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Paleogeodynamics

SECULAR CHANGES IN RELATIONSHIPS BETWEEN PLATE-TECTONIC AND MANTLE-PLUME ENGENDERED PROCESSES DURING PRECAMBRIAN TIME

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Abstract: Paradoxically, the lists of "proxies" of both plate- and plume-related settings are devoid of even a mention of the high-grade metamorphic rocks (granulite, amphibolite and high-temperature eclogite facies). However, the granulite-gneiss belts and areas which contain these rocks, have a regional distribution in both the Precambrian and the Phanerozoic records. The origin and evolution of the granulite-gneiss belts correspond 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 pyroclastic flow deposits; and metamorphism of lower and middle crustal complexes under conditions of granulite and high-temperature amphibolite facies that spreads over the fill of rift depressions also. Granulite-gneiss complexes of the East European Craton form one of the main components of the large oval intracontinental tectonic terranes of regional or continental rank. Inclusion of the granulite-gneiss complexes from Eastern Europe, North and South America, Africa, India, China and Australia in discussion of the problem indicated in the title to this paper, suggests consideration of a significant change in existing views on the relations between the plate- and plume-tectonic processes in geological history, as well as in supercontinent assembly and decay. The East European and North American cratons are fragments of the long-lived supercontinent Lauroscandia. After its appearance at ~2.8 Ga, the crust of this supercontinent evolved under the influence of the sequence of powerful mantle plumes (superplumes) up to ~0.85 Ga. During this time Lauroscandia was subjected to rifting, partial breakup and the following reconstruction of the continent. The processes of plate-tectonic type (rifting with the transition to spreading and closing of the short-lived ocean with subduction) within Lauroscandia were controlled by the superplumes. Revision of the nature of the granulite-gneiss complexes has led to a fundamental new understanding of: a more important role than envisaged previously for mantle-plume processes in the juvenile additions to the continental crust, especially during the Neoarchaean-Proterozoic; the existence of the supercontinent Lauroscandia from ~2.80 to 0.85 Ga; the leading role of mantle plumes in the interaction of plate- and plume-tectonics in the Neoarchaean-Proterozoic history of Lauroscandia and perhaps of the continental crust as a whole. We propose that the evolution of the geodynamic settings of the Earth's crust origin can be represented as a spiral sequence: the interaction of mantle-plume processes and embryonic microplate tectonics during the Palaeo-Mezoarchaean (~3.8–2.8 Ga) → plume-tectonics and local plume-driven plate-tectonics (~2.80–0.55 Ga) → Phanerozoic plate tectonics along with a reduced role of mantle plumes.

Key words: Precambrian; plate-tectonics; plume-tectonics; granulite-gneiss belts; supercontinent; intracontinent oval orogen

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Длиннопериодные изменения в соотношении процессов тектоно-плитного и мантийно-плюмового происхождения в докембрии

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Аннотация: Как ни парадоксально, в списках главных свидетельств обстановок как тектоники плит, так и тектоники плюмов отсутствуют любые упоминания о породах, подвергшихся метаморфизму при высоких параметрах, таких как гранулиты, амфиболиты, высокотемпературные эклогиты. Между тем гранулито-гнейсовые пояса и ареалы, которые включают такие породы, имеют региональное распространение как в докембрии, так и в фанерозое. Возникновение и эволюция гранулито-гнейсовых поясов связаны с мантийно-плюмовой активностью, которая выражается в интенсивном прогреве континентальной коры, внутриплитном магматизме, формировании рифтогенных депрессий, заполнявшихся осадками, ювенильными лавами и отложениями пирокластических потоков, в метаморфизме пород нижней и средней коры в условиях гранулитовой и высокотемпературной амфиболитовой фации, который распространяется также и на породы, заполнившие рифтогенные депрессии. Гранулито-гнейсовые комплексы Восточно-Европейского кратона представляют собой один из главных компонентов обширных овальных внутриконтинентальных тектонических структур регионального или континентального уровня. Благодаря вовлечению в обсуждение гранулито-гнейсовых комплексов Восточной Европы, Северной и Южной Америки, Африки, Индии, Китая и Австралии стала очевидной необходимость значительного пересмотра существующих взглядов на соотношения процессов тектоно-плитного и тектоно-плюмового типа, равно как и процессов сборки и распада суперконтинентов, в геологической истории. Показано, что Восточно-Европейский и Северо-Американский кратоны являются фрагментами долгоживущего суперконтинента Лавроскандия. После его возникновения около 2.8 млрд лет назад кора этого суперконтинента эволюционировала при воздействии последовательности мощных плюмов (суперплюмов) вплоть до ~0.85 млрд лет. За это время Лавроскандия неоднократно подвергалась рифтингу, частичному расколу и последующему восстановлению целостности континента. Процессы тектоно-плитного типа (рифтинг с переходом в спрединг с последующим закрытием короткоживущего океана с участием субдукции) во внутренней области Лавроскандии контролировались суперплюмами. Переоценка природы гранулито-гнейсовых комплексов привела к ряду принципиально новых заключений: о существенно более значительной, чем предполагалось раньше, роли мантийно-плюмовых событий в приросте ювенильной континентальной коры, особенно в неоархее – протерозое, о существовании суперконтинента Лавроскандия с ~2.85 до 0.85 млрд лет, о ведущей роли мантийных плюмов во взаимодействии тектоники плит и тектоники плюмов в неоархейской – протерозойской истории Лавроскандии и, возможно, в истории континентальной коры в целом. Мы предполагаем, что эволюция геодинамических обстановок в истории корообразования может быть представлена в виде спиральной последовательности: взаимодействие мантийно-плюмовых процессов и эмбриональной тектоники плит в палеомезоархее (~3.8–2.8 млрд лет) → плюм-тектоника и локально проявленная тектоника плит, инициированная плюмами (~2.80–0.55 млрд лет) → фанерозойская тектоника плит при редуцированной роли мантийных плюмов.

Ключевые слова: докембрий; тектоника плит; тектоника плюмов; гранулито-гнейсовый пояс; суперконтинент; внутриконтинентальный овальный ороген

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1. Introduction

The direct observations of geological processes on the modern Earth, as well as the data from the Mesozoic-Cenozoic record were the starting points for the creation and development of the plate-tectonics model that was initially defined in a series of papers between the late 1950s and early 1960s [e.g., *Dietz, 1961; Hess, 1962; Wilson, 1963*]. At present, lithospheric plate evolution is widely accepted as the principal framework within which much of the geological record has been produced. The basic concept of mantle plumes was originated in the 1960s and 1970s [*Wilson, 1963; Morgan, 1971*]. Hotspots and mantle plumes hold the key to several crucial issues of mantle dynamics, and geoscientists are now aware of their geodynamic importance [e.g., *Maruyama, 1994; Condie, 2001, 2004; Ernst, Buchan, 2003; Foulger, 2010*]. However, direct evidence for actual plumes is sparce, and many questions remain unanswered. Many scientists have been disappointed at the limited ability of observational data to confirm the predictions based on the plume hypothesis (see discussion in [*Foulger, Jurdy, 2007*]).

According to the existing models, mantle plumes are expected to rise from the lowermost mantle in the form of cylindrical conduits with a radius of 50–150 km. This range of values is based on fluid dynamics, whereas conduit radius estimated from seismological data is much broader. The morphology of a thermal plume is controlled by the viscosity contrast between the hot plume material and the mantle above it. The mushroom shape is favored, because a hot plume is likely to be a hundred times less viscous than the surrounding mantle, owing to the strong temperature dependence of viscosity [*Farnetani, Hofmann, 2011*, and references therein]. Numerical experiments, where heterogeneous thermal conditions at the top surface mimic the presence of a continent, show that within the convecting fluid, a vertical upwelling of deep hot material is always locked beneath the central part of the 'continent' [*Guillou, Jaupart, 1995; Trubitsyn, Trubitsyn, 2005; Guillou-Frottier et al., 2012*]. The size of the head depends mainly on the starting depth. For plumes originating at the base of the mantle, the final diameter upon arrival at the base of the lithosphere is about 1000 km. Flattening of the plume against the lithosphere doubles the diameter to about 2000–2500 km. For plumes starting at the 660-km boundary, the diameter upon reaching the base of the lithosphere is about 250 km, and the final diameter of the flattened plume head is about 500 km [*Ernst, Buchan, 2003*, and references therein]. The class of larger events termed *superplumes* (or *superplume events*) of several-thousand-kilometer diameters is also distinguished [*Condie, 2001*].

It seemed initially that plates and plumes were complementary, each related to and possibly driving a distinct form of mantle convection, and they seemed to operate independently [*Hill et al., 1992*, and references therein]. On the basis of global tomographic images and three-dimensional models of the seismic velocity structure of the Earth's interior, *Maruyama* [*1994*] and *Maruyama et al.* [*2007*] proposed *plume tectonics* to explain not only the superficial lithospheric layer but also dynamics in the whole Earth. Maruyama's model of plume tectonics emphasizes the dominant role of plumes and particularly superplumes in the dynamic evolution of the Earth: superplumes control the Earth's dynamic system, and plate tectonics is a subordinate element in the whole-Earth system driven by superplumes.

Regardless of their global interaction, the manifestations of plate- and plume-tectonics are more or less independently recorded in geological time and space and are characterized by distinct sets of indicators, which include peculiarities of sedimentary, igneous and metamorphic processes, lithological, geochemical, structural, and some other characteristics of the resultant rocks. For a long time, using more subtle features and characteristics, many researchers sought more clear definitions of the products of both types of processes, more accurate assessments of their role in the formation of tectonic structures at the global, regional and local levels, and attempted to clarify their relative roles in geological history. It has been shown that it is possible to derive a history of plume and superplume volcanism on the surface of the Earth with reference to plume-generated features that survive in the geological record.

The most important criteria for recognizing plate-tectonic processes are widely accepted: ophiolites, blueschists, ultrahigh-pressure terranes, eclogites, passive margins, subduction-related batholiths, arc-type igneous rocks, isotopic evidence of recycling, and palaeomagnetic constraints. In turn, the most diagnostic manifestation of mantle plumes seems to be the Large Igneous Provinces (LIPs) that consist of tholeiitic to transitional flood basalts, minor alkali rocks and picrites, variable amounts of felsic rocks, plumbing systems of dikes, sills and layered intrusions, and associated mafic underplating. Additional proxies comprise anorogenic felsic-intermediate magmatic rocks, carbonatites, kimberlites, lamproites, and peaks in BIF and black shale abundances [*Isley, Abbott, 1999*; *Condie, 2001, 2004; Ernst, Buchan, 2001; Ernst, Bleeker, 2010*]. Dike swarms that radiate from the plume head centre can transport magma laterally for up to 3000 km away from the plume centre and can feed lava flows, sill provinces, and layered intrusions at any distance along their extent [*Ernst, Buchan, 2003*]. Here we note an important omission: regardless of the high-temperature character of the mantle plume-related processes, researchers have not attempted yet to discuss high-temperature metamorphism among prominent 'plume proxies'.

Thus, the currently most popular geotectonic theories and models accept that the geological record contains fixed evidence of the operation of two major geodynamic mechanisms that determined the formation of the crust and lithosphere of our planet in geological history. With the growth and improvement of the geological database, researchers found it challenging to reconstruct the secular evolution of plate tectonics and plume tectonics in geological history. Initially, it seemed that the transition from an earlier plume-dominated Earth to one where plate tectonics became paramount occurred more or less gradually as the temperature decreased within Earth's interior. For further development of ideas on secular trends in Earth's evolution, two results were particularly important: (1) reconstruction of the supercontinent Pangaea and understanding the very high likelihood of supercontinent creations in Precambrian history, (2) discovery of the episodic nature of crust evolution, recorded by U-Pb ages of zircons from Precambrian granitoids and sediments of major rivers, which are grouped into a series of major peaks at about 2.7, 2.5, 2.1, 1.9, 1.1 and 0.7 Ga. The possibility of a widespread magmatic shutdown between 2.45 and 2.2 Ga based on coincidence of a variety of observations in this time window was noted as well [*Condie et al., 2009b*].

The perceived antiquity of the Recent–Phanerozoic style of plate tectonics remains contentious. Some authors argue for a very early (3.9–3.8 Ga or earlier) onset of plate regimes [*de Wit, 1998; de Wit, Hynes, 1995; Mints, 1999; Komiya et al., 1999; Komiya, 2007; Furnes et al., 2007; Shirey et al., 2008*], whereas others assume that this transition has come significantly later (at ~3.0 Ga) [e.g., *Smithies, Champion, 2000*], at ~2.0 Ga [e.g., *Hamilton, 1998*], at ~1.0 Ga with earlier episodes of proto-plate tectonics at 2.7–2.5 and 2.0–1.8 Ga [*Stern, 2008*]. *Condie* [*1998*] and *Isley* and *Abbott* [*1999*] have presented arguments that superplume events have been important throughout Earth history. Although the meaning of the term "superplume event" varies in the scientific literature, these authors constrain the term to refer to a short-lived mantle event (~100 m.y.), during which many superplumes, as well as smaller plumes bombard the base of the lithosphere. During a superplume event, plume activity may be concentrated in one or more mantle upwellings, as during the Mid-Cretaceous superplume event of some 100 m.y., when activity was focused mainly in the Pacific mantle upwelling. Precambrian superplume events at 2.7 and 1.9 Ga correlate with maximums in worldwide production rate of juvenile crust and thus, may not have been confined to one or two mantle upwellings. *Arndt* and *Davaille* [*2013*] concluded that initial destabilization of the basal layer in the mantle at 2.7 Ga generated a large swarm of mantle plumes that produced a global peak of crust formation. Subsequent episodes were smaller and more dispersed and the final episodes at about 1.1 and 0.7 Ga marked the transition to the modern plate tectonics.

When interpreting age peaks, researchers are mostly guided by geochemical and petrologic evidence and partly rely on lithology [*Condie, 1998, 2001; Condie et al., 2001; Hawkesworth, Kemp, 2006; Arndt, Davaille, 2013*]. In contrast, we focus our attention on interpretation of large tectonic structures, e.g., granulite-gneiss belts and orogens as whole entities. In particular, we do not suggest *a priori* that granulite-facies metamorphism is related to intercontinental collision or to growth or breakdown of supercontinents. The discrepancies in the existing interpretations demonstrate the essential ambiguity of existing ideas about what are the necessary and sufficient evidences for plate- and plume-tectonic processes. We need to detect more confidently and reliably the magmatic, metamorphic, and tectonic fingerprints of the pre- and non-plate tectonic Earth in the geological record [*Stern, 2008*].

A somewhat hidden contradiction in the current understanding is that, despite the important role of plume-related processes in geodynamics, the assessment of their contribution to the origin of juvenile continental crust remains problematic. Most investigators agree that cratons are the end-product of collisional orogenesis (resulting from termination of the Wilson cycle), and continents are made of collisional orogens of various ages [*Condie, 2005*, and references therein]. On the other hand, the episodic growth of juvenile crust can be explained through the episodicity of mantle-plume events [*Condie, 1998*]. Based on field observations on exposed lower crustal sections and seismic data, it was suggested that great volumes of plume-related juvenile magmas had been ponded beneath the crust-mantle boundary. According to an approximate estimate based on the distribution of flood basalts, dike swarms, high-Mg rocks, layered mafic-ultramafic intrusions, the largest Precambrian superplume events were planetary in scale, involving at least 14–18 % of the Earth's surface area [*Abbott, Isley, 2002*]. But how complete is the list of the plume-related rock types? And what is the significance of other plume-related complexes?

Although the signatures of the above-listed mantle plume-related and plate-tectonic-related processes, in general, are recognized by most investigators, their application in specific situations leaves a vast arena for debate. Paradoxically, the lists of "proxies" of both plate- and plume-related settings are deprived of even a mention of the high-grade metamorphic rocks (granulite, amphibolite, high-temperature eclogite facies). However, the granulite-gneiss belts and areas, which contain these rocks, have a regional distribution in both the Precambrian and the Phanerozoic records. An origin of the granulite-gneiss belts and areas is one of the main problems in continental crust provenance studies [*Mints, 2014*], because granulites and related high-grade rocks are the backbones of continents [*Collins, 2002*]. In this regard, the proportion of the granulite-facies juvenile rocks and, in particular, khondalites (one of the major components of granulite-gneiss belts) in the continental crust has to be rather high, but a correct estimate has not been made yet. This uncertainty is aggravated by the mystery of the source of the juvenile greywackes metamorphosed under granulite-facies conditions [*Taylor, Kalsbeek, 1990; Kalsbeek, Nutman, 1996; Van Kranendonk, 1996; Kalsbeek et al., 1998; Nutman et al., 1999; Scott, 1999; Mints, 2007; Mints, 2015b*]. Anorthosite and gabbroanorthosite intrusions, which almost always occur in granulite-gneiss belts, are, as a rule, not regarded as constituents of LIPs, and their contribution to the continental crust remains unvalued. Meanwhile, abundant xenoliths of these rocks carried up from the lower crust underplated by mantle magma show that they are a substantial constituent of the continental crust [*Rudnick, Fountain, 1995*]. The analysis of an extensive database on the structure and evolution of the East European Craton (EEC) has led to the conclusion that the EEC granulite-gneiss complexes form one of the main components of the large oval intracontinental tectonic terranes of regional and continental rank [*Mints, 2014, 2015b, 2015c*]. Granulite-gneiss complexes indicate significant vertical displacements, as a result of which deep-crustal rocks reached or approached the palaeotopographic surface. Introducing the granulite-gneiss complexes in the discussion of the problem stated in the title to this paper suggests a significant revision of the current views on the relations between the plate- and plume-tectonic processes in geological history, including supercontinent assembly and breakdown.

The palaeogeographic reconstructions of the Early Precambrian are hampered significantly by natural limitations of applying palaeomagnetic methods and by significant hiatuses in the palaeomagnetic database [*Buchan et al., 1998*]. The reconstructions of the Early Precambrian supercontinents are based primarily on the correlation of geological structures and events investigated on the present-day continents. The fact that the models for the origin and evolution of supercontinents in the Proterozoic [*Gaál, 1992; Rogers, 1996; Condie, 1998; Li et al., 2008*] are fundamentally different, is a consequence of gaps in understanding the nature of the Early Precambrian rocks and geological structures.

The antiquity of the supercontinent cycle is the subject of a parallel debate, equally unresolved [*Unrug, 1992; Rogers, 1996; Aspler, Chiarenzelli, 1998*]. The inherent link [*Zhong et al., 2007; Santosh et al., 2009*] between this unresolved question and the mantle plume concept [*Condie, 2004; Condie et al., 2001*] suggests that the supercontinent cycle was generated by the interaction of both mantle thermal and plate-tectonic regimes.

Although the idea of a simple and gradual transition from plume- to plate-tectonic domination with a starting point somewhere along a continuous trend is the prevailing hypothesis today, it bears intrinsic contradictions. Problems with strict interpretation in the frame of uniformitarian plate-tectonic models, when applied to the Neoarchaean and Proterozoic records, have been noted repeatedly [*Hoffman, 1989; Bickford, Hill, 2007; Stern, 2008*]. Significant variations in geological processes in space and time argue against such a simple view on the evolution of the Earth's geodynamics.

Our previous study and the results presented in this paper are based on a systematic review and analysis of the geological relationships engendered by plate-tectonic and mantle-plume processes in the Precambrian lithologies of the East European and North American cratons, termed Lauroscandia as a whole. The evolutionary trends of the Early Precambrian crust characteristic of Lauroscandia are also typical of other continents. The volume of this publication does not allow us to involve the full body of available information, so we will mention only a few striking examples beyond Lauroscandia.

2. Reassessment of the nature of granulite-facies metamorphism, granulite-gneiss complexes, and intracontinental orogens

Granulite-gneiss complexes are an essential component of the Early Precambrian crust within all ancient cratons, East European and North American, in particular. The nature of the heat source and the tectonic control responsible for high to ultrahigh temperatures (from 750 °C to >1000 °C) characteristic of granulite-facies metamorphism, especially at moderate pressures, are particularly contentious. What are the relative roles of extension and mafic magma underplating in the intracontinental plume-related events and/or accretionary arc – back-arc systems, as opposed to crustal thickening during collisional orogeny in the formation of these rocks? Are granulites closely linked to the generation of new crust or do they simply manifest intracrustal reworking during continental collision and assembly?

Typically, the high-grade metamorphic processes are related to the plate-tectonic environments. Formation of high-pressure granulites and high-temperature eclogites, which frequently occur at the base of thick granulite-gneiss complexes, is commonly not only a result of accretionary and/or collisional settings but also primary evidence for their reality [*Percival, 1994; Daly et al., 2006; Brown, 2007, 2009; Li et al., 2008; Gower, Krogh, 2002*; and many others]. It is usually accepted that regional metamorphism mostly occurs in plate boundary zones. *Brown* [*2007, 2009*] pointed out that lower thermal gradients are associated with subduction and early stages of collision, whereas higher thermal gradients are characteristic of back-arc and orogenic hinterlands. It was suggested also that fractional crystallization of basalts in the lower crust may produce anorthosites, and losses of fluids from the thickened lower crust may leave behind granulite-facies mineral assemblages [*Condie, 2005*, and references therein]. The regional granulite-gneiss belts are regarded as hot collisional orogens [*Rivers, 2009; Hynes, Rivers, 2010*], or as cores of these orogens [*Daly et al., 2006*]. In some cases, granulite-facies metamorphism is related to the back-arc extension, develops beneath sedimentary basins [*Collins, 2002*], or is initiated by subduction accompanied by slab breakup that ensures high heat flux from the mantle [*Kramers et al., 2001*]. *England* and *Thompson* [*1984, 1986*] concluded that the only way to reach low-*P*–high-*T* conditions of granulite- and high amphibolite facies would be transport of extra heat to upper crustal levels by granitic magma produced by partial melting in the thickened lower sialic crust. In addition, the metamorphism under granulite and high amphibolite facies conditions is typical of the intracontinental setting and is associated with an increase in heat flux from the mantle accompanied by extension and crustal thinning followed by collisional compression [*Thompson, 1989; Thompson et al., 2001*]. *Harley* [*1989*] and *Sandiford* [*1989*] argued that high-grade metamorphism needs a mantle heat source. The thermal model of *Depine et al.* [*2008*] demonstrates that long-lived melt migration must result in a quasi-steady-state geotherm with a rapid temperature increase in the upper crust and nearly isothermal conditions in the middle and lower crust. Granulite-gneiss complexes worldwide are accompanied by massifs of anorthosite–charnockite–mangerite–granite (AMCG) and/or anorthosite–rapakivi granite (ARG) complexes. Recent studies of the EEC have shown that the Archaean and Proterozoic granulite-gneiss complexes together with AMCG and ARG intrusions are the typical components of the intracontinental magmatism and orogeny [*Mints, 2015b, 2015c*].

Recently recognized intracontinental oval orogens are oval-shaped intracontinental tectonic ensembles of regional rank from 600–1000 to 2500–3000 km in diameter, and partly characterized by concentric structure and metamorphic zoning or bowl-shaped crustal structure [*Mints, 2011, 2014, 2015b, 2015c*]. Involvement of the granulite-gneiss complexes in the intracontinental oval orogens indicates significant vertical displacements of the deep-seated rocks to a higher level of the crust or directly to the surface. Such movements are accompanied by large-scale deformations of the Earth's surface and mountain building. In addition to granulite-gneiss complexes, the intracontinental oval orogens comprise juvenile and crust-contaminated mafic igneous rocks, layered mafic-ultramafic intrusions, 'dry', high-temperature within-plate granites, enderbites and charnockites, as well as low-grade sedimentary-volcanic belts.

The oval or oval-concentric structure rules out the origin of intracontinental oval orogens as a result of the processes at convergent plate boundaries. The size, morphology, and influx of mantle heat make the intracontinental oval orogens comparable with oceanic plateaus and large igneous provinces (LIPs) at continents. The emergence of oceanic plateaus and continental LIPs is usually associated with mantle plumes (superplumes) [*Condie, 1998; Ernst, Buchan, 2001; Abbott, Isley, 2002; Ernst et al., 2004; Kerr, Mahoney, 2007; Sharkov, 2010*]. The Karelian–Belomorian and Kola areas of high-temperature metamorphism and magmatism and the Volgo-Uralia granulite-gneiss area are typical examples of Neoarchaean intracontinental oval orogens [*Mints, 2014, 2015b, 2015c*]. The recently identified Palaeoproterozoic intracontinental oval orogen, two approximately equal halves of which are situated in Eastern Europe (the Lapland–Mid-Russia–South Baltia orogen) and North America (Trans-Hudson, Taltson-Thelon, Penokean orogens, etc.), was named the *Lauro-Russian intracontinental oval orogen* [*Mints, 2014, 2015c*]. The Meso-Neoproterozoic Grenville–Sveconorwegian orogen (GSNO) along with associated Keweenawan-Midcontinent rift system, AMCG and ARG complexes, are also referred to the intracontinental oval orogen type (see [*Mints, 2014*] for detailed discussion). Fragments of the intracontinental oval orogens, such as the Lapland–Mid-Russia–South Baltia Orogen, can be considered as linear or arcuate intracontinental collisional orogens (close to understanding and application of the term by *Cawood et al.* [*2009*]).

In addition to the granulite-gneiss complexes and associated intrusions, juvenile complexes are concentrated within volcanic-sedimentary and volcanic-plutonic belts composed of low-grade metamorphic rocks of greenschist to low-temperature amphibolite facies. These belts can be interpreted largely as former continental rifts, some fragments of which are similar to modern sutures and include not only riftogenic, but also oceanic and island-arc complexes formed in short-lived oceans of the Red Sea type (e.g., Pechenga–Varzuga, Circum-Karelian, Circum-Superior, Trans-Hudson and other Palaeoproterozoic belts). Other low-grade belts are similar to marginal accretionary orogens consisting of oceanic, island-arc, and back-arc terranes accreted to continents (e.g., Svecofennian and Wopmay orogens). It has been shown that the origin and evolution of the granulite-gneiss belts correspond 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 pyroclastic flow deposits. Metamorphism of lower and middle crustal complexes under conditions of granulite and high-temperature amphibolite facies also affects the fill of rift depressions [*Mints, 2007, 2014, 2015b*].

Thus, the granulite-gneiss belts are distinct intracontinental tectonic structures composed of sedimentary and igneous rocks undergoing high-grade metamorphism. These belts are a characteristic component of plume-related intracontinental oval orogens [*Mints, 2015b, 2015c*]. Therefore, the Precambrian granulite-gneiss belts should not be regarded as counterparts of the Phanerozoic sutures.

3. Geological record of the relationships between plate-tectonic and mantle-plume engendered complexes in the Precambrian crust of the East European and North American cratons and beyond Lauroscandia: a review

To date, an extensive database allows us to describe in detail the relationships between plate-tectonics and plume-tectonics in the history of Precambrian crustal evolution. The most complete pattern can be presented based on the geological record of the North American and East European cratons. In order to substantiate more comprehensively our findings, we will also include, briefly, the geology of some other continents beyond Lauroscandia.

3.1. Enigmatic Earth (>3.8 Ga ago)

Information about the initial stages of the Earth's crust formation is scarce. Our knowledge on geological events of that time is to a significant extent or mostly (for Hadean) based on indirect evidence. Until recently, it was suggested that soon after accretion, our planet underwent intense heating, which predetermined all main parameters of geological processes of that time, including origination and differentiation of magmatic ocean, prevalence of high-temperature metamorphism up to granulite facies conditions, the Venusian environment at the terrestrial surface, etc.

The discovery of Hadean detrital and inherited zircons in the Palaeoarchaean (3.28–3.20 Ga) metasedimentary rocks of the *Narryer Complex* in the Jack Hills area (the Yilgarn Craton, southwestern Australia) [*Wilde et al., 2001; Nelson, 2002; Iizuka et al., 2010*] implies that extensive continental crust existed even in the Hadean, indicating early differentiation of the upper mantle. Metaquartzite and metaconglomerate in the Jack Hills area contain a significant amount of detrital zircons, the age of which varies in a wide range. The oldest estimated ages are 4.37–4.01 Ga [*Harrison et al., 2005*]. Quartz, K-feldspar, and monazite inclusions in zircons along with enrichment in LREE, as well as results of oxygen-isotope thermometry of zircon crystallization show that relatively low-temperature (~700 °C) hydrous granitic magmas already existed at that time. Generation of such magmas, in turn, indicates that water-bearing sediments and correspondingly water proper existed at the Earth's surface. The temperature at the Earth's surface was lower than 200 °C as early as the Hadean [*Valley et al., 2002*]. The initial $^{176}Hf/^{177}Hf$ isotope ratios in detrital zircons from Jack Hills zircons provide evidence for the onset of continental crust growth as early as 4.5 Ga [*Harrison et al., 2005*], i.e. immediately after completion of the Earth's accretion as a planet.

Fragments of the oldest known terrestrial sialic continental crust (~4.28 Ga) are retained in the *Nuvvuagittuq supracrustal belt* of the northern Superior Province [*O'Neil et al., 2011*]. These are supracrustal mafic amphibolites and ultramafic rocks, BIF intermixing with coarse-grained ferruginous quartz–pyroxene rocks and quartz–biotite schists, which resemble polymictic metaconglomerates. The trondhjemitic orthogneiss sheets locally have intrusive contacts with supracrustal rocks. The high-Ti mafic rocks share geochemical characteristics with tholeiitic volcanic suites with low Al_2O_3 and high TiO_2 contents and are considered to be a product of crystal fractionation at low pressures under dry conditions. In contrast, the low-Ti rocks are geochemically similar to boninitic and calc-alkaline volcanic suites fractionated at elevated water pressures [*Cates, Mojzsis, 2007; O'Neil et al., 2008*]. Transition from tholeiitic to calc-alkaline magmatism established in the Nuvvuagittuq greenstone belt is a typical feature of many younger Archaean greenstone belts.

Another example of the oldest (4.2–3.7 Ga) sialic continental crust is orthogneiss from the *Acasta Gneiss Complex* in the Northwest Territories of Canada [*Bowring, Williams, 1999; Komiya, 2007*, and references therein]. According to field observations and chronology of the rocks, the Acasta Gneiss comprises at least four intrusive phases (3.97–3.94, 3.74–3.73, 3.66, and 3.59–3.58 Ga) which are tonalitic to granitic in composition [*Iizuka et al., 2007*].

Thus, the available data show that by ~3.8 Ga, the early Earth had a hydrosphere, and the crust was divided into granitic and mafic domains [*Nutman et al., 1984; Mints, 1999; Bohlar et al., 2004*]. The geochemical features suggest that embryonic plate-tectonic and mantle-plume processes operated as early as the Hadean, however, geodynamic interpretations of the oldest rocks remain controversial as yet.

3.2. Suggested co-existence and interaction of plume- and embryonic plate-tectonics (3.8-3.5 Ga)

3.2.1. Evidence of embryonic plate-tectonics

The Eoarchaean *Isua Greenstone Belt* (southwest Greenland) includes the *Isua Ophiolite Complex* (~3.8 Ga) that provides evidence for the oldest seafloor spreading and the earliest intact terrestrial complex of this type. Geochemistry of the rocks points to intraoceanic island-arc and mid-ocean ridge-like settings; the oxygen isotopes suggest a hydrothermal alteration of oceanic floor [*Furnes et al., 2007*]. Detailed mapping of the Isukasia area in the Isua greenstone belt has shown that the reconstructed lithostratigraphy was quite similar to Phanerozoic oceanic plate stratigraphy, and consisted of pillowed-massive lava units, bedded cherts, and turbidites. The Isua greenstone belt includes also voluminous mafic metavolcanic rocks that are composed primarily of pillow basalts intercalated with ultramafic units. The geochemical features of these rocks are comparable to Phanerozoic boninites. Recent boninites occur in the intraoceanic subduction-related environment, and it can be suggested that intraoceanic subduction zones were operating as early as 3.7–3.8 Ga ago [*Polat et al., 2002*]. The discovery of duplex structures and oceanic-plate stratigraphy indicates that the Isua supracrustal belt is the oldest known Archaean accretionary complex. This assumes underthrusting of oceanic crust and horizontal motion of oceanic plates [*Komiya et al., 1999*].

3.2.2. Evidence of plume-tectonics

Friend and *Nutman* [*2005*] noted that after the juvenile crustal accretion events dated at 3.85–3.69 Ga, the evolution of the *Itsaq* (*Amîtsoq*) *Gneiss Complex* (southwest Greenland) was followed by superimposed crustal-reworking events between 3.66 and 3.55 Ga, including intrusion of several generations of geochemically diverse granites, monzonites, quartz diorites and subordinate ferrogabbros and ferrodiorites. Within-plate granite chemistry has been established for the augen gneiss suite (~3.64 Ga). Multiple episodes of amphibolite to granulite facies metamorphism (630–700 °C at 8–10 kb) were also noted [*Griffin et al., 1980*]. Inclusions of metavolcanic and metasedimentary rocks in the Itsaq Gneiss are dominated by banded and commonly skarn-bearing amphibolites; lesser amounts of BIF, marble, siliceous rocks and calc-silicate rocks have been identified as well. Taken together, these inclusions are named the *Akilia association*. The geochemical and geochronological studies of these rocks indicate that they were formed in epicontinental basins, which had been between 3.65 and 3.60 Ga (from dating of the youngest detrital zircons) and were inverted and metamorphosed by ~3.57 Ga (dating of oldest *in situ* metamorphic overprint). A single body of the Akilia association is composed of layered gabbroanorthosite. As suggested, granulite-facies metamorphism of the Itsaq (Amîtsoq) Gneiss Complex is related to collisional orogenic events that mark termination of crustal accretion. Underplating of the crust by mafic magma caused deep crustal melting with generation of variegated suites of granites, monzonites and Fe-rich mafic rocks [*Friend, Nutman, 2005*, and references therein]. However, with allowance for the important role of epicontinental sedimentary and igneous rocks subject to high-grade metamorphism up to granulite-facies conditions, it is inferred to be more probable that the association of the rocks and the rock-forming events between 3.66 and 3.55 Ga might correspond to mantle plume-related orogeny (see discussions in [*Mints, 2014, 2015b, 2015c*]).

3.3. Mainly plume-related development of the crust (3.50–2.85 Ga)

3.3.1. Origin of the continental embryos of the East European craton (EEC): a plume-related start

The earliest events in the history of the Early Precambrian crust of the EEC (3.5–2.93 Ga) resulted in the emergence of spatially isolated areas of independent origin of the continental crust showing early development of the “pre-greenstone” continental blocks which dimensions did not exceed a few hundred kilometers (Fig. 1) [*Mints et al., 2015a*]. The origin of the pre-greenstone TTG gneisses, paragneisses, and amphibolites of the *Vodlozero microcontinent* embraced the interval from 3.24 to ~3.15 Ga. Geochemical features of the TTG gneisses and the predominance of CO_2 fluid inclusions in ancient zircon cores can be regarded as evidence for their formation due to high-temperature (granulite or high amphibolite facies) melting of the pre-existing TTG type crust of significant thickness, the provenance of which remains unknown [*Lobach-Zhuchenko et al., 1993; Chekulaev et al., 2009; Mints et al., 2015a*].

The pre-greenstone basement of the *Inari-Kola microcontinent* was formed by a TTG complex, the existence of which is fixed at 2.93 Ga. From 2.88–2.87 Ga, cores of zircon crystals of the Archaean TTG from the Kola super-deep well (7622–12262 m depth), isolated primary inclusions of liquid CO_2 and melt inclusions were extracted [*Chupin et al., 2006, 2009*]. With a high degree of confidence, it can be concluded with reference to the glass inclusions that the protolith of the TTG gneisses was mostly formed as a product of volcanic eruptions of high-temperature rhyodacite-rhyolite magmas saturated with CO_2. With allowance for dry melting of mafic substrate, the geodynamic

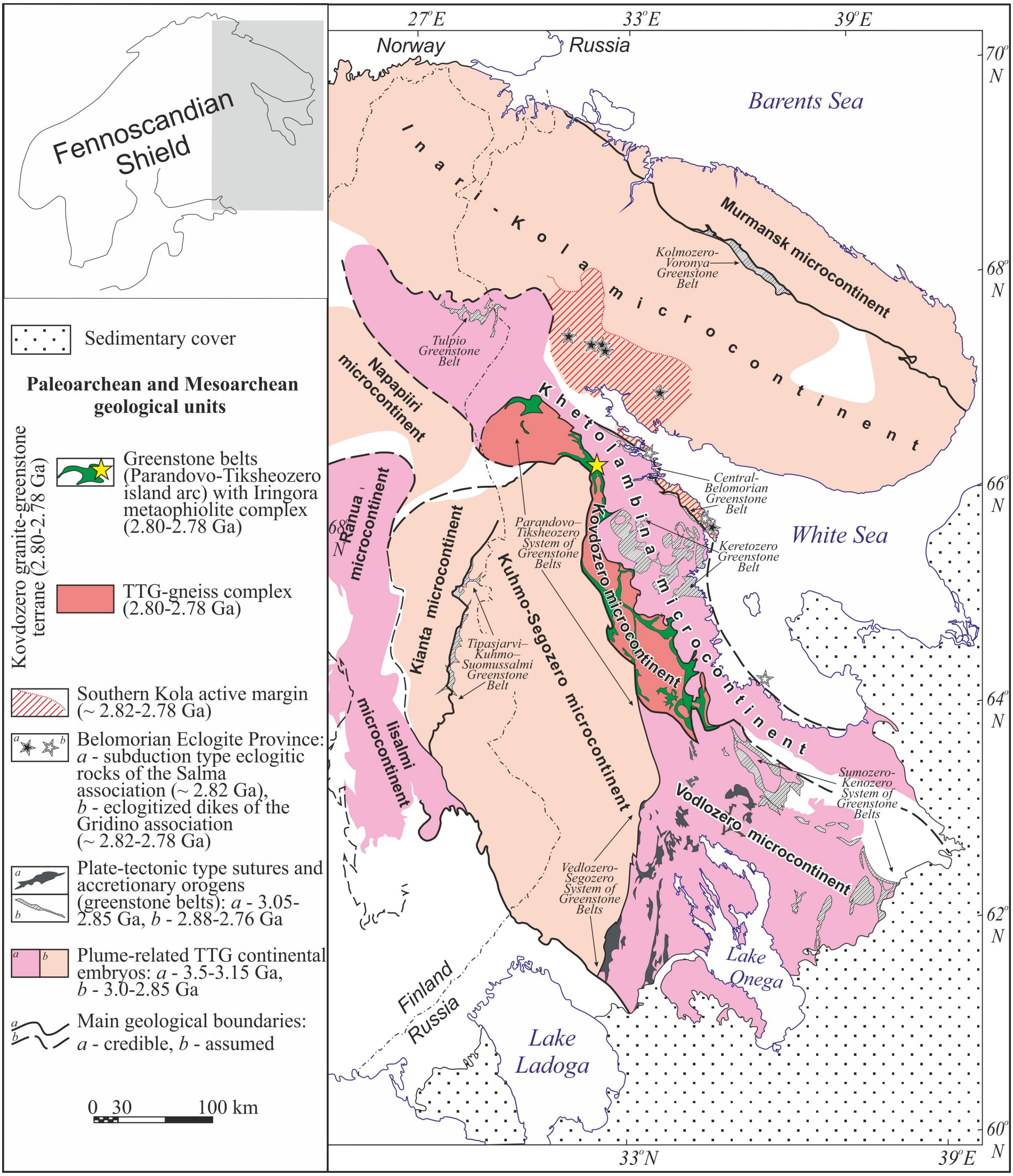

Fig. 1. The Palaeoarchaean and Mesoarchaean units in the eastern Fennoscandian Shield: the microcontinents formed by TTG gneisses mainly of the mantle-plume nature, are combined by greenstone belts (paleo-sutures). Slightly modified after [*Mints et al., 2015a*].

Рис. 1. Палеоархейские и мезоархейские комплексы восточной части Фенноскандинавского щита: микроконтиненты, образованные корой ТТГ-гнейсового типа преимущественно мантийно-плюмовой природы, объединены зеленокаменными поясами (палеосутурами). С небольшими изменениями по [*Mints et al., 2015a*].

setting of magmatic activity could have hardly been related to subduction. An intracontinental plume-type setting, with intense heating, underplating by mafic magma, and subsequent melting of mafic rocks with formation of TTG melts, is more probable [*Mints et al., 2015a*].

3.3.2. Succession of granulite-facies metamorphic events in West Greenland

As indicated above, the earliest granulite-facies metamorphism in West Greenland rocks occurred between 3.66–3.60 Ga in the *Itsaq Gneiss Complex*. Later, the Fiskenæsset anorthosite intruded the Palaeoarchaean crust and was metamorphosed at ~2.97–2.95 Ga. High-grade events followed repeatedly at 2.89–2.88 and ~2.82 Ga, and the last high-pressure granulite-facies metamorphism occurred at 2.75–2.70 Ga [e.g., *Windley et al., 1981; Baadsgaard et al., 1984; Nutman et al., 1989, 1993, 2009; Keulen et al., 2010*]. According to *Polat et al.* [*2011*], as the Archaean layered anorthosite complexes are typically associated with tonalite–trondhjemite–granodiorite (TTG) gneisses, the petrogenesis of both suites appears to have been controlled mainly by slab melting. However, according to the recent view of particularly competent researchers of the TTG problem, "…any environment allowing hydrous mafic rocks to melt at high pressure would be equally suitable to form "TTG-like" magmas …" [*Moyen, Martin, 2012, p. 20*]. In general, following the plume model of granulite-facies metamorphism [*Mints, 2014; Mints et al., 2015a*], we assume that in the West Greenland Palaeo-and Mesoarchaean, the close relationship in time and space of the intracontinent-type anorthosite intrusions and granulite-facies metamorphism may be seen as evidence that both were linked with mantle plume activity, like their Neoarchaean and Proterozoic counterparts.

3.3.3. The intracontinental development beyond Lauroscandia

Within the period of 3.5–2.9 Ga, intracontinental development was widespread within many distant areas beyond Lauroscandia. It is usually accepted that the Kaapvaal and Zimbabwe cratons in Southern Africa were formed ca. 3.1 Ga ago as a result of initial intra-oceanic processes and subsequent amalgamation of oceanic and arc terranes with extensive granitoid magmatism at the later stages [*de Wit et al., 1992*]. Alternatively, *Eriksson et al.* [*2009*] postulate that a global superplume event at ca. 3.0 Ga included a plume beneath the Kaapvaal cratonic nucleus, thus halting any subduction around that terrane due to the thermal anomaly.

The Limpopo granulite-gneiss belt, which is a classic Archaean high-grade gneiss terrane, straddles the border between the Kaapvaal and Zimbabwe cratons [*Van Reenen et al., 1992a; de Wit et al., 1992*]. The geological map [*Brandl, 1992*] shows that the Limpopo Belt is a synformal thrust and nappe structure, as is usual for large granulite-gneiss complexes with southwestern centroclinal closure. The belt is subdivided into three main subterranes, the Central Zone, the Southern and Northern Marginal Zones. Rocks within the marginal zones belong to granite-greenstone assemblages of the Kaapvaal and Zimbabwe cratons metamorphosed under granulite-facies conditions. Retrograde granulites of the Southern Marginal Zone occur as isolated nappes (klippen) above prograde granite-greenstone rocks with inverted metamorphic zoning up to at least 50 km south of the thrust front [*Roering et al., 1992a, 1992b; Van Reenen et al., 2011b*]. Despite the large database, researchers still disagree on many aspects of the geology and evolution of the Limpopo Complex (see review by van *Reenen et al.* [*2011a*] and references therein). In particular, this debate involves the significance of the Neoarchaean anatectic events that affected the high-grade rocks of the Central Zone of the Limpopo Complex, and the status of the ca. 2.61 Ga Bulai pluton as an important tectono-metamorphic time marker in the Central Zone. The lack of consensus is clearly demonstrated by the variety of models proposed for the evolution of the Limpopo Complex, such as the Himalayan-type evolution model [e.g., *Roering et al., 1992b*], the gravitational redistribution model [*Gerya et al., 2000*], and the Turkish-type accretion model [*Barton et al., 2006*]. Such significant discords in interpretation of the available data are mainy caused by the initial idea of relating the Limpopo Unit to collision orogenic belts. The resumption of thermal events, each regarded as an immediate consequence of collision, could not and cannot be coordinated with a general model of collisional orogens. Meanwhile, the geological map of the Limpopo Belt and its framework [*Brandl, 1992*] shows that this belt is a synform with southwestern centroclinal closure. In combination with resumption of thermal events and lithology of sedimentary and igneous protoliths of granulites, such a structure provides evidence for the intracontinental nature of the Limpopo Belt related to plume activity. The following thermal events in the history of this belt are not interrelated and correspond to consecutive global manifestations of superplume activity. The oldest granulite-facies metamorphic events are documented in the Central Zone.

The Central Zone of the belt is dominated by granodioritic and granitic gneisses intercalating with the *Beit Bridge Complex* of metasedimentary rocks. This complex consists of mainly epicontinental metasediments, including metapelite, paragneiss, marble, mafic granulite, quartzite, and BIF [*Van Reenen et al., 1992b; Laurent et al., 2011*]. Protoliths of mafic granulites (metatholeiites) are dated at ~3.3 Ga [*Chudy et al.,* *2008*]. Combined U–Th–Pb and Lu–Hf isotope datasets indicate that tholeiitic magmas and TTG with close to chondritic εHf_{in} (−2.0 to +1.4) were emplaced in the Palaeoarchaean, between 3.40 and 3.27 Ga, and that these igneous rocks are products of recycling of older rocks. This is also suggested by U–Th–Pb–Lu–Hf isotopic data on detrital zircon grains found in the metasedimentary rocks deposited after emplacement of the Sand River Gneiss at 3.24–3.0 Ga. These sediments contain detrital zircon grains as old as 3.9 Ga [*Zeh et al., 2010*]. A part of the metasedimentary rocks of the Beit Bridge Complex was intruded at ~3.36 Ga by the Messina gabbroanorthosite layered intrusion and Sand River TTG gneisses, all being affected by granulite-facies metamorphism (~10 kbars and >800 °C) at ~3.14 Ga [*Horrocks, 1983; Van Reenen et al., 1992b; Barton, 1996; Kröner et al., 1999; Zeh et al., 2007; Chudy at al., 2008; Laurent et al., 2011*]. Subsequent metamorphism within all three zones of the Limpopo Belt includes high-grade events between 2.74 and 2.57 Ga and between 2.03 Ga and 1.97 Ga (see below).

Another important example is the *epicontinental Witwatersrand Basin* formed between 3.07 and 2.71 Ga on the Kaapvaal Craton. Pulses of sedimentation and volcanism were apparently episodic, at 3.09–3.07 Ga (Dominion Group), 2.97–2.91 Ga (West Rand Group), and 2.89–2.71 Ga (Central Rand Group). The Ventersdorp bimodal volcanic association of felsic to mafic lavas and pyroclastics with subordinate basal komatiites (2.71 Ga) overlies the Witwatersrand succession. It was suggested by *Robb* and *Meyer* [*1995*] that both the Dominion and the Ventersdorp volcanics were erupted during periods of extensional tectonic activity and that the main stage of the Witwatersrand basin evolution was linked with collision of the Zimbabwe and Kaapvaal cratons. A more recent model [*Catuneanu, 2001*] supports a complex flexural foreland basin evolution resulting from collision of two ocean-displaced arc terranes [*de Wit et al., 1992*] with the Kaapvaal cratonic nucleus. The macrodiamonds, found together with kimberlite indicator minerals in palaeoplacers pertaining to the 2.89 to 2.82 Ga Central Rand Group (Witwatersrand Supergroup), emphasize that the Witwatersrand Basin rests on the oldest continental plate that was also the first known plate to be large and thick enough to develop clear intraplate magmatism including ~2.9 Ga kimberlite events [*Gurney et al., 2010*] and the oldest foreland basin system.

Mt. Narryer and Jack Hills metasedimentary rocks in the Narryer Gneiss Complex of the Yilgarn Craton, Western Australia, deposited between 3.28 and 3.20 Ga, are of particular importance because they contain Hadean detrital zircons. The Mt. Narryer supracrustal belt dominantly comprises epicontinental-type assemblages: (1) quartzite with various amounts of garnet, cordierite, sillimanite, plagioclase, pyroxene and amphibole, (2) quartz-pebble metaconglomerate and (3) polymictic metaconglomerate with quartz pebbles, BIF and paragneiss. The Mt. Narryer metasediments experienced at least two metamorphic events: one at ~3.3–3.2 Ga and another at 2.7–2.6 Ga ago, when peak granulite-facies metamorphism was accompanied by extensive recrystallization. The Jack Hills supracrustal belt has undergone a complex and long-term deformation history; however, it did not experience high-grade (granulite-facies) metamorphism comparable to that of the Mt. Narryer metasedimentary rocks [*Nutman et al., 1991; Iizuka et al., 2010*; and references therein].

3.3.4. The plume-related origin of harzburgitic rocks in the subcontinental lithospheric mantle (SCLM)

The origin of the thick, refractory roots of the SCLM underlying most of the Archaean cratons is a key question for understanding the crust–mantle interaction during the Palaeoarchaean. The extensive information on composition, structure, and age of the SCLM has been obtained from the study of deep mantle xenoliths, including diamond-bearing varieties, which were entrained by kimberlite and lamproite pipes. The diamond-bearing harzburgites and eclogites are approximately equal in abundance as host rocks of economic diamonds. The harzburgitic SCLM represents buoyant residues of high-pressure, high-degree melting that have risen from below having been produced by large plumes or mantle overturns [*Griffin, O'Reilly, 2007a, 2007b*; and references therein]. Harzburgite composed of high-Mg olivine and orthopyroxene represents a solid residue left after partial melting of the mantle as a result of thermal impact of anomalously hot mantle plume. A relatively low density of high-Mg lithospheric matter ensured formation of stable lithospheric roots at the base of the Archaean continental domains. The subsequent plume effects facilitated strengthening of the lithospheric roots [*Arndt et al., 2009*]. Further, high pressure and relatively low temperature at the base of the roots corresponded to the diamond-facies conditions. Multiple isotope studies define 3.52–3.20 Ga as the oldest age for the Archaean harzburgitic diamonds recognized in the SCLM beneath the Kaapvaal, Siberian, and Slave cratons, three of Earth's oldest continental nuclei [*Gurney et al., 2010*, and references therein]. The age of these diamonds dates termination of the origin of SCLM. The Re–Os isotopic systematics of sulphide inclusions incorporated into diamonds assumes that harzburgite was modified later as a result of percolation of C–H–O–S fluids generated in the Archaean subduction zones.

It is necessary to accentuate that the existence of the SCLM keels (roots) of the continents clearly demonstrates that the concept of the 'hotter' Archaean mantle is conditional: obviously, heating of the mantle was extremely irregular, and the mantle included 'cold' areas beneath the continents. The mantle plumes that produced komatiitic magma were evidently 'hotter' than more recent plumes. Logically, the temperature within rising plumes could contrast sharply with the temperature of the surrounding mantle.

3.4. Mainly plate-tectonic type development (origin, evolution and accretion of ancient island arcs, collision of microcontinents) (3.05–2.75 Ga)

3.4.1. The eastern Fennoscandian Shield

In the Karelia region, the remnants of a number of island arc systems accreted to the margins of the continental 'embryos' were identified based on geological, geochemical and geochronological data [*Mints et al., 2015a*, and references therein] (Fig. 1). The 3.05–2.85 Ga *greenstone belts of the Vedlozero-Segozero system* in southeastern Karelia are the derivatives of the oldest Mesoarchaean island arc systems in the region, which were formed and accreted to the western and northwestern margins of the Vodlozero microcontinent. The greenstone belts contain three characteristic sedimentary–volcanic associations: (1) basalt–andesite–rhyolite (3.05–2.90 Ga) of active margins type, (2) komatiite-tholeiite (2.94–2.86 Ga) formed in a back-arc basin, and (3) an andesite-dacite association (2.86–2.85 Ga) that completed the evolution of the active margin [*Svetov et al., 2001; Svetov, 1997*]. In turn, the *Sumozero-Kenozero greenstone belts* are the main part of the 2.88–2.85 Ga accretionary orogen formed along the eastern margin of the Vodlozero microcontinent. The base of the succession is dominated by tholeiitic basalts and komatiites that are the 'pure product' of the mantle that avoided contamination by an ancient sialic crust [*Puchtel et al., 1999*]. The second volcano-sedimentary assemblage is composed of island-arc type dacite-rhyolite tuffs, tholeiites and metasediments [*Popov et al., 1979*]. The *Tipasjärvi-Kuhmo-Suomussalmi greenstone belt* can be interpreted as an accretionary orogen along the western margin of the Kuhmo–Segozero microcontinent [*Hölttä et al., 2012; Mints et al., 2015a*, and references therein] formed during the time span of 2.81–2.74 Ga. The *Kovdozero granite-greenstone terrane* that separates the Khetolambina terrane and Karelia palaeocontinent is formed by TTG-granitoids and gneisses hosting metasediments and metavolcanics of several greenstone belts, which belonged to the Parandovo-Tiksheozero island arc that existed from c.2.81 to 2.78 Ga [*Mints et al., 2015a*]. The *Iringora greenstone belt* includes the ophiolite complex of the same name with the age of 2.78 Ga [*Shchipansky et al., 2004*]. The origin of the island arc system inferred for the *Kolmozero-Voronia greenstone belt in the Kola Peninsula* is dated at 2.87–2.83 Ga; at 2.78–2.76 Ga, the island arc together with a fragment of oceanic plateau were accreted to the margin of the Murmansk microcontinent [*Mints et al., 2015a*].

The recently discovered *Meso-Neoarchaean* (*2.87–2.82 Ga*) *Belomorian eclogite province* in the eastern Fennoscandian shield (Fig. 1) [*Mints et al., 2010, 2014, 2015a; Konilov et al., 2011; Dokukina, Konilov, 2011; Dokukina et al., 2014*] contains two eclogite associations distributed within TTG gneisses: (1) the subduction-type Salma association, and (2) the Gridino eclogitized mafic dikes. The protolith of the Salma eclogites is thought to have been the sequence of gabbro, Fe-Ti gabbro and troctolites, formed at ~2.9 Ga in the slow-spreading ridge (analogous to the Southwest Indian Ridge). The main subduction and eclogite-facies events occurred between ~2.87 and ~2.82 Ga. The Salma eclogite association is structurally linked with the Central Belomorian greenstone belt separating the Inari-Kola and Khetolambina tectonic units. It is formed by a mafic-ultramafic ophiolite type sequence dated at 2.88–2.85 Ga [*Bibikova et al, 1999; Slabunov et al., 2006*]. Mafic magma injections into the crust of the active margin, which led to formation of the Gridino dike swarm, were associated with immersion of a mid-ocean ridge in the subduction zone, starting from ~2.87 Ga. Crustal delamination of the active margin and subsequent involvement of the lower crust in subduction 2.87–2.82 Ga ago, led to high-pressure metamorphism of the Gridino dikes that reached eclogite-facies conditions during the collision event between 2.82 and 2.78 Ga. As a result, the Karelia, Kola, and Khetolambina blocks were consolidated, and the Mesoarchaean Belomorian accretionary–collisional orogen was formed. Detailed study indicates complicated post-eclogite history of the Belomorian eclogite province with mantle plume-related reactivation of thermal and fluid processes mainly at 2.72–2.70, ~2.4 and ~1.9 Ga. The eclogite assemblages were transferred to mid-to-upper crustal depths at ~1.7 Ga. Erosion or younger tectonic events were responsible for final exhumation to the surface. The high-temperature conditions of the eclogite-facies metamorphism can be ascribed to subduction of the mid-ocean ridge [*Konilov et al., 2011; Mints et al., 2014*]. To date, the subduction-related Salma eclogites provide the most complete and meaningful information on the nature of plate-tectonics in the Archaean from ocean floor spreading to subduction and collision.

3.4.2. Subduction-related modification of the subcontinental lithospheric mantle (SCLM)

Close to the beginning of the Mesoarchaean, SCLM was for the first time modified by subduction. The geochemical and isotopic signatures of the eclogite xenoliths from kimberlites are compatible with those of oceanic crustal rocks. The progressively metamorphosed coesite- and diamond-bearing eclogitic assemblages and coesite inclusions in eclogitic diamonds may be viewed as an expression of Archaean and Proterozoic ultrahigh-pressure (UHP) metamorphic events [*Helmstaedt, 2013*]. The earliest age for these processes, according to dating of diamondiferous eclogite xenoliths, is ~2.9 Ga [*Gurney et al., 2010*, and references therein]. The appearance of eclogitic UHP metamorphic assemblages at ~2.9 Ga implies that the microcontinental nuclei were coupled with lithospheric roots, which were sufficiently thick and cool to reach the diamond stability field [*Helmstaedt, 2013*]. The closeness of these dates to the ~2.9 Ga age of the Salma eclogites raises a question whether this is coincidence with or evidence for the onset of plate-tectonically driven regime of crustal evolution? To answer such a question, it is necessary to continue comprehensive investigations.

3.5. The first prominent superplume event recognizable worldwide (interaction of plume- and plate-tectonic processes initiated by the global superplume) (~2.79–2.50 Ga)

An age of ~2.7 Ga is the first and the most prominent peak in the U–Pb zircon age spectra of juvenile crust formation in the Earth's history, followed by lower peaks at 2.0–1.7 and 1.2–1.0 Ga, and suggests a sharp and sudden turning point in crust formation in the Neoarchaean (e.g., Fig. 1 in [*Condie, 1998*], and Fig. 2 in [*Hawkesworth, Kemp, 2006*]). What process was responsible for this vast addition of juvenile crust? To quote the latter authors, "It is difficult to conceive how global pulses of crustal growth might occur in response to conventional Wilson-style plate-tectonics, … . An alternative is that periodicity in igneous activity reflects rapid crust formation during thermal pulses associated with the emplacement of mantle plumes" [*Hawkesworth, Kemp, 2006, p. 814*]. The same peak is plotted in a number of diagrams that display secular changes in the evolution of crustal rock associations [*Bradley, 2011*], in particular, distribution of granitic rock ages [*Condie et al., 2009a*], granulites and ultrahigh- temperature metamorphic rocks [*Brown, 2007*], carbonatites [*Woolley, Kjarsgaard, 2008*], LIPs [*Prokoph et al., 2004*], komatiites [*Isley, Abbott, 1999*] and orogenic gold deposits [*Goldfarb et al., 2001*].

3.5.1. Intracontinental areas of sedimentation, magmatism, and high-temperature metamorphism (hot regions) in the eastern Fennoscandia

Specific areas of intracontinental thermal and tectonic activity have been identified recently in the eastern Fennoscandian Shield [*Mints, 2015a*].

The Karelian–Belomorian area is characterized by concentric oval structure incorporating several zones (Fig. 2). (1) The central zone of the area is characterized by the early manifestation of the tectonomagmatic and thermal activity expressed in granulite-gneiss complexes (2.76–2.73 Ga in age), epicontinental greenstone belts (2.76–2.73 Ga, possibly up to 2.70 Ga), and sanukitoid massifs within an arc-shaped zone (2.76 to 2.70 Ga). (2) The arcuate zone formed by the *Varpaisjärvi, Pudasjärvi and Notozero-Chupa granulite-gneiss belts* is about 600 km long and has a larger radius. Granulite-facies metamorphism is dated at 2.70–2.62 Ga. (3) The outer zone is formed by an array of 'younger' granite bodies (2.69–2.58 Ga) and local manifestations of granulite-facies metamorphism. Thus, the main development trend was rapid expansion of the area [*Mints, 2015a*, based on data of [*Bibikova et al., 1996, 2005; Halla, 2002, 2005; Heilimo et al., 2007; Käpyaho, 2006; Mutanen, Huhma, 2003; Lobach-Zhuchenko et al., 1993, 2005; Lobach-Zhuchenko, Chekulaev, 2007; Samsonov et al., 2004*].

The Kola area is distinguished by a block-linear geometry (Fig. 2) and probably continues in the southeastern direction, being overlain by the platform sedimentary cover. The emergence and evolution of rock associations belonging to the Kola area correspond to a long time interval from 2.79–2.76 to 2.61–2.55 Ga. The peaks of metamorphic and magmatic activity mutually correlate in age from 2.79–2.71 to 2.67–2.63 Ga. The intracontinental volcanic–sedimentary sequences of greenstone belts originated and evolved within the time interval of 2.76–2.66 Ga. Late metamorphic reworking is dated at 2.58–2.55 Ga [*Mints, 2015a*, and references therein].

The *Central Kola granulite-gneiss belt* within the Kola area exhibits a synformal structural ensemble superimposed on the Inari–Kola granite–greenstone terrane. The thickness of the tectonostratigraphic succession amounts to 7–8 km. In the history of the belt, there were three metamorphic events, (1) 2.79–2.71 Ga, (2) 2.66–2.64 Ga, and (3) 2.58–2.55 Ga [*Bibikova, 1989; Avakyan, 1992; Glebovitsky, 2005; Petrovskaya et al., 2007; Myskova, Milkevich, 2005; Levchenkov et al., 1995; Myskova et al., 2008; Bayanova, 2004*].

The extensive and poorly dated felsic calc-alkaline volcanism within the 2.87–2.78 Ga age bracket [*Pushkarev et al., 1978; Bayanova, 2004*] was combined with alkaline volcanism (2.67–2.63 Ga, probably up to 2.76–2.75 Ga [*Bayanova, 2004*]) in the Keivy structure of the Kola area. The latter is the first manifestation of alkaline magmatism in the Earth's record. These volcanic rocks were transformed into calc-alkaline and alkaline gneisses [*Mints, Tson, 1997; Mints, 2015a*]. Sill-shaped gabbroanorthosite bodies occur along the borders of the Keivy structure.

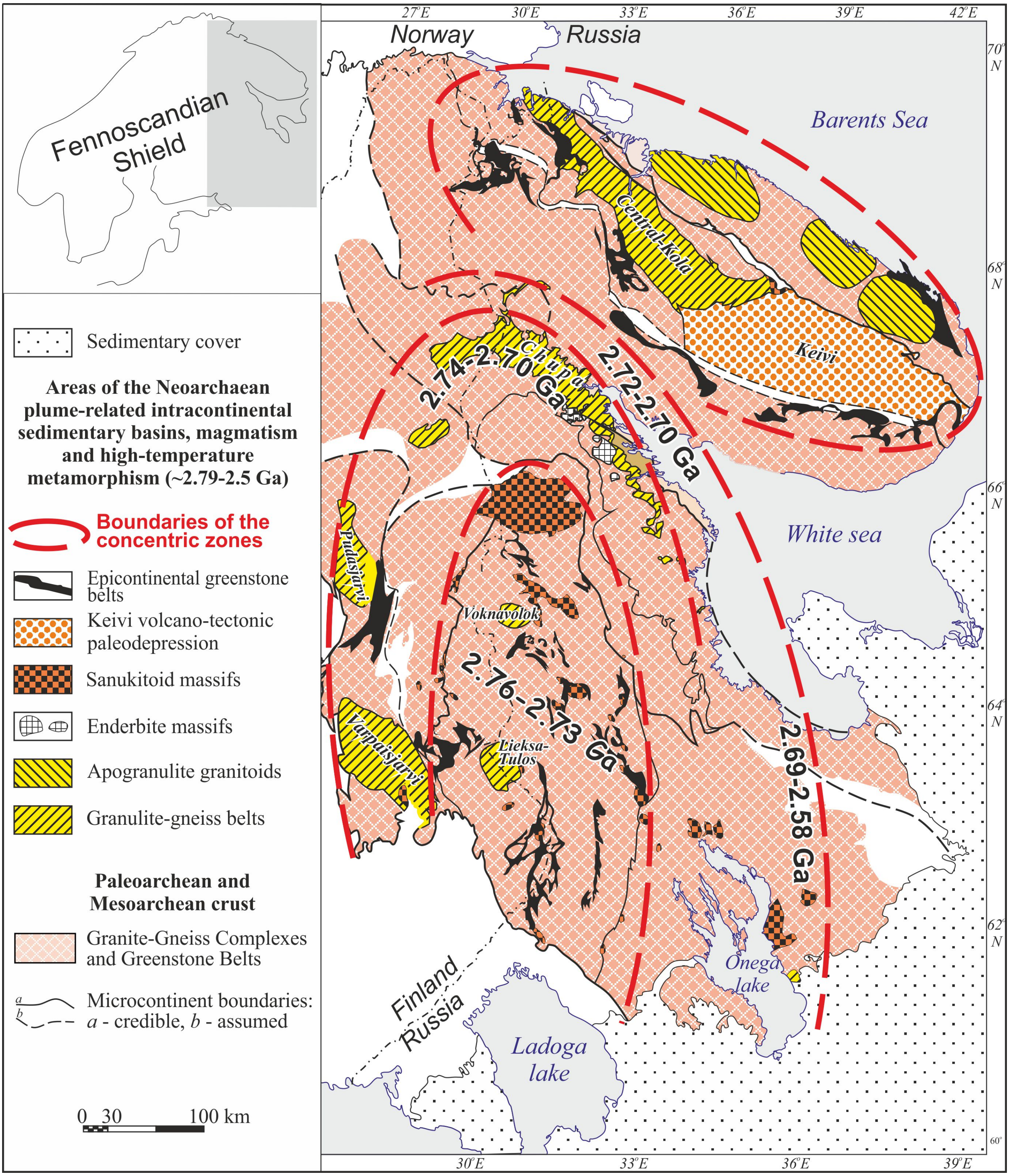

Fig. 2. The Neoarchaean intracontinental areas of high-temperature metamorphism, magmatism, and sedimentation (intracontinental oval orogens) in the eastern Fennoscandian Shield. Slightly modified after [*Mints et al., 2015a*].

Рис. 2. Неоархейские внутриконтинентальные ареалы высокотемпературного метаморфизма и магматизма и осадконакопления (внутриконтинентальные овальные орогены) в восточной части Фенноскандинавского щита. С небольшими изменениями по [*Mints et al., 2015a*].

According to the above data, the main events in the history of the Karelo-Belomorian and Kola areas were almost synchronous. The concentric oval structure and migration of the endogenic activity from centre to periphery are characteristic only of the Karelo-Belomorian area. The main features of the endogenic processes indicate conditions similar to the anorogenic environment and provide evidence for a powerful heat influx as a major event of the mantle-plume type.

The *Siilinjärvi carbonatite massif* in the western part of the Karelia Craton dated at 2.61 Ga [*Patchett et al., 1981; Zozulya et al., 2007; O'Brien et al., 2015*] strengthens evidence for the intracontinental mantle-plume nature of magmatism and metamorphism in the East European Craton at the end of the period under consideration.

3.5.2. Volgo-Uralia granulite-gneiss area

The extensive Volgo-Uralia granulite-gneiss area, 2.74–2.59 Ga in age [*Mints et al., 2015b*] (Fig. 3) with special tectonic structures called *ovoids* [*Bogdanova, 1986; Bogdanova et al., 2005*] plays the main role in the regional crustal structure. Ovoids are bowl-shaped crustal blocks 300–600 km across, which bases lie at the crust–mantle boundary. Ovoids are built up of mafic granulites with inclusions of gabbro, gabbro-anorthosite and ultramafic rocks. They are bounded by conical surfaces, along which the ovoid-forming rocks are thrust over the framework. Smaller oval synforms between ovoids are mainly filled with metasedimentary and mafic granulites. Metamorphism proceeded at the lower and middle levels of the crust in high *PT* conditions, at temperatures obviously exceeding the maximum estimates of 940–950 °C, and pressures above 9.5 kbar recorded in the samples from well cores. The protoliths dated at 3.4–3.2 to 3.1–3.0 Ga underwent granulite-facies metamorphism and subsequent retrogression at 2.74–2.70 and 2.62–2.59 Ga [*Bogdanova, 1986; Mints et al., 2015b*; and references therein]. In general, we interpret the structure and metamorphic evolution of the Volgo-Uralia region as derivatives of the 2.7 Ga superplume presented here in the form of a sequence of short-lived mantle events (each shorter than 50 Ma), when smaller plumes bombarded the base of the lithosphere. It was suggested that ovoids and oval synforms evolved where depressions formed due to extension of the crust above plumes. The depressions were rapidly filled with sediments and volcanic rocks that very soon thereafter experienced high-temperature metamorphism.

3.5.3. Superior Craton

The Superior Province in Canada is the largest of the retained Archaean cratons of the Earth. The origin and provenance of the Superior Craton are mainly the result of amalgamation of microcontinents and oceanic terranes to form a composite Superior Superterrane during a very short time span between ~2.75 and ~2.68 Ga; that is directly within that period, which is the subject of this section. A number of contradictions still remains in the understanding of its origin and provenance. According to the almost 'pure' plate-tectonic model, the Superior Craton resulted from a succession of events in the convergent-margin setting similar to the Pacific Rim. The distinct differences (e.g., abundance of komatiites and tonalitic plutons) are attributed to the hotter Archaean mantle [*Card, 1990; Jackson et al., 1994; Calvert et al., 1995; Daigenault et al., 2004*]. Although the accretionary tectonic model has been in the focus of many works concerning the Superior Province, there is no single model of the Archaean plate tectonics that would be universally accepted [e.g., *Hamilton, 1998*].

The *Abitibi greenstone belt* within the Superior Craton was formed between 2.75 and 2.70 Ga. It is widely known as a classic example of Neoarchaean juvenile continental crust formed in accordance with the plate-tectonic model. However, an old crust (3.8–2.8 Ga) became known recently at depth beneath the Neoarchaean granite-greenstone assemblages [*Percival et al., 2006; Ketchum et al., 2008*]. Besides, sialic crust of 2.93 Ga age occurs in the adjacent Wawa subprovince. The detailed geochemical and geochronological study of plutonic and volcanic rocks from the southwestern Abitibi Belt provides evidence for the involvement of the older crust in magma generation near the Abitibi–Wawa boundary. The isotopic data on felsic volcanic and plutonic rocks assume their dominantly juvenile source, although in some cases they demonstrate attribuites of significant crustal contamination. The available data taken together suggest that an approximately 75-km wide belt of the Abitibi crust was underlain by the older Wawa crust. Thus, it is thought that the Abitibi subprovince developed at the edge of the Wawa protocraton, perhaps as a result of rifting in this older continental block [*Ketchum et al., 2008*]. The discovery of diamondiferous lamprophyres and kimberlites provides direct evidence that the pre-2.7 Ga Superior Craton has a thick root reaching the field of diamond stability [*Stone, Semenyna, 2004; Barnett et al., 2007; Kopylova et al., 2011*]. The seismic image of the crust beneath the Abitibi greenstone belt, which was initially interpreted as direct evidence of Archaean subduction [*Calvert et al., 1995*], is linked to the north–south collision-related crustal shortening [*Bellefleur et al., 1998*].

The Superior Province hosts a number of high-grade metamorphic terranes. Some of them occur at the margins of the craton, others are localized between the granite–greenstone terranes, and the Kapuskasing

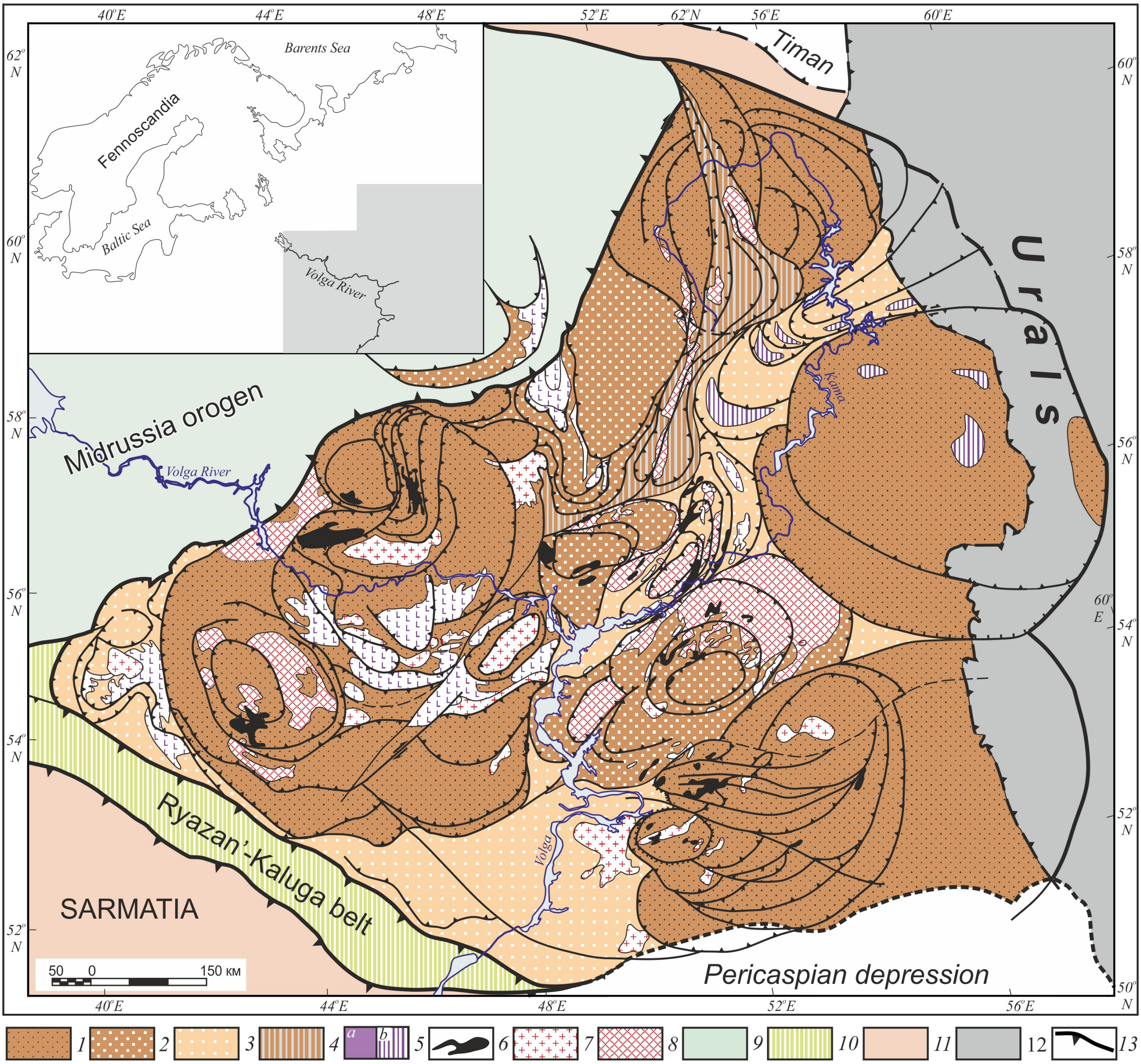

Fig. 3. The Neoarchaean Volgo-Uralia granulite-gneiss area (platform cover removed).

Neoarchaean (mainly 2.74–2.59 Ga): *1–4* – granulite-gneiss complexes: *1* – ovoids (mainly mafic granulites and subordinate metasedimentary granulite-gneisses), *2* – oval synforms (mainly metasedimentary granulite-gneisses), *3* – interovoid zone (alternating mafic granulites and metasedimentary granulite-gneisses), *4* – intensively deformed complexes of (1) and (3) types; *5–8* – igneous rocks: *5* – anorthosite and gabbroanorthosite: *a* – drilling and geophysical data, *b* – geophysical data only, *6* – mafic and ultramafic rocks, *7* – enderbite, charnockite, and garnet-bearing granite, *8* – granitoids; *9–10* – Palaeoproterozoic: *9* – Mid-Russia sector of Palaeoproterozoic Lapland–Mid-Russia–South Baltia orogeny, *10* – Ryazan-Kaluga volcanic–sedimentary belt; *11* – Archaean granite-gneiss complexes; *12* – Phanerozoic Urals Orogen; *13* – thrust, tags point to upthrown block.

Рис. 3. Неоархейский внутриконтинентальный гранулито-гнейсовый ареал (платформенный чехол удален).

Неоархей (преимущественно 2.74–2.59 млрд лет): *1–4* – комплексы гранулито-гнейсов: *1* – овоиды (преимущественно мафитовые гранулиты, подчиненно метаосадочные гранулиты), *2* – овальные синформы (преимущественно метаосадочные гранулиты), *3* – межовоидная зона (чередующиеся мафитовые и метаосадочные гранулито-гнейсы), *4* – интенсивно деформированные комплексы (1) и (3) типов; *5–8* – изверженные породы: *5* – анортозиты и габбро-анортозиты: *а* – по данным бурения и геофизики, *b* – исключительно по геофизическим данным, *6* – мафиты и ультрамафиты, *7* – эндербиты, чарнокиты, гранатовые граниты, *8* – гранитоиды; *9–10* – палеопротерозой: *9* – Среднерусский сектор палеопротерозойского Лапландско–Среднерусско–Южно-Прибалтийского орогена, *10* – Рязано-Калужский вулканогенно-осадочный пояс; *11* – архейские гранито-гнейсы; *12* – фанерозойский Уральский ороген; *13* – надвиги, бергштрихи указывают на надвинутый блок.

granulite-gneiss belt cross-cuts greenstone belts. The *Ashuanipi complex and Minto block* in the northeastern and the *Pikwitonei Domain* in the northwestern part of the Superior Province are composed mainly of orthopyroxene-bearing granitoids, including igneous charnockitic bodies and metapelite gneiss remnants corresponding to the low-pressure, high-temperature amphibolite- and granulite-facies metamorphism (up to 880 °C and 10.4 kbar in the Split Lake Block) dated between 2.71 and 2.64 Ga [*Heaman et al., 2011*, and references therein].

Two main alternative evolutionary models have been suggested for the Minto Block. According to the first model, the synformal structure of the Minto Block resulted from bilateral subduction beneath the Mesoarchaean crust [*Percival et al., 1992, 1994; Percival, Skulski, 2000*]. The second model is based on the data from sapphirine-bearing xenoliths in the enderbite of the Minto granulite complex, which suggests a pressure between 7.5 and 14 kbar (24–46 km) and a very high temperature (755–1260 °C). The peak temperature recorded in the xenoliths reflects the influx of mantle heat required to generate enderbitic magmas [*Cadéron et al., 2005*].

The high-grade metamorphic history of the *Pikwitonei granulite domain* at the northwestern margin of the Superior Province includes at least four episodes, at 2.72, 2.69, 2.68, and 2.64 Ga. According to the estimations, temperature reached 830–880 °C, and maximum pressures amounted to 9.3–10.4 kbar [*Heaman et al., 2011*, and references therein]. According to [*Mezger et al.,1990*], these events correspond to the emplacement of hot (>1100 °C) orthopyroxene-bearing enderbitic magma.

The 1.9 Ga *Kapuskasing granulite-gneiss belt* transects the central Superior Province over a distance of 500 km. It has exposed a crustal section through a palaeodepth of 25–30 km. The low-crustal granulite-facies ortho- and paragneisses underwent metamorphism at 750–790 °C and 8–11 kbar [*Percival, 1994; Percival, West, 1994*]. These rocks host the *Shawmere Anorthosite Complex*, one of the largest (~800 km^2) among Archaean anorthosites. The age of intrusion is constrained by the time span of 2.68–2.62 Ga, while granulite facies metamorphism occurred between 2.66 and 2.64 Ga [*Percival, Krogh, 1983; Krogh, 1993; Ashwal, Myers, 1994*]. It is noteworthy that metamorphism of the low-crustal rocks took place soon after formation of the granite-greenstone association at the upper crustal level (2.73–2.67 Ga) [*Corfu, 1987; Corfu, Ayres, 1991; Davis, 2002; Moser et al., 2008*]. Sedimentary rocks dated at 2.48 Ga unconformably overlie Superior Province granites, indicating that erosion developed prior to ~2.5 Ga.

According to *Wyman* and *Kerrich* [*2010*], the rapid growth of crust of the Abitibi–Wawa Orogen ~2.7 Ga ago was related to the interplay between plate-tectonics and mantle plumes. A low to moderate pressure recorded in the high-grade complexes and isobaric cooling are not consistent with tectonic thickening of crust during continent–continent or arc–continent collision; an external heat source is required to produce these high-grade conditions [*Heaman et al., 2011*].

The above data allow us to interpret the Neoarchaean evolution of the Superior Craton in terms of interacting plume- and plate-tectonic processes initiated by a global superplume ~2.7 Ga ago, as inferred for other continental domains in the same time period.

3.5.4. Northern Baffin Island

The *Mary River Group (Committee belt)*, 2.76–2.72 Ga in age, is represented by a sedimentary basin-fill, which underwent high-grade metamorphism up to granulite-facies conditions. The group consists of the lower sequence of metamorphosed bimodal metavolcanic rocks, including spinifex-bearing komatiite, and the upper sequence of pelite–graywacke turbidite affinity. The stratigraphic positions of quartzite, conglomerate, Algoma-type iron formation (one of the highest-grade primary and secondary iron deposits in the world), and minor marble vary on a regional scale. Mafic and ultramafic intrusions are abundant, whereas anorthosites occur locally. Deposition of the group is related to rifting in the large, unstable, and volcanically active basin [*Jackson, Berman, 2000*, and references therein].

3.5.5. Superplume-related evolution beyond Lauroscandia

As mentioned above, the *Witwatersrand epicontinental Basin* was formed between 3.07 and 2.71 Ga ago on the Kaapvaal Craton. The final pulse was expressed in formation of the Ventersdorp bimodal volcanic association (felsic and mafic lavas, pyroclastic rocks, and subordinate basal komatiites dated at 2.71 Ga). These rocks overlie the Witwatersrand sequence. It is suggested that Ventersdorp volcanics were erupted during a period of extensional tectonic activity [*Robb, Meyer, 1995*].

The *Limpopo granulite-gneiss belt.* The geological map [*Brandl, 1992*] shows that the Limpopo Belt is a synform with southwestern centroclinal closure (Fig. 4). The early history of the *Central Zone* of the Limpopo belt included the ~3.14 Ga granulite-facies metamorphic event (~3.14 Ga). Subsequent metamorphism of the rocks in all the three zones of the Limpopo Belt comprised high-grade events between 2.74 and 2.57 Ga and between 2.03 and 1.97 Ga [*Rigby, 2009; Blenkinsop, 2011; Laurent et al., 2011*].

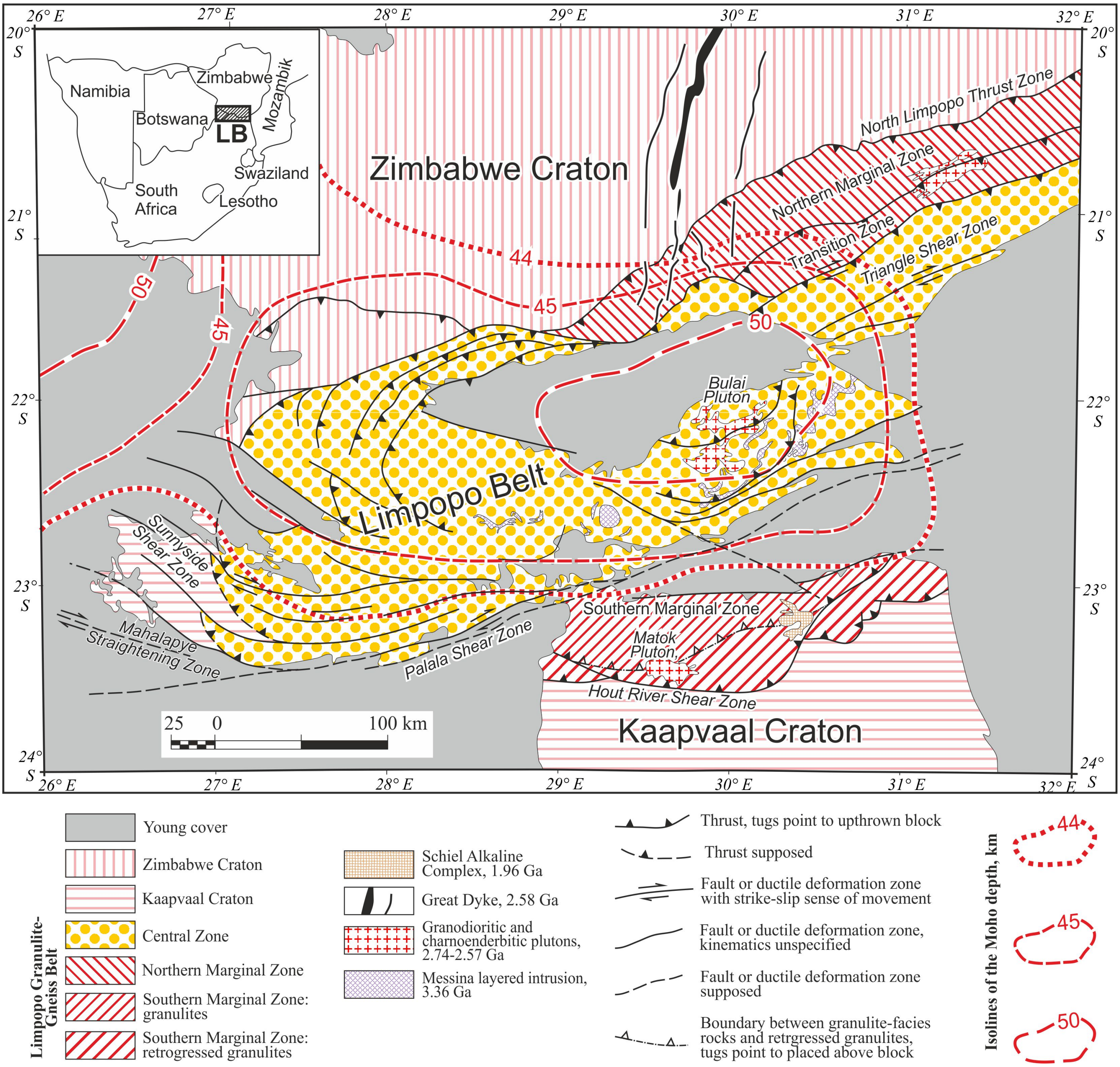

Fig. 4. The Limpopo granulite-gneiss belt, South Africa. Based on *Brandl* [*1992*], *Gore et al.* [*2009*], *Boshoff et al.* [*2006*].

Рис. 4. Гранулито-гнейсовый пояс Лимпопо, Южная Африка. По данным *Brandl* [*1992*], *Gore et al.* [*2009*], *Boshoff et al.* [*2006*].

The origin, evolution, and the thrust-controlled exhumation of the Southern Marginal Zone of the Limpopo Belt are factors related to the Neoarchaean high-grade event that commenced with crustal thickening ca. 2.72 Ga ago and ended at ca. 2.62 Ga [*Van Reenen et al., 2011b*]. Plutons of igneous enderbite and charnoenderbite were intruded synchronously with a peak of metamorphism in the Central and Northern Marginal zones [*Barton, Van Reenen, 1992; Blenkinsop, 2011*]. The Neoarchaean (2.61–2.58 Ga) Bulai pluton in the Central Zone is made up of large bodies of porphyritic granodiorites with subordinate inclusions and monzodiorite, enderbite, and granite dikes, as well as xenoliths of strongly deformed metamorphic country rocks, which underwent a high-grade metamorphic overprint at peak *P-T* conditions

(830–860 °C and 8–9 kbar) 2.64 Ga ago, followed by a decrease in temperature to 750 °C and pressure to 5–6 kbar. Later on, the pluton underwent a high-grade thermal overprint at 2.03–1.97 Ga [*Laurent et al., 2011*].

According to the currently prevailing understanding, the Limpopo Belt corresponds to a complexly built Himalayan-type orogenic belt and was formed as a result of continental collision between two Archaean blocks, the Zimbabwe Craton at the north and the Kaapvaal Craton at the south [*Van Reenen et al., 1992b; Laurent et al., 2011*]. The epicontinental metasediments and emplacement of the layered gabbroanorthosite intrusion indicate intracontinental development. In addition, it seems impossible to link the long metamorphic history of the Limpopo Belt that also contains evidence for a prominent succession of high-grade repetitions with a single collisional event, as supported by a body of more recent literature [*Boshoff et al., 2006; Zeh et al., 2007: Gerdes, Zeh, 2009; van Reenen et al., 2008; Millonig et al., 2008; Perchuk et al., 2008; Rigby et al., 2008; Rigby, 2009; Rigby, Armstrong, 2011*]. In our view, the time of tectonic compression, extrusion of metavolcanic–metasedimentary fill of the palaeobasin, and formation of synformal thrust-nappe ensemble can be most reliably constrained by intersection of the northern synform limb by the Great Dyke. This event took place earlier than 2.575 Ga [*Oberthür et al., 2002*], most likely immediately after the granulite-facies event dated at 2.64 Ga.

Charnockites and garnet-biotite gneisses of the *Central domain in the Mozambique granulite-gneiss belt*, Tanzania were formed ~2.7 Ga ago, then metamorphosed ~2.6 Ga ago, 50–100 m.y. after their emplacement. These rocks underwent isobaric cooling to 800–850 °C at 12–14 kbar soon after the post-peak metamorphic stage [*Johnson et al., 2003*].

The above-mentioned *Mt. Narryer metasediments*, Western Australia, experienced at least two metamorphic events, one ~3.3–3.2 Ga ago and another at 2.7–2.6 Ga, when the peak granulite-facies metamorphism was accompanied by extensive recrystallization. The Archaean rocks of *the Gawler Craton*, Australia, are mainly high-grade metamorphic gneisses [*Parker, 1990*]. Extensive felsic orthogneiss (~2.65 Ga) and mafic granofels are interpreted as synorogenic intrusions that cut through granulite-facies paragneiss sequences that include BIF, calc-silicate rocks and migmatites [*Daly et al., 1998*].

The Neoarchaean (2.7–2.5 Ga) was also an important period of crustal growth in the *North China Craton*. During this period, a large amount of mantle materials was added to form the juvenile crust: tonalite–trondhjemite–granodiorite gneisses, potassic granite and sedimentary–volcanic rocks. The Neoarchaean magmatism and contemporaneous metamorphism were interpreted as related to a mantle plume [*Diwu et al., 2013*].

Thus, the geological processes from ~2.79 to ~2.52 Ga are almost universally characterized by high heat influx from the mantle, extension and subsidence associated with activity of a mantle plume. The only notable exception is formation of the Superior Craton, where the plate-tectonic and plume-related processes were closely interrelated in space and time.

3.6. Superplume and failed rifting of the Archaean continental crust (2.53–2.30 Ga)

3.6.1. Evidence of initial rifting

The *initial rifting in the Kola-Karelia region* was marked by emplacement of mantle-derived magma and formation of mafic dikes, layered mafic-ultramafic and gabbroanorthosite intrusions (Fig. 5, *a*). Ages of the layered peridotite–gabbronorite intrusions localized in the upper crust of the Kola Peninsula and Northern Karelia are estimated at 2.53–2.49 and 2.44–2.43 Ga. The gabbroanorthosite bodies in the lower and middle crust were formed at virtually the same time (2.51–2.43 Ga) [*Balagansky et al., 2001; Mints, 2007* and references therein]. The intrusions were accompanied by intense volcanism. The earliest manifestations of mafic–ultramafic volcanic rocks in association with acid volcanic breccia and arkoses are noted at 2.53–2.51 Ga [*Mutanen, Huhma, 2001; Mutanen, 1997*]. Formation of the contrasting volcanic series (trachyandesite–basalts, komatiitic basalts, rhyolitic ignimbrites) followed at 2.45–2.42 Ga, and locally up to 2.32 Ga. 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. High-Mg basalts (basaltic komatiites) are less abundant. The section is completed with dacite and rhyolite dated at 2.44–2.42 Ga and 2.32 Ga subaerial andesite [*Melezhik, Sturt, 1994; Smolkin et al., 1995; Amelin et al., 1995; Mints et al., 1996; Puchtel et al., 1996; Melezhik et al., 2012*]. The geochemistry of the mafic volcanic rocks indicates relations to plume or T-MORB types. The above-mentioned magmatic bodies make up the Early Palaeoproterozoic LIP in eastern Fennoscandia [*Sharkov, Smolkin, 1997*].

The initial rifting in the framework of the Superior Craton is fixed by the Huronian Supergroup that unconformably overlies the Archaean basement. The lowermost portion of the Huronian section comprising plateau basalts, felsic lavas, and arkosic metasediments, is cut through by gabbroanorthosite, mafic–ultramafic rocks, and granitoid intrusions of 2.49–2.46 Ga age, as well as by mafic dikes of the Matachewan swarm

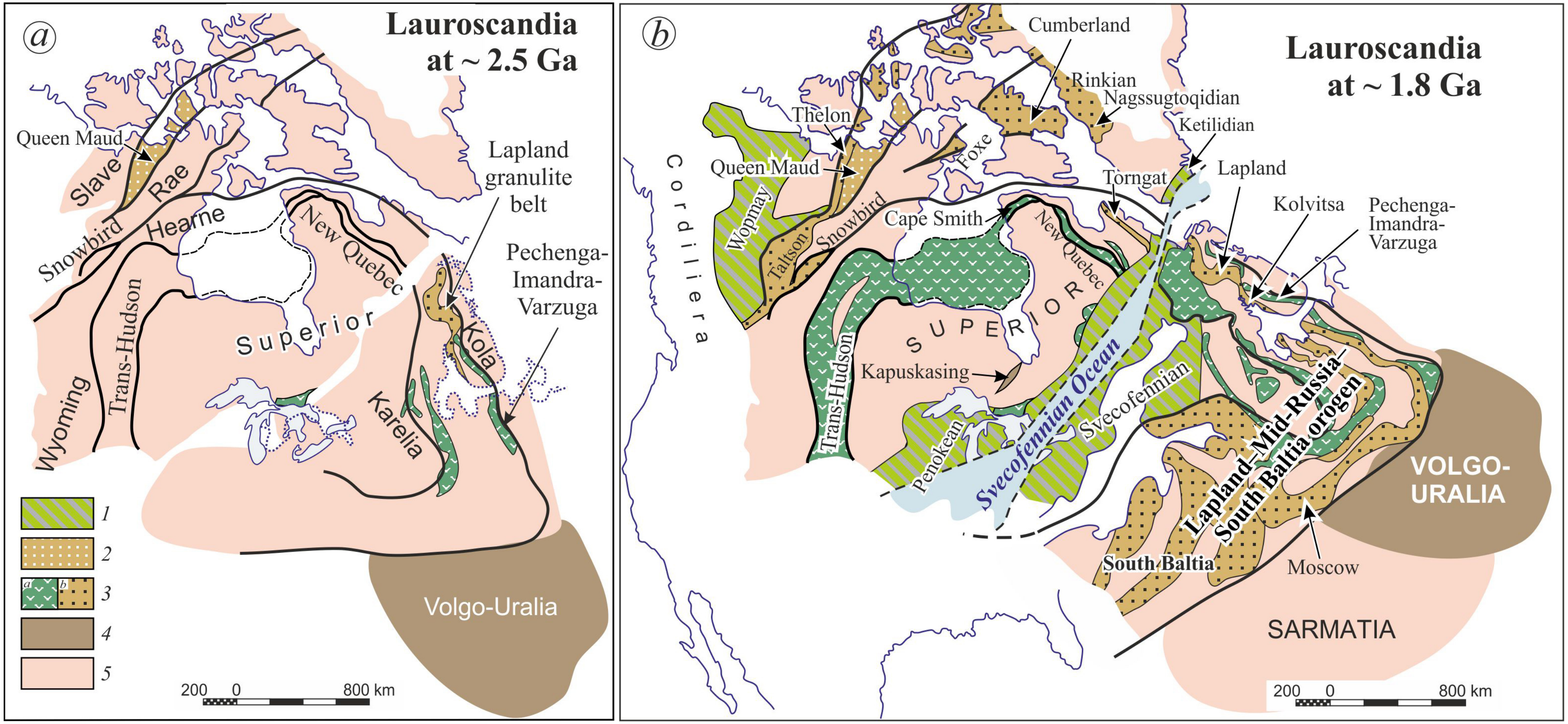

Fig. 5. Reconstruction of the Palaeoproterozoic evolution of Lauroscandia: Lauro-Russian intracontinental oval orogen: *a* – at ~2.5 Ga, *b* – at ~1.8 Ga.

1–3 – Early Palaeoproterozoic: *1* – accretionary complexes, *2* – granulite-gneiss complex of Queen Maud Block, *3* – rocks marking the evolutionary onset of the intracontinental collisional orogen: *a* – volcanic-sedimentary belts, *b* – granulite-gneiss belts; *4–5* – Archaean (3.3–2.5 Ga): *4* – Volgo-Uralia granulite-gneiss area, *5* – granite–greenstone terranes with inclusions of Archaean complexes partially reworked in late Palaeoproterozoic. Names of Archaean tectonic units are shown in Arial typeface, Palaeoproterozoic units appear in Times New Roman.

Рис. 5. Реконструкция палеопротерозойской эволюции Лавроскандии: Лавро-Русский внутриконтинентальный овальный ороген: *a* – ~2.5 млрд лет, *b* – ~1.8 млрд лет.

1–3 – ранний палеопротерозой: *1* – аккреционные комплексы, *2* – гранулито-гнейсовый комплекс блока Куин-Мод, *3* – породы, маркирующие начало эволюции внутриконтинентального коллизионного орогена: *a* – вулканогенно-осадочные пояса, *b* – гранулито-гнейсовые пояса; *4–5* – архей (3.3–2.5 млрд лет): *4* – Волго-Уральский гранулито-гнейсовый ареал, *5* – гранит-зеленокаменные области с включениями архейских комплексов, частично переработанных в палеопротерозое. Названия архейских тектонических подразделений даны шрифтом Arial, палеопротерозойских подразделений – шрифтом Times New Roman.

of the southern Superior Province, which is dated at 2.47–2.45 Ga. The upper part of the Huronian section is composed of tillite, shale, quartzite, and carbonate rocks and crowned by cross-bedded sandstone. *Ernst* and *Bleeker* [*2010*] argue that the Huronian Supergroup is not related to breakup of the southern Superior Craton and therefore represents an epicratonic basin, rather than a passive margin.

The tonalite gneiss of the Sask Craton (~3.1.Ga) in the *axial part of the Trans-Hudson Orogen* underwent high-temperature melting, recrystallization and was intruded by charnockite melts 2.53–2.47 Ga ago [*Bickford et al., 2005*].

All these events can be viewed as evidence for initial rifting associated with insignificant mantle-plume activity, which resulted in further separation of the Superior and Hearne cratons during the Palaeoproterozoic.

3.6.2. The high-grade metamorphic event

The origin of mafic rocks and the first manifestation of granulite-facies metamorphism (860–960 °C and 10.3–14.0 kbar) at the base of the *Lapland-Kolvitsa granulite belt in the Kola Peninsula* (Fig. 5, *a*) are recorded immediately after gabbroanorthosite intrusions (M0 event at 2.46–2.43 Ga) [*Mints et al., 2007*, and references therein; *Balagansky et al., 2001; Fonarev, Konilov, 2005*]. The deepest granulites are retained at the base of the crust. Fragments of this crust are available as the deep-seated xenoliths captured by the Devonian lamprophyre dikes and pipes outcropping along the Kandalaksha Bay coast. Xenoliths are predominantly composed of the metanorite–gabbroanorthosite rocks (garnet granulites) dated at 2.47 Ga and comparable to the same rocks from the Lapland granulite complex, which underwent the earliest metamorphism at 2.41 Ga [*Kempton et al., 2001; Downes et al., 2002; Vetrin et al., 2009*]. The *Topozero charnockites* intruded mid-crustal rocks 2.45–2.43 Ga ago [*Mints et al., 2015c,* and references therein].

High-temperature processes within the North American craton dated at 2.6–2.5 Ga are known only in a few places along the boundaries of the Trans-Hudson Orogen [*Rayner et al., 2005; Dahl et al., 2006*].

In *Northern Baffin Island,* a ~2.55–2.43 Ga tectonothermal event which affected much of the Archaean *Committee Belt* is marked by the Bylot charnockite batholith in northwestern Baffin Island and the orthogneisses on southern Devon Island [*Jackson, Berman, 2000*].

The Proterozoic tectonic history of the *Queen Maud Block of Arctic Canad*a includes: (1) widespread magmatism derived from Neoarchaean source rocks 2.50–2.46 Ga ago, (2) the origin of a sedimentary belt dated at 2.44–2.39 Ga and containing detritus 2.50–2.45 Ga in age, and (3) ~2.39 Ga regional granulite-facies metamorphism [*Schultz et al., 2007*].

3.6.3. Superplume-related evolution beyond Lauroscandia

Late Neoarchaean–early Palaeoproterozoic rifting of *the Zimbabwe craton (southern Africa)* is marked by the Great Dike, 550 km long and up to 11 km wide, which intruded 2.575 Ga ago [*Oberthür et al., 2002*]. The age of satellite dikes falls within the 2.51–2.41 Ga range [*Söderlund et al., 2010*]. Rocks that underwent HT and UHT granulite-facies metamorphism in the early Palaeoproterozoic 2.56–2.42 Ga ago are known from cratons in *India* (*Nilgiri Complex*)*, Antarctica* (*Napier Complex*), and elsewhere [*Raith et al., 1990, 1999; Harley, 1998; Konilov et al., 2001; Asami et al., 2002; Yang et al., 2015*]. At the passive margins of the *Pilbara Craton in Australia* and *the Kaapvaal Craton in South Africa*, the time interval from 2.51 to 2.45 Ga was characterized by a peak of metamorphism and the origin of BIFs under extension related to the impact of a global plume [*Barley et al., 2005*].

The Palaeoproterozoic evolution of *the North China Craton* has been thoroughly studied in recent years. In almost all publications, the Palaeoproterozoic history of the North China Craton is interpreted in terms of the plate-tectonic model (subduction and orogeny), which is based on geochemistry of rocks, as well as on direct and unequivocal relations between granulite-facies metamorphism and collisional swelling of the crust. The endogenic events (TTG magmatism; emplacement of granite, charnockite, and gabbro; formation of protolith for mafic granulites; metamorphism under conditions of greenschist to granulite facies) dated back to 2.57–2.35 Ga have been documented in the North China Craton, although geodynamic interpretation of these events remains a matter of debate. According to the predominant viewpoint, the Neoarchaean evolution was completed by amalgamation of the eastern and western blocks of the Archaean crust and by formation of the Trans-North China Orogen between these blocks [*Zhao et al., 2005; Kusky, Li, 2003; Guo et al., 2005; Kröner et al., 2005; Yang, Santosh, 2015; Shan et al., 2015*].

The reappraisal of the geodynamic setting of granulite-facies metamorphism and its role in formation of granulite-gneiss belts makes it possible to reconstruct the main processes and settings in the Palaeoproterozoic history of this craton in a new model. The tectonothermal events from 2.57 to 2.35 Ga mark the onset of Palaeoproterozoic evolution in the North China Craton, which is surprisingly close to coeval evolution of the East European and North American cratons in respect to succession and types of magmatic activity, sedimentation, and metamorphism, as well as to synchronization of these events in all three cratons. The more definite pattern has been drawn for the middle and late Palaeoproterozoic events, which are considered below.

Thus, taking a broad view, the character of magmatism, high-grade metamorphism, and sedimentation from ~2.58 to ~2.32 Ga indicates extensional conditions and an influx of mantle heat to the inner domain of the rather large continent. The proximity in time of the processes proceeding in distant areas shows that these areas belonged to a single and the same continent (probably, a supercontinent that was much larger then Lauroscandia). The emplacement of mantle-derived magma in combination with high-grade metamorphism allows suggesting that these processes were related to the mantle-plume phenomena. The widespread occurrence of these processes is consistent with the idea of a superplume.

3.7. Quiescent within-plate activity and diffuse rifting (2.3–2.1 Ga)

The time from ~2.3 to 2.1 Ga is characterized by limited to moderate tectonic and magmatic activity in Lauroscandia, somewhat younger in comparison with the global assessment of such conditions proposed at 2.45–2.2 Ga [*Condie et al., 2009b*]. In the riftogenic belts of the *Kola Peninsula,* the alkali basalt and trachybasalt in association with alkali picrite and rhyolite dated at ~2.2 Ga are intercalated with epicontinental red beds, stromatolitic limestone and dolomite, phosphate- and manganese-bearing interlayers. Most areas of the *Karelia Craton,* 2.3 to 2.1 Ga in age, were submerged below sea level and were covered by shelf-type sediments. A new stage of rifting is marked by emplacement of metadolerite sills, 2.2–1.97 Ga ago [*Vuollo, 1994*].

Rifting of the North American Craton is recorded in the 2.22–1.99 Ga age range by emplacement of dike swarms [*Ernst, Buchan, 2001* and references therein]. *Ernst* and *Bleeker* [*2010*] used formation of the LIP for more precise definition of the margins that experienced aborted breakup.

Rifting of the Superior Craton took place between 2.23 and 2.03 Ga. Riftogenic volcanics similar to the above-mentioned rocks in Kola and Karelia are known at the passive margin, but complicated by rifting. Fragments of these sequences are retained in the New Québec and Trans-Hudson belts [*Hoffman, 1989*].

Pre-2.1 Ga rifting of the Hearne Province is marked by a succession of mafic flows intercalated with sandstone and overlain by conglomerate of the Spi Group. The diabase dikes of the Kaminak swarm are probably related to the mafic flows. Recognition of widespread rifting in this part of the Hearne Province constrains tectonic modeling of the pre-orogenic history of the Trans-Hudson Orogen [*Patterson, 1991*].

3.8. The second prominent superplume event recognizable worldwide, and plume-driven local plate-tectonic processes (2.2–1.8 Ga)

3.8.1. Rifting and partial disruption of Lauroscandia

Rifting of the Kola-Karelia continent. The renewal of tectonic activity in the middle Palaeoproterozoic is recorded by powerful magmatism, which was partly similar to the early Palaeoproterozoic initial magmatism dated at 2.53–2.40 Ga. In contrast to the previous event, the recurrent initial magmatism over the time interval from 2.1 to 1.92 Ga, and possibly up to 1.88 Ga, was localized in the long separate linear structures of the Pechenga–Imandra–Varzuga, North and East Karelian and Kalevala–Onega belts. Magmatism was associated with interaction between two sources related to the mantle plume and the continental sialic crust. In comparison with the similar event of the early Palaeoproterozoic, the role of the crustal source declined markedly. As in the first case, the area of initial magmatism corresponds to the LIP [*Ernst et al., 2004*]. The igneous assemblages of this younger period are as follows: (1) mafic volcanic rocks of the continental rift type and, in some cases, of oceanic type combined with bimodal volcanic rocks of the rhyolite–picrite series close to the geochemical type of oceanic island magmas, subvolcanic gabbro-wehrlite intrusions and dikes (2.11–1.92 Ga), (2) sedimentary–volcanic complex of the Onega depression (~1.98 Ga), (3) Kimozero kimberlites (~2.0 Ga), (4) gabbroanorthosite intrusions (2.0–1.95 to 1.88 Ga), (5) mafic–ultramafic massifs (1.98 Ga), (6) alkaline intrusions (1.97–1.95 to 1.88 Ga), and (7) granitoid massifs (about 1.95 Ga) [*Mints, 2007; Mints et al., 2015c*, and references therein].

The intrusive processes were associated with reactivation of the rapidly downwarping intracontinental basins, which were rapidly filled with sedimentary, volcanic, and pyroclastic materials of predominantly juvenile origin. As before, these rocks were then affected by granulite-facies metamorphism.

Evidence for *rifting of the Superior Craton* is supported by the dike swarms. The Maguire, Senneterre, and Klotz dikes form a roughly radiating pattern and may represent a quadrant of the giant radiating dike swarm centred southeast of Ungava Bay, whose focus marks the location of a mantle plume responsible for ~2.22 Ga breakup along the eastern margin of the Superior Province [*Buchan et al., 1998*], that might have initiated opening of the Svecofennian Ocean. The dikes of 2.09 Ga age from the Cauchon Lake area close to the western margin of the Superior Craton probably represent a continental rifting episode that preceded opening of the Manikewan Ocean in the future site of the Trans-Hudson Orogen [*Halls, Heaman, 2000*].

Rifting that transformed locally to oceanic spreading in the regions of the future collisional orogens (*2.1–1.9 Ga*). Spreading-type events are recorded in the fragmented ophiolitic sections of Purtuniq (2.0 Ga) [*Scott et al., 1991*] and Jormua (1.95 Ga) [*Kontinen, 1987; Peltonen et al., 1996, 1998*] (the Cape Smith Belt in 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 (2.00 Ga) [*Hanski et al., 1998*]. The T-MORB-type tholeiitic pillow lavas in the Pechenga Structure (2.11–1.92 Ga) are combined with picritic lava flows and felsic ash flows that geochemically resemble the products of volcanic eruptions on oceanic islands. Subvolcanic gabbro–wehrlite bodies related to the activity of these volcanoes intruded the lower portion of the tholeiitic sequence, but are more abundant as lenticular tectonic boudins in the imbricate complex of sulphidated terrigenous rocks [*Melezhik, Sturt, 1994; Mints et al., 1996; Sharkov, Smolkin, 1997; Skuf'in, Theart, 2005; Melezhik et al., 2012*]. This complex, known as a productive layer of the Pechenga Cu–Ni ore field, was interpreted as a fragment of the accretionary prism [*Mints et al., 1996*]. The youngest (1.87–1.86 Ga) rock complexes of the Pechenga–Varzuga Belt include 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 [*Melezhik, Sturt, 1994; Sharkov, Smolkin, 1997*].

The most significant disruptions of the Lauroscandian crust are marked by the Svecofennian and Trans-Hudson Orogens. Rifting and dike injections at ~2.2 Ga initiated the Svecofennian Ocean opening and the origin of the Svecofennian passive margin of Kola–Karelia 1.99–1.91 Ga ago. The lower part of the passive margin succession is composed of quartzite, quartzitic sandstone, conglomerate, metavolcanics, dolomite and calcite marbles, with subordinate sulphidated black shale [*Glebovitsky, 2005; Lahtinen et al., 2008*].

Starting with *Hoffman* [*1989*], it is usually suggested that the Trans-Hudson Orogen formed due to collision between two separate continental blocks. Palaeomagnetic studies testify to significant displacements of the boundaries of the Trans-Hudson Orogen [*Gala et al., 1998; Halls, Heaman, 2000; Symons, Harris, 2000*]. The Manikewan Ocean separated fragments of the Archaean continent in the area of the future Trans-Hudson Orogen, and could have reached a width of ~4000 km at ~1.84 Ga, as supported by a large body of palaeomagnetic data [*Gala et al., 1998*]. However, the above-mentioned evidence of Neoarchaean-Palaeoproterozoic (2.6–2.5 Ga) activization, which was clearly confined to the periphery and the inner region (Sask Craton) of the Trans-Hudson Orogen in its present-day configuretion, as well as a strict parallelism of the orogen borders, and some other features, which we discuss below, allow us to postulate that the Trans-Hudson Orogen records the location of disruption of initially a single continent and its subsequent recovery. The maximum extension of the continental crust, passing into the oceanic floor spreading, was possibly confined to the marginal zones of the orogen [*Baird et al., 1996; White et al., 2005; Eaton, Darbyshire, 2010*].

3.8.2. Plume-related evolution beyond Lauroscandia

The Bushveld (Layered) Complex in South Africa contains the world's largest PGE, Cr, and V reserves. The plutonic rocks were emplaced and cooled over within less than one million years, between 2.056 and 2.055 Ga [*Zeh et al., 2015*]. Formation of the Bushveld Complex can be regarded as evidence for local but vigorous mantle-plume activity.

In the *Trans-North China Orogen located in the axial zone of the North China Craton*, numerous mafic–ultramafic bodies cut through the Neoarchaean Wutai granite-greenstone complex. Magmatic zircons from gabbro are dated at ~2.19 Ga [*Wang et al., 2010*].

3.9. Interaction of plate- and plume-tectonics and formation of intercontinental collisional orogens (2.0–1.7 Ga)

The recently identified Palaeoproterozoic *Lauro-Russian intracontinental oval orogen* is subdivided into two approximately equal halves, one of which is situated in Eastern Europe (the Lapland–Mid-Russia–South Baltia orogen) and the second one encompasses almost the entire territory of North America (Trans-Hudson, Taltson-Thelon, Penokean orogens, etc.) [*Mints, 2014, 2015c*].

3.9.1. Lapland–Mid-Russia–South Baltia Orogen

The geochronology of sedimentary and volcanic protoliths that were subsequently transformed into the *Lapland–Kolvitsa granulite-gneiss belt* (Fig. 5, *b*) shows that a vast sedimentary basin was formed at ~2.0 Ga. Volcanism and sedimentation were completed at ~1.9 Ga [*Balagansky et al., 1998; Bridgwater et al., 2001; Daly et al., 2001; Huhma, Meriläinen, 1991; Sorjonen-Ward et al., 1994*]. The morphology of detrital zircon grains shows that granite–granodiorite provenance is most probable, although corresponding plutons in the vicinity of the granulite belt are not known. The granulite facies metamorphic reworking of the volcanic-sedimentary succession started almost contemporarily with its deposition and lasted approximately 100 m.y. The first metamorphic event, M1 at 2.0–1.95 Ga [*Mitrofanov, Nerovich, 2003; Mints et al., 2007; Kaulina, 2010*] was characterized by *PT* conditions of 860–960 °C and 10.3–14.0 kbar. The M2 *PT* parameters ranged from 860 to 800 °C and from 12.4 to 5.8 kbar. The M3 event was characterized by *PT* parameters between 770 °C at 10.7 kbar and 640 °C at 4.8 kbar. The high-temperature metamorphism corresponding to M2 and M3 events affected both igneous and sedimentary rocks at ~1.95–1.92 and 1.92–1.90 Ga, respectively [*Bibikova et al., 1993; Daly et al., 2001, 2006; Mints et al., 2007; Kaulina et al., 2004*]. The collision-related transport of hot tectonic sheets composed of granulite complexes to the erosion level at 1.89–1.87 Ga [*Bibikova et al., 1993; Daly et al., 2001; Kaulina, 2010; Tuisku, Huhma, 2006; Tuisku et al., 2006*] was the immediate cause of the last metamorphic event, M4 (650–550 °C and 8.4–4.5 kbar). As noted above, the deepest granulites are retained in the lower crust and 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 norite–gabbroanorthosite composition similar to the rocks of the Lapland granulite complex [*Vetrin et al., 2009* and references therein]. The parameters of metamorphism in the composite crustal section that comprises granulites from the Lapland Belt and from the xenoliths determine its total crustal thickness as approximately 70 km. This crust was affected by granulite-facies metamorphism within a depth interval from 70 to ~25 km [*Mints et al., 2007*].

Geophysical data make it possible to trace the Lapland Belt to the East European Platform beneath sedimentary cover. In addition, the system of poorly studied granulite-gneiss belts surrounds the Karelian Craton in the east and the south [*Mints, 2015b; Mints et al., 2015c*]. The leuconorite and enderbite of the Moscow granulite belt were emplaced at 1.98 Ga; ultrahigh-temperature granulite-facies metamorphism (~1000 °C and 10–12 kbar [*Bogdanova et al., 1999; Mints et al., 2015c*]) took place soon after intrusion.

The basement of the East European Platform in the South Baltia region (Fig. 5, *b*) has previously been regarded as a part of the Svecofennian accretionary orogen [*Gorbatschev, Bogdanova, 1993; Bogdanova et al., 2005*]. In contrast to 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 rocks pertaining to greenschist, epidote-amphibolite, and high amphibolite–granulite facies of metamorphism. The volcanic and intrusive complexes within high-grade areas were formed from 2.1 to 1.8 Ga, somewhat earlier than in the neighboring Svecofennian Orogen, and become younger westward. Granulite-facies metamorphism at temperatures up to 900 °C and pressures of 8–10 kbar [*Skridlaite, Motuza, 2001; Taran, Bogdanova, 2001*] was related to the events dated at 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) was probably accompanied by anorogenic magmatism [*Baltybaev et al., 2004; Glebovitsky, 2005; Mints, 2007*]. Terrigenous protoliths of granulite associations contain detrital zircon dated at 2.45–1.98 Ga; an insignificant admixture of Archaean populations was noted [*Claesson et al., 2001*].

The formation of the *Tiksheozero alkaline mafic–ultramafic complex with carbonatites* in northern Karelia 1.85 Ga ago [*Belyatsky et al., 2000*] provides strong support for the plume-related nature of the late Palaeoproterozoic processes in the EEC.

3.9.2. Intracontinental orogens within the North American Craton

The *Taltson–Thelon Orogen* that extends in a submeridional direction across the North American Craton (Fig. 5, *b*) is similar in many respects to the Lapland granulite-gneiss belt. This 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. Earlier, it was suggested that the genesis of the Taltson Magmatic Zone was initiated by the oceanic crust subduction beneath the margin of the Archaean Churchill Province 1.99–1.96 Ga ago, and that it was completed with collision between this continent and the 2.4–2.0 Ga Buffalo Head Terrane [*Hoffman, 1989; McDonough et al., 1995; Ross, Eaton, 2002*]. The subsequent geochemical study has shown that granitoids of both I- and S-types are the products of intracrustal melting [*De et al., 2000*]. These data, along with high-grade metamorphism, indicate that the rocks of the Taltson Magmatic Zone were formed under intraplate conditions [*Chacko et al., 1994, 2000; Farquar et al., 1996*] and probably with participation of a plume-type heat source.

The ~1.9 Ga granulite-gneisses and high-temperature eclogites known within the *Snowbird tectonic zone* are considered to be a branch of the Taltson–Thelon Orogen [*Baldwin et al., 2004*]. Earlier, these high-T and high-P mylonitic quartz–feldspar gneisses and gneiss-hosted metabasic rocks were suggested to be Archaean [*Snoeyenbos et al., 1995*]. The high-T eclogites are emplaced at the bottom of the granulite-gneiss tectono-stratigraphic section of the oval Athabasca Synform. Their metamorphic *PT* parameters of 920–1000 °C at a minimum pressure of 18–20 kbar exceed the corresponding parameters in associated mafic granulites (890–960 °C and 13–19 kbar). The metamorphic processes proceeded in the inland environment; hence, the Snowbird zone is not a palaeosuture [*Baldwin et al., 2003, 2004*].

The *northern extension of the Thelon Orogen* (*the Boothia Peninsula – Somerset Island and Devon–Ellesmere Island terranes*) comprises 1.9 Ga complexes of orthogneisses, graphite-bearing paragneisses, marbles, and BIF with lenses of mafic and ultramafic granulites of similar metamorphic grade (740–900 °C and 6–8 kbar) [*Kitsul et al., 2000; Frisch, Herd, 2010*].

The origin of the *branching Palaeoproterozoic belts* (*Nagssugtoqidian, Rinkian, Foxe, Torngat* in Arctic Canada and Greenland [*Hoffman, 1989*]) and orthogneiss complexes in the structurally interrelated tectonic units (the Lake Harbour Group, the Narsajuaq Arc, and the Ramsay River [*St-Onge et al., 1999*]) may be interpreted in the same way. In these belts, the granulite-facies gneisses metamorphosed at temperatures up to 950 °C and pressures of ~4 to ~12 kbar. The protoliths of granulites in the lower part of the sections mainly consisted of metasedimentary rocks of platform- and rift types with the participation of evaporites, mafic and ultramafic volcanic rocks, anorthosite sills. The age of detrital zircons indicates that juvenile Palaeoproterozoic rocks of unknown origin dated at 2.4–1.93 Ga were a source of the sediments. A significant contribution of Archaean rocks is noted as well. Sedimentation started about 2.0 Ga ago and terminated at 1.95–1.93 Ga. The oldest manifestation of granulite-facies metamorphism is recorded in charnockite of the Sisimiut Complex (1.92–1.90 Ga), i.e., synchronously with the end of sedimentation or immediately after its cessation. The age of the main phase of granulite facies 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 [*Jackson, Berman, 2000; Kalsbeek, Nutman, 1996; Kalsbeek et al., 1998; Nutman et al., 1999; Scott, 1999; Kalsbeek, Taylor, 1999; Van Kranendonk, 1996; St-Onge et al., 2007*].

The core of the orogen is the huge, triangular *Cumberland Batholith* largely composed of orthopyroxene-bearing monzonite and charnockite plutons 1.87–1.85 (1.87–1.81) Ga in age. In many cases, similar batholiths have been regarded as continental arc type. The Cumberland Batholith includes rocks of arc, within-plate (A-type) and post-collisional geochemical affinity and volumetrically minor mafic rocks with both arc and non-arc features. Thus, the Cumberland Batholith likely encompasses various magmatic suites generated at deep- to mid-crustal depths. It is interpreted as a batholith resulting from large-scale lithospheric mantle delamination followed by the upwelling of hot asthenospheric mantle giving rise to voluminous crustal partial melting [*Whalen et al., 2010*].

The charnockite bodies within the *Dexterity granulite belt* (Northern Baffin Island) are considered to be ~1.82 Ga, that is the age of granulite-facies metamorphism (8.0–9.2 kbar, 710–850 °C) and deformation [*Jackson, Berman, 2000*, and references therein].

The *Kisseynew Domain within the Trans-Hudson Orogen* originated as a sedimentary basin filled with turbidities 1.86–1.84 Ga ago. Rocks of this domain were thrust over and structurally interbedded with 1.9 Ga island arc and ocean-floor assemblages during convergence of the Trans-Hudson Orogen and the Archaean Superior craton 1.84–1.81 Ga ago. These turbidites were interpreted to have been deposited in a back-arc basin behind a retreating subduction boundary [*Ansdell et al., 1995*]. The central Kisseynew Domain was metamorphosed at a uniform high grade (700–800 °C and 5–6 kbar). According to *Kraus* and *Menard* [*1997*], the low- to medium-pressure and high-temperature metamorphism in the Kisseynew Domain was related to heat advected by sheets of peraluminous granitic rocks. The mobilization of granitic magma in the lower crust of the Kisseynew Domain is explained by these authors by the high basal heat-flow resulting from convective thinning of the lithosphere due to thickening of the crust.

The *Trans-Hudson Orogen* as a whole is similar in scale and tectonic style to the modern Himalayan–Karakorum orogen. During collision, the lobate shape of the indentor (Superior Craton) formed an orogenic template, which along with the smaller Sask Craton exerted a persistent influence on the tectonic evolution of the region and ensured anomalous retention of juvenile Proterozoic crust [*Eaton, Darbyshire, 2010*].

3.9.3. Closure of the intracontinental ocean via subduction

Subduction-related rocks are widespread in the Trans-Hudson 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) [*Hoffman, 1989; Stern, Lukas, 1994*]. Boundary structures of the Trans-Hudson Orogen plunge beneath the framing structural units in both east and west. Fragments of pre-Palaeoproterozoic crust (Sask Craton) overlain by Palaeoproterozoic subduction-related complexes are retained beneath the axial zone of the Trans-Hudson Orogen [*Lucas et al., 1993; Baird et al., 1996; White et al., 2005*].

The history of the *arc-continent and continent-continent collisions along the northwestern margin of the Trans-Hudson Orogen* includes an accretion phase (1.92–1.84 Ga) and a collision phase (1.84–1.80 Ga). The collision zone between two large Archaean cratons (Hearne and Superior provinces) includes also a smaller Archaean to Palaeoproterozoic microcontinent (Sask Craton) and an orogenic internide (Reindeer

Zone), which consists of accreted juvenile arcs, oceanic plateaus, and associated sedimentary basins that were formed during the whole interval 1.92–1.80 Ga [*Corrigan et al., 2005*, and references therein].

According to *Chiarenzelli et al.* [*1998*], the Sask Craton is rifted from the Superior and/or Hearne Provinces rather than representing an exotic continental fragment. In such a case, the Trans-Hudson Orogen can be understood as an internal orogen.

3.9.4. Two types of accretionary orogens

First type. The orogenic area, which occupies the central part of the Fennoscandian shield, is usually considered as a classical *Svecofennian accretionary orogen* [*Windley, 1992*]. However, as shown above, the orogen was initiated by rifting and subsequent breakup of Lauroscandia. Thus, this orogen probably reunited the temporarily separated North American and East European fragments of Lauroscandia. The Svecofennian accretionary orogen is mainly 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 [*Gaál, Gorbatschev, 1987; Pharaoh, Brewer, 1990; Korsman et al., 1999; Lahtinen et al., 2005*]. The high-grade metamorphism (up to 800 °C at 4–5 kbar) of turbidites deposited in back-arc basins is dated at 1.89–1.81 Ga [*Hölttä, 1998; Korsman et al., 1999*]. The emplacement of early and late orogenic granitoids intensified at 1.90–1.87 and 1.83–1.77 Ga and was followed by minor post-orogenic granite intrusions [*Gaál, Gorbatschev, 1987; Lahtinen et al., 2005*; and references therein].

Second type. The *1.9 Ga Wopmay accretionary orogen* (Coronation geosyncline) along the western margin of the Slave craton was among the first Precambrian continental margins to be identified. It is the only known real accretionary orogen that has been formed along the outer margin of Lauroscandia. The Wopmay orogen includes the Great Bear magmatic arc (1.88–1.84 Ga), the Hottah arc terrane accreted to the Coronation margin at 1.88 Ga, and the Asiak foreland thrust-fold belt comprised of the rift, the passive margin and foredeep deposits deformed by the Hottah–Slave collision. The Vaillant basalt-dominated rift assemblage, which underlies the passive margin sequence, Lac de Gras dike swarm, and related Booth River intrusive suite were formed during the initial rift-forming event. The age of 2.01 Ga dates the rift-to-drift transition and 1.88 Ga is the age of arrival of the passive margin at the trench bordering the Hottah terrane [*Hoffman et al., 2011*, and references therein].

3.9.5. Superplume-related evolution beyond Lauroscandia

The tectonic and thermal events of this period (2.0–1.7 Ga), which are close to those described in Lauroscandia, have been established more or less distinctly for almost all continents. Here we will restrict our discussion to the most striking examples. It should be emphasized that the authors of those publications from which we have taken factual data, interpret these data in terms of plate-tectonic models of the subduction–collision type.

The North China Craton (Fig. 6) is composed of igneous and sedimentary rocks metamorphosed under conditions of greenschist to granulite, in particular,

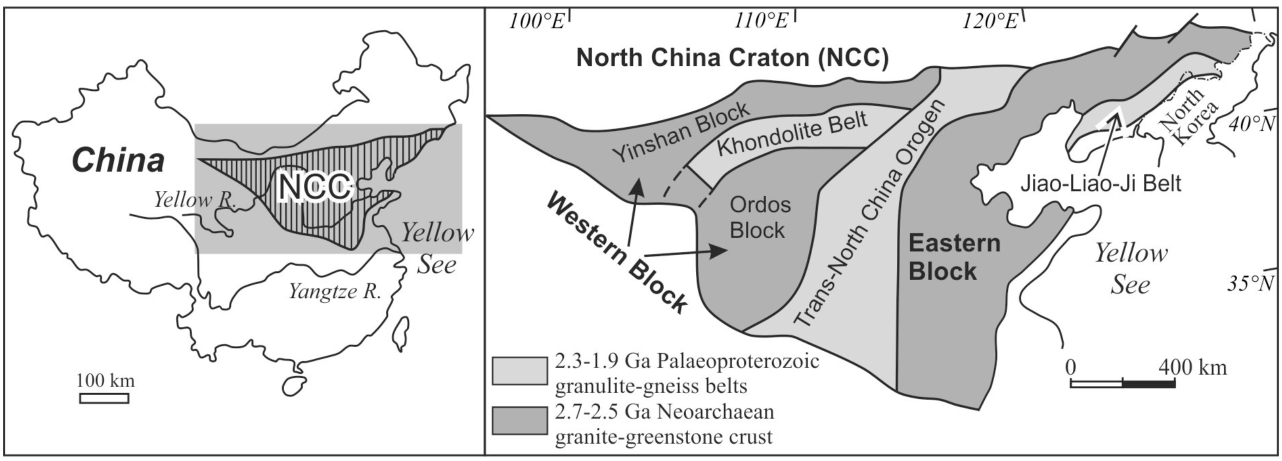

Fig. 6. North China Craton. Modified after *Guo et al.* [*2002*], *Zhao et al.* [*2005*].

Рис. 6. Северо-Китайский кратон. С изменениями по [*Guo et al., 2002; Zhao et al., 2005*].

high-pressure facies. The crustal architecture of this craton was constructed in the Palaeoproterozoic, as was deduced from collision of the major crustal blocks [*Zhao et al., 2005; Santosh et al., 2015*, and references therein]. According to these authors, the following tectonic events took place 2.1–1.8 Ga ago in the North China Craton: (1) creation of the Inner Mongolian Suture Zone between Yinshan and Ordos blocks also known as *the Khondalite Belt* to form the Western Block (1.95–1.90 Ga); (2) intracontinental rifting within the Eastern Block (2.2–1.9 Ga), forming the Jiao-Liao-Ji Belt, and (3) emergence of *the Trans-North China Orogen* (*TNCO*) also known as *the Central Orogenic Belt* (ca. 1.85–1.80 Ga). The TNCO is the N–S-trending collisional orogeny, 100–300 km wide and 1000 km long, running across the central part of the North China Craton.

The Khondalite Belt within the Inner Mongolia Suture Zone consists of khondalite series, TTG gneiss with minor mafic granulites, and syntectonic charnockites and S-type granites. It is important to note that the khondalite series is composed of graphite-bearing sillimanite–garnet gneiss, garnet quartzite, calc-silicate rock and marble, which were deposited in the stable continental environment. The available isotopic data furnish evidence that protoliths of the khondalite series were deposited in the Palaeoproterozoic 2.3 to 1.9 Ga ago; the age of metamorphism is estimated at 1.9–1.8 Ga [*Zhao et al., 2005,* and references therein]. Voluminous granites were emplaced at ca. 2.15–2.00 Ga into the sedimentary sequence. Subsequently, all rock units were intruded by S-type granitic plutons and gabbronorite intrusions and subjected to high- and ultrahigh-temperature and locally high-pressure granulite-facies metamorphism of 1.95–1.90 Ga age [*Guo et al., 2012; Zhang et al., 2014; Dan et al., 2014*]. The peak temperature of regional metamorphism of khondalites is 700–820 °C, whereas the peak temperature of sapphirine granulites locally occurring in the Khondalite Belt reaches 910–1000 °C at ~10 kbar (UHT conditions) [*Guo et al., 2012; Santosh et al., 2012*].

In the Trans-North China Orogen, at least some of the protolith rocks of granite and sillimanite gneisses from the Fuping Complex formed 2.1–2.0 Ga ago [*Zhao et al., 2005; Santosh et al., 2015*]. The Nd isotopic data suggest that significant amounts of basement rocks older than 2.6 Ga are lacking in the Central Zone, although minor amounts of rocks dated at 2.7 Ga probably existed in local areas, representing fragments of the reworked Archaean crust. The extensional setting was documented by injection of mafic dikes, which are known from the Hengshan Complex in the central part of the Trans-North China Orogen, ~1.92 Ga ago. Metamorphic zircons are dated at 1.89–1.85 Ga. The youngest gneissic granite has an age of 1.87 Ga [*Kröner et al., 2006*]. These dikes now occur as boudins and deformed sheets within migmatitic gneisses and contain relics of high-pressure granulite or even high-temperature eclogite-facies mineral assemblages [*Kröner et al., 2006*]. The peak conditions of metamorphism are 13.4–15.5 kbar and 820–840 °C. The inclusions pertaining to the earlier mineral assemblage were presumably formed at a pressure of 17–18 kbar corresponding to the conditions of high-temperature eclogite facies [*Zhao et al., 2001*].

With reference to geodynamic reconstructions and conditions of Palaeoproterozoic granulite-facies metamorphism, recently published researchers prefer plate-tectonic terminology and the model of collisional orogeny [*Kusky, Li, 2003; Kröner et al., 2006; Santosh et al., 2012*]. Meanwhile, the aforementioned age determinations, lithologies, and metamorphism of rocks in the North China Craton are close in chronology to Middle and Late Palaeoproterozoic evolution of the East European Craton and Lauroscandia in general. The experience of studying granulite-gneiss complexes of the East European Craton [*Mints, 2014; 2015b*] allows us to interpret the data briefly characterized above in terms of a plume model of intracontinental granulite-facies metamorphism, including formation of ultrahigh-temperature and high-pressure granulites and high-temperature eclogites. It should be noted than in contrast to the East European Craton, the gabbroanorthosite intrusions associated with Palaeoproterozoic granulites are not numerous in the North China Craton. These are the Damiao anorthosite complex in the northern part of the Trans-North China Orogen with an exposed area of ~80 km^2, Changsaoying K-feldspar granite, Lanying anorthosite and quartz syenite, and Gubeikou K-feldspar granite (these complexes yield intrusion ages between 1.75 and 1.68 Ga [*Zhang et al., 2007; Zhao et al. 2009*]), and the Sancheong–Hadong anorthosite complex ~250 km^2 in area within the Yeongnam Massif (upper amphibolite- and granulite-facies gneisses) in South Korea, which consists of massive anorthosite, leucogabbro, leuconorite, and oxide-bearing leucogabbro dated at 1.87–1.86 Ga [*Lee at al., 2014*].

The Itabuna–Salvador–Curaçá Orogen located in the São Francisco Craton, Brazil is another prominent example. This orogen was subject to granulite facies metamorphism at ultrahigh temperatures of 900–1000 °C and pressures of 7.0–8.0 kbar at ca. 2.08–2.05 Ga. Tonalitic and enderbitic granulites from the lower part of succession show two distinct zircon populations dated at 2.7–2.6 and 2.08–2.07 Ga (cores and rims of zircon grains, respectively). The older estimates are interpreted as the age of magmatism [*Silva et al., 2002; Barbosa et al., 2006; Leite et al., 2009*]. A series of small Palaeoproterozoic (ca. 2.1 Ga) gabbroanorthosite massifs occurs along the western boundary of the granulite belt [*Cruz et al., 2000*].

The oldest Gavião Block in the WSW part of the São Francisco Craton is composed essentially of Archaean granitoid rocks (3.4–3.2 Ga) and associated greenstone sequences (3.3–3.2 Ga). As in most orogens with granulite-gneiss belts, the repeated manifestations of the Archaean granulite-facies metamorphism are known in marginal zones of the Itabuna–Salvador–Curaçá Orogen. Along the western boundary, there is the Jequié Block characterized by heterogeneous granulitic migmatites with supracrustal inclusions and several charnockitic intrusions that represent the older sequence dated at 3.0–2.9 Ga, together with younger granodiorite and granite, 2.8–2.6 Ga in age [*Barbosa, Sabaté, 2002*]. The eastern margin of the orogen consists of orthogneisses, migmatites, and tonalites of the Serrinha (Uauá) Block, 3.2–2.9 Ga in age. It consists mainly of banded gneisses intruded by tonalitic bodies and mafic-ultramafic complexes, most of them metamorphosed under granulite facies conditions and later retrogressed to amphibolite grade; it is bounded to the west by a sequence of steeply dipping quartzites, metapelites, migmatites, deformed granitoids and mafic rocks, collectively called the Caldeirão Belt, along with the 3.16 Ga Lagoa da Vaca anorthosite–leucogabbro layered complex [*Paixão, Oliveira, 1998; Barbosa, Sabaté, 2002*]. Thus, the Itabuna–Salvador–Curaçá Orogen is characterized by synformal structure (Fig. 7) and is mostly composed of metasedimentary rocks, peraluminous granites [*Leite et al., 2009*] and involves gabbroanorthosite massifs. All this corresponds in full measure to the intracontinental granulite-gneiss belts related to plumes. In general, the geological history of the Itabuna–Salvador–Curaçá Orogen almost completely mimics the history inherent to the Limpopo Belt (Kaapvaal craton, southern Africa), although formation of the Itabuna–Salvador–Curaçá Orogen was related to the youngest high-grade metamorphic event (2.08–2.05 Ga).

3.10. Accretion of multiple island arcs and/or mainly intracontinental development (1.8–1.2 Ga)

3.10.1. Lauoroscandia evolution

The Palaeoproterozoic evolution of both the East European and the North American cratons (as we postulate, both were fragments of a single Lauroscandia continent that maintained its integrity during the Proterozoic) was followed by an extended period of about 600 m.y., the character of which is the subject of discussion in this section.

According to the commonly held point of view, the main characteristic of this period was formation of accretionary orogens, i.e., *sequential accretion of island-arc terranes to the southeastern edge of Laurentia and the western margin of the EEC* accompanied by magmatism at the active continental margins of these continents. As a result, at the site of the future Grenville Orogen, the Labradorian (1.71–1.60 Ga), Pinvarian (1.52–1.45 Ga) and Elzevarian (1.25-1.22 Ga) orogens were formed in sequence. The 1.46–1.23 Ga interval, which is known by its magmatism, was named *Elsonian*. The rocks formed between 1.81 and 1.71 Ga ago are regarded as *Pre-Labradorian* (Fig. 8a, 8b). In the southwestern Grenville Province, a similar nature is attributed to the rocks of the Allochthonous Monocyclic Belt and the Composite Arc Belt, formed 1.74 to 1.30–1.25 Ga ago. At ~1.16 Ga, rocks of the Frontenac-Adirondack belt were accreted from the east to the Allochthonous Monocyclic Belt [*Carr et al., 2000; Gower, Krogh, 2002*, and references therein]. Metamorphosed under granulite-facies conditions, the volcano-sedimentary deposits formed ~1.2 Ga ago and closely preceded the Grenvillian Orogeny proper are retained in the Central Metasedimentary Belt (CMB) in the southwestern part of the Grenville Province [*Corriveau, van Breemen, 2000*]. The CMB consists of three domains: (1) marble-predominating, (2) with predominance of quartzite, (3) with predominance of gneisses. Metapelites, calc-silicate rocks, and amphibolites are of subordinate abundance. Associations of metasedimentary rocks of similar type and age (gneisses and their sulphidated varieties, marble, quartzite, including deposits of graphite (1.2–1.1 Ga)) are also revealed in the Grenvillian inliers within the Appalachian Orogen. This association, 1.08 Ga in age, has undergone granulite-facies metamorphism (680–760 °C and 4.1–5.0 kbar) [*Volkert et al., 2000*].

Lithological and geochemical characteristics of the CMB rocks can be regarded as supporting a riftogenic association formed as a result of failed back-arc extension of the continental margin [*Dickin, McNutt, 2007*]. An important indicator of the geodynamic settings related to the final stages of the Grenville–Sveconorwegian Orogeny is the association of carbonatite dikes and veins formed 1.17–1.09 Ga ago, which was discovered and studied within the CMB [*Moecher et al., 1997*].

It is commonly suggested that the lithosphere of southern North America was built up by progressive addition of a series of dominantly juvenile volcanic arcs and oceanic terranes accreted along a long-lived southern margin of Laurentia between ~1.8 and 1.3 Ga: Pembine-Wausau and Elves Chasm arcs (1.84–1.82 Ga), the Yavapai province (1.80–1.70 Ga), the Mazatzal province (1.70–1.65 Ga), the Granite–Rhyolite province (1.50–1.30 Ga) and associated intracratonic A-type granitic magmatism (1.45–1.30 Ga). Each episode of addition of juvenile crust and its transformation into stable continental crust was facilitated by voluminous granitoid plutonism that stitched new and previously existing orogenic boundaries. Slab rollback created

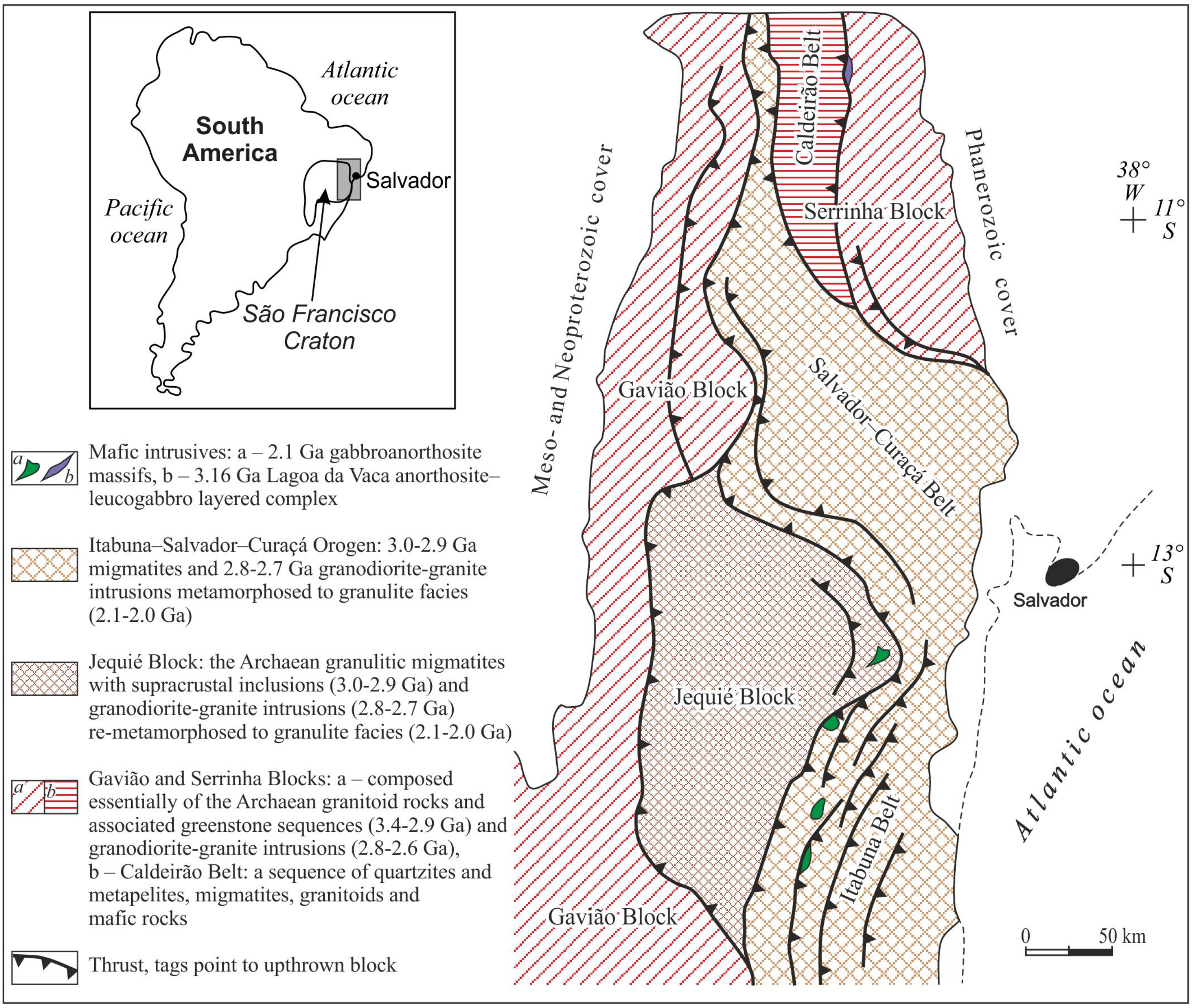

Fig. 7. Itabuna–Salvador–Curaçá Orogen, South America. Based on *Paixão* and *Oliveira* [*1998*], *Cruz et al.* [*2000*], *Barbosa* and *Sabaté* [*2002*], *Leite et al.* [*2009*].

Рис. 7. Ороген Итабуна-Сальвадор-Кюраса, Южная Америка. На основе *Paixão* and *Oliveira* [*1998*], *Cruz et al.* [*2000*], *Barbosa* and *Sabaté* [*2002*], *Leite et al.* [*2009*].

transient extensional basins (1.70 and 1.65 Ga), where Palaeoproterozoic quartzite–rhyolite successions were deposited [*Whitmeyer, Karlstrom, 2007*]. Between 1.73 and 1.48 Ga, the accretionary crustal growth along the southwestern edge of the East European Craton (EEC) was directed toward the west. The NS-trending crustal belts become younger westward [*Bogdanova et al., 2008*, and references therein]. It is assumed that the accretionary orogens and their extensions to Australia and Baltica make up a long-lived orogenic system more than 10000 km in extent, which retains a record of 800 million years of magmatism and tectonism at a convergent margin [*Gower et al., 1990; Karlstrom et al., 2001; Holm et al., 2005*].

There are a lot of potentially disputable problems in the above interpretation. *Bickford* and *Hill* [*2007*] pointed out that the Palaeoproterozoic crust of southeastern Laurentia is more complex than would be expected from simple accretion of island arcs. The accretionary model does not explain obvious evidence that older crust was involved in the formation of many rocks. Despite numerous features suggesting that the rocks were formed in the arc environment, the contrasting bimodal volcanic suites formed at the rifted continental margins, such as the Kenya Rift and the Basin and Range Province, rather than in island or continental arcs [*Bickford, Hill, 2007*, and references therein] (Fig. 8a, 8b).

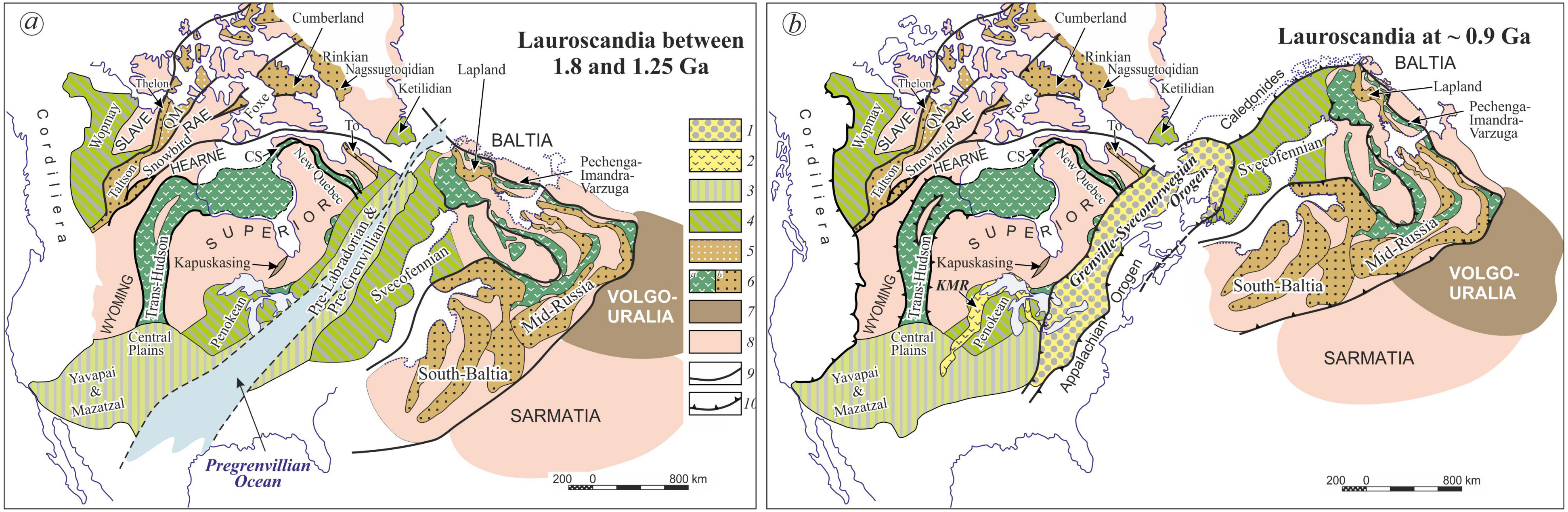

Fig. 8. Reconstruction of the Mesoproterozoic-Neoproterozoic evolution of Lauroscandia: Grenville-Sveconorwegian orogen: *a* – between 1.80 and 1.25 Ga, *b* – at ~0.9 Ga.

1–3 – Mesoproterozoic-Neoproterozoic: *1* – granulite-gneiss complexes of the Grenville- Sveconorwegian orogen, 2 – sedimentary-volcanic rocks of the Keweenawan Midcontinent Rift (KMR), 3 – Pre-Labradorian and Pre-Grenvillian accretionary complexes; *4–6* – Palaeoproterozoic: *4* – accretionary complexes, *5* – granulite-gneiss complex of Queen Maud Block, *6* – intracontinental collision orogen: *a* – volcanic-sedimentary belts, *b* – granulite-gneiss belts; *7–8* – Archaean (3.3–2.5 Ga): *7* – Volgo-Uralia granulite-gneiss area, *8* – granite–greenstone terranes with inclusions of Archaean complexes, partially reworked in late Palaeoproterozoic; *9–10* – main tectonic boundaries of: *9* – unrecognized type, *10* – thrust type; tags point to upthrown block. Names of Archaean tectonic units are shown in Arial typeface, Palaeoproterozoic units appear in Times New Roman. Abbreviations in figure: CS – Cape Smith Belt, KMR – Keweenawan–Midcontinent rift system, QM – Queen Maud Block, To – Torngat Orogen.

Рис. 8. Реконструкция мезо-неопротерозойской эволюции Лавроскандии: Гренвилл-Свеконорвежский ороген: *a* – между 1.80 и 1.25 млрд лет, *b* ~ 0.9 млрд лет.

1–3 – мезопротерозой – неопротерозой: *1* – гранулито-гнейсовые комплексы Гренвилл-Свеконорвежского орогена, *2* – осадочно-вулканогенные породы рифта Кьюиноу Мидконтинент (KMR), *3* – Пре-Лабрадорский и Пре-Гренвиллский аккреционные комплексы; *4–6* – палеопротерозой: *4* – аккреционные комплексы, *5* – гранулито-гнейсовый комплекс блока Куин-Мод, *6* – внутриконтинентальный коллизионный ороген: *a* – вулканогенно-осадочные пояса, *b* – гранулито-гнейсовые пояса; *7–8* – архей (3.3–2.5 млрд лет): *7* – Волго-Уральский гранулито-гнейсовый ареал, *8* – гранит-зеленокаменные области с включениями архейских пород, переработанных в палеопротерозое; *9–10* – главные тектонические границы: *9* – неопределенного типа, *10* – надвиги, бергштрихи указывают на надвинутый блок. Названия архейских тектонических подразделений даны шрифтом Arial, палеопротерозойских подразделений – шрифтом Times New Roman. Аббревиатуры: CS – пояс Кейп-Смит, KMR – рифт Кьюиноу-Мидконтинент, QM – блок Куин-Мод, To – ороген Торнгат.

When considering this topic, it is important to bear in mind that *anorthosite, gabbro-norite, norite and potassic granitoid (including rapakivi granite) massifs* are widely, though nonuniformly, distributed within the framework and in the vicinity of the Grenville–Sveconorwegian Orogen (GSNO) (Fig. 9). Some of these intrusions situated nearly 1000 km from the GSNO were previously regarded as *anorogenic*. In the northwest, outside the Grenville Province and partially directly within the Grenville Orogen, the autochthon contains older massifs dated at ~1.65, ~1.45 and 1.33–1.30 Ga [*Hamilton et al., 2004*, and references therein]. Numerous ARG complexes are known in the western part of the East European Craton [*Sharkov, 2010*]. The late Palaeoproterozoic–Mesoproterozoic (1.8–1.5 Ga) massifs composed of the AMCG and ARG complexes occur around the GSNO [*Mints, 2014*]. Some massifs are at a considerable distance from the orogen. In the absence of distinct zoning by age, the main trend is quite clear: up till the end of the Mesoproterozoic, intrusions shift back to the GSNO axial zone and then are recorded only in the orogen interior, where magmatism was accompanied by high-temperature metamorphism. *Wiebe* [*1980*] and *Sharkov* [*2010*] pointed out that thickness of the crust increases by up to 50–60 km in the area where the AMCG and ARG complexes occur. In some cases, these intrusions are associated with semicircular sedimentary basins [*All et al., 2006*]. These depressions are filled with epicontinental conglomerate, arkosic to subarkosic sandstone, and shale, all intercalating with mafic and felsic igneous rocks [*Kohonen, Rämö, 2005*]. Among the igneous formations situated within the Sveconorwegian sector, the granitoid batholiths of the Transscandinavian Igneous Belt (TIB) occupy a special position. This belt, which extends for 1400 km along western edge of the Svecofennian accretionary orogen, directly borders the Sveconorwegian sector of the GSNO [*Högdahl et al., 2004*]. The TIB is largely composed of coarse-grained porphyritic rocks varying from monzodiorite to granite, had been formed in the time span from 1.85 to 1.65 Ga and was reworked and deformed due to the Grenville–Sveconorwegian Orogeny 1.1–0.9 Ga ago. Three genetic models of TIB magmatism have been proposed [*Högdahl et al., 2004*, and references therein]: (1) Andean-type convergent plate boundary, (2) intracontinental extension, and (3) post-collisional collapse of the thickened crust with subsequent extension. Most researchers prefer the first alternative. In our opinion, a close structural and temporal proximity of TIB granites and ARG, as well as high alkalinity of rocks testify in favor of the second (intracontinental) model.

Epicontinental intracratonic sag basins form one more characteristic type of tectonic structures of this age. We have mentioned above the sedimentary basins interrelated with AMCG and ARG complexes. More extensive basins are known in the periphery of the area with the AMCG and ARG complexes in Lauroscandia. These are the Thelon Basin (1.72–1.00 Ga) and the Athabasca Basin (1.74–1.50 Ga) in the central and western parts of the North American Craton [*Jefferson et al., 2007; Hiatt et al., 2010; Coleman, Cahan, 2012; Allen et al., 2015*], the Onega Depression (from ~2.0 to 1.86 Ga) and the Peri-Onega Depression (ca. 1.77 Ga) in the southeastern part of the Fennoscandian Shield [*Mints et al., 2015c*, and references therein]. Like the sedimentary basins, which are directly linked with AMCG and ARG complexes, these basins have semi-oval outlines (Fig. 9).

The above features of the AMCG and ARG complexes together with accompanying sedimentary basins clearly indicate their intracontinental origin without any links to subduction and/or collisional processes and events.

3.10.2. Epicontinental intracratonic basins beyond Lauroscandia

The Espinhaço Basin, one of the largest Precambrian intracratonic basins in the Brazilian Shield, comprises three tectonostratigraphic megasequences deposited at 1.80–1.68, 1.60–1.38, and 1.2–0.9 Ga, respectively [*Guadagnin et al., 2015*].

The Cuddapah Basin is one of a series of Proterozoic basins that overlie the cratons of India. The terrigenous sequences filling this basin were deposited during several time spans from 1.89 to 1.1 Ga [*Collins et al., 2015*].

3.11. The third prominent superplume event recognizable worldwide (origin of the Grenville–Sveconorwegian orogen) (1.2–0.9 Ga)

3.11.1. Superplume-related evolution within Lauroscandia

The Grenville Province that extends along the southeastern margin of the Canadian Shield (Fig. 8, *b*, see 9) includes the orogen of the same name, as well as tectonic structures within its framework. The Sveconorwegian orogen in southwestern Scandinavia is regarded as a direct continuation of the Grenville Orogen to the northeast [*Hoffman, 1991; Dalziel, 1997; Andersson et al., 2008*]. In general, *the Grenville–Sveconorwegian Orogen* (*GSNO*) is a combination of crustal fragments that differ from one another in structure and formation history, which were assembled in a single tectonic unit over a vast territory under the influence of vigorous pulses of Meso–Neoproterozoic thermal, magmatic, and tectonic activity from ~1.19 to 0.96 Ga [*Wardle et al., 1986; Cosca et al., 1998; Tollo et al., 2004; Cawood et al., 2007; Andersson et al., 2008; Rivers, 2009; Mints, 2014*]. The Grenville Orogen contains both Allochthonous and

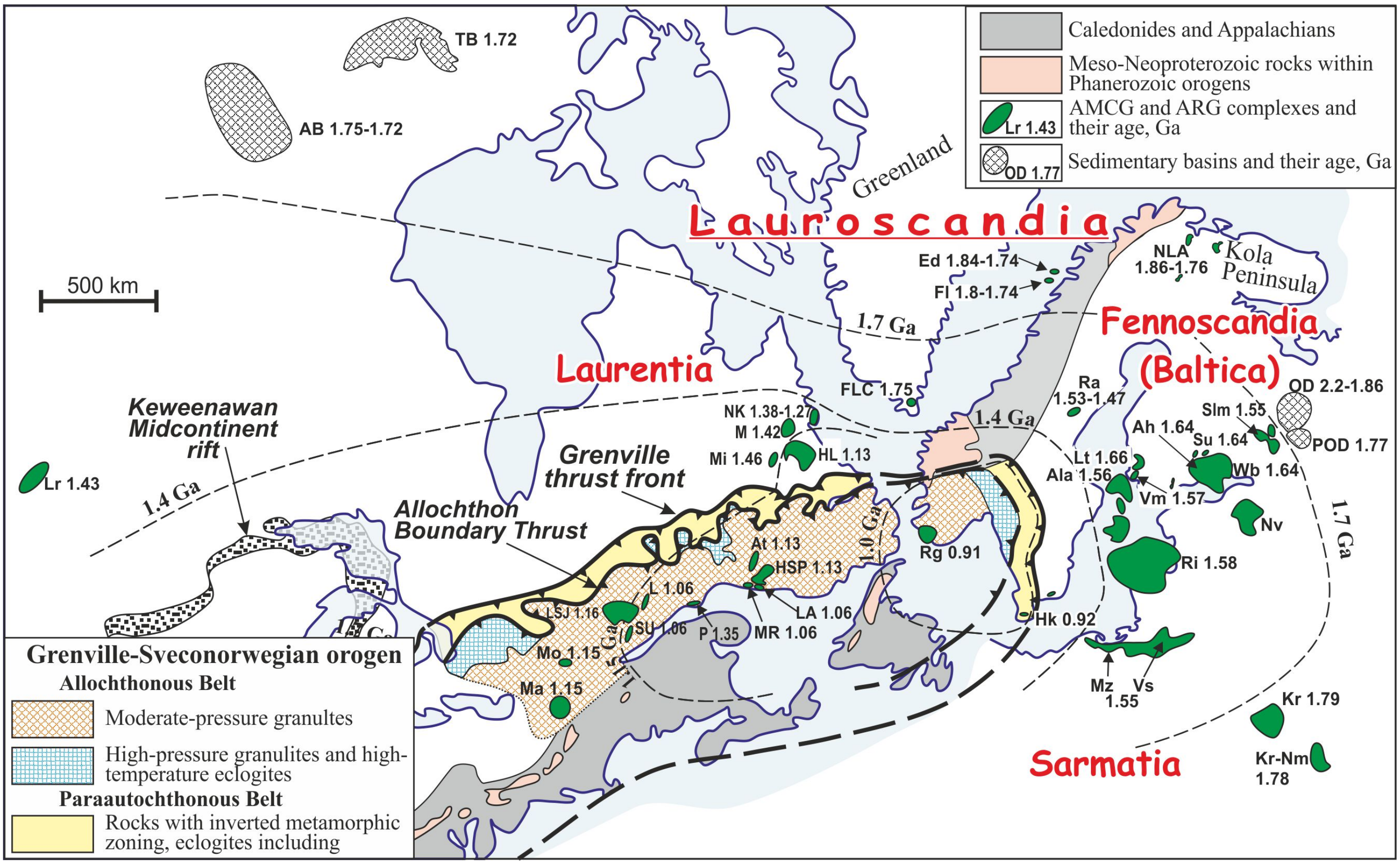

Fig. 9. Reconstruction of Grenville–Sveconorwegian Orogen at ~0.8 Ga, emplacement and age of the Mesoarchaean-Neoarchaean anorthosite–mangerite–charnockite–granite (AMCG) and anorthosite–rapakivi granite (ARG) complexes within and in framework of orogen. Modified after *Mints* [*2014*].

Map is based on the data of *Wardle et al.* [*1986*], *Andersson et al.* [*2008*] and *Hynes* and *Rivers* [*2010*]. References to geochronological data are given in [*Mints, 2014*]. Designations employed: Kr – Korosten; Kr-Nm – Korsun-Novomirgorod; Ri – Riga; Nv – Novgorod; Slm – Salmi; Ul – Ulyaleg; LP – Lodeinoe Pole; Ala – Aland; Lt – Laitila; Vm – Vehmaa; Ah – Ahvenisto; Su – Suomenniemi; Wb – Wiborg; Ra – Ragunda; Mz – Mazur complex; Vs – Veisiejai Complex; Ma – Marcy; Mo – Morin; LSJ – Lac-Saint-Jean; HSP – Havre-Saint-Pierre; LA – Lac Allard; MR – Magpie River; At – Atikonak; HL – Harp Lake; L – Labrieville; M – Mistastin; Mi – Michikamau; NK – Nain–Kiglapait; P – Pentecôte; SU – St. Urbain; Lr – Larami; FLC – Flat Lying Complex; Rg – Rogaland; Hk – Hakefjorden; Fl – Flakstadǿy; Ed – Eidsfjord.

Рис. 9. Реконструкция Гренвилл-Свеконорвежского орогена для возраста ~0.8 млрд лет, показан возраст внедрения мезо-неоархейских анортозит-мангерит-чарнокит-гранитных (АМЧГ) и анортозит-рапакиви-гранитных (АРГ) комплексов внутри и в обрамлении орогена. С изменениями по *Mints* [*2014*].

Карта базируется на данных *Wardle et al.* [*1986*], *Andersson et al.* [*2008*] и *Hynes* and *Rivers* [*2010*]. Ссылки на источники геохронологических данных даны в [*Mints, 2014*]. Использованы наименования массивов: Kr – Коростеньский; Kr-Nm – Корсунь-Новомиргородский; Ri – Рижский; Nv – Новгородский; Slm – Салминский; Ul – Улялегский; LP – Лодейного поля; Ala – Аландский; Lt – Лайтила; Vm – Вермаа; Ah – Ахвенисто; Su – Суоменниеми; Wb – Выборгский; Ra – Рагунда; Mz – Мазурский комплекс; Vs – Веиседжаи комплекс; Ma – Марси; Mo – Морин; LSJ – Лейк-Сен-Джин; HSP – Хавьер-Сен-Пьер; LA – Лэк Аллар; MR – Мегпи Риверг; At – Атиконак; HL – Харп Лейк; L – Лебривилл; M – Мистасин; Mi – Мичикамау; NK – Нейн–Киглапайт; P – Пентекот; SU – Сан-Урбан; Lr – Ларамийский; FLC – Флет-Лайинг комплекс; Rg – Рогалан; Hk – Хейкефьорден; Fl – Флакстодой; Ed – Ейдсфьорд.

Para-autochthonous belts. The subduction-type eclogite complex (1.15–1.12 Ga) at the southwestern edge of the Grenville Province within the Llano Uplift in Texas [*Carlson et al., 2007; Mosher et al., 2008*] and the sedimentary–volcanic association of the Keweenawan-Midcontinent Rift System (1.12–1.09 Ga) extending for about 2.000 km [*Van Schmus et al., 1982; Vervoort et al., 2007*] reveal close spatio-temporal relations to the GSNO.

In the Grenvillian sector of the GSNO, three discrete stages of thrusting and high-grade metamorphism have been identified: ~1.19–1.14, 1.08–1.02 and 1.01–0.98 Ga [*Rivers, 1997; Rivers et al., 2002*]. Manifestations of the latest stage are localized within the Para-autochthonous belt. High-grade metamorphism and thrusting are usually considered to be the main components of the Grenvillian continent-continent type collision that terminates the almost 700 Ma evolution of the Laurentian continental margina. The granulite-facies metamorphism is regarded as a direct result of the collisional orogeny and crustal thickening.

The age and intensity of the peak deformation and metamorphism vary considerably within the GSNO. The age of metamorphism becomes younger in the northeastern direction along the Grenville sector and in the eastern direction within the Sveconorwegian sector. The manifestations of granulite-facies metamorphism were repeatedly recorded long before the Grenville Orogeny. The granitoids and paragneisses crystallized 1.81–1.71 Ga ago then underwent high-grade metamorphism (1.67–1.66 Ga) [*Korhonen, 2006*, and references therein]; gneisses and schists together with the AMCG Petit Mecatina Suite underwent granulite-facies metamorphism at 1.47–1.45 Ga. The gabbronorite, mangerite, and granite of the Matamek Complex intruded 1.38–1.37 Ga ago and then underwent granulite-facies metamorphism almost synchronously, between 1.37 and 1.35 Ga. The next manifestation of the granulite-facies metamorphism (1.20–1.18 Ga) is recorded in the Bondy gneiss complex in the western Grenville Province [*Boggs, Corriveau, 2004*].

The early-stage metamorphic events at various structural levels are characterized by high temperatures ranging from 800 to 900 °C. Along with a syn-metamorphic mafic dike swarm, this indicates a vigorous heat influx from a mantle source. The next stage (onset of tectonic exhumation) corresponds to the temperature range from 700 to 800 °C and pressures of 10–17 kbar [*Cox et al., 2002; Indares, Dunning, 2004*]. In the southwestern Grenville Province, the highest pressure and temperature are confined to the bases of tectonic slices, and lower values are related to their inner regions [*Wodicka et al., 2000*]. The metamorphism of rocks in the Para-autochthonous belt proceeded, as a rule, at a moderate pressure, whereas locally occurring high-pressure granulite and eclogite mineral assemblages were formed at pressures exceeding 12 kbar.

In addition to high-grade metamorphism and thrust-nappe structures, *the Grenville–Sveconorwegian orogen* (*GSNO*) is distinguished by specific intrusive magmatism, which is expressed in numerous massif-type anorthosite, gabbroanorthosite, charnockite, and potassic granitic plutons, including rapakivi granite. Taken together, they make up anorthosite–mangerite–charnockite–granite (AMCG) complexes (Fig. 9). The AMCG complexes of Grenvillian age are found only in the inner region of the GSNO, where they are associated with high-temperature metamorphic rocks. According to *Corrigan* and *Hanmer* [*1997*], within the Grenville sector, the AMCG complexes were formed twice, 1.16–1.13 and 1.09–1.05 Ga ago. The Late Grenvillian mafic intrusive bodies including dikes and the subalkaline granitic plutons with rapakivi-like structure are dated at 0.99 and 0.96 Ga [*Gower, Krogh, 2002*]. All AMCG massifs underwent granulite-facies metamorphism at temperatures of 800–900 °C. The norite–anorthosite bodies within the Sveconorwegian sector of the GSNO were emplaced a little later at 0.93–0.92 Ga. They cut through the gneisses that underwent granulite-facies metamorphism ~1.00 Ga ago and then a succeeding high-temperature contact metamorphism [*Åreback, Stigh, 2000; Möller et al., 2003*].

For a long time, the thrust and fold structures, intensive deformation and the high-grade metamorphism in the Grenville Province have been regarded as evidence for the accretionary-collisional nature of the Grenville Orogen [*Dewey, Burke, 1973*]. It is assumed that the Grenvillian Orogeny represents the final stage of the orogenic cycle that culminated in the continent-continent collision. It was assumed also that most of the GSNO crust was created as a result of subduction [*Gower, Krogh, 2002*, and references therein]. The Himalayan-Tibetan collision orogen has been proposed as the closest analogue of the GSNO [e.g., *Rivers, 2009; Hynes, Rivers, 2010; Jamieson et al., 2007*]. However, only one of a couple of colliding continents, i.e., Laurentia is known. In addition, the geochemical signature of the magmatism and, in particular, unusually extensive occurrence of anorthosite massifs, which is unique in the geological record, can be referred to the subduction environment only with great reservations.

3.11.2. Superplume-related evolution beyond Lauroscandia

The Grenvillian tectonothermal events are manifested to one or another degree in almost all continents. Here we give only a few characteristic examples.

The Musgrave Inlier in Central Australia is a part of the Musgrave–Albany–Fraser Orogen (MAFO), which provides evidence for the Grenvillian (1.3–1.1 Ga) tectonothermal event in Australia. This orogen is characterized by a synformal thrust and nappe structure, as is inherent to large granulite-gneiss complexes (Fig. 10). The Musgravian Gneiss that is interpreted as a volcano-sedimentary sequence comprises predominantly felsic orthogneiss, subordinate mafic interlayers, and sporadic metapelite, quartzite, and calc-silicate rocks. The protoliths to these rocks were emplaced or deposited in the time interval between 1.60–1.54 Ga. According to *Tucker et al.* [*2015*, and references therein], the Grenvillian tectonism is characterized by regional-scale deformation, continual and voluminous high-temperature A-type felsic magmatism, and amphibolite–granulite-facies metamorphism dated at 1.22–1.14 Ga. The peak metamorphic conditions ranged from ~650 to 800 °C at 5–6 kbar in the East Musgrave Inlier and ~750–850 °C at 5–6 kbar in the West Musgrave Inlier, to ~700–800 °C at 9–11 kbar in the Musgrave Ranges. Mafic magmatism in the Musgrave

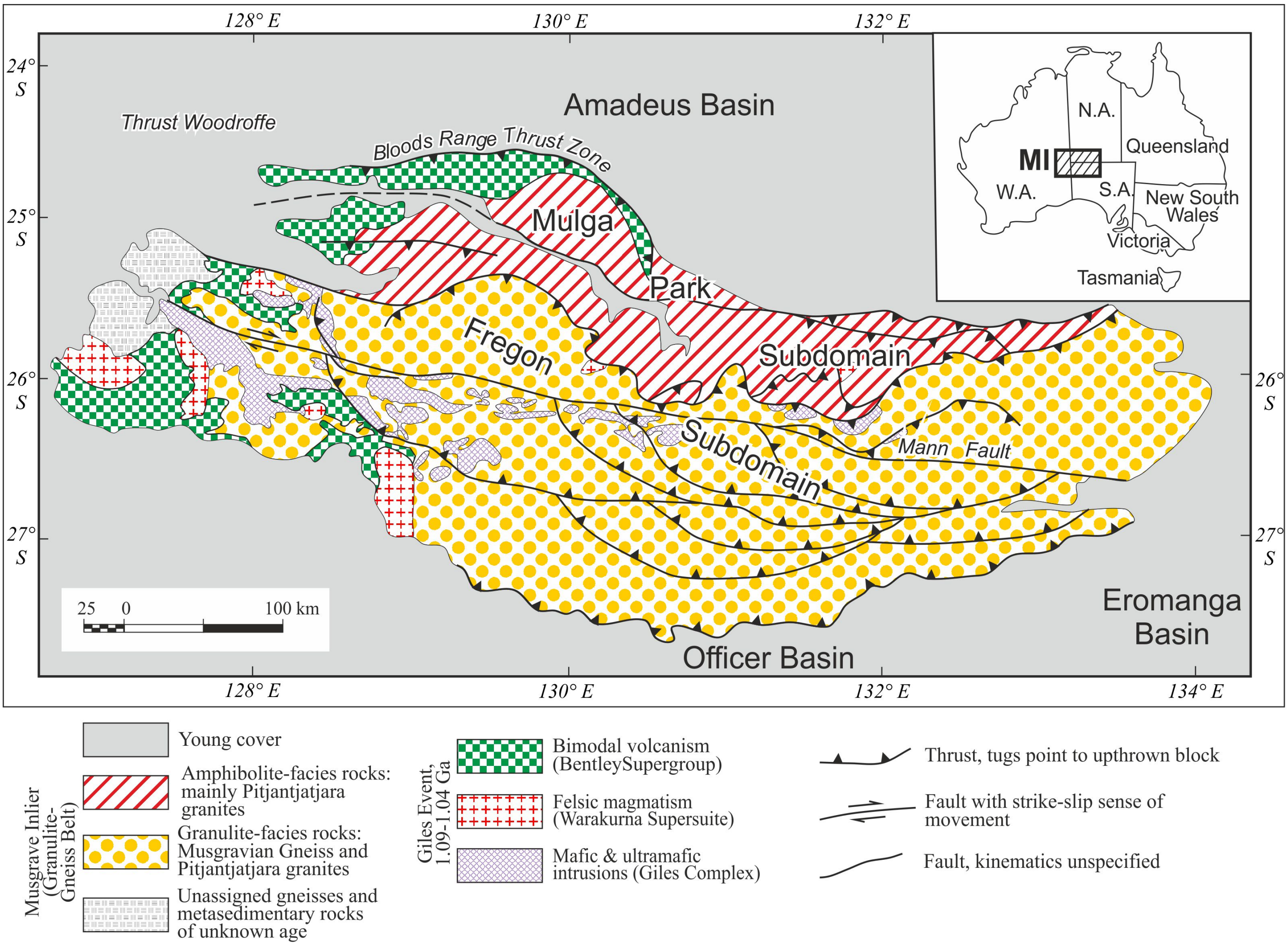

Fig. 10. Musgrave Inlier in Central Australia. Modified after *Tucker et al.* [*2015*].

Рис. 10. Выступ Масгрейв в Центральной Австралии. С изменениями по *Tucker et al.* [*2015*].

Inlier is represented by two major events. Although the MAFO is widely interpreted to have formed during the amalgamation of Proterozoic Australia, *Smithies et al.* [*2011*] and *Howard et al.* [*2015*] suggest that much of the Musgrave Orogen was formed as a result of intracratonic and extensional processes.

The Musgrave Inlier was significantly reactivated during the Petermann (Pan-African) Orogeny (0.63–0.53 Ga), which is characterized by intracratonic high-pressure granulite-facies mylonitization, development of ultramylonite and pseudotachylite zones [*Scrimgeour, Close, 1999*].

The Namaqua-Natal metamorphic belt extends along the southern margin of the Kaapvaal Craton in South Africa. High-temperature garnet and garnet-free leucogranite and charnockite were formed within this belt. Amphibolite- to granulite-facies metamorphism is dated at 1.20 and 1.02 Ga [*Eglington, 2006; Colliston et al., 2015; Mendonidis et al., 2015*].

3.12. Superplume event mainly recognizable in Gondwana areas (origin of the Pan-African orogen) (0.87–0.55 Ga)

The term *Pan-African* is used to describe tectonothermal activity of Neoproterozoic to earliest Palaeozoic age, restricted mainly within the Gondwana supercontinent. The orogenic events of this period occur in Lauroscandia only extremely rarely. Within the Pan-African domains, two broad types of tectonic belts can be distinguished. One type consists predominantly of Neoproterozoic low-grade supracrustal and magmatic assemblages that are similar to those in Phanerozoic collisional and accretionary belts. Such belts include the Arabian–Nubian shield of Arabia and northeastern Africa, the Damara–Kaoko–Gariep Belt and Lufilian Arc of southern-central and southwestern Africa, the West Congo Belt of Angola and Congo Republic, the Trans-Sahara Belt of West Africa, and the Rokelide and

Mauretanian belts along the western part of the West African Craton. Belts of the other type generally contain polydeformed high-temperature and in some cases ultrahigh-temperature granulite-facies metamorphic associations. The protoliths consisted largely of much older Mesoproterozoic to Archaean continental crustal rocks, which were strongly reworked in the Neoproterozoic. Well-studied examples are the Mozambique Belt of East Africa, including Madagascar and extensions into western Antarctica, the Zambezi Belt of northern Zimbabwe and Zambia, and some little known terranes [*Kröner, Stern, 2004*, and references therein].

A combination of the two aforementioned types of tectonic belts with allowance for widespread occurrence of the second-type of belts coincides with a similar feature of Palaeoproterozoic orogens in Lauroscandia. This allows us to suggest a similar nature for the Pan-African Orogeny with a critical role for mantle plumes.

4. Discussion

The list of geological units presented above, which characterize the main trends in geodynamic evolution during certain periods of the geological chronicle is far from being exhaustive. We have listed only the prominent examples, which display (1) systematic links of LIPs, and of formation history of intracontinental sedimentary basins and granulite-facies metamorphism, to peak U–Pb zircon ages of juvenile rocks; and (2) synchronism of these phenomena in almost all continents. These two sets of circumstances cannot be explained by plate-tectonic reconstructions and presuppose the existence of a large Neoarchaean–Palaeoproterozoic supercontinent like that suggested by *Piper* [*2015*] on the basis of recent paleomagnetic data.

The above review demonstrates the validity of our interpretation of the granulite-gneiss complexes and of the evaluation of their role in continental crust formation, which are adopted in this paper (see *Mints* [*2014, 2015b, 2015c*] for more details). Nevertheless, some aspects require further discussion.

As mentioned above, the granulite-facies metamorphism is accepted by a number of researchers not only as the result of, but also as direct evidence in favour of geological events related to accretion and/or collision. In many papers on the history of the emergence and disintegration of supercontinents, almost all the above-described granulite-gneiss belts were viewed as collisional orogens [e.g., *Percival, 1994; Daly et al., 2006; Brown, 2007, 2009; Li et al., 2008; Gower, Krogh, 2002*]. At the same time, as reviewed above, there is good evidence for the participation of granulite-gneiss complexes in intracontinental orogeny.

The data and discussion presented by *Mints* [*2014, 2015c*] emphasize the need for a new view on Early and Late Precambrian geodynamics, not only in regard to the origin of granulite-gneiss belts, but also related to the history of supercontinents and the supercontinent cycle, the possible uniqueness of Neoarchaean–Proterozoic geodynamics, the origin of anorthosites and some other problems. Consideration of most of these issues goes far beyond the scope of this paper. Below we will briefly touch on only some of them.

The granulite-gneiss complexes of various ages along with their shared features have certain inherent differences, which may occur due to differences in mantle-plume processes. In particular, the alternation of concentric zones, especially characteristic of the Karelian–Belomorian and Lauro–Russian intracontinental oval orogens, involves concentric change in the characteristics and/or morphology of the mushroom-shaped head of a mantle plume. Most high-temperature granulite-facies metamorphism is located not in the center of the round-oval areas, but within a more or less clearly defined arcuate zone. Apparently, the high temperatures were characteristic for the central part of the Karelian–Belomorian intracontinental oval orogen, but the vast interior region also fits low- or moderate-grade metamorphism (see Fig. 3). As a whole, the Karelian-Belomorian intracontinental oval orogen is characterized by relatively low temperatures, limited development of granulite-facies rocks and the lack of AMCG and ARG complexes, which are so characteristic of all other orogens of this type. The highest level of metamorphism is characteristic of the Volga-Uralia granulite-gneiss area (see Fig. 4), where overall concentric zoning is absent, and the elements of the internal zoning are associated with numerous local centres within the ovoids and structures of the inter-ovoid region [*Mints et al., 2015b*].

On the other hand, the differences in the level of erosion of orogens seems obvious. In particular, the abundance of intrusive bodies of the AMCG complex in the inner region of the Grenville-Sveconorwegian orogen (GSNO) (see Fig. 9), in combination with participation of high-pressure granulites and involvement of high-temperature eclogite at the base of the tectonostratigraphic section, provide evidence for the deep level of erosion. In fact, within the GSNO the metamorphosed rocks of the '*basement*' of the sedimentation basin are exposed. The filling of the Basin was transformed later in the synformal structure of the GSNO. In contrast, the significant thickness of metasedimentary and metaigneous granulites and limited erosion of the gabbroanorthosite intrusions are characteristic of the Lapland granulite-gneiss belt.

4.1. The Intracontinental Oval Orogens

As demonstrated above, the intracontinental oval orogens are large (600–1000 to 2500–3000 km in diameter) oval-shaped intracontinental tectonic ensembles of the regional rank. At least some of them are characterized by concentric structure and metamorphic zoning or contain internal bowl-shaped crustal structures [*Mints, 2014; 2015b, 2015c; Mints et al., 2015b*]. The intracontinental oval orogens involve granulite-gneiss complexes, derivatives of juvenile but crust-contaminated mafic magmas (gabbroanorthosites and layered mafic–ultramafic rocks), intrusions of "dry" high-temperature within-plate granites, enderbites, and charnockites, and low-grade sedimentary-volcanic belts. An emplacement of the granulite-gneiss complexes within the intracontinental oval orogens indicates significant vertical displacements of the deep crustal associations to a higher level in the crust or directly to the erosion level.

The lithological features of the metasedimentary successions and the specific character of intrusive and volcanic associations, which are encompassed in granulite-gneiss, as well as in the low-grade volcanic belts in the intracontinental oval orogen, directly point to predominance of mantle-plume processes. We described above a number of characteristic examples of such orogens, the Archaean Karelian-Belomorian, Kola, Volgo-Uralian and the Proterozoic Lauro-Russian and Grenville-Sveconorwegian orogens. The Palaeoproterozoic Lauro-Russian Orogen, approximately equal-sized fragments of which are retained in the European and North American continents, is of special interest and significance. The distinct feature of this orogen is emplacement of juvenile volcano-sedimentary and volcano-plutonic complexes in its inner region, which are traditionally considered to be a part of accretionary orogens: Svecofennian, Labradorian, Pinvarian, Elzevarian, Penokean, Yavapai, Mazatzal etc. The above review shows that only the Svecofennian orogen can lay claim to the pure meaning of the term *accretionary*. The rest of these "orogens" include large amounts of previously created continental crust and juvenile intrusive complexes that were formed in the intracontinental environment. We suggest that the formation of these orogens can be viewed as a result of the interacttion of plume- and plate-tectonics; however, there is no doubt that this issue needs further study.

We should also pay attention to the fact that the above listed "accretionary" orogens, as well as the Trans-Hudson collisional orogen, during almost the entire time of their development, remained in the interior of the temporally disrupted supercontinent. As a result, the complexes of rocks formed within these orogens were incorporated into the Lauroscandian crust. We can conclude that, for a long period of time, the plate-tectonic processes at least within Lauroscandia were most likely plume-driven.

We attributed the studied orogens to the intracontinental type, with application of a convention. This implies that (i) these orogens mainly originated and evolved in the intracontinental area; (ii) their evolution was completed in the intracontinental area (e.g., Volgo-Uralia, Karelo-Belomorian, Kola regions); (iii) in some cases, the evolution involved short-lived local ocean opening that closed more or less rapidly; the integrity of the continent was not compromised (Lapland–Mid-Russia–South Baltia Orogen); (iv) in the most complete version of evolution, we reconstructed a combination of plate-tectonic and plume-tectonic processes (Lauro-Russian orogen as a whole). In the latter case, the integrity of the Precambrian supercontinent was also restored and plate tectonics was driven by the superplume.

4.2. The Lauroscandia supercontinent

The above data and their interpretation show that a very large single continent (supercontinent) existed from ~2.8 to 0.85 Ga (probably, up to 0.55 Ga). Lauroscandia could have been a part of this supercontinent or one of several vast continents stable in the Neoarchaean and Palaeoproterozoic. [*Mints, 2007; Mints et al., 2015c*]. A reconstruction of the Palaeoproterozoic (from ~2.5 Ga to ~1.8 Ga) evolution of Lauroscandia was shown above in Fig. 5. Reconstruction of the further Lauroscandian history in the late Palaeoproterozoic, Mesoproterozoic and Neoproterozoic (from ~1.8 Ga to ~0.9 Ga) is presented in Fig. 8.

Analyzing the geological features of the eastern Fennoscandian Shield, many researchers have concluded that the Palaeoproterozoic processes in this region gave rise only to local and temporary disruptions of the continental lithosphere. However, it seems that the emergence of the Svecofennian accretionary orogen fixes final separation of Lauroscandia into European and American portions. The above-described new interpretation of the GSNO granulite-gneiss complexes indicates restoration of Lauroscandia's integrity. In the Meso- and Neoproterozoic, the extensive Grenville–Sveconorwegian Orogen in the Lauroscandian interior was formed. The massif-type anorthosite bodies confined to the 2.7–0.9 Ga time interval indicate the unique Proterozoic supercontinental environment. Arguments for incomplete retention or insufficient depth of erosion to explain this specific fact seem to be weak, and some degree of non-uniformitarianism can indeed be postulated [*Ashwal, 1993, 2010*]. *Mints* [*2014*] has shown that the geological data clearly indicate the intracontinental formation of the AMCG and ARG complexes of Lauroscandia.

The plume initiated tectonic events, such as rifting with local transition to spreading and the formation of short-lived oceans that did not lead to the final separation of the supercontinent fragments, can be classified as "unsuccessful" attempts to break up the Archaean supercontinent. The fundamental changes in the geological evolution of the Earth, which, as we can now state, date back to the beginning of the Neoarchaean, when the Archaean microplate-tectonics gave way to the Neoarchaean-Proterozoic supercontinent tectonics. Formerly it seemed that the supercontinent tectonics was confined to the Palaeoproterozoic [*Mints, 1998, 2007*].

It is noteworthy that the style and features of the tectonic processes and geodynamic settings of plate tectonics during the Neoarchaean-Proterozoic time differ from both the Archaean and the Phanerozoic. At the same time, paradoxically, the Archaean miniplate tectonics has many more similar features as compared to the Phanerozoic plate tectonics, rather than to the Neoarchaean-Proterozoic supercontinent tectonics.

4.3. The episodicity of juvenile crust formation

U–Pb zircon age peaks at 3.1–2.9, 2.8–2.7, 2.55–2.45, 2.00–1.65, 1.2–1.0, 0.87–0.55 Ga and peaks at 0.35–0.20 Ga from granitoids and sediments of major rivers, which are grouped into a series of major peaks at about 2.7, 2.5, 2.1, 1.9, 1.1 and 0.6 Ga [*Condie, Aster, 2010; Abbott, Isley, 2002; Hawkesworth, Kemp, 2006; Voice et al., 2011; Bradley, 2011*, and references therein], mostly correspond to Lu–Hf, Sm–Nd and Re–Os depleted-mantle model ages. At the first order, they reflect juvenile crust production and related rapid crustal growth [*Hawkesworth, Kemp, 2006; Condie et al., 2009a; Lee et al., 2011*]. The most prominent peak in the U–Pb data, at 2.8–2.7 Ga, followed by successively lower peaks at 2.0–1.7 and 1.2–1.0 Ga [*Hawkesworth, Kemp, 2006*], suggests an abrupt turning-point in crust-forming processes in the Neoarchaean (Fig. 11, 12).

Peaks in the age diagram may be associated with periods of rapid mantle convection, which could manifest in two ways: (1) at least some of these peaks might be associated with hot superplumes; (2) accelerated plate motion and subduction that had to be accompanied by periods of supercontinent assembly. Creation of the supercontinents, in turn, is strongly influenced by mantle convection. In the second version, age peaks do not correspond to true crustal growth, but result from enhanced retention of the continental crust. Assessing the credibility of these interpretations, we note that it is difficult to imagine how global pulses of crustal growth could be controlled by plate tectonics. A model that explains the periodicity in magmatic activity and the rapid crustal growth at the heat pulses accompanied by emplacement of mantle plumes (superplumes) seems to be more realistic. Such pulses are much less obvious in the last one billion years of the Earth's history, for which formation and reworking of continental crust in the divergent and convergent zones along plate boundaries are much more typical [*Arndt, Davaille, 2013; Hawkesworth, Kemp, 2006*].

All the major Precambrian tectonic structures of Lauroscandia have been briefly described above. Analysis of the geological record shows a periodicity of the mainly plume-related geological events, which is almost coincident with the geological history as shown by the distribution of age peaks in the course of juvenile crustal growth. This coincidence suggests that episodicity and the main facets of Lauroscandian evolution agree with the global patterns in space and time.

The following phases/ stages can be distinguished in the Precambrian history of Lauroscandia:

(1) >3.8 Ga: Enigmatic Earth (dominating plume tectonics?);

(2) 3.8–3.5 Ga: Co-existence and interaction of plume tectonics and embryonic plate tectonics;

(3) 3.5–2.85 Ga: Mainly plume-related development (probable existence of the first stable continents);

(4) 3.05–2.75 Ga: Mainly plate-tectonic development (origin, evolution and accretion of ancient island arc systems, collision of microcontinents);

(5) 2.79–2.5 Ga: The first prominent superplume event recognizable worldwide (interaction of plume- and plate-tectonic processes initiated by the global superplume);

(6) 2.53–2.3 Ga: Superplume and failed rifting of the Archaean continent;

(7) 2.3–2.1 Ga: Quiescent within-plate development, diffuse rifting;

(8) 2.2–1.8 Ga: The second prominent superplume event recognizable worldwide (plume-driven plate-tectonic processes);

(9) 2.0–1.7 Ga: Interaction of plate tectonic and plume tectonics, and formation of intercontinental collisional orogens;

(10) 1.8–1.2 Ga: Accretion of multiple island arcs and/or mainly intracontinental development;

(11) 1.2–0.9 Ga: The third prominent superplume event recognizable worldwide (origin of the Grenville–Sveconorwegian Orogen);

(12) 0.87–0.55 Ga: Superplume event mainly recognizable in Gondwana areas (origin of the Pan-African Orogen).

We thus arrive at the conclusion that reassessment of the nature of granulite-facies metamorphism, granulite-gneiss complexes, intracontinental orogens, and involvement of the intracontinental oval orogens in the discussion, suggests a significant revision of the current views on the relationship of plate tectonics and plume tectonics in geological history, as well as views on supercontinent assembly and breakup.

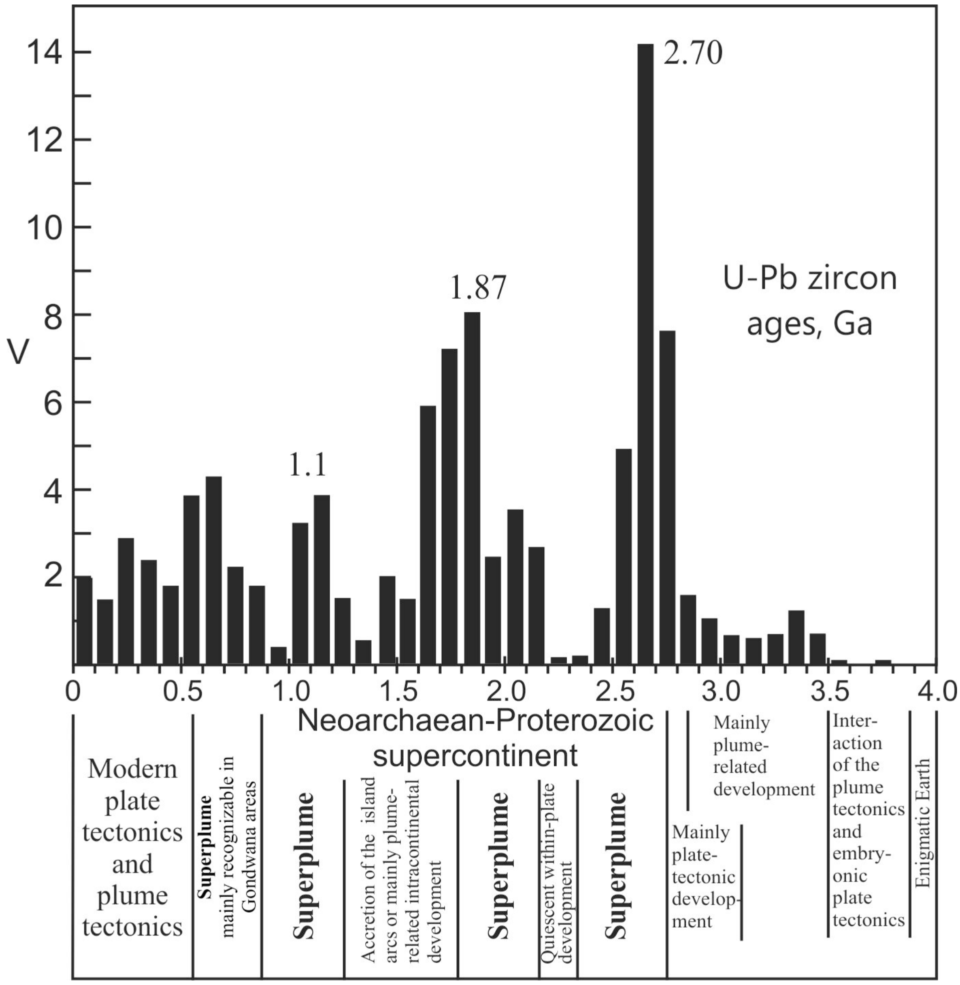

Fig. 11. Distribution of juvenile continental crust ages based on *Condie* and *Aster* [*2010*], correlated with secular changes in relations between plate-tectonic and plume-tectonic processes in the history of Lauroscandia. Modified after *Eriksson et al.* [*2013*].

Рис. 11. Распределение возрастов ювенильной континентальной коры по *Condie* and *Aster* [*2010*], сопоставленное с длиннопериодными вариациями соотношений тектоники плит и тектоники плюмов в истории Лавроскандии. С изменениями по *Eriksson et al.* [*2013*].

4.4. Secular changes in relationships between plate tectonic and mantle-plume engendered processes during Precambrian time, exemplified from East European and North American cratons

The above review is the basis for a 'spiral' model of Precambrian crustal evolution, which might be capable of explaining both the cardinal secular changes in the geological record and supercontinent cyclicity. Employing the data on the Neoarchaean, Palaeoproterozoic and Meso–Neoproterozoic of Lauroscandia, with additional evidence from other continents, we can state that the main role in the evolution and growth of the continental crust might have been played by a long-lived supercontinent (or group of large continents) and processes therein over the long interval from ~2.80 to ~1.00 Ga or maybe even to ~0.55 Ga (between the Meso-Neoarchaean and the Neoproterozoic–Palaeozoic boundaries) (see Fig. 12).

Development of accretionary orogens along the outward-facing margins of the inferred supercontinent is inferred to be of subordinate importance. The plume-related riftogenic and spreading processes within the supercontinent can be ascribed to weak or incomplete attempts to disrupt the supercontinent.

5. Conclusions

(1) The revision of protoliths, formation conditions, and history of the granulite-gneiss complexes has shown that these complexes represent an independent type of geological association. In addition to the high grade of metamorphism at high and ultrahigh

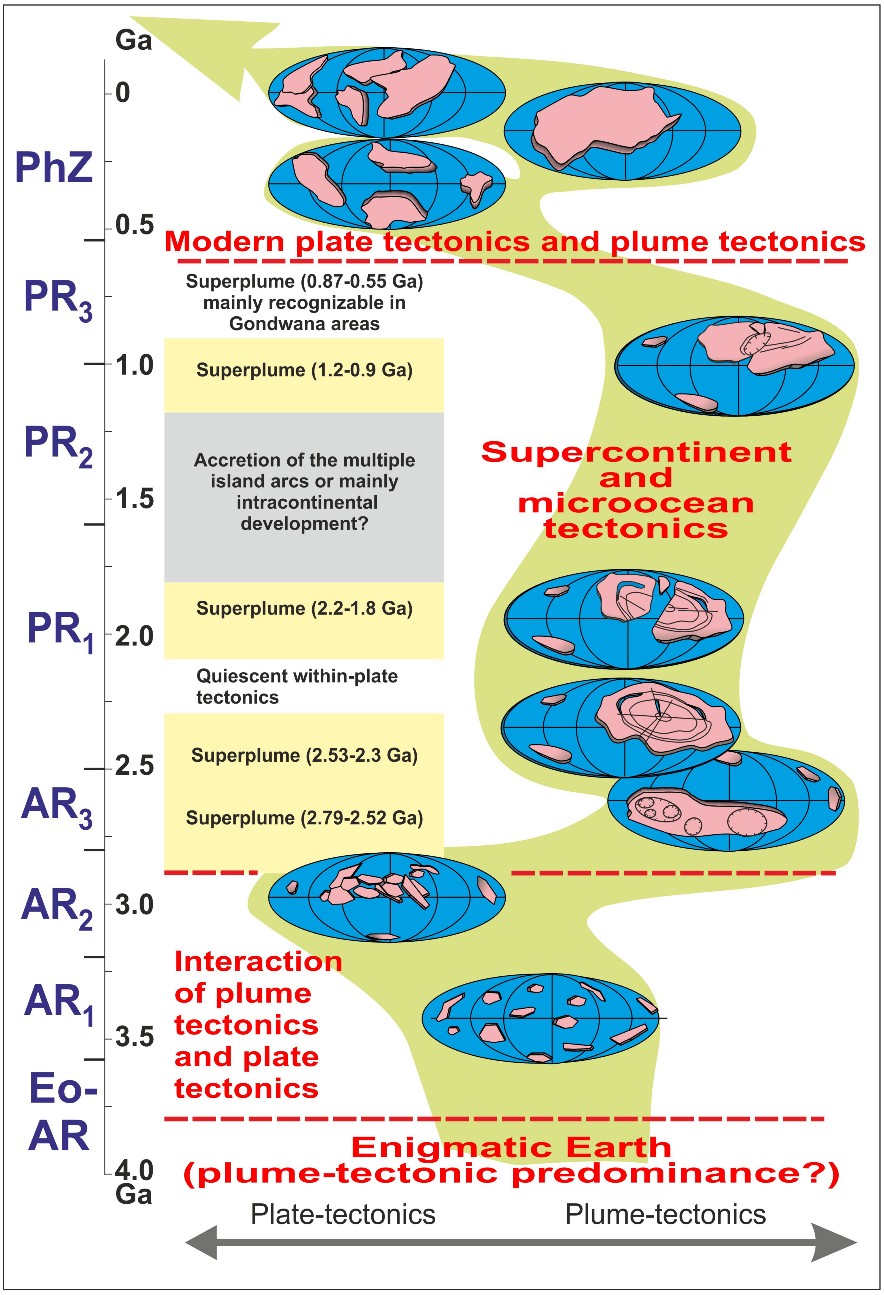

Fig. 12. Secular changes of relations between plate- and plume-tectonic processes in the history of Lauroscandia.

Рис. 12. Длиннопериодные вариации соотношений тектоники плит и тектоники плюмов в истории Лавроскандии.

temperatures, the specificity of these complexes is associated with their emergence as a result of mantle-plume activity expressed in:

– intracontinental (in some cases, a back-arc) origin and evolution;

– oval and concentric-oval morphology of the intracontinental orogens composed of granulite-gneiss complexes;

– juvenile source of the crust-contaminated mafic and partially felsic magmas.

(2) The geological record of the East European and North American cratons displays the extraordinary role of granulite-gneiss belts, intracontinental orogens as a whole and intracontinental oval orogens, in particular, in the history and structure of the Early Precambrian crust. The review of tectonic structures important for geodynamic reconstructions from other continents agrees in full measure with the results obtained for Lauroscandia.

(3) The intracontinental oval orogens formed as a result of mantle-plume activity are built up by a combination of high-grade granulite-gneiss and low-grade sedimentary-volcanic belts.

(4) The East European and North American cratons

are fragments of the long-lived Lauroscandia supercontinent. After formation of these cratons ~2.8 Ga ago, the crust of the supercontinent evolved under the influence of vigorous mantle plumes (superplumes) up to ~0.85 Ga. During that time, Lauroscandia was subject to rifting, partial breakup, and succeeding reconstruction.

(5) The reappraisal of the tectonic role of granulite gneiss belts and fields creates the basis for an alternative model that suggests the existence of a long-lived Neoarchaean-Proterozoic (~2.8–0.55 Ga) supercontinent. The plate-tectonic processes in the supercontinent were controlled by superplumes (rifting with transition to spreading and closure of short-lived oceans by means of subduction).

(6) Episodic growth of Precambrian juvenile continental crust in Lauroscandia and in other continents corresponds to the global trend. The plot of the peaks in global ages [*Condie, Aster, 2010*] corresponds to the maximums of superplume activity.

(7) The geodynamic processes resulting in formation of the Earth's crust are represented as a spiral sequence: interaction of mantle plumes and embryonic microplate tectonics in Palaeo- and Mesoarchaean (3.8–2.8 Ga) → predominance of mantle-plume activity within supercontinent in Neoarchaean and Proterozoic (2.8–0.55 Ga) accompanied by local manifestations of plate tectonics initiated by mantle plumes → Phanerozoic plate tectonics along with a reduced role of mantle plumes. This evolution model drastically differs from the conventional model of discrete forward crust formation.

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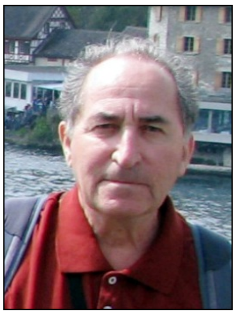

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