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# THE VOLGA-DON COLLISIONAL OROGEN IN THE EAST EUROPEAN CRATON AS THE PALEOPROTEROZOIC ANALOGUE OF THE HIMALAYAN-TIBETAN OROGEN

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**ABSTRACT.** The ca 2.0 Ga Volgo-Don fold-and-thrust belt, about 500 km in width and at least 600 km in length, covering an area of about 300000 square kilometers intervenes between the Archean Sarmatian and Volgo-Uralian proto-cratonic blocks of the East European Craton, both of which are coupled with 200–300 km thick sub-continental lithospheric mantle keels. The focus of this paper is the elucidation of its nature in order to answer the basic question how this and other thrust-and-fold belts could be formed in the Paleoproterozoic, and whether they are the same as or different from modern collision orogens. The active Himalayan-Tibet orogen is commonly thought of as the most extensively studied large, bi-verging fold-and thrust belt continental collision zone which may provide insight into key tectonic mechanisms for an understanding of orogenic processes in the Earth's geological past. Precambrian orogens are tentatively perceived yet as something that was distinct from recent orogenic styles and was due to the initial elevated geotherm and higher radiogenic heat production in the early Earth.

In this paper we report for the first time the revealation of the large, slightly eroded divergent Paleoproterozoic Volgo-Don orogen which is mostly composed of juvenile metasediments and comprises well-preserved patterns of the crustal orogenic architecture which are characteristic of the archetypal Himalayan-Tibet collisional orogen rather than of hot/ultra-hot Precambrian orogens based on numerical modeling.

**KEYWORDS:** Early Precambrian; lithosphere; thermal model TC1; orogen; crustal architecture; collision

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# ВОЛГО-ДОНСКОЙ КОЛЛИЗИОННЫЙ ОРОГЕН ВОСТОЧНО-ЕВРОПЕЙСКОГО КРАТОНА КАК ПАЛЕОПРОТЕРОЗОЙСКИЙ АНАЛОГ ГИМАЛАЙ-ТИБЕТСКОГО ОРОГЕНА

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**АННОТАЦИЯ.** Волго-Донской складчато-надвиговый пояс, возникший около 2.0 млрд лет тому назад, занимает площадь около 300000 км$^2$ (~500 км в ширину и ~600 км в длину) и располагается между архейскими протократонными Сарматским и Волго-Уральским блоками Восточно-Европейского кратона, которые подстилаются мощными, 200–300 км, сублитосферными мантийными килями. Целью настоящей статьи является выяснение природы его происхождения, для того чтобы ответить на фундаментальный вопрос о том, как этот и другие складчато-надвиговые пояса могли формироваться в палеопротерозое и был ли стиль орогенеза того времени схожим с таковым современных коллизионных орогенов или отличным от него. В качестве тектонотипа коллизионной геодинамики принято рассматривать хорошо изученный дивергентный Гималайско-Тибетский орогенический пояс, особенности развития которого, как правило, служат основой для расшифровки орогенических процессов в геологической истории Земли. Однако для раннего докембрия широко распространены представления о том, что орогенические процессы того времени должны были сильно отличаться от современного орогенеза вследствие высокого геотермического градиента в коре, обусловленного повышенной радиоактивной теплогенерацией.

В статье авторы детально рассматривают глубинную тектонику палеопротерозойского Волго-Донского орогена, реконструкция которого свидетельствует о том, что он представляет собой слабо эродированную орогеническую постройку дивергентной архитектуры; она сложена преимущественно ювенильными метаосадками, фазы ее развития сопоставляются с историей становления Гималайско-Тибетского коллизионного орогена, но не согласуются с представлениями о «горячем/ультрагорячем» стиле орогенеза в раннем докембрии, базирующимися, прежде всего, на результатах численного моделирования.

**КЛЮЧЕВЫЕ СЛОВА:** ранний докембрий; литосфера; термальная модель TC1; ороген; глубинная структура земной коры; коллизия

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## 1. INTRODUCTION

The mechanisms of the Precambrian orogeny still remain debatable. Tectonic settings of numerous Precambrian fold-and-thrust belts are comparable in many ways to those established for present-day orogenic belts (e.g. [Windley, 1992; Cawood et al., 2009; Condie, Kröner, 2013]). However, there are also some speculations based on the fact that in consequence of the initial elevated geotherm and higher radiogenic heat production in the Precambrian, an orogenic convergence should mostly have resulted in uniform lithospheric deformation and consequently would have given rise to the formation of low topographic hot/ultra-hot orogens [Cagnard et al., 2006, 2011; Chardon et al., 2009; Gapais et al., 2009; Sizova et al., 2014; Perchuk et al., 2018]. On the other hand, [McKenzie, Priestley, 2008, 2016] have highlighted the resemblance of the lithospheric structure of Tibet and the surrounding mountain ranges to that of Archean-Proterozoic cratons. They have argued that the depleted thick (200–300 km) cratonic lithosphere could have been also produced by shortening in order (i) to provide a crustal thickness of 60–80 km, and (ii) to transport depleted mantle material downwards to form cratonic keels. In addition to that, the thermal structure of the Tibetan middle-lower crust is not too different from that deduced for the Precambrian high-grade terrains, which were formed at pressure of 0.8–1.0 GPa and temperature of 800–1000 °C despite the fact that the rate of crustal heat generation in the Precambrian was greater than it is now.

However, as was recognized by [Griffin et al., 2009], the mantle lithosphere beneath the Archean cores of the Precambrian cratons world-wide was the most depleted whereas in succeeding geological eras it became progressively less depleted. Based on the well-studied suites of xenoliths and xenocrysts from sub-continental lithospheric mantle, this inference is in good agreement with the results from the global thermal modeling of the Precambrian lithosphere. I.M. Artemieva and W.D. Mooney [Artemieva, Mooney, 2001] have revealed a clear global trend in a progressive thinning of the continental lithosphere from 250 ±70 km in the Archean lithosphere to 200±50 km in the Paleoproterozoic and to ~140±40 km in the Mesoproterozoic lithosphere.

The role of the lithospheric mantle heterogeneity in tectonic interpretation of the Precambrian cratons, which are largely buried under the post-cratonic sedimentary basins, is poorly constrained. The purpose of this paper is to re-examine the evolution of the continental lithosphere of the

East European Craton (EEC), with highlighting the nature of the Paleoproterozoic orogenic belts which join the main crustal segments of the EEC to each other. Our key object is the Volgo-Don orogen and its crustal architecture to examine how collisional thrust-and-fold belts were formed in the Paleoproterozoic, and in to what extent they are identical to or different from modern collision orogens.

## 2. TECTONIC FRAME OF THE EAST EUROPEAN CRATON

### 2.1. General

The EEC is built up from a tectonic collage of the Archean and Paleoproterozoic rocks, which forms a crystalline basement of the vast East European (Russian) Platform (EEP). As of now, there is a common consensus that the EEC evolved from an amalgamation of three individual Precambrian crystal segments, each of which has its own tectonic history: Fennoscandia in the northwest and north, Sarmatia in the south and southwest, and Volgo-Uralia in the east [Bogdanova, 1993; Gorbatschev, Bogdanova, 1993] (Fig. 1). Over the years this subdivision was generally adopted as the working model for tectonic studies of the EEC and its realm. An important prerequisite for this dividing was the spatial pattern of the Meso-Neoproteroic transcratonic rift-aulacogen systems which tentatively tend to follow the Paleoproterozoic junction zones of the three major

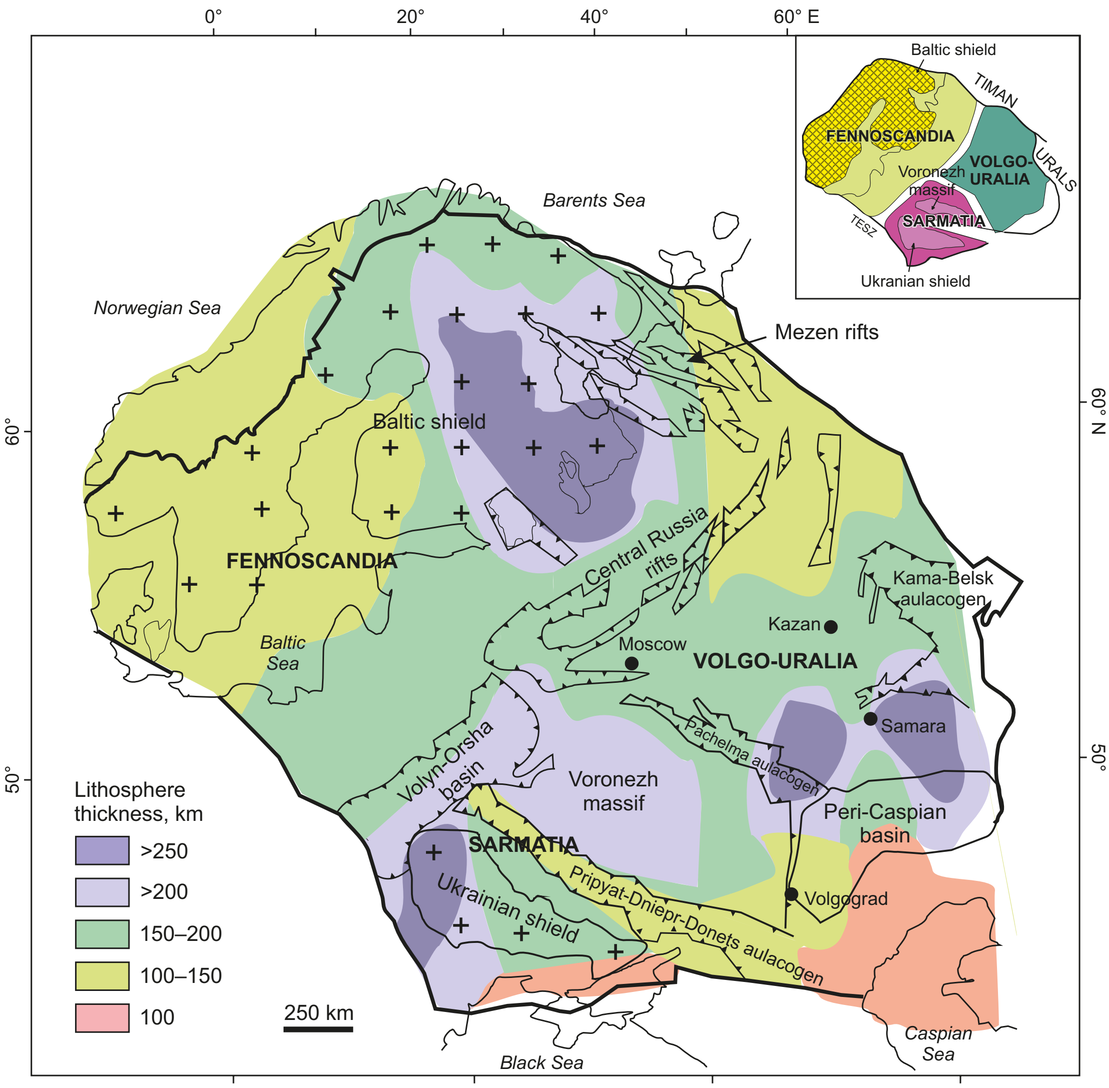


**Fig. 1.** Distribution patterns of the Meso-to-Neoproterozoic rifts and aulacogens beneath the sedimentary cover of the Russian Platform [Bogdanova et al., 2016] on the schematic map of thermal thickness of the lithosphere of the East European Craton (EEC) (modified from [Artemieva, 2007]). The Pripyat-Dnieper-Donets aulacogen is of Late Devonian age. See text for explanation. The inset presents a sketch illustrating the three basic crustal segments of the EEC [Bogdanova, 1993]. TESZ – Trans-European suture zone.

**Рис. 1.** Расположение мезо- и неопротерозойских рифтов/авлакогенов под осадочным чехлом Русской плиты [Bogdanova et al., 2016] на схематизированной карте термальной мощности литосферы Восточно-Европейского кратона (ВЕК) по [Artemieva, 2007]. Припятско-Днепрово-Донецкий авлакоген формировался в позднем девоне. Пояснения см. в тексте. На врезке показано схематическое изображение трех основных сегментов земной коры ВЕК [Bogdanova, 1993]. TESZ – Трансъевропейская шовная зона.

crustal segments [Bogdanova et al., 1996]. This statement is principally true for the Central Russian rift system and the Pachelma aulacogen, and yet it is irrelevant to the slanting Mezen rifts, as is clear also from the comparative examination of their deep crustal architectures attending to different mechanisms of their origin [Kostyuchenko et al., 1999]. The overwhelming majority of the EEP rift-aulacogen systems began to expand at ca 1.4 Ga and were slowly developing for a long time during the Meso- and Neoproterozoic eras, i.e. more than 400 Ma [Bogdanova et al., 2008]. An important point is that these systems are avolcanic and filled with terrigenous and carbonate deposits of several kilometers in thickness.

The 1500-km-long Pripyat-Dnieper-Donets Aulacogen (PDDA) which bisects the Sarmatian crustal segment and separates the Ukrainian shield from the Voronezh massif, stands out against a riftogenous background. The PDDA evolved during the Late Devonian and is marked by saliferous-clastic-carbonate and effusive sequences with a total thickess of 4 km, whereas the post-rift thermal subsidence accommodated to the Carboniferous volcanic-free sediments (limestone, siltstone, salts, coals) with a thickness of up to 15 km [Stovba et al., 1996]. The syn-rift volcanic activity occurred in the Late Devonian as a result of two major phases of the Sarmatian crustal stretching, in Late Fransnian and Late Famenian. Chemically, the volcanic community includes alkali basalts, basanites, nephilinites, picrites, trachytes, and rhyolites that derived from high-degree partial melting of an enriched deep mantle or plume source at potential mantle temperatures that were at least 200 K above those in ambient mantle [Wilson, Lyashkevitch, 1996].

Two groups of the riftogenous structures, i.e. Meso-Neoproterozoic and Phanerozoic, also differ in deep seismic crustal construction. Beneath the PDDA, the crustal basement thickness is has currently been decreased from 7 to 20 km, as compared with that in the adjacent Ukraine shield and Voronezh massif [Ilchenko, 1996]. Even more thinning of the crystalline crust occurs in the central part of the Peri-Caspian depression which is thought of as being genetically related to the PDDA [Artemieva, 2007]. On the contrary, the crustal thickness along the Meso-Neoproterozoic rifts within the intersegment junction zones does not show noticeable decrease as compared with their flanks. The regional structural patterns suggest the Phanerozoic PDDA stretching due to pure shear (β (stretching) factor of up to 1.3) [Kusznir et al., 1996], whereas the Meso-Neoproterozoic rifts were developed in accordance with a simple shear model (β<1) [Bogdanova et al., 2008].

Since, as opposed to the Phanerozoic plume-related rifting, the intersegment Meso-Neoproterozoic rifting could not largely affect lithospheric behaviors of the EEC (cf. [Artemieva, 2003]), there is a need to re-examine the origin of the junction zones between its major crustal segments. The nature of the junction zones has been previously considered either as a result of rigid Archean plates collision [Bogdanova et al., 2008] or a product of an intracontinental orogenic plume-related process [Mints et al., 2015]. In both cases it is expected that the lithosphere of the EEC has been chiefly formed in the Archean.

To verify these models, we used the global thermal model TC1 developed by [Artemieva, 2006] to account for the EEC tectonic evolution [Artemieva, 2007]. Furthermore, we have compiled recent isotope-geochronological data derived mostly from deep drilling cores in order to clarify how the age-depending thermal model TC1 is in line with geological observations. Such methodology coupled with the available data on seismic profiling provides insight into the nature of the junction zones of the EEC, as for tectonics of the Precambrian collisional orogeny as a whole.

### 2.2. Archean *vs.* Paleoproterozoic lithosphere

As can been seen from Fig. 1 and Fig. 2, the supposedly Archean lithosphere of the EEC with a thickness of more than 200 km occupies a floor space of about one third of the total area of the craton. Unlike the thermal model TC1, the P- and S-wave seismic tomography modeling of cratonic lithosphere (Fig. 2, left inset) provides only a rough idea about the lithosphere and gives no way to efficiently differentiate crustal terrains by age and composition. Perhaps, the best illustration for the validity of the TC1 model is the Fennoscandian/Baltic shield whose area was being extended west-southwards by intense growing of a juvenile Paleoproteroic continental crust due to successive accretion of several volcanic arc systems to the Archean core between 1.9 and 1.7 Ga and that prolonged tectonic event is collectively named the Svecofennian orogeny (e.g. [Korja, Heikkinen, 2005; Nironen, 1997; Bogdanova et al., 2016]), and the Svecofennian orogen has been envisaged as the key object for classifying orogens [Windley, 1992; Cawood et al., 2009]. In keeping with the rejuvenation of crustal ages, the thickness of the lithosphere is decreases from >200 km to <150 km, so that the lithospheric frontiers coincide to a first approximation with geological crustal borders.

The Archean core of the Fennoscandian shield is the largest and most pronounced in the EEC. The largest values of lithospheric thickness, 250–300 km, are found for the Karelian granite-greenstone terrain, or the Karelian protocraton. This Archean core is fringed with a slightly thinner lithosphere and giving way to an even thinner Paleoproterozoic lithosphere. The location of the currently known kimberlite pipes within the Fennoscandian shield correlates well with that inference (Fig. 2). All but one of the diamond-bearing kimberlite pipes occur on the margins of the Archean core with thick lithosphere that is typical for other cratonic cores world-wide (e.g. [McKenzie, Priestly, 2008]). By this is meant that the Archean core is coupled with a thick subcratonic mantle keel which extends deep into the field of diamond stability, up to 200–275 km [Artemieva, 2006; Lehtonen, O'Brian, 2009]. The lithosphere of margins of the Archean core can be thinned by as much as 50 km that can be related to emergence of a new uppermost mantle layer that resulted from the Paleoproterozoic metasomatism of a previous upper mantle depleted by slab-derived fluids [Peltonen, Brügmann, 2006].

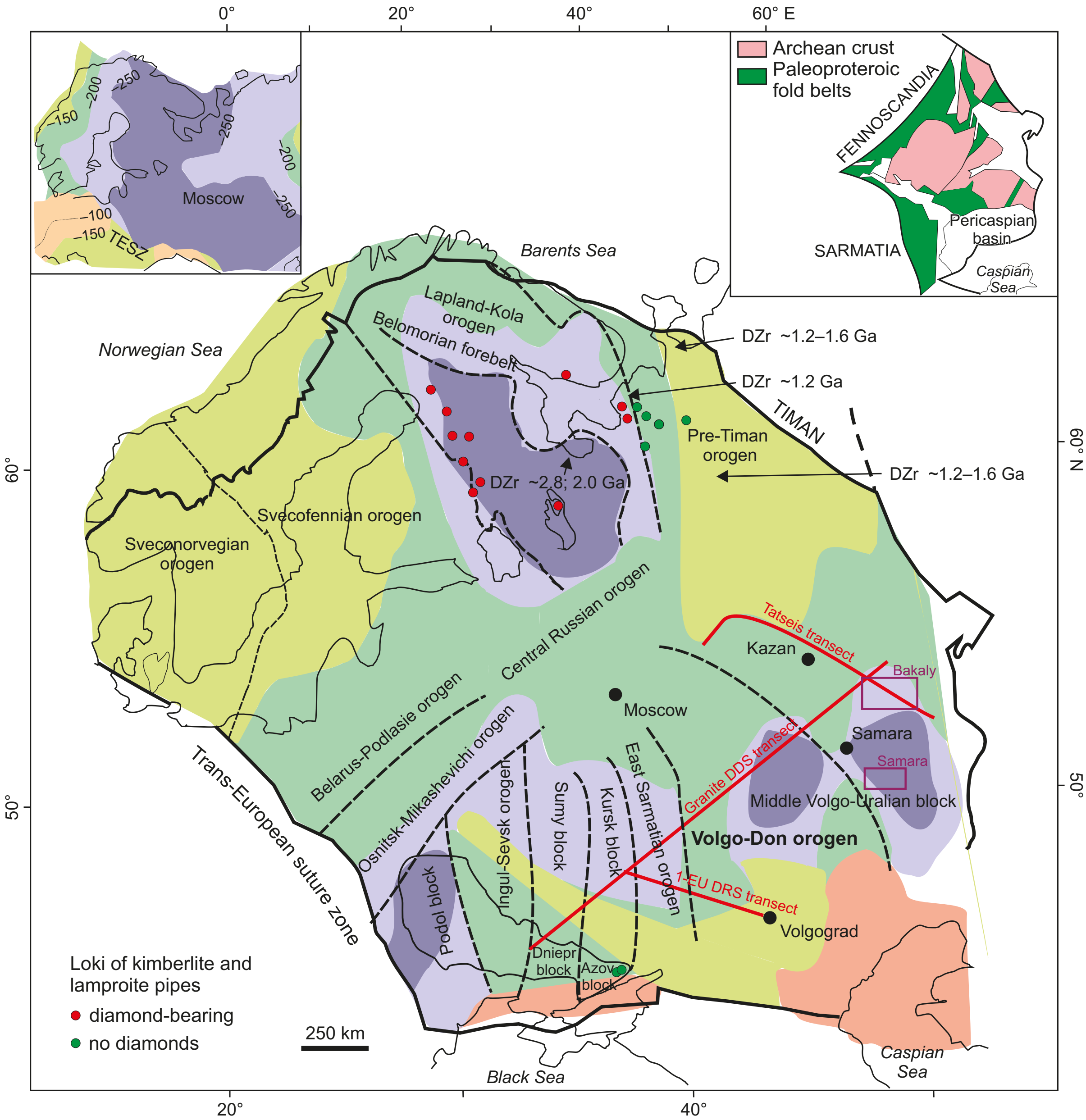


**Fig. 2.** Sketch illustrating locations of main tectonic divisions of the EEC on the map shown in Fig. 1.
The lithospheric boundaries coincide to a first approximation with the known tectonic boundaries, thus providing efficient differentiation between the Paleoproterozoic fold-and-thrust orogenic belts and the Archean proto-cratonic blocks. Left and right insets show the tomography-based EEC lithosphere model ([Artemieva, 2007], and references in this publication) and the inferred major tectonic subdivisions of Volgo-Uralia [Bogdanova et al., 2016], correspondingly. Small boxes within the Volgo-Uralian block display locations of deep drilling whose core samples have yielded the Archean SIMS U-Pb zircon ages. DZr – detrital zircon ages; the arrows mark localities of drill core samples with studied detrital zircons after [Kuznetsov et al., 2014]. The boundary between the diamond-bearing and diamond-barren kimberlite pipes in eastern Fennoscandia is drawn after [Mahotkin et al., 2000]. The red lines exhibit the seismic transects discussed in the text.

**Рис. 2.** Схема, иллюстрирующая положение главных тектонических подразделений ВЕК на карте термальной модели ВЕК (рис. 1).
Границы литосферы разной мощности в первом приближении совпадают с известными тектоническими границами, разделяя палеопротерозойские орогенические пояса и архейские протократонные блоки. Левая и правая врезка показывают образ ВЕК в сейсмической томографии [Artemieva, 2007; и ссылки в этой работе] и предполагаемые тектонические блоки Волго-Уральского сегмента ВЕК [Bogdanova et al., 2016] соответственно. Малые прямоугольники на схеме показывают площади глубинного бурения, при котором из керна скважин был получен архейский возраст циркона SIMS U-P методом локального датирования. DZr – возраст детритовых цирконов; стрелками показаны места отбора проб на выделение детритовых цирконов по [Kuznetsov et al., 2014]. Граница, проходящая в восточной части Фенноскандии между алмазоносными кимберлитовыми трубками и кимберлитовыми трубками, не пригодными для промышленного использования, приведена по [Mahotkin et al., 2000]. Красными линиями показаны сейсмические трансекты, рассматриваемые в статье.

The cratonic core edges of this sort are in agreement with the field observations indicating a strong Paleoproterozoic tectonic reworking of the Archean crust. In paleotectonic terms, these edges fit into forebelt/foreland patterns of the evolution of the Paleoproterozoic orogens.

The northern and eastern lithospheric fringes of the Archean Karelian core, known on the crustal surface as the Belomorian high-grade belt, are largely considered as being involved with the Lapland–Kola collisional orogeny at about 1.94–1.90 Ga (e.g. [Daly et al., 2006; Bogdanova et al., 2016; Lahtinen, Huhma, 2019]). The eastern delimitation of the Belomorian forebelt beyond the shield remained under the question (cf. [Mints et al., 2015; Bogdanova et al., 2016]). Yet referring to Fig. 2, the Archean lithospheric keel is quickly eradicated eastwards and is replaced by the substantially thinner lithosphere, as much as it occurs in the western part of the shield. This is well evidenced from the locus of the Devonian diamond-bearing and diamond-barren kimberlite pipes [Mahotkin et al., 2000] implying that the northern part of the Volgo-Uralia crustal segment is rather lacking in discernable portions of the Archean crust underlain by the mantle keel (cf. Fig. 2, right inset). Because of a very thick sedimentary cover, little is known about the basement tectonism of this part of Volgo-Uralia. Results from the deep seismic sound (DDS) profiling across therein have shown that the basement is only 30–32 km thick and contains abundant mafic rocks, which may be attributed to crustal rifting of a Precambrian age [Kostyuchenko et al., 1999]. On the other hand, main lithospheric features of the Pre-Timan province correspond closely with those of the Mesoproterozoic Sveconorwegian orogen on the west of Fennoscandia. In that instance, this province can be ascribed to Mesoproterozoic accretionary orogen, whose collapse, induced by the Pre-Timan tectonism, led to its collapse was followed by prolonged tectonic thermal subsidence attended by rifting of an adjacent foreland which involved both Paleoproteroic and Archean crust. This inference is corroborated by the detailed dating of detrital zircons (DZr) collected from the deepest horizons of the Cambrian platform cover and Neoproterozoic Mezen-rift sediments until which rare deep wells have been drilled. Kuznetsov and coauthors [Kuznetsov et el., 2014] have revealed that a great majority of the zircon grains dated yields the Mesoproterozic ages of ca 1.2–1.6 Ga stemmed from a juvenile crustal source, mostly dioritic in composition, with $\varepsilon_{Hf}(t)>0$. Just a few units of the zircon grains have Archean ages suggesting their distal provenance from the Archean core (Fig. 2). Tectonic crustal boundaries between the distinct lithospheric blocks are uncertain in the northern part of Volgo-Uralia (cf. [Mints et al., 2015; Bogdanova et al., 2016]). One might expect that they are faced with thrust-sense shear zones characteristic of recent accretionary-type orogenic belts (e.g. [Gee et al., 2006]). A related question is how far rocks of the Pre-Timan Province are extended to the south. As may be inferred from Fig. 1 and Fig. 2, the lithosphere with the Meso- to Neoprotroterozoic characteristics extends well beyond current ideas of the Volgo-Uralia crustal volume.

The rejuvenation of the lithosphere within the Sarmatian crustal segment was going another away. The total thickness of lithosphere in the core of this crustal segment is lesser than that of the Fennoscandian crustal segment (see Fig. 1). As was mentioned above, at the Late Devonian Sarmatia was cut up by the PPDA into two parts, the Ukrainian shield and the Voronezh massif, without any sensible displacement of the basement boundaries that trend mostly N-S [Shchipansky, Bogdanova, 1996]. The Ukrainian shield demonstrates a glowing example of how later plume impingements have influenced the lithosphere of Precambrian cratons. Here, the eradication of the former lithospheric keel occurs throughout the shield, across its intracrustal boundaries along with the course of the PDDA.

Until recently, it has long been thought that the Precambrian crust of the Ukrainian shield consists mainly of Archean rocks which extend from the west to the east implying that a thick lithosphere keel should have occurred underneath the whole shield (cf. Fig. 2, left inset). The implication of the thermal model TC1 for understanding of tectonism in the Ukrainian shield yields better fit to available geochronological data. The Archean lithosphere keel of 200–250 km in thickness preserved solely in the westernmost part of the shield that is in excellent agreement with the available zircon data from high-grade tonalite (enderbite) of the Podol block which yield the Paleo- to Eoarchean isotopic ages in the range 3.65–3.75 Ga [Claesson et al., 2006]. The oldest crust was largely reworked in the Mesoarchaen and Paleoproterozoic as judged from the zircon ages but this reworking has failed to eradicate the Archean lithosphere properties. Further to the east, the thickness of lithosphere decreases correspondingly to the Ingul-Sevsk building block composed mainly of a 2.3–2.1 Ga juvenile orogenic crust (e.g. [Bogdanova et al., 2016]). The next Middle-Dnieper block has a dome-and-keel structural pattern and represents the East Pilbara-type, i.e. Mesoarchean granite-greenstone terrain formed at 3.2–3.0 Ga. The easternmost block of the Ukrainian shield herein referred to as the Azov block evolved likewise the crustal growth in the Podol block [Bibikova et al., 2010], but it displays a substantially thinner lithosphere. This suggests that an Archean lithospheric keel was essentially removed from underneath the most part of the Ukrainian shield by underplating due to asthenospheric upwelling during the Late Devonian rifting. Indeed, as has been inferred from the petrological studies on mantle xenocrysts from diamond-barren kimberlite pipes occurred in the easternmost edge of the Azov block, the lithosphere mantle essentially involved peridotites enriched both in Fe and LREE elements while depleted peridotites were only a small part of the mantle variety [Ashchepkov et al., 2021]. Noteworthy that the mantle xenocrysts formed at peak pressures about 4.2 GPa which correspond to a depth of about 140 km, and the lithosphere thickness inferred from these data matches well with the thermal model C1 (Fig. 2).

The northern counterparts of the Ukrainian crustal tectonic subdivisions form the Precambrian crust of the Voronezh massif. As evident from Fig. 2, its lithosphere maintains the Archean attributes implying some tectonic processes later on. The crystalline basement of the Voronezh massif is buried beneath the Phanerozoic platform cover ranging in thickness from tens of meters in the anticline hinge to thousands of meters on its slopes. Thus, the principal sources of information on the Precambrian geology of this region are subsurface samples obtained from a great deal of wells and detailed geophysical maps. Recent advances in isotope-geochemical studies on drill core samples acquired from a wealth of the Voronezh massif provided vital data about the timing of major tectonic events that led to the formation of an the Archean crust as an essential precursor to subsequent Paleoproteroic collisional events [Savko et al., 2021]. As has been found, in the Kursk Block early crust-forming processes took place mainly in the Paleoarchean and in the Mesoarchean just as it happened in the Azov Block.

The most enigmatic part of the EEC is the Volgo-Uralian crustal segment which is hidden entirely beneath later cover deposits of profound thickness. So far, it is anticipated that the Archean crust should occur within very large (≥300 km across) ovoid negative magnetic anomalies which form domal crustal structures and are surrounded by tight curved or linear positive magnetic anomalies [Bogdanova, 1986; Bogdanova et al., 2016; Mints et al., 2015]. A large discrepancy exists between the afore-mentioned models concerning both the timing of dome-forming events and their tectonic origin, i.e. Archean vs. Paleoproterozoic, or plume-related tectonism vs. orogenic reworking [Bogdanova et al., 2021]. Turning to the thermal model of the lithosphere of the EEC, one can easily see that the only tectonic block that displays the Archean lithospheric properties is herein referred to as the Middle-Uralian Block. It is noteworthy that all radiologic data showing the Archean ages have been only obtained from two small parts of this block named as the Bakaly and Samarian locations [Bogdanova et al., 2010, 2021] (Fig. 2). We cannot rule out that some small Archean crustal terranes can be found within large areas with the Paleoproterozoic lithospheric properties. In any case, the Volgo-Uralian lithospheric mantle attributes testify that the tectonic frame of Volgo-Uralia consists of the Paleoproterozoic rather than of the Archean crust.

An intriguing issue from the thermal model of lithosphere for the EEC is as a follows: what would constitute the crustal structure at the junction zone between the two Archean protocratons which both are coupled with thick mantle keels. Although this junction zone has been tentatively assigned to Paleoproterozoic collisional tectonics [Bogdanova et al., 2008, 2016], many important details of that process have remained unclear, especially when compared to modern collisional orogens. Currently this junction zone is referred to as the Volga-Don orogen because of its geographic occurrence between the largest rivers of the EEP [Bibikova et al., 2009; Bogdanova et al., 2016; Terentiev et al., 2020]. As opposed to other fold-and-thrust junction zones of the EEC, the Volga-Don orogen has been better studied in the context of both age dating and seismic profiling to constrain tectonic implications.

## 3. HOW THE VOLGA-DON OROGEN WAS DEVELOPED

### 3.1. Geological background

A pre-history of the Volga-Don orogen goes back to the early Paleoproterozic or even late Neoarchean (Fig. 3) when the eastern part of Sarmatia and the western part of Volgo-Uralia experienced a mantle plume impingement and, perhaps, rifting of their margins at 2.6–2.5 Ga that corresponds to the timing of the Kenorland supercontinent break-up [Savko et al., 2019; Bogdanova et al., 2021].

The early Paleoproterozoic Kursk group 3.0–4.0 km thick, associated with banded iron formations (BIFs), lay unconformably on the Archean basement, and tectonically evolved from a passive margin to fold-and-thrust ferruginous linear belts (Fig. 4). A clear similarity in the lithostratigraphic BIF-bearing sequences from the Voronezh massif and Ukrainian shield attests that a vast shallow-water sedimentary basin developed atop the Archean peneplained stable basement over a long Siderian period of the Paleoproterozoic [Shchipansky, Bogdanova, 1996]. Then this basin was tectonically transformed into dismembered structures that preserved as fold-and-thrust belts produced by later Paleoproterozoic (Orosirian) orogenic events [Shchipansky et al., 2007].

The first evidence of Sarmatia and Volgo-Uralia plate convergence is found in the eastern edging of Sarmatia which evolved as an active continental margin. In domestic terms it is known as the volcanic Lipetsk-Losev belt, about 150 km wide and at least 450 km long (Fig. 4). It is largely composed of low- to medium grade metamorphosed tholeiites, felsic volcanics and related granites, as well as volcano-sedimentary lithologies. Geochemically, the tholeiites display an arc-related affinity indicated by some enrichment in LILE and LREE and an appearance of negative Nb anomalies while Nb-enriched basalts also occur. The felsic volcanics are dacite to rhyolite in composition and correspond to the calc-alkaline volcanic series. Their trace element patterns exhibit highly fractionated REE patterns with $La/Yb_N$>10, a high Sr/Y ratio (>40), and elevated Ni and Cr contents that matches well with the composition field of adakites [Martin et al., 2005]. Spatially associated plagiogranite intrusions are adakite-like trondhjemites in chemical compositions and may be thought of as the magmatic counterpart of felsic volcanics. Thus the volcanism of the Lipetsk-Losev belt was dominantly bimodal as is typical for volcano-plutonic edifices of active continental arcs (e.g. [Defant et al., 1992; Defant, Drummond, 1993]). At the eastern Sarmatian edge the age of the crustal growth onset is assessed upon zircon dates at around 2.1 Ga [Terentiev et al., 2016]. These zircon dates go well together with the results earlier obtained from the elucidation of Rb-Sr and Sm-Nd isotope characteristics of the Lipetsk-Losev volcano-plutonic assemblage which evidence a juvenile source of crustal growing at that time [Shchipansky et al., 2007]. In later times at about 2.07 Ga large granitoid batholiths, differentiated compositionally from diorite through monzonite to normal granite, were emplaced along the boundary of the Lipetsk-Losev belt with the Archean crust of the Kursk block and entrained older crustal components in

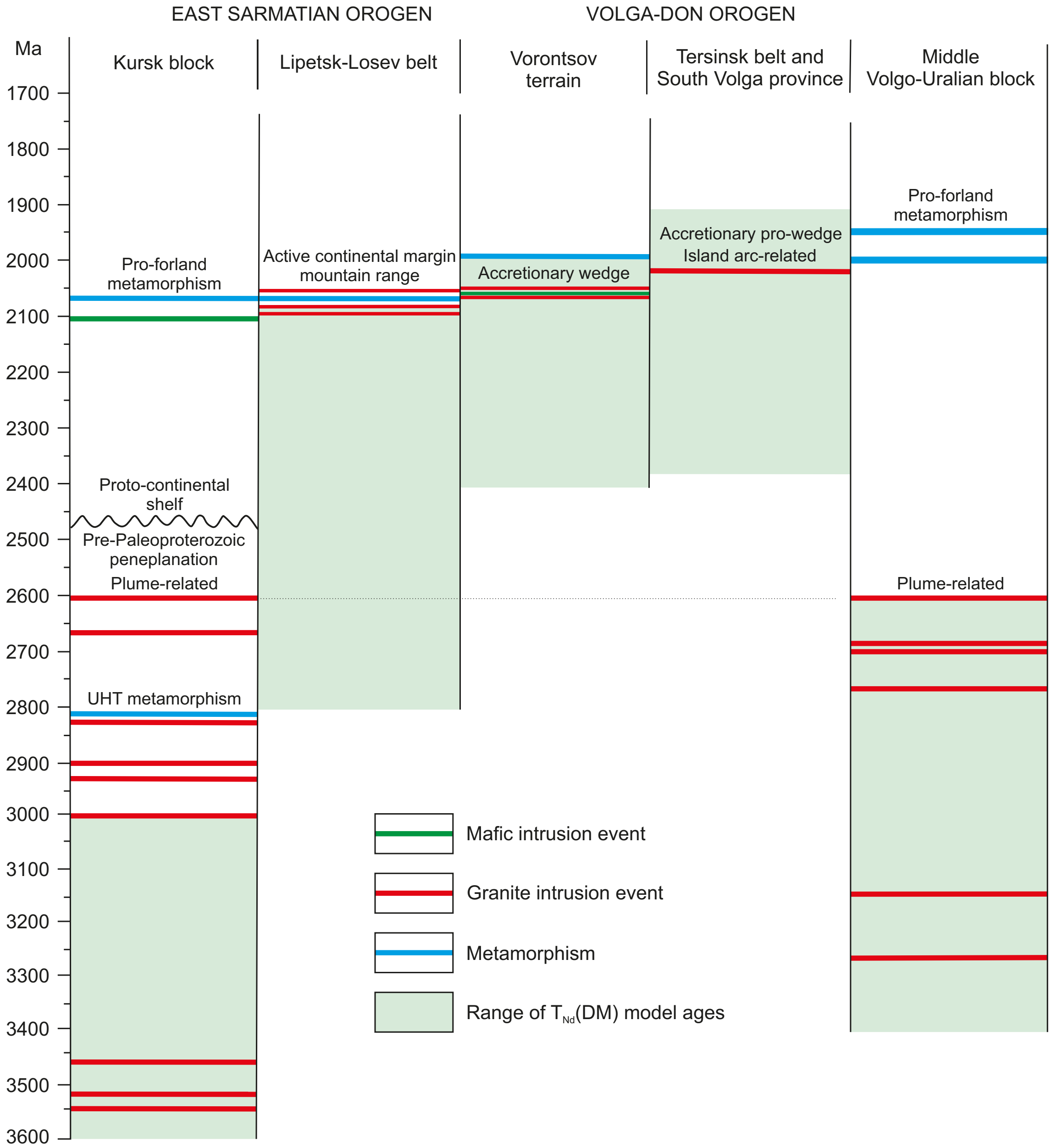


**Fig. 3.** Time–space plot for the Volga-Don orogeny (after [Bibikova et al., 2009, 2015; Bogdanova et al., 2010, 2021; Savko et al., 2011, 2014, 2018, 2019, 2021; Shchipansky et al., 2007; Terentiev et al., 2016, 2020; Fedotova et al., 2019]).

**Рис. 3.** Геохронологическая корреляция событий развития Волго-Донского орогена (по [Bibikova et al., 2009, 2015; Bogdanova et al., 2010, 2021; Savko et al., 2011, 2014, 2018, 2019, 2021; Shchipansky et al., 2007; Terentiev et al., 2016, 2020; Fedotova et al., 2019]).

their petrogenesis [Terentiev et al., 2016]. Available scarce data on metamorphic assemblages from the Losev-Lipetsk rocks suggest that the peak metamorphism reached temperature of 500–700 °C and pressure of 0.4–0.6 GPa [Savko et al., 2019] which roughly corresponds to a burial depth of about 20 km. Taking into account that the current crustal thickness is about 40 km at this area, the crust could have been reached a maximum thickness of about 60 km during the mountain building similar to the Andean mountain range at the average elevation of about of 4000 m where adakite volcano-plutonic suites are also of widespread occurrence (e.g. [Martin et al., 2014]).

A unique feature of the eastern edge of Sarmatia is that a revealing case of a Paleoproterozoic forearc sedimentary basin can be found east of the Losev-Lipetsk Belt of the continental arc affinity. This is known as the Vorontsovka/Vorontsov terrain covering an area of about 300000 km$^2$, 500 km in width, and 600 km in length. At present it is adjacent to the Losev-Lipetsk belt via the large regional-scale Losev-Mamon fault. The terrain is almost exclusively formed in flysch-type shale-sandstone sedimentary deposits combined into the Vorontsov Group. This group was deformed and underwent the Barrovian-type zonal metamorphism in the T-P range of 490–750 °C and 0.3–0.5 GPa [Savko

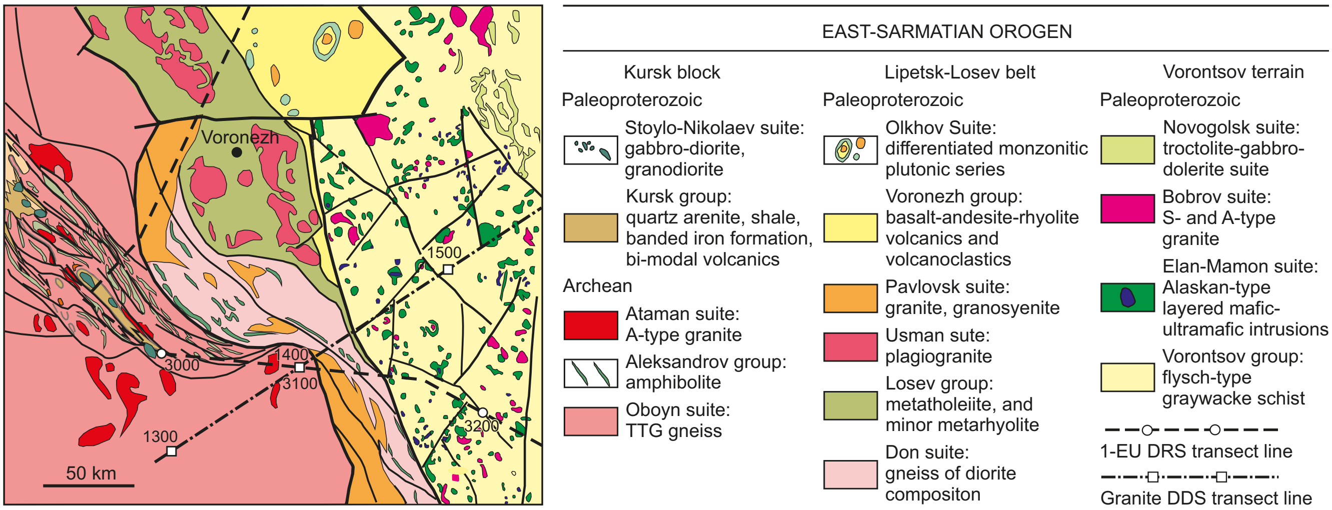


**Fig. 4.** Plot illustrating the geological background of the southern part of the East-Sarmatian orogen with the seismic transect lines therein.

**Рис. 4.** Схематизированная геологическая карта южной части Восточно-Сарматского орогена с линиями сейсмических профилей этой части.

et al., 2018]. As has been highlighted, the highest-grade metamorphism at ca 2.07 Ga is documented nearby the Losev-Mamon fault at its southwestern edge while the lowest-grade metamorphic mineral assemblages are observed around mafic-ultramafic massifs implying that these massifs were tectonically emplaced into the upper crust. The western part of the Vorontsov Province stands out by an abundance of the 2.06–2.08 Ga old small Elan-Mamon differentiated PGE-Ni-Cu-bearing peridotite-gabbronorite intrusions scattered along a vast corridor 50–130 km wide and up to 400 km long [Chernyshov et al., 1990]. Such a type of mafic-ultramafic intrusion belongs to the Alaskan-type which occurs in recent arc-related settings [Shchipansky et al., 2007]. At the same time, the small Bobrov granite intrusions of both S-type and A-type were injected into the Vorontsov flysch sequence [Savko et al., 2014]. The semi-simultaneous emplacement of these intrusive suites suggests a pulse of magmatism, supposedly produced by magma underplating at the base of an accretionary wedge. Depleted mantle Nd model ages from the Vorontsov group fall into the range of 2.4–1.9 Ga, with two exceptions which have shown the Archean Nd model ages [Shchipansky et al., 2007; Bibikova et al., 2009]. By this is meant that a high mountain range should have existed at the current location of the Lipetsk-Losev Belt impeding a significant sedimentary input from the neighboring Archean crust into the forearc basin. In other words, the syn-orogenic sedimentation was predominated by material derived from the mountain range of the Eastern Sarmatian Andean-type orogen. At the same time, the Archean Kursk block acted as a pro-foreland part of the orogen involving the back-arc extension followed by crustal shortening which led to the formation of ferruginous linear belts and to tectonic reworking of their Archean basement.

The Vorontsov terrain has been previously speculated to be unified sedimentary basin. Authors of [Bibikova et al., 2009] were the first who established that the Vorontsov-type sediments occurring in the deeply buried eastern part of the province differ from those in the shallow buried part in which these sediments are distinctly richer in iron. Furthermore, the available radiological ages tend to decrease from the west to the east whereas metamorphic grade tends to increase. The border between the lower and upper parts of the Vorontsov terrain has been delimited by a wide tectonic zone herein referred to as the Tersinsk Thin-Skinned belt where high-grade and low-grade metamorphic rocks are intercalated with each other and intruded by calc-alkaline and normal granites (Fig. 5).

Temperatures of about 750 °C and pressures of 0.7 GPa have been reported from this part of the orogen, which also shows evidence of partial melting [Bibikova et al., 2009]. The zircon dating of calc-alkaline tonalite yielded an age of ca 2.04 Ga whereas the normal granite was dated at 2.02 Ga. East of the Tersinsk belt, the Vorontsov-type metasedimentary rocks extend across to the Archean core of Volgo-Uralia for a distance of about 250 km. An outstanding feature is that the Paleoproterozoic granulite-grade metasedimentary rocks are of widespread occurrence at that area. Such a framework gave ground to separate the eastern part of the Vorontsov group occurrence into a self-contained lithostratigraphic unit named as the South Volga group [Bibikova et al., 2009]. Previously, the high-grade granulites in the Volga-Uralia crustal segment have been traditionally ascribed to the Archean constituents referred to as the Bolshecheremshan formation (e.g. [Bogdanova, 1986; Bogdanova et al., 2008; Mints et al., 2015]). In such a situation, the basic challenge is to clarify their tectonic settings and nature of high-grade metamorphism. In the absence of opportunities for direct field observations, a deep seismic profiling serves as a vital technique to gain a better insight into tectonic mechanisms of the orogen formation throughout the Earth's geological history.

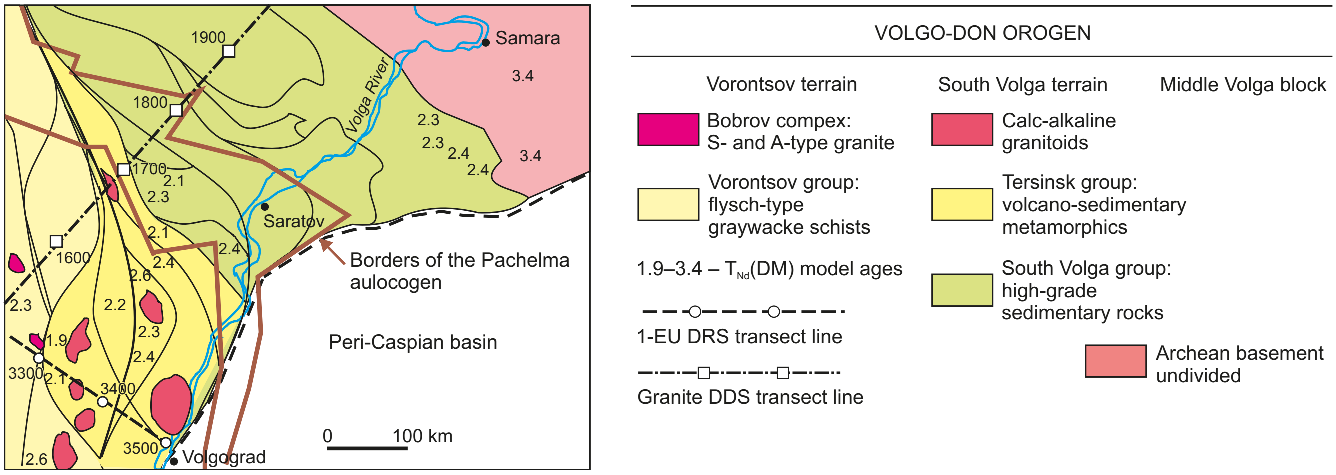


**Fig. 5.** Plot illustrating the geological background of the hinterland (core) of the Volga-Don orogen (modified from [Bibikova et al., 2009]).

**Рис. 5.** Схематическая геологическая карта, иллюстрирующая строение хинтерланда (ядра) Волго-Донского орогена (модифицирована из работы [Bibikova et al., 2009]).

### 3.2. Crustal Architecture of the Volga-Don Orogen

The Volga-Don orogen was intersected by two seismic profiles/transects (see Fig. 2). The first is a part of interest of the "Granite" transcratonic DDS profile 4500 km in length oriented from the central Ukrainian shield through the Voronezh massif and Volgo-Uralia up to the Western-Siberian plate [Sokolov, 2002]. The second is the southern part of the transcratonic 1-EU vibroseiss-source, deep seismic reflection (DRS) profile 4080 km in length [Mints et al., 2015]. The both parts of these transects were oriented across the Volgo-Don orogen that enables us to consider its geological background as viewed from deep seismic images. The DRS survey was performed with the use of the common midpoint method (CMP) described in detail by [Mints et al., 2015] and provided high-resolution seismic images of crustal sections. The previous interpretation of the Volgo-Don orogen in the above cited work favored an intacratonic setting which involves the hypothetical Archean Khoper craton buried entirely under the Vorontsov group metasediments.

Our explanation of seismic data has been based on susceptible approaches for interpreting seismic images obtained from geotransects through orogenic belts of the Precambrian cratons and modern orogens. These approaches include (i) an analysis of reflective patterns, a detection of reflector displacement, and constraints both from regional geology and isotope dating advances. Moreover, we have assumed that pronounced reflectivity in a cratonic crustal deep is commonly related to zones of tectonic ductile flows, mafic sill-like intrusions, and bimodal gneisses' fabrics whereas seismically isotropic areas are largely interpreted as granite intrusions. In addition, we have used the interpretation of the "Granite" DDS profile [Sokolov, 2002] to keep a check on the validity of our explanation of the DRS profile across the Volgo-Don orogen (Fig. 6).

As can be seen from Fig. 6, both seismic techniques revealed that there is a clear similarity in the behavior between refractory and reflective patterns which tend to change in accordance with subsurface geological borders. In either case, the Losev-Mamon fault is defined as a steep fault that extends downwards through the upper and middle crustal levels. The Kursk block is not too different from the Lipetsk-Losev belt upon reflective patterns while these tectonic units differ significantly in refractory crustal characteristics. The seismic image of the Vorontsov province stands out by scattered patchy horizons of high reflectivity within the accretionary wedge-shaped medium of low dispersive reflectivity. No large-scale seismically isotropic areas occur throughout the Vorontsov terrain, thus implying a lack of large granite intrusions. Such a situation checks well with geological frame of the province in which the Alaskan-type mafic-ulramafic intrusions are widespread amongst flyschoid metasediments. Thus the patchy zones of high reflectivity can be interpreted as folded feeders for the high-level Elan-Mamon intrusions. A series of west-verging, east-dipping thrusts can be observed along the "Granite" DDS profile from the southwest to the northeast up to the Tersinsk belt. The construction of this belt is determined by a single, steep, west-dipping fault zone at mid- to lower-crustal levels which splays upwards into a transpressive flower structure defined by steep to flat-laying, northwest- or southeast-dipping thrusts. This observation suggests that the Tersinsk belt represents a major suture zone stemmed from a final convergence of the Sarmatian and the Volgo-Uralian lithosphere plates.

Although both Vorontsov and South Volga terrains formed in similar metasediments, their seismic images differ drastically. First of all, the South Volga province displays about 10 km thick sub-horizontal band of high reflectivity which extends over 200 km southeastwards at the base of the crust up to the border of the Pre-Caspian depression. Secondly, a vergence of thrusting shows the reverse, east-verging course as compared to that at the Vorontsov province. Furthermore, the seismic image of the easternmost part of the 1-EU transect DRS profile indicates pronounced crustal-scale duplexing clearly implying that during the orogenic convergence the South Volga province acted as a pro-wedge ground and suffered a sizeable crustal stacking. Unfortunately, the 1-EU transect was not acquired through the Archean core of Volgo-Uralia (see Fig. 2), and we cannot well constrain interactions between the South Volga metasediments and the Archean basement. However, we can say with reasonable confidence that the Archean Middle Volgo-Uralian core was largely overthrust by ductile nappes, transported northeastwards from the orogenic hinterland up to several hundreds of kilometers (see Fig. 5; Fig. 6, a). The "Granite" DDS profile also confirms an orogenic construction of the Volga-Don fold-and thrust

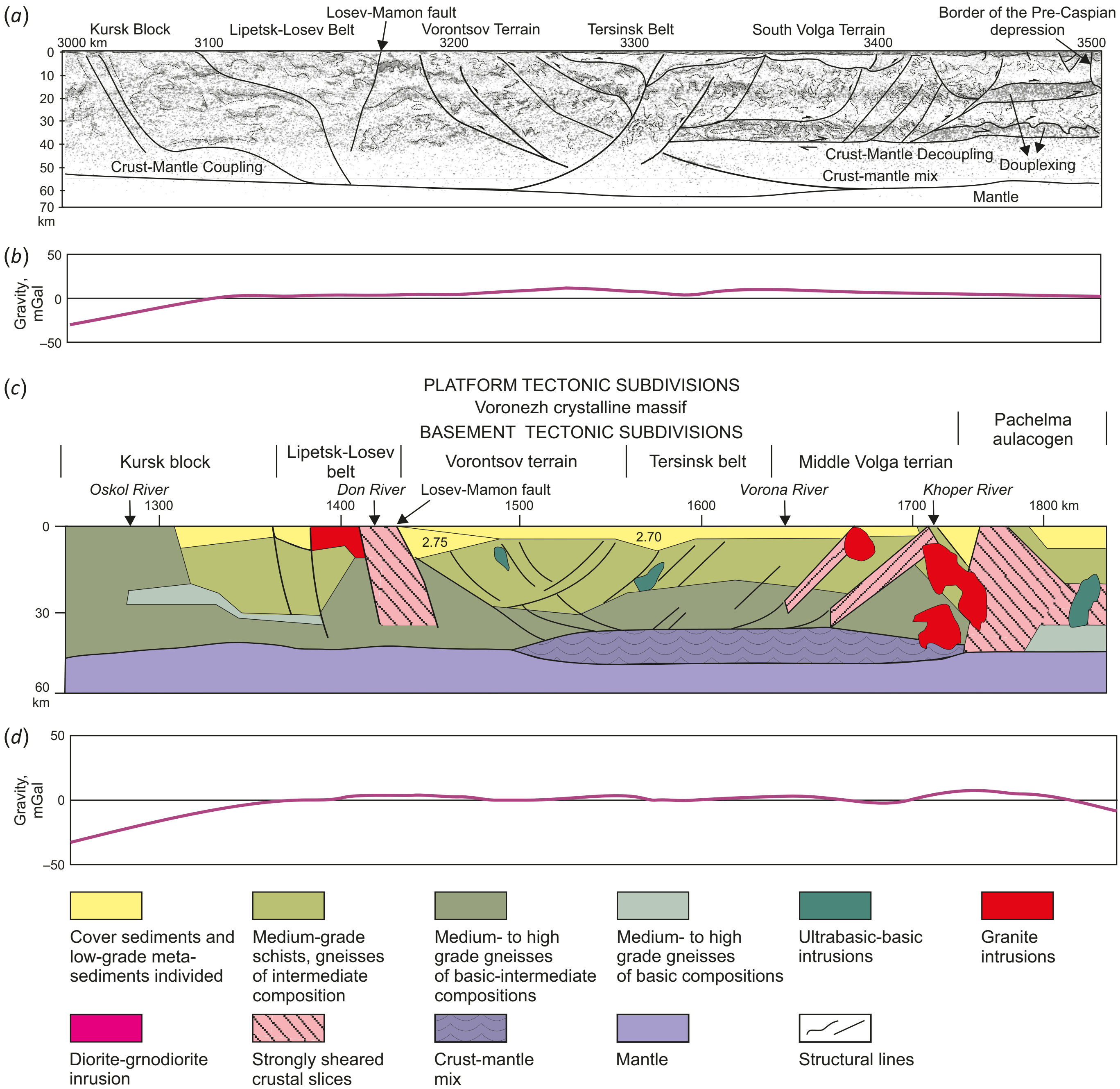


**Fig. 6.** Crustal architecture of the Volga-Don orogen along the EU-1 DRS transect (*a*–*b*) and along the Granite DDS transect (*c*–*d*). (*a*) – seismic section interpreted based on reflection patterns and subsurface geology; (*b*) – gravity profile along the transect based on the gravity map [Mints et al., 2010, 2015]; (*c*) – interpretation of seismic section along the Granite DDS transect (slightly modified from [Sokolov, 2002]); (*d*) – gravity profile along the transect.

**Рис. 6.** Глубинная структура Волго-Донского орогена вдоль профиля 1-ЕВ (*a*–*b*) и вдоль профиля «Гранит» (*c*–*d*). (*a*) – интерпретация сейсмического разреза, базирующаяся на анализе распределения отражающих площадок и геологических данных строения фундамента; (*b*) – гравитационный профиль вдоль трансекта, построенный на основе карты поля силы тяжести [Mints et al., 2010, 2015]; (*c*) – интерпретация сейсмического разреза по профилю «Гранит» (по [Sokolov, 2002] с небольшими изменениями); (*d*) – гравитационный профиль вдоль трансекта.

junction zone. Its deep refractory crustal structure exhibits a roughly divergent structure which descends beneath from the Tersinsk hinterland in which there occurs a mantle-crust mixture lens a few kilometers thick with compressional velocities of 7.5 to 8.0 km/sec (Fig. 6, c). It is vital to note that along the "Granite" DDS profile beneath the Ural Collisional orogen there is yet another, more distinct pattern of emergence of such a kind of mantle-crust mix domain.

To reveal the north-eastern offset of the Volgo-Don orogen, we have used an extension of the "Granite" DDS profile from 1800 to 2500 km points up to the border of the Volgo-Uralian crustal segment (Fig. 7, a). A seismic image obtained previously from the TATSEIS DRS transect [Trofimov,

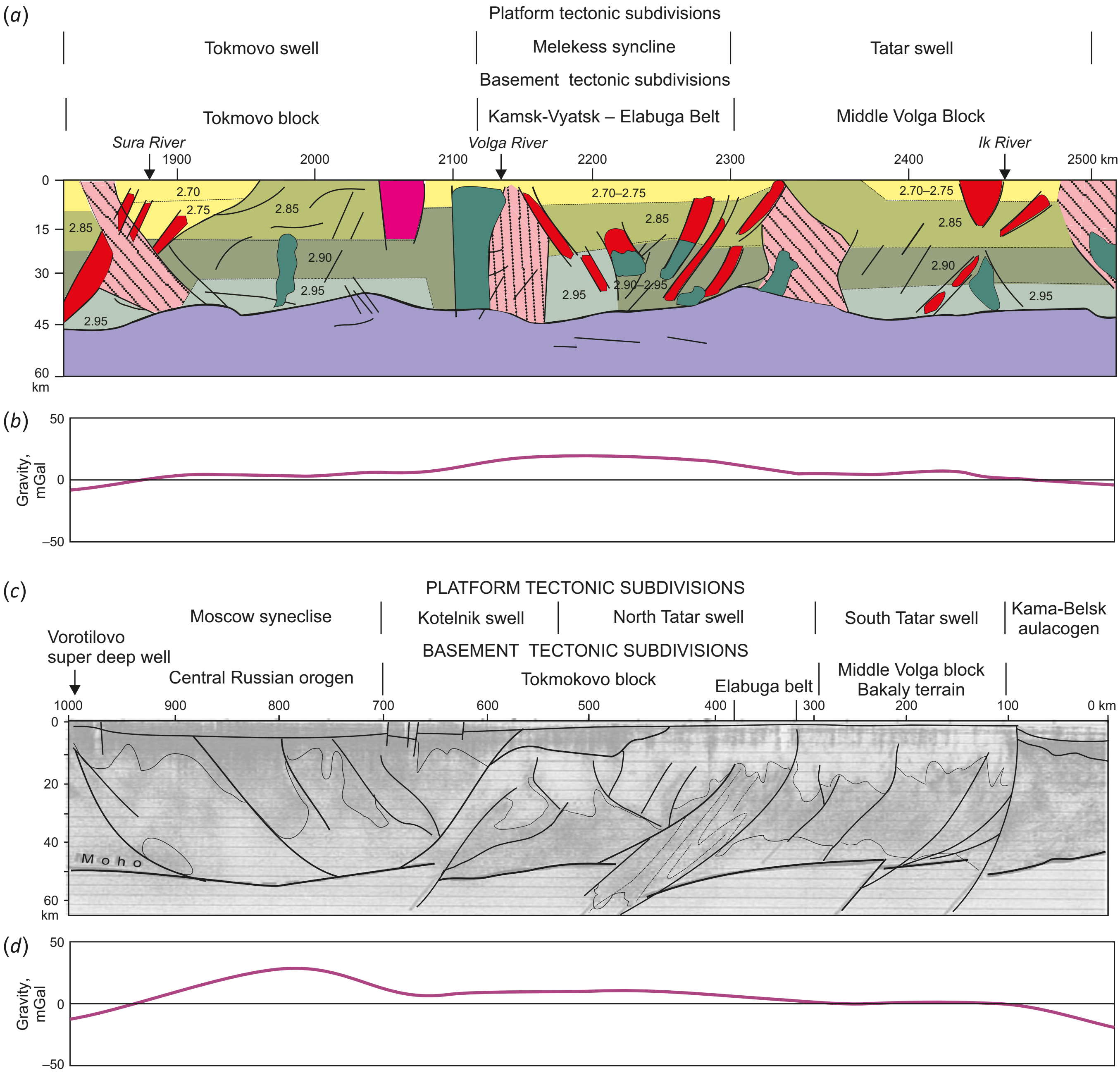


**Fig. 7**. Crustal architecture along the northeastern margin of the Volga-Don orogen along the Granite DDS transect (*a*–*b*) and along the TATSEIS DRS transect (*c*–*d*) (see Fig. 2).
(*a*) – interpretation of seismic section; (*b*) – gravity profile along the transect; (*c*) – interpretation of seismic section, modified from [Trofimov, 2006]; (*d*) – gravity profile along the transect. The legend is the same as in Fig. 6.

**Рис. 7.** Глубинная структура коры вдоль северо-восточной окраины Волго-Донского орогена по профилю «Гранит» (*a*–*b*) и по профилю ТАТСЕЙС (*c*–*d*) (см. рис. 2).
(*a*) – интерпретация сейсмического разреза; (*b*) – гравитационный профиль вдоль трансекта; (*c*) – интерпретация сейсмического разреза с некоторыми изменениями [Trofimov, 2006]; (*d*) – гравитационный профиль вдоль трансекта. Условные обозначения см. на рис. 6.

2006] has been added for comparison between different types of crustal growing to suit changing lithospheric properties (Fig. 7, c).

Along the "Granite" DDS profile, the east border of the Volgo-Don orogen would be expected nearby the 1950 km point where the Paleoproterozoic metasediments overthrust the northern edge of the Archean Volgo-Uralian core towards the east. This area is referred to in literature as the central part of a large, about 600 km wide Tokmovo ovoid-type block which is presumably Archean in age [Mints et al., 2015; Bogdanova et al., 2016]. Little is known about lithologies that make up this block. Scarce wells yielded two-pyroxene gneisses or charnockites and enderbites. The eastern border of the Tokmovo block nearby the 2100 km point is distinct because of a pronounced rheological difference between the Archean and Proterozoic lithospheres (see Fig. 2; Fig. 7, a). Indeed, the deep crustal structure from 2100 to about 2300 km point shows a large amount of mafic intrusions, as indicated also by a clear positive gravity anomaly (Fig 7, b). Data from seismic survey and thermal modeling agree well with each other implying that a previously unknown N-S striking fold-and-trust belt of about 200–250 km wide can be discerned between the Archean and Proterozoic lithospheric parts of the Archean Volgo-Uralian core (see Figs 1, 2). This belt might conceivably represent the southern extension of the Kama-Vyatka or Elabuga belts (cf. [Bogdanova et al., 2016]). It is particularly noteworthy that the borders of this belt closely matched those of the overlying Melekess deep syncline filled with platform sediments. The eastward part of the "Granite" DDS profile crosses the east-northernmost edge of the Archean Volgo-Uralian core manifested as the Tatar swell in terms of platform tectonics. It is significant that the base of the Volgo-Uralian crust displays detectable topography undulations with amplitude of up 10 km as against the almost flat seismic Moho typical for the base of the crust of the Volgo-Don orogen.

The 1000 km TATSEIS profile crosses the central part of the Volgo-Uralian crustal segment from the southeast to northwest where the lithosphere reveals mostly the Paleoproterozoic properties (see Fig. 2). The geological and radiological data on the Archean crust, derived solely from the Bakaly terrain in the range of about 100 to 300 km points of the TATSEIS profile, were reliable [Bogdanova et al., 2010; Bibikova et al., 2015]. Contrary to the Archean Sarmatian core, a seismic image of the Archean Middle Volga-Uralian core exhibits the other type of deep crustal structure (Fig. 7, c). It involves roughly the low reflective upper crust and high to middle reflective middle and lower crustal levels. They show clear topography undulations with amplitude of up 10 km in line with the undulation of the Moho surface. The profile line between about 300 and 400 km passes seemingly through the Elabuga belt where the high reflective lower crustal slice penetrates into the mantle for a distance of up to 10 km. Furthermore, Moho depths in the southeast are about 5 km deeper than in the northwest. It is conceivable that the Elabuga belt involves a subduction zone which dips to the southwest. However a crustal underthrusting mechanism that could have operated during crustal shortening should not be ruled out. The following line of the profile from about 400 to 600 km displays a seismic image of the three-layer deep crustal sandwich-type structure where the mid-crustal level of low reflectivity occurs between the upper- and lower crustal levels of high reflectivity. Note also a vergence-sense change that takes place within the crust of the Tokmovo block. The profile line between about 600 and 1000 km points crosses a part of the Central Russian orogen that is beyond the scope of this paper.

### 3.3. Tectonic implication

The renewed understanding of the seismic images presented in this study provides a compelling evidence that the junction zone between the Archean Sarmatian and Volgo-Uralian lithospheric plates was derived from their orogenic convergence rather than a plume-related stretching of a single Archean continental massif (cf. [Mints et al., 2015]). Furthermore, coupled with available data on radiologic ages and results from crystalline basement drilling, the seismic data make possible to generate a new tectonic model to explain the Volgo-Don orogen as an exemplary case of the Paleoproterozoic collisional orogens (Fig. 8).

As evident from Fig. 3, the beginning of the plate convergence is manifested in an emergence of the Losev bimodal volcanism at ca 2.1 Ga on the eastern outskirts of the Sarmatian crystal segment. As a marginal mountain range was increased, the fore-wedge filled with the Vorontsov flysch-type sediments was propagated eastwards, thus forming the Paleoproterozoic vast fore-arc basin which was then deformed during a continued west-verging contraction. Due to a flat geometry of subduction typical of the Andean active margins, a compressive push at the back of the wedge gave birth to a large lithospheric reworking of the Kursk-Azov block which involved stretching with subsequent shearing and folding of both the Archean basement and early Paleoproterozoic BIF-bearing shallow marine sequences. At that time this back-arc continental terrain was related to the pro-foreland of the Eastern Sarmatian Orogen.

At ca 2.05 Ga, a drastic change in tectonic development of the Eastern Sarmatian orogen occurred through the surfacing of the intra-oceanic Tersinsk volcanic arc which served as a key source of deposition of the South Volga turbiditic pelites and carbonaceous greywackes. The initiation of the Tersinsk volcanic arc can be explained by two alternative factors. First, this arc originated on its own on the Volgo-Uralian pro-plate. Second, its origin was concerned with the eastward retreating slab followed by slab break-up during the collision between the Sarmatian and Volgo-Uralian lithospheric plates, after which the entire oceanic lithosphere had been consumed. Since then, the tectonic patterns of the Eastern Sarmatian orogen became the retro-side whereas on the east, pro-tectonic side, there began the development of the hinterland (Tersinsk belt), the pro-wedge (South Volga province), and pro-foreland basin on the Archean Middle Volga block. Note that we use the terms

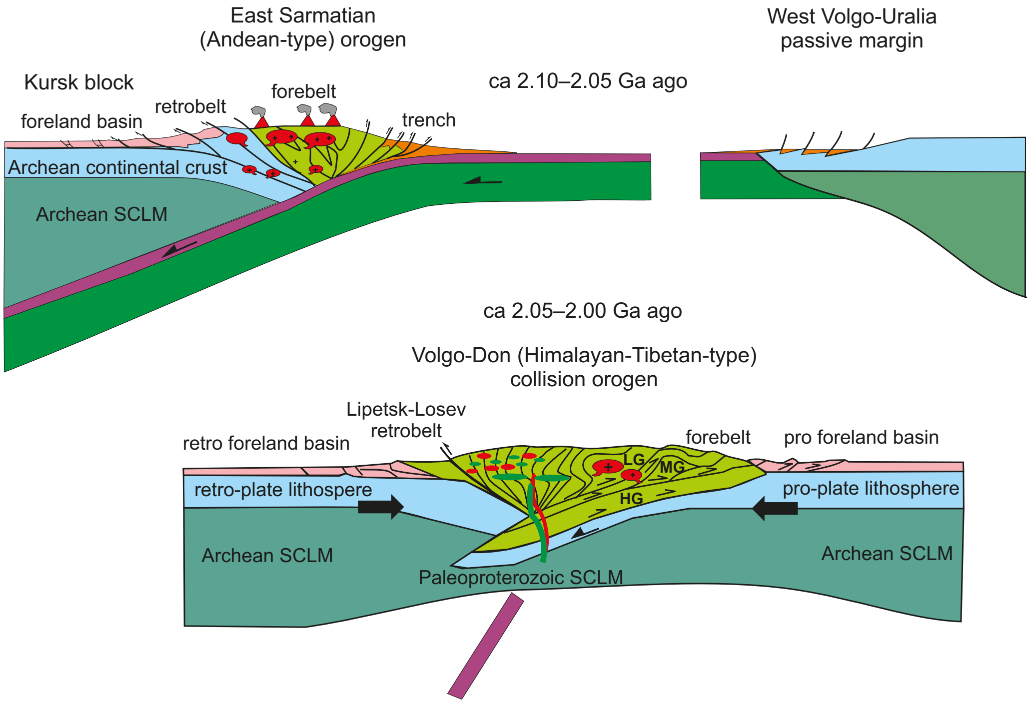


**Fig. 8.** Schematic illustrations of the geodynamic evolution of the Volga-Don orogen. SCLM – subcontinental lithosphere mantle, HG, MG, and LG – high-grade, mid-grade, and low-grade metamorphism, respectively.

**Рис. 8.** Схематическая модель геодинамической эволюции Волго-Донского орогена. SCLM – субконтинентальная литосферная мантия. HG, MG, LG – метаморфизм высокой, средней и низкой ступени, соответственно.

"pro-side" and "retro-side" after [Willett et al., 1993; Willett, Beaumont, 1994; Jamieson, Beaumont, 2013].

The orogenic development led to the formation of the divergent Paleoproterozoic Volgo-Don orogen which is akin both in structure and in concept to the modern Himalayan-Tibetan mountain range. Fig. 6 shows a distinct difference in crustal structures between the western and eastern parts of the Volgo-Don orogen, highlighting the substantial disparity in rheological attributes of the orogenic lithosphere formed therein. The Eastern Sarmatian orogenic crust of the Volgo-Don orogen including the Vorontsov province does not show clear evidence of a décollement at the Moho depth whereas the orogenic crust on the easternmost pro-plate flank of the orogen displays a laminated near-horizontal high reflectivity band in the lower crust which is largely interpreted as a result of horizontal shearing from the viewpoint of current tectonics, i.e. progressive decoupling of crustal material from the lithospheric mantle. In such an event, lower crustal flow about 10 km thick have to occur in the region about 100 km long while the topographic relaxation in crustal thickness occurs quickly and the large lateral variations in crustal thickness cause fluid to develop a steep faulting front [McKenzie et al., 2000]. In addition, the temperatures required for flow in the lower crust may be as high as 800–850 °C whereas such a kind of reflectors in the middle crust could be frozen from when the mid-crustal depths were at a higher temperature [Hyndman, 2017].

The surprising thing is that these constraints deduced from both analytical and geophysical studies on the modern high elevated mountain ranges of the North America Cordillera bordering the cratonic lithosphere on the east may be well used to explain the crustal architecture of the eastern part of the Paleoproterozoic Volgo-Don orogen. It cannot to be too highly stressed that in terms of geodynamics the South Volga province acted as a back-arc basin occurring between the Tersinsk arc and the Middle Volgo-Uralian protocratonic block, just as was the case with the recent history of the North America Cordillera [Hyndman, Currie, 2011; Hyndman, 2017]. The model of back-arc mobile belt cited above implies the following: (i) the orogenic heat should have originated from a preexisting hot back-arc rather than the orogenic process itself, (ii) a through-going basal detachment in the lower crust should have separated the entire crustal section from the underlying lithosphere leading to thrusting of juvenile crustal material over the stable craton, and (iii) the estimated duration of high temperature regime in former back-arcs is a few tens of millions of years after subduction termination when during subsequent crustal shortening there occurred a high-grade regional metamorphism, ductile crustal deformation, and orogenic plutonism.

A further indication of difference in rheological attributes of the recent orogenic lithosphere has been highlighted by [Ryan, Dewey, 2019] who have proposed to recognize two main modes of the arc-continent collision at early stages of the continent-continent collision. The first is defined as bulldozing when much of the shortening occurs due to inversion of a continental arc margin. The second involves the collision between a hot buoyant arc/forearc and an old hyperextended sediment-covered margin, with the former ridden over the latter. In the case of the Volgo-Don orogen, the role of the obducting "hot iron" should have been performed by hot granulitic lower crustal rocks instead of ophiolitic nappes. Depending upon different modes of the collision events, the foreland flanks of the orogen display a clear dissimilarity. As opposed to the bulldozed Archean crust of the Kursk block, the Middle Volgo-Uralian block was largely subjected to overthrusting by high-grade metasedimentary rocks. It may be safely suggested that the orogenic loading on the Archean lower density lithosphere of the Middle Volgo-Uralian block should have resulted in a domal uprising of the Archean crust, thus forming foredeep, forebulge, and backbulge zones in the pro-foreland just as it happens in modern collision orogens [DeCelles, Giles, 1996]. Indeed, as evidenced from the "Granite" DDS profile (see Fig. 7, a), the foredeep zone, about 20 km deep and about 100 km long, borders the Tokmovo forebulge (previously termed the Tokmovo oval/ovoid) on the east. The "ovoid" hypothesis, however, met uncertainties in regard to the ages of formation and origin of these geophysical structures [Bogdanova et al., 2016]. This setback is attributable to the evolution of the foreland lithosphere in the context of substantial horizontal shortening of orogenic lithosphere in recent collision systems, not to the evolution of a specific mechanism which operated in the Precambrian times. Yet another evidence of the orogenic impact on the Volgo-Uralian pro-foreland is expressed by the flexural wave through its lithosphere, testifying that the foreland lithosphere migrated towards the Tersinsk belt (see Fig. 7, a). An important point is that the Moho topography is smooth throughout the proper Volgo-Don orogen whereas the flexural lithosphere of the Volgo-Uralian crustal segment reveals a short wavelength of about 150 km, suggesting that the pro-foreland of the orogen could be substantially compressed after the onset of continent-continent collision. This development led to a strong structural and metamorphic reworking of the Archean Volgo-Uralian core at ca 1.95 Ga ago under conditions of T=700–750 °C and P=0.4–0.5 GPa [Bogdanova et al., 2021]. Taking into account that the onset of collision roughly agrees with an age of 2.02 Ga from the S-type granite in the vicinity of the Tersinsk belt [Bibikova et al., 2009], an age of 1.95 Ga for the rocks of the pro-foreland area, where high-grade metamorphism predominates without exposed plutons, might favor a model of progressive build-up of radiogenic heat in the middle and lower crust after a shortening event [England, Thomson, 1984].

Notice also that throughout the Paleoproterozoic Volgo-Don orogen, significant positive gravity anomalies are lacking thus attesting to the absence of a large mafic constituent that is characteristic of a hypothetical plume-related orogeny (see Fig. 6, b, c). On the contrary, the architecture of this orogen may be well depicted in terms of large modern collisional orogens implying that the modern-type plate tectonics operated since the Paleoproterozoic, at least when large Archean proto-cratonic continental masses became vital actors in the history of evolving continents.

## 4. DISCUSSION

One of main problems of orogenesis consists in whether there were any secular changes in tectonic processes in the Earth's history (e.g. [Stern, 2005; Brown, 2009]). In most of the numerical simulation models of Precambrian orogenic tectonism it is postulated that the secular changes have a direct relationship to the initially elevated heat flux from the mantle and higher radiogenic heat production as compared with modern orogens, which should have led to a common occurrence of high-grade lithologies throughout the ancestral orogenic settings. This in turn necessitates virtual scenarios that involve very hot and weak asthenosphere which lacks a stiff mantle (e.g. [Gerya, 2014; Sizova et al., 2014; Perchuk et al., 2018]). In other words, it is assumed that at that time the mantle was depleted as much as the modern mantle and, therefore, there was no relevant secular change in the lithospheric mantle (cf. [Griffin et al., 2009]).

A comparative study of natural orogens throughout the Earth's history from the ca 1.90–1.80 Ga Paleoproterozoic Trans-Hudson orogen up to the modern Himalayan-Tibetan collisional orogen shows no significant differences for most of their characteristics involving time spans of collision events, recorded durations of metamorphism, and structural geometries following continent-continent collision [Weller et al., 2021]. The only difference between them is that the present-day Himalayan-Tibetan orogen consists mostly of sedimentary rocks, with a narrow sliver of high-grade metamorphic rocks and associated partial melts expressed as the high-level leucogranite intrusions while eroded orogens are sediment-poor.

This gap is well overcome by the data presented herein. Indeed, the Volgo-Don Collisional orogen is mostly filled with fertile low- to medium grade metasediments as is typical for modern large orogens, and, therefore, may be regarded as a slightly eroded, well preserved Paleoproterozoic collisional orogen. The conservation of this orogen was apparently related to its armoring from below by a refractory, stiff lithospheric mantle and from both sides by the Archean core blocks coupled with strongly depleted mantle keels. A recent thermo-mechanical modeling of development of collisional orogens has shown, inter alia, that orogenic growth involving the lithospheric depletion is largely independent on the negative or positive buoyancy of a subducting lower lithosphere, and solely a function of internal crustal loading and, therefore, large divergent orogen emerge within a time span of 70 Ma [Wolf et al., 2021]. In that event the inflow of depleted lithospheric material is balanced by a small distributed outflow in a sub-lithospheric mantle.

Besides the clear structural and lithological similarity between the present-day Himalayan-Tibetan orogen and the Paleoproterozoic Volgo-Don orogen (Fig. 9), it is amazing that both orogens reveal also close time spans of the main tectonic phases occurred during their building. Prior to the terminal collision with India in the Eocene, the Tibetan part (Lhasaplano) of the Himalayan orogen developed as an Andean-type orogen since at least the Late-Cretaceous time, i.e. during ca 50 Ma [Kapp, DeCelles, 2019; Kapp et al., 2003]. The same time span can be recognized from the Andean-type East Sarmatian orogen (see Fig. 3). The onset of collision between the Indian and Asian plates is estimated to have occurred 50±10 Ma ago although the underthrusting of the Indian plate is still in progress with cumulative evidence of metamorphism from ca 50 to <10 Ma at different locations along the belt [Weller et al., 2021]. We cannot constraint with confidence an age of termination of the Sarmatian-Volgo-Uralian collision but it should be emphasized that the youngest metamorphic zircons from metasedimentary rocks of the Volgo-Uralia yielded

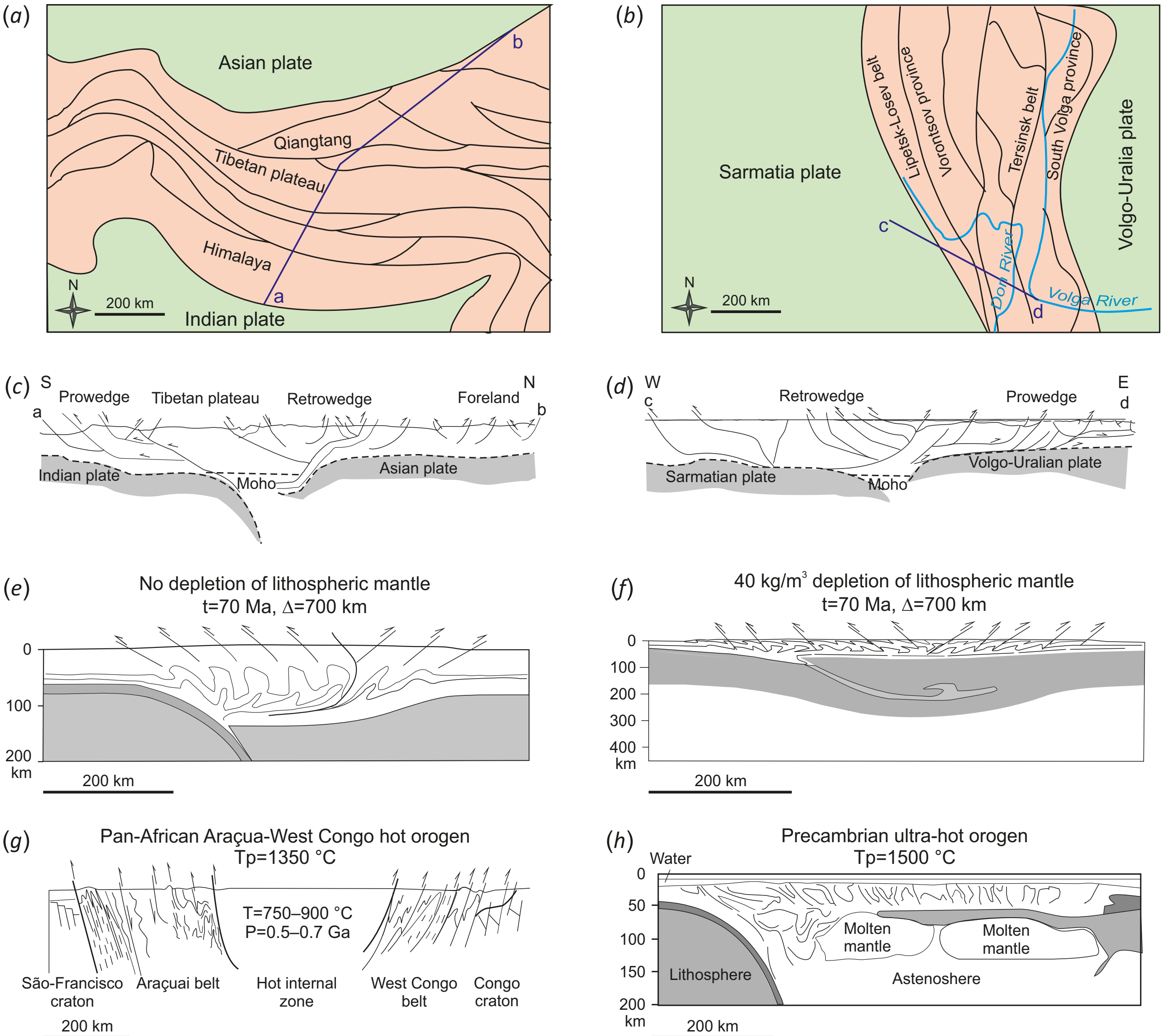


**Fig. 9.** Natural vs. model orogens.
(*a*–*b*) – size and structural comparison between the Himalayan-Tibetan and Volga-Don orogens. Both orogens have the same width. The Quingtang terrane is a retro-forebelt. (*c*–*d*) – schematic geologic cross-sections across the Himalayan-Tibetan orogen [Yin, Harrison, 2000] and the Volga-Don orogen (this study). Locations of the cross-sections are shown on the upper panel. (*e*–*f*) – schematic structural model of continent-continent collision for undepleted (*e*) and depleted mantle conditions (*f*) [Wolf et al., 2021]. (*g*–*h*) – a comparison between structural styles of ultra-hot orogens based on the geological observations (*g*) [Fossen et al., 2017] and numerical modeling (*h*) [Perchuk et al., 2018]. See text for explanation.

**Рис. 9.** Сравнительная характеристика природных оргенов и орогенов по результатам численного моделирования.
(*a*–*b*) – сравнение размеров Гималайско-Тибетского и Волго-Донского орогенов. Ширина и характер структуры обнаруживают заметное сходство. Террейн Цинъян представляет собой фрагмент ретрофорланда. (*c*–*d*) – схематические геологические разрезы Гималайско-Тибетского [Yin, Harrison, 2000] и Волго-Донского орогенов (настоящая работа). На верхней панели показаны локации разрезов. (*e*–*f*) – результаты моделирования коллизии континент – континент для условий недеплетированной (современной) мантии (*e*) и деплетированной мантии (*f*) [Wolf et al., 2021]. (*g*–*h*) – сравнение структурных стилей ультрагорячих орогенов по данным геологических исследований (*g*) [Fossen et al., 2017] и компьютерного моделирования (*h*) [Perchuk et al., 2018]. Объяснения см. в тексте.

ages of ca 2.0 Ga and ca 1.95 Ga [Fedotova et al., 2019]. By analogy with the Himalayan-Tibetan orogen, the timing of post-collision convergence for the Volgo-Don orogen could be considered as roughly the same and shortening of the orogen could have accommodated a few thousands kilometers of plate convergence.

Most of the numerical geodynamical models for the Precambrian orogenesis postulate a decisive role of high ambient upper mantle temperature expressed as potential mantle temperature (*Tp*) during the Archean and Paleoproteroic eons to produce hot/ultra-hot orogens at that times. Such an approach met with inconsistencies between model and natural orogens. The simplest example is the comparison to the large Ediacaran Araçuaí-West Congo ultra-hot orogen [Fossen et a., 2017], which formed when *Tp* of ambient mantle was roughly equal to the modern one as being taken 1350 °C [Herzberg et al., 2010], with a model ultra-hot orogen developed under *Tp* which increased up to 150 K above the present-day value (Fig. 9, g, h). This implies that the secular change in ambient upper mantle temperature should not be considered as a straightforward ground to thermo-mechanical understating of the orogenic crustal evolution without taking into account the secular change in physic-chemical properties of the lithospheric mantle. As can be seen from Fig. 9, f, h, the thick, stiff refractory lithosphere mantle that evolved from partial melting of the Paleoproterozoic ambient mantle at *Tp* ~1500 °C during the formation of contemporaneous thick oceanic crust should have protected the orogenic crust against high-temperature inflow from the asthenospheric mantle. Thus, in the Earth's history the secular change in the ambient mantle temperature was apparently balanced by change in both thickness and composition of orogenic lithosphere mantle instead of drastic change in the mode of development of orogenic crust.

## 5. CONCLUSION

(1) The global thermal model TC1 of the lithosphere [Artemieva, 2006] is fruitful and helps to place constraints on mechanisms of the Precambrian crustal growth not only for the EEC but also for other cratons.

(2) The frame of the Volgo-Uralian lithospheric mantle behaviors testifies that it consists of the Paleoproterozoic crust rather than of the Archean. Within the Volgo-Uralian crustal segment there is the only firmly recognized Archean crustal block, namely, Middle Volga-Uralian block, whereas the northern part of this segment shows no evidence for any presence of the Archean lithosphere. In the northern part of Volgo-Uralia, the Paleoproterozoic lithosphere is predominant.

(3) The Volgo-Don orogen intervenes between the two Archean protocratons, both of which are coupled with the thick mantle keels, and thus provides the principal interest to understanding a mechanism of the continent-continent collision in the Paleoproterozoic. We have shown that the crustal architecture of the Volgo-Don orogen may be well depicted in terms of development of the archetypal Himalayan-Tibet Collisional orogen involving their dimensions, time spans of collision events, recorded durations of metamorphism, and structural geometries following the continent-continent collision.

(4) The close similarity between these collisional orogens which differ radically in age suggests that the modern-type plate tectonics operated since the Paleoproterozoic, at least when the large Archean proto-cratonic continental masses became vital actors in the history of evolving continents.

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## 7. CONTRIBUTION OF THE AUTHORS

Both authors made an equivalent contribution to this article, read and approved the final manuscript.

## 8. DISCLOSURE

Both authors declare that they have no conflicts of interest relevant to this manuscript.

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