment of the Circum-Superior Belt) [129]. The subsequent extension took place ~1.96 Ga ago (alkaline volcanic rocks of the Povungnituk Group and the Cape Smith Belt) [115]. The paleomagnetic studies testify to significant displacements of the boundaries of the Trans-Hudson Orogen [77, 83, 139]. The subduction-related rocks are more widespread in the Trans-Hudson Orogen than in the Lapland–Peribaltic Orogen. The intense island-arc magmatism of ~1.92 to 1.88 Ga gave way to the accretion of fragments of the oceanic lithosphere (1.87 Ga) and vigorous plutonism that included the emplacement of the Wathaman–Chipewyan calc-alkaline batholith (1.86–1.83 Ga) [89, 136]. In contrast to the Lapland–Peribaltic and Taltson–Thelon orogens, the crust of the boundary structures of the Trans-Hudson Orogen plunges beneath the framing structural units in both the east and the west. Fragments of pre-Paleoproterozoic crust overlain by the Paleoproterozoic subduction-related complexes are probably retained in the axial zone of the Trans-Hudson Orogen [47, 152]. In other words, like in the northern branch of the Lapland–Peribaltic Orogen, the maximum extension of the continental crust passing into the spreading of the oceanic bottom was confined to the marginal zones.

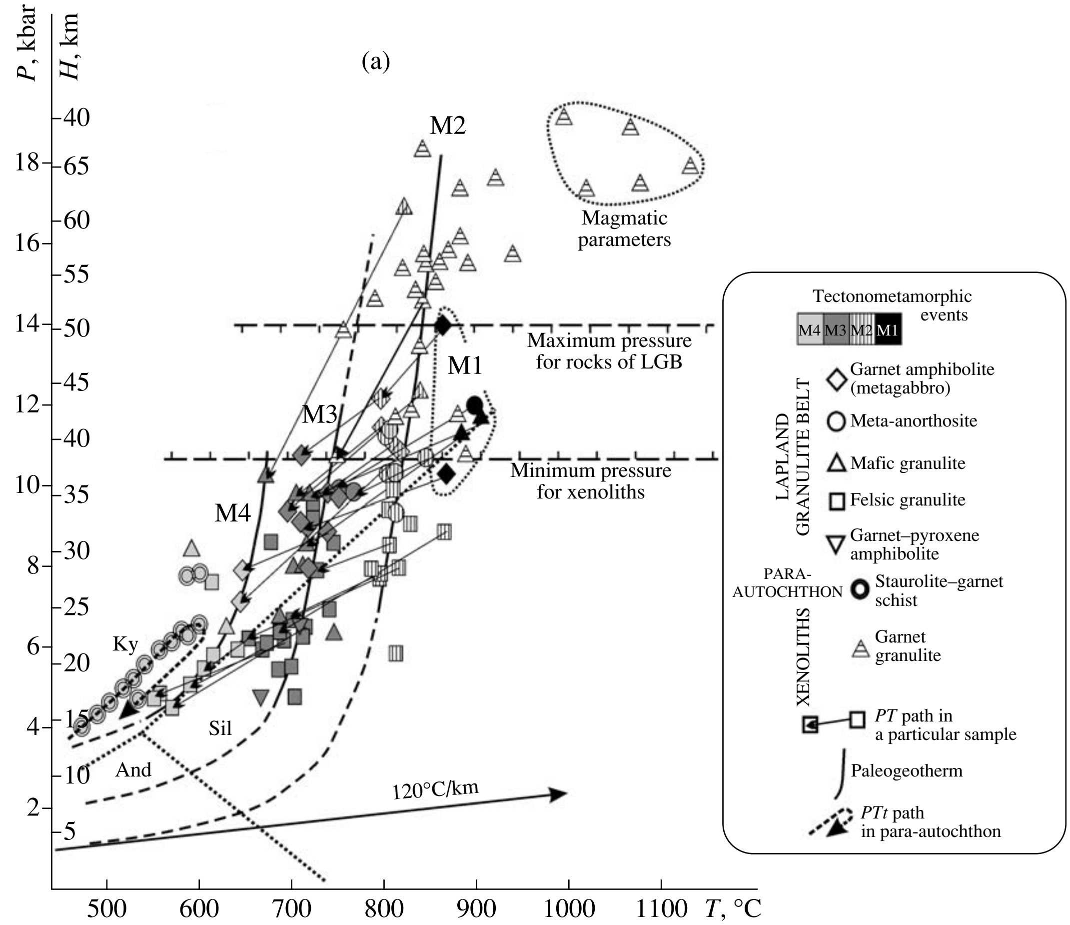


Fig. 4. Metamorphic evolution of the Lapland Granulite Belt: *PT* parameters estimated for (a) various rocks and (b) mineral assemblages.

Continental Margins (Accretionary Orogens)

The Svecofennian Orogen. As was mentioned above, the western slope of the early Paleoproterozoic Kola–Karelian continent was a passive margin complicated by rifting. Afterward, the rock complexes and tectonic units of the Svecofennian Accretionary Orogen were formed during a short time interval. This orogen is composed of subduction-related mafic, intermediate, and felsic volcanics (1.93–1.86 Ga); terrigenous and volcaniclastic sediments; and large granitoid plutons. The locally occurring high-Ti intraplate tholeiites are regarded as an indication of rifting in mature island arcs [76, 101, 114 and references therein]. The high-grade metamorphism (up to 800°C at 4–5 kbar) of turbidites that were deposited in backarc basins is dated at 1.89–1.81 Ga [90, 101]. The emplacement of early and late orogenic granitoids intensified 1.90–1.87 and 1.83–1.77 Ga ago and was followed by small postorogenic granitic intrusions [76]. Gabbroanorthosite and rapakivi granite plutons dated at 1.70–1.54 Ga may be regarded as manifestations of anorogenic magmatism.

The Penokean, Makkovik–Ketilidian, Labradorian, and Yavapai–Mazatzal orogens. The Penokean Accretionary Orogen is localized along the southern margin of the Neoarchean Superior Craton (Fig. 3) and composed of island-arc terranes and fragments of backarc basins that consist of volcanic and sedimentary rocks and coeval gabbroic and granitic plutons that were emplaced 1.89–1.84 Ga ago. The Niagara and Eau Plaine suture zones that connect this orogen with the Archean continent arose 1.86 and 1.835 Ga ago. In turn, the structure of the Makkovik–Ketilidian and Labradorian orogens that extend along the eastern margin of the Archean composite continent fits the Andean model. In the lower part of the section, the terrigenous rocks and basaltic flows are cut through by metadolerite dikes 2.13 Ga in age and granodiorite dated at 1.91 Ga. The felsic and intermediate tuffs and lava flows that erupted 1.86–1.81 Ga ago dominate in the upper part of the section. The collision accompanied by emplacement of granodiorite (1.81–1.80 Ga) and tonalite ended ~1.79–1.76 Ga ago. The depressions that arose owing to the gravitational collapse of the thickened crust were filled with volcanic–sedimentary sequences. Anorogenic bimodal magmatism, including rapakivi granite (1.76 Ga) and gabbro–anorthosite–monzonite complexes,

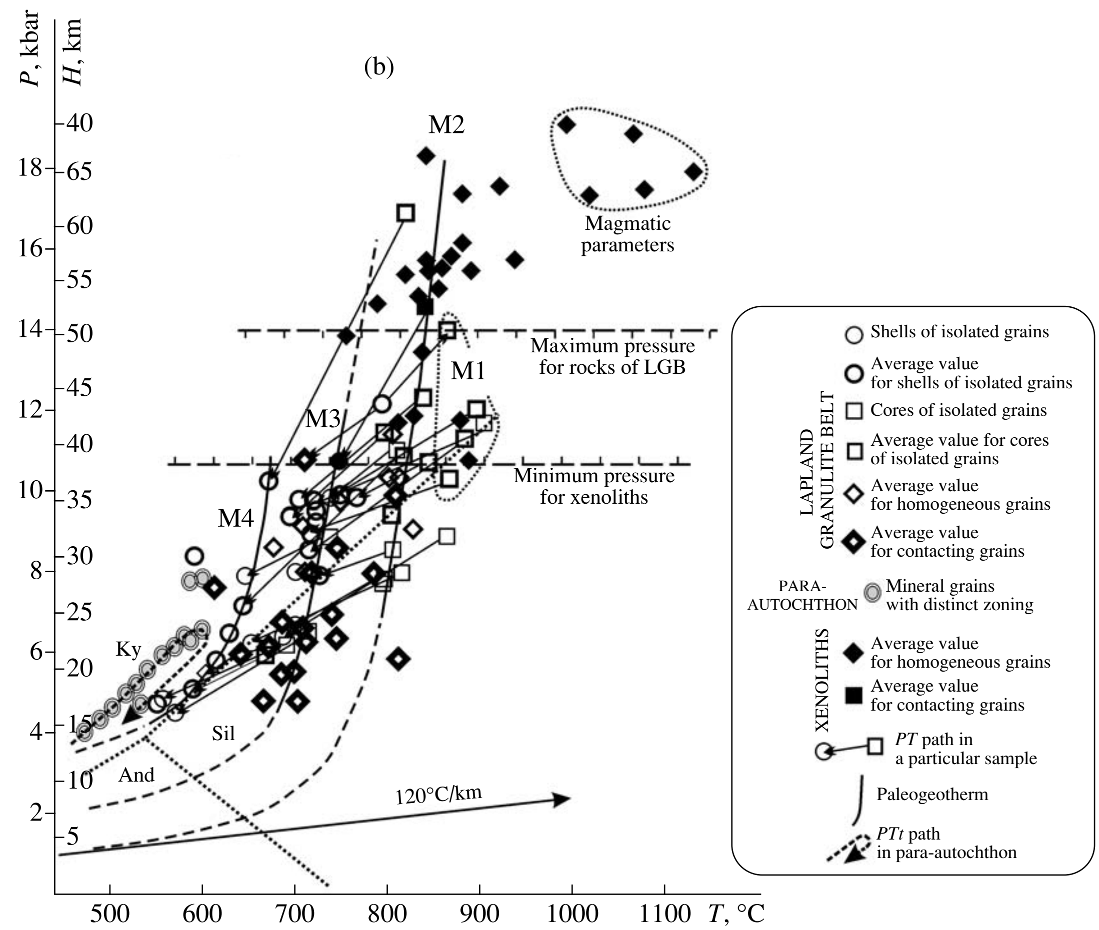


Fig. 4. (Contd.)

was accompanied by metamorphism of country rocks under conditions of granulite facies 1.71–1.63 Ga ago [89].

The Paleoproterozoic juvenile crust in the southern midcontinent and in the southwest of the United States (the Yavapai–Mazatzal–Midcontinent Orogen) was formed during two stages. At the first stage, dated at 1.8 Ga, the calc-alkaline volcanic–plutonic complexes that formed in island arcs and backarc basins were attached to the margin of the Archean continent 1.79–1.71 Ga ago. The occurrences of subaerial felsic volcanism are younger; posttectonic granites were emplaced 1.64–1.62 Ga ago. The Precambrian evolution ended with Meso- and Neoproterozoic anorogenic magmatism [89].

The Wopmay Orogen that extends along the western margin of the Archean Craton of the Slave Province is made up of poorly studied complexes dated at 2.4–2.0 Ga. The Hottah and Buffalo Head terranes composed of these complexes were accreted to the margin of the Archean continent 2.4–2.3 and 1.9–1.7 Ga ago, respectively [62, 81, 124]. The sedimentary wedge of the passive margin localized along the western boundary of the Slave Craton and affected by tectonic compression participates in the structure of the accretionary orogen. The age of the initial shelf sedimentation is ~1.97 Ga. The rift-related bimodal volcanism at the outer edge of the shelf is dated at 1.90–1.88 Ga. The island-arc terranes that formed to the west of the passive margin (in present-day coordinates) were accreted to the craton margin 1.95–1.91 and 1.88–1.86 Ga ago. In the present-day structure, they are underlain at least partly by crust inaccessible to observation. According to indirect evidence, this crust is 2.4–2.0 Ga in age. (A narrower interval of 2.3–2.1 Ga is acceptable). Postorogenic syenogranite was emplaced soon after the completion of accretion (1.86–1.84 Ga). Diorite, gabbro, anorthosite, and syenite intrusions were formed 1.71 Ga ago synchronously with rifting at the northwestern boundary of the Wopmay Orogen [145]. The stabilization of the craton was related to the completion of the deformation that developed from 1.84 to 1.66 Ga ago [89, 125].

Geological Evolution of the Siberian Craton

Siberia markedly differs from Lauroscandia by the nearly complete absence of the Paleoproterozoic low-grade volcanic–sedimentary belts in the inner domain of the Siberian Craton (Fig. 5). The Akitkan Belt, com-

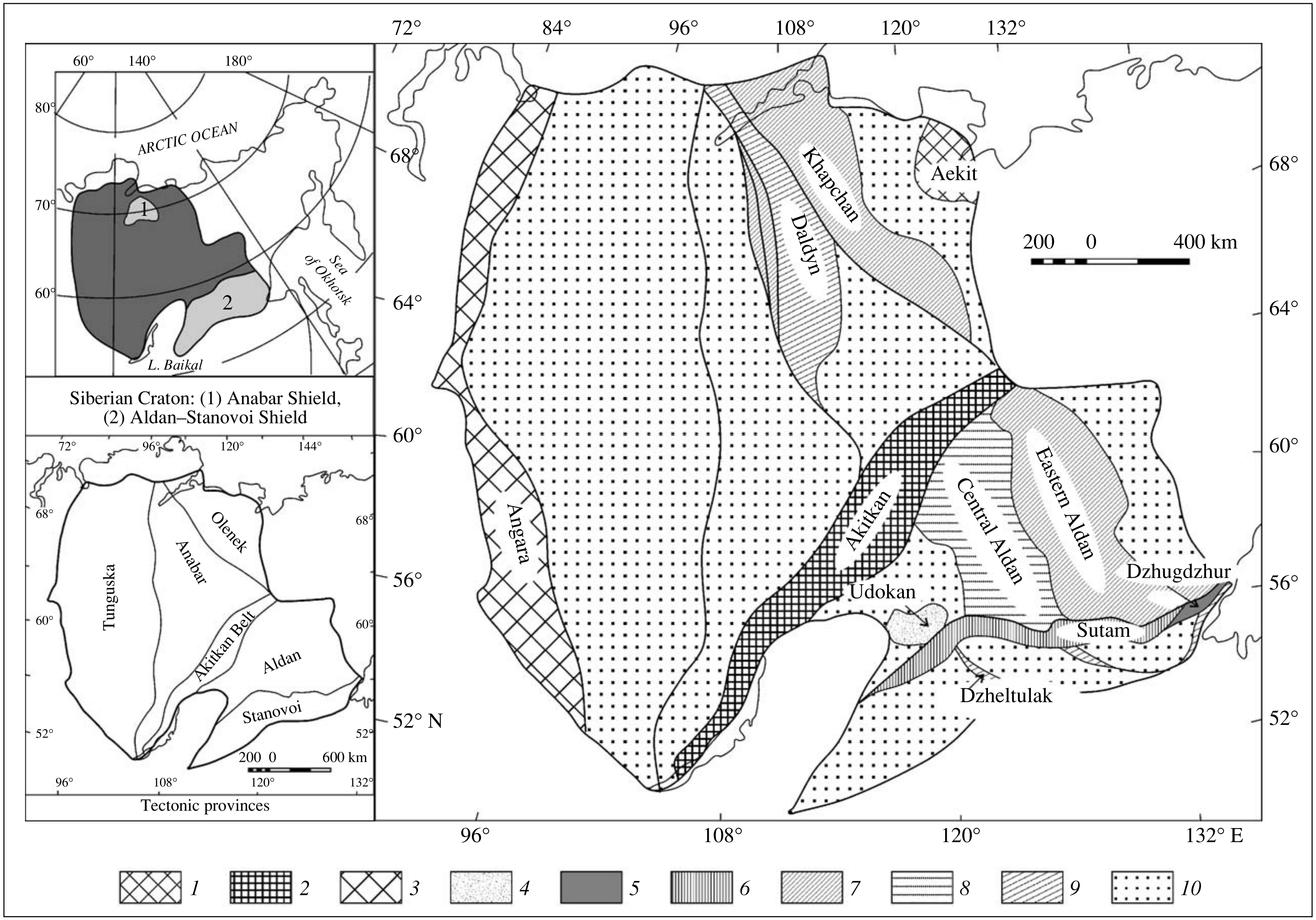


Fig. 5. Paleoproterozoic tectonic belts of the Siberian Craton, modified after [31, 123]. (*1*) Inferred accretionary orogens, 1.91–1.71 Ga; (*2*) intraplate volcanic–plutonic belt, 2.03–1.82 Ga; (*3*) inferred magmatic arcs of active margins, 2.2–1.78 Ga; (*4*) sedimentary basins, 2.2–1.83 Ga; (*5–9*) granulite–gneiss complexes with different ages of granulite metamorphism: (*5*) gabbroanorthosite, 1.74–1.70 Ga; (*6*) Archean rocks reworked in the Paleoproterozoic, 2.2–1.70 Ga; (*7*) Paleoproterozoic juvenile complexes, 2.04–1.76 Ga; (*8*) Archean rocks reworked in the Paleoproterozoic and Paleoproterozoic juvenile complexes, unspecified, 2.04–1.76 Ga; (*9*) Archean rocks reworked in the Paleoproterozoic, 2.02–1.76 Ga; (*10*) Archean granite–greenstone and granite-gneiss domains.

posed of rock complexes dated at 2.03–1.82 Ga, is the only exception [9, 28, 123]. At the same time, the Paleoproterozoic granulite belts of the Siberian Craton are no less important than their counterparts of Lauroscandia. In the opinion of O.M. Rosen and his colleagues, the origin of these belts is related to the amalgamation of Archean microcontinents and the collisional stacking of the crust. The frontal areas of granulite belts, including the extended blastomylonitic zones, are interpreted as sutures [30, 123] (Fig. 5). This concept is broadly consistent with the model of the Paleoproterozoic evolution of the North American Craton developed by Hoffman [89]. According to [30, 123], the Anabar and Olenek microcontinents (provinces) are made up of both Archean and Paleoproterozoic rocks. The ages of the magmatic processes responsible for the formation of the Paleoproterozoic complexes in the Daldyn and Khapchan granulite belts are 2.55 Ga (gabbroanorthosite) and 2.42 Ga (volcanic rocks transformed into mafic granulites). Protoliths of metasedimentary granulites in the Khapchan Belt (graywackes and carbonate rocks) were deposited on the shelf of the passive continental margin between 2.44 and 2.08–1.97 Ga. According to the Sm–Nd data, the Paleoproterozoic mature continental crust unknown at the present-day ground surface was a provenance [154]. High-grade metamorphic events recorded in both Archean and juvenile Paleoproterozoic rocks are dated at 2.18, 1.97, 1.94–1.90, and 1.80–1.76 Ga. The subalkali granitoids that completed the regional evolution are dated at 1.84–1.80 Ga [31, 123].

The onset of the Paleoproterozoic evolution of the Aldan Province is marked by emplacement of intraplate A-type granitoids into the Archean crust 2.49–2.40 Ga ago [30, 33, 123 and references therein]. Other events that could be referred to as early Paleoproterozoic are not dated reliably. However, the model Nd age of Paleoproterozoic metasedimentary granulites and igneous rocks falls within an interval of 2.5 to 2.0 Ga [15, 16, 31, 32]. The younger Udokan Group 9–12 km thick was deposited 2.18–1.95 Ga ago in an intracontinental basin or at a passive margin. The sequence consists of Cu-bearing quartz arenites with interlayers of black shale, marine carbonate, and molassoid conglomerate. The Katugin alkali granite pluton (2.01 Ga) is chronologically and spatially related to the deposition of the Udokan Group.

Granulite complexes are localized in the central and eastern portions of the Aldan Shield. Granulite-facies metamorphism (800–970°C and 7.0–10.7 kbar) affected both Paleoproterozoic and Archean rocks. The results of Sm–Nd measurements show that both Archean and juvenile Paleoproterozoic rocks served as sources of calc-alkaline and subalkali volcanic rocks and sediments that underwent granulite metamorphism 2.10–1.92 Ga ago soon after the cessation of sedimentation and volcanic activity. Granulite metamorphism was accompanied by emplacement of tonalite–trondhjemite magmas (2.04–2.01 Ga) and granitoids of variable composition (1.99–1.90 Ga). The events listed above were followed by thrusting driven by tectonic compression in the latitudinal direction (in present-day coordinates). The coeval mafic and ultramafic dikes (1.92 Ga), A-granite (1.9–1.8 Ga), and late granulite metamorphism (1.87–1.77 Ga) are known. The Paleoproterozoic evolution ended with intraplate magmatism. The rapakivi-like granitic batholiths, K-rich rhyolitic pyroclastic flows, and alkali granite plutons were formed 1.74–1.70 Ga ago at the eastern flank of the Aldan Province [15, 16, 30–33, 123].

In the Stanovoi Range Province (Fig. 5), the Paleoproterozoic juvenile rocks remain almost unknown. The Dzhugdzhur gabbroanorthosite pluton that intruded 1.74–1.70 Ga ago is the only exception [25, 37]. The fragmented linear belts and blocks of Archean rocks affected by Paleoproterozoic granulite metamorphism 2.2–2.0 and 1.98–1.94 Ga ago have been mapped in Archean granite–greenstone complexes [33 and references therein].

The lateral shortening of the crust in the near-latitudinal direction (in present-day coordinates) predated the second metamorphic event. The Paleoproterozoic evolution ended no later than 1.7 Ga ago after the northward overthrusting of the Stanovoi Range Province and its juxtaposition with the Aldan Province. The peak parameters of metamorphism (up to 1000–1100°C and 9.5–10.0 kbar) [12] that correspond to the deep crustal levels have been noted in rocks of the Sutam Complex localized along the boundary with the Aldan Province. The metamorphism is referred to the interval 1.98–1.84 Ga; however, the last metamorphic event likely was coeval with the emplacement of the Dzhugdzhur anorthosite pluton and thus occurred directly before the junction of the Aldan and Stanovoi Range Province.

The Sm–Nd data testify that all or at least the majority of the Paleoproterozoic juvenile igneous rocks studied in the Siberian Craton, including gabbroanorthosite, crystallized from the melts appreciably contaminated with the Archean crustal material [15, 16, 31, 32].

Geological Evolution of Other Continental Domains

The review of rock associations from Lauroscandia and the Siberian Craton presented above illustrates the main evolutional trends and the most important Paleoproterozoic events. The tectonic belts of both recognized types that comprise the rocks of low–moderate and high grades of metamorphism and that formed 1.95–1.90 Ga ago or somewhat later are known in Asia, Australia, Africa, and South America. However, their role in the structure of the above continents is much more modest in comparison with Lauroscandia and Siberia [44, 73, 102, 117, 142, 151]. The rocks that underwent high-grade metamorphism in the early Paleoproterozoic 2.56–2.42 Ga ago are known from cratons of southern Australia, China, India, and Antarctica [46, 49, 87, 120, 121]. At the passive margins of the Pilbara Craton in Australia and the Kaapvaal Craton in South Africa, the period from 2.51 to 2.45 Ga was characterized by a peak of metamorphism and the formation of banded iron formations under extension related to the impact of a global plume [49]. The period from 2.44 to ~2.0 Ga was the time of the quiescent within-plate development not only in Lauroscandia but also in the African Craton [72] and, probably, over the entire planet [3, 66].

GRANULITE (GRANULITE–GNEISS) BELTS: GEODYNAMIC INTERPRETATION

As follows from the review presented above, the granulite belts from various provinces, being characterized by high-grade metamorphism, have a number of other specific properties that are not inherent to sedimentary–volcanic and volcanic–plutonic belts composed of low-grade metamorphic rocks.

(1) Episodes of high-grade metamorphism are commonly predated or accompanied by the emplacement of gabbroanorthosite magma appreciably contaminated with materials of the continental crust, intraplate "dry" granite intrusions, enderbite, and charnockite.

(2) The lower portions of tectonostratigraphic sections are often composed of rift-related metavolcanic rocks and meta-arenites that are products of erosion of the underlying Archean crust. In contrast, the Paleoproterozoic juvenile rocks of unclear origin were provenances of most metasedimentary granulites (khondalites or metagraywackes) in the upper portions of the sections. The interpretations of these provenances remain a matter of debate. The age of detrital zircon grains and the Sm–Nd data inevitably indicate the existence of large bodies of juvenile felsic igneous rocks markedly contaminated with materials of the Archean continental crust. These rocks were formed in the early and middle Paleoproterozoic and served as sources of the inferred sedimentary protoliths. As can be seen from the above review, in virtually all studied cases, such source rocks are unknown not only close to granulite belts but also in distant territories.

(3) The exhumed granulite complexes make up nappe and thrust assemblies. The metamorphism of the underlying para-autochthonous complexes reveals inverse metamorphic zoning, which was formed as a result of heating from above, i.e., from the side of relatively hot tectonic sheets.

(4) The total thickness of the continental crust that was affected by granulite-facies metamorphism during one metamorphic event could have reached 40 or more kilometers.

(5) Volcanic and sedimentary rocks underwent granulite metamorphism virtually immediately after sedimentation or in a short time span after its cessation.

(6) In most cases, the nappe and thrust assemblies, i.e., granulite belts proper, rapidly followed metamorphism or were formed during that time when the crust remained hot. However, this statement is not universal. The collisional thrusting may have developed much later in connection with a different tectonic setting. Granulite complexes are left deep in the crust for a long time, and in many cases, forever, as indicated by deep xenoliths [126]. The Kapuskasing Belt in the Superior Craton is an example of such a situation. Here, the granulite metamorphism is dated at 2.66–2.63 Ga, whereas the thrusting related to collision occurred much later, between 2.04 and 1.89 Ga [113].

The specific features listed above show that the intense heating that initially affected the lower crustal rocks spread over the volcanic and sedimentary fill of rapidly sagging basins. The completion of sedimentation, high-grade metamorphism, and thrusting were developed jointly during a short time interval of tens of million years but no longer than 50 Ma.

The tectonic and geodynamic implications of granulite belts have remained controversial until now. In most cases, the granulite belts are regarded as analogues of Phanerozoic sutures differing from the latter by a higher grade of metamorphism or as collisional orogens. The modeling of the thermal evolution of collisional orogens demonstrates a considerable heating of the thickened crust up to the granulite-facies conditions [71, 143]. However, the specific features of granulite belts summarized above come into conflict with such a geodynamic model. Harley [86] pointed out that increase in thickness of the crust caused by collisional stacking is not able to provide the extremely high temperature that is typical of the metamorphism of most granulite belts, because of limitations imposed by heat conductivity, heat generation, and heat flow in the crust. According to the Harley's estimates, regional granulite metamorphism requires much greater influxes of heat than could have been provided by an increased thickness of the crust. A relatively slow cooling documented by the study of many granulite complexes has compelled researchers to suggest that an additional heat source existed beyond the crust that experienced metamorphism. In most cases, the geological data show that the interrelated magmatism and high-grade metamorphism most likely are consequences of thermal events of the lithospheric rank [86]. The arguments put forward by Harley are especially convincing with allowance for a great thickness of the crust affected by high-grade metamorphism during a single metamorphic event [20, 108, 109 and references therein].

A low activity of water is one more factor crucial for the interpretation of the geodynamic setting of granulite metamorphism. The "dry" mineral assemblages are characteristic of metasedimentary and metavolcanic sequences as well as of derivatives of enderbite–charnockite magmas [146]. The high water activity typical of suprasubduction processes provides the partial melting of rocks at a relatively low temperature and the general stabilization of the temperature field in the crust at a level of amphibolite-facies conditions. Dewatering of considerable crustal bodies in suprasubduction zones is thought to be unrealistic. However, the "dry" conditions are quite possible under conditions of backarc spreading. While analyzing the probability of granulite-facies metamorphism in the crust under conditions of continental rifting, Harley [86] and Sandiford [127] drew a conclusion that, at present, granulites can be formed in the crust that underlies the Basin and Range Province in North America. Another example is an association of Early Cretaceous basic and intermediate granulites in the Fiordland Province of New Zealand [78, 79] and the Paleozoic Black Giants gabbroanorthosite pluton that formed under conditions of backarc extension [57]. The aforementioned prevalence of rift-related protoliths in the lower portions of granulitic tectonostratigraphic sections is consistent with suggestions on the rift or backarc setting of the currently and recently formed granulite complexes.

In summary, it should be stated that granulite-facies metamorphism most likely is induced by vigorous influxes of mantle heat. The formation of granulite complexes is not connected directly with collisional thrusting and thickening of the crust. At the same time, the granulite metamorphism was interrupted in many cases by the collision that provided the fast upward transportation of hot crustal sheets.

The recently obtained data on the age of detrital zircons from metasedimentary granulites laid the foundation for a new genetic model of granulite complexes. Two variants of the explanation of the absence of the Paleoproterozoic granitoid rocks, which could have been a source of these zircons in the crustal domains adjacent to granulite belts, may be offered:

(1) The source rocks were localized close to sedimentation basins undergoing rapid subsidence accompanied by virtually complete destruction of the basin slopes and a supply of detrital material into the area of sedimentation.

(2) The sedimentary protholiths could have at least partly included the deposits of pyroclastic ash flows that filled vast calderas and volcanotectonic depressions.

The first scenario is suitable to a greater extent for backarc basins. However, it is known that many Paleoproterozoic granulite belts occur far from subduction-related igneous rocks. Furthermore, in some regions, the volcanic–sedimentary and volcanic–plutonic belts of respective age are either not abundant or completely absent (Siberia, southern India). The second scenario seems more attractive because many attributes of the crustal structure, geodynamic settings, and lithology of granulite protoliths are fairly similar to those of pyroclastic flows. In particular, both are related to backarc extension and intracontinental rifting [41]. The great bodies of pyroclastic material fill calderas and depressions over a short time [133] and under "dry" and high-temperature (up to 940°C [138]) conditions. In addition, it should be noted that pyroclastic flows carry up deep xenoliths that underwent granulite metamorphism [132] and ortho- and clinopyroxene, olivine, and garnet crystals commonly occur in deposits of felsic pyroclastic flows and in accompanying intrusions [50, 138].

Thus, there is much evidence in favor of the formation of granulite belts as the following sequence of events: intense heating of the thick crust by mantle thermal sources (plumes) $\longrightarrow$ origin of rifts and volcanotectonic depressions $\longrightarrow$ filling of these basins with rift-related sediments and juvenile lavas and pyroclastic flows contaminated with crustal material $\longrightarrow$ high-grade metamorphism of rocks in the lower and middle crust, including the fill of basins and depressions in the inner continental domain or in the backarc extension setting $\longrightarrow$ delamination of the crust and thrusting under conditions of general compression (collision) that gives rise to exhumation of the rocks affected by granulite metamorphism. These events resulted in the formation of regional intraplate granulite belts and the incorporation of granulite complexes into the structure of accretionary and collisional orogens.

DISCUSSION

The suggested conditions and settings of granulite metamorphism provide for the intraplate origin of large (regional) granulite belts related to the plume activity. It is reasonable to suppose that the initial stages (extension, intracontinental magmatism and sedimentation) could have been the same for granulite belts and volcanic–sedimentary basins distinguished by a low or moderate grade of metamorphism. Further, the transition from continental rifting to oceanic spreading of the Red Sea type created favorable conditions for the rapid dissipation of deep heat and the termination of high-grade metamorphism in the adjacent continental crust. The causes of the retained integrity of the crust affected by extension in one case and breakup in another case remain unknown; however, it is evident that these causes were crucial for the evolution of large crustal domains. If the proposed genetic model of granulite belts is valid, these belts should not be regarded as analogues of Phanerozoic sutures.

The onset of the Paleoproterozoic evolution of Lauroscandia was related to the activity of a superplume 2.51–2.44 Ga ago that was responsible for the division of the North American Craton and Fennoscandia (Fig. 6). Further evolution gave rise to the origination of the present-day system of arcuate tectonic belts in the North American and East European cratons.

The consideration of the above review widens our knowledge on the nature and mutual relations of intracontinental collisional orogens of those two types, which are opposed to each other in this paper. Orogens of the first, Trans-Hudsonian type composed of low-grade metasedimentary–metavolcanic complexes were formed in the framework of the full Wilson cycle, including large-scale opening of ocean and its subsequent closure involving subduction beneath the adja-

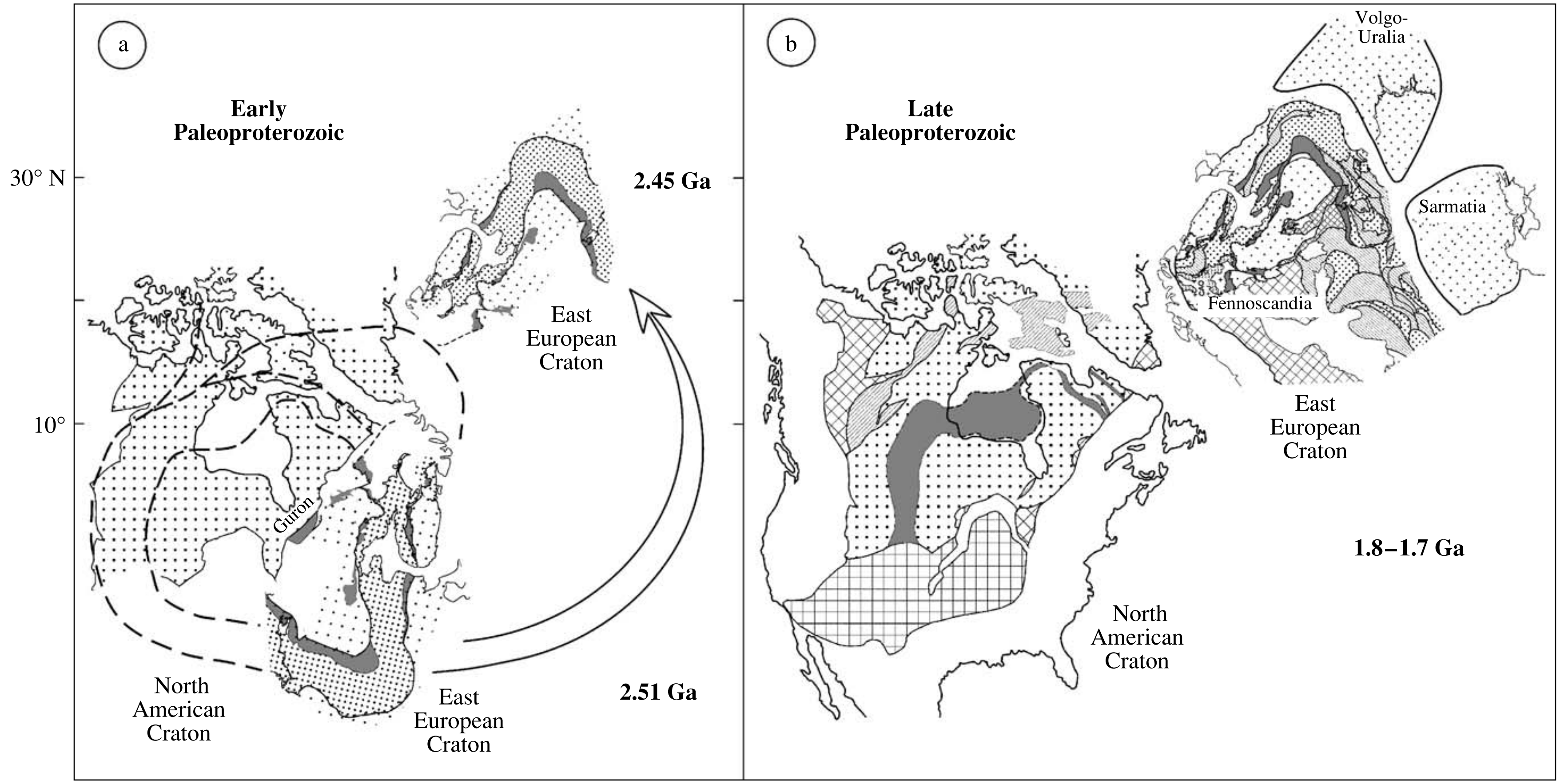


Fig. 6. Reconstruction of the main paths of the Paleoproterozoic evolution of Lauroscandia: (a) early Paleoproterozoic and (b) late Paleoproterozoic. The relative position of the North American and East European Craton at 2.45 Ga ago is shown after [105]. An inferred location of the Huronian Group relative to the reconstructed paleocontinent is indicated. See Figs. 2 and 3 for legend.

cent continents. Orogens of the second, Lapland–Peribaltic type included high-grade metamorphic complexes and were formed under conditions of intense plume-related heating of the crust along with short-term and spatially restricted spreading in marginal zones. The formation of orogens belonging to both types ended with collision and, subsequently, plunging of the adjacent continental crustal blocks (at least on one side) beneath the orogen and extrusion of the axial zone of the orogen. In both types of orogens, the maximum extension (up to spreading of the oceanic bottom) was confined to the marginal portions of the growing orogens. The active stage of formation of the orogens pertaining to both types lasted for 50–70 Ma or less (in any case, less than 100 Ma).

The post- and anorogenic events (bimodal magmatism, formation of gabbroanorthosite–rapakivi granite complexes, and granulite metamorphism) occurred in different portions of the renascent Paleoproterozoic supercontinent at a different time soon after the complete formation of accretionary and collisional orogens, largely later than 1.7 Ga ago. In a certain sense, these events may be regarded as the beginning of a new, Meso- and Neoproterozoic evolutional stage.

The Paleoproterozoic tectonics and crust formation were controlled to a great degree by global plume-related events (superplumes) dated at 2.51–2.44 and 2.0–1.95 Ga. As was stated above, the magmatism, high-grade metamorphism, and sedimentation that occurred 2.51–2.44 Ga ago in continental domains and that is now localized in both hemispheres are indications of the extension and influx of mantle heat into the inner region of a rather vast continent. The synchronism of these processes in currently distant continental domains indicates that they once were related to a common continent or supercontinent. The emplacement of mantle-derived magmas, along with high-grade metamorphism, likely came about by plume-related processes. The widespread occurrence of such phenomena allows the inference of superplumes. The coeval manifestations of quiescent within-plate development (2.44–2.0 (2.11) Ga) in separate domains likewise indicates their relations to a common continent. While discussing the diminished magmatic activity within the interval 2.44 to 2.2 Ga, Condie [65] made a special note that the heat engine of our planet could not have stopped and that the heat generation in the Earth's interior could not have decreased markedly during this period. Moreover, because the ocean ridges are the main elements in the system of deep-heat abstraction, they should have functioned in a normal regime. In other words, there are no sufficient grounds to suggest that the plates ceased to move and afterward plate tectonics resumed again. The suggested intense recycling of the juvenile oceanic crust formed in the "oceanic hemisphere" and the virtually complete consumption of this crust by subduction might be a suitable explanation for the period of relative stability of the Paleoproterozoic supercontinent and the restriction of subduction-related magmatic processes along its boundaries.

Tectonic events initiated by plumes, in particular, rifting locally passing to spreading and formation of the short-lived oceans of the Red Sea type, which, as a rule, did not result in the eventual fragmentation of supercontinent, may be classified as failed attempts to break up the Archean supercontinent [106]. The main changes of the geological evolution of the Earth at the Archean–Proterozoic boundary may be interpreted as a change of the Archean tectonics of microplates by the Proterozoic tectonics of the supercontinent or the tectonics of microoceans with allowance for the restricted dimensions of Red Sea–type oceans that arose within the partly broken supercontinent. The origin of the Earth's first supercontinent, which embraced a considerable portion of the global surface by the end of Archean, obviously gave rise to the rearrangement of the system of convective cells in the underlying mantle. It is noteworthy that, in terms of the style of tectonic processes and geodynamic settings, the Paleoproterozoic differs from both the Archean and the Phanerozoic. Paradoxically enough, the Archean tectonics of numerous microplates resembles the Phanerozoic plate tectonics to a greater degree than the Paleoproterozoic tectonics of the supercontinent.

CONCLUSIONS

(1) The formation of granulite belts included the following sequence of events: intense heating of the thick crust by mantle thermal sources (plumes) ⟶ origin of rifts and volcanotectonic depressions ⟶ filling of these basins with rift-related sediments and juvenile lavas and pyroclastic flows contaminated with crustal material ⟶ high-grade metamorphism of rocks in the lower and middle crust, including the fill of basins and depressions in the inner continental domain or in the backarc domains ⟶ delamination of the crust and thrusting under conditions of general compression (collision) that gives rise to the exhumation of the rocks affected by granulite metamorphism and the origination of intraplate granulite–gneiss belts incorporated into the structure of accretionary and collisional orogens. The Early Precambrian granulite belts should not be regarded as analogues of Phanerozoic sutures.

(2) The main distinguishing feature of the Paleoproterozoic evolution of the supercontinent, or a few large continents, formed by the end of Archean was the concentration of tectonic events in its inner domain, which could have been partially broken up. The formation of accretionary orogens along the outer boundaries of the supercontinent covered a short time interval at the end of the Paleoproterozoic and was accompanied by recovery of integrity of the reorganized supercontinent that accommodated considerable bodies of the juvenile Paleoproterozoic crust.

(3) The Paleoproterozoic geological evolution of Lauroscandia may be subdivided into five periods: a

superplume and the division of the North American Craton and Fennoscandia (2.51–2.44 Ga); the quiescent development within an intracontinental domain locally complicated by plume and plate tectonics (2.44–2.0 (2.11) Ga); a superplume that initiated tectonic, magmatic, and metamorphic events (2.0–1.95 Ga); the interaction of the plume and plate tectonics, partial breakup of the continental crust, formation of accretionary orogens along the margins of the supercontinent, collision in its inner domain, and eventual recovery of the supercontinent's integrity (1.95–1.75 (1.71) Ga; and post- and anorogenic magmatism (<1.75 Ga).

(4) Large fragments of the Early Precambrian continental crust in the Siberian Craton and other cratons evolved following a similar scenario. However, chronological boundaries of particular stages were somewhat shifted and the intensity of their occurrence was markedly variable.

(5) Two types of intracontinental collisional orogens are recognized: (i) the Trans-Hudsonian-type orogens composed of low- and moderate-grade metamorphic rocks and formed in the framework of the complete Wilson cycle with the involvement of subduction beneath the adjacent continents and (ii) the Lapland–Peribaltic-type orogens made up of high-grade metamorphic rocks and related to plume activity along with short-term and spatially restricted occurrence of spreading in marginal zones. Both types of orogens started to evolve with extension and heating of the crust under the effect of plumes or superplumes. In the case of the complete breakup of the continental crust, the subsequent evolution resulted in the formation of the first type of orogens. When the continental crust underwent only a partial breakup, the orogens of the second type were formed. Their evolution ended in the collision setting with plunging of the adjacent continental crust on at least one side and extrusion of the axial zone of the orogen upward. In both types of orogens, the maximum extension up to the spreading of the oceanic bottom was confined to the marginal zones of the growing orogens.

(6) The active stage of the formation of Paleoproterozoic collisional and accretionary orogens lasted for <100 Ma.

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