

Early evolution stages of the arctic margins (Neoproterozoic-Paleozoic) and plate reconstructions

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ABSTRACT

In this paper we offer paleoreconstructions for key structures of the Arctic based on the synthesis of geosstructural, geochronological and new paleomagnetic data bearing upon the Late Neoproterozoic and the Paleozoic histories of the Taimyr fold belt and Kara microcontinent. These tectonic features are part of a greater continental mass that we term “Arctida”, with an interesting history of breakup and reassembly that is constrained by our new data and synthesis. In the Central Taimyr accretionary belt fragments of an ancient island arc (960 Ma) have been discovered, and the paleomagnetic pole for the arc approximates the synchronous (950 Ma) pole for the Siberian paleocontinent. For the Kara microcontinent we demonstrate its evolution in the Early Paleozoic and its collision with Siberia in the Late Paleozoic. These data along with an extensive published paleomagnetic database for the cratons of Laurentia, Baltica, Siberia, and Gondwana are the basis for the presented paleotectonic reconstructions. The migrations of those Arctida tectonic blocks that lack paleomagnetic data are reconstructed based on geologic information.

INTRODUCTION

The current structure of the Arctic Ocean is determined by the position of the Amerasian (Canadian) and Eurasian basins, whose formation took place as a result of significant tectonic processes in the Late Mesozoic – Cenozoic. However it is impossible to understand relatively recent and modern tectonic displacements without analyzing previous tectonic events.

The discovery of Precambrian metamorphic complexes among the main structures of the

Arctic Region led to the suggestion that in the Late Precambrian a paleocontinent – termed “Arctida” – existed between Laurentia, Baltica and Siberia (Zonenshain, Natapov, 1987). In the classic presentation it is composed of several blocks of continental crust, whose relicts are now located in the Arctic (Fig. 1): the Kara block, the New Siberian block (the New Siberian Islands and the adjacent shelf), the North Alaska and Chukotka blocks, as well as small fragments of the Inuit Fold Belt in northern Greenland (Peary Land, the northern part of Ellesmere and Axel Heiberg islands) and the blocks of the underwater Lomonosov and Alpha-Mendeleev Ridges (Zonenshain, Natapov, 1987; Zonenshain et al., 1990). In the modern interpretation, aside from these fragments, Arctida also includes parts of Barentsia, which includes the structures of the Svalbard and the Timan-Pechora plates (Vernikovsky, 1996; Kuznetsov et al., 2007).

Late Precambrian and Paleozoic global tectonic history is defined by the breakup of Rodinia, the evolution of newly formed oceanic basins and the formation of Pangea as a result. Many paleotectonic schemes and reconstructions have been composed for the Late Precambrian – Paleozoic stages of the plates interactions (Scotese and McKerrow, 1990; Dalziel, 1991,1997; Hoffman, 1991; Powell et al., 1993; Condie and Rosen, 1994; Torsvik et al., 1996; Golonka, 2002; Golonka et al., 2003; Cocks and Torsvik, 2002; Lawver et al., 2002; Li et al., 2008; Pisarevsky et al., 2008, Metelkin et al., 2012). However, when dealing with the details of the evolution of separate lithosphere segments, including those of the Arctic Region, there are still many unsolved, debatable and ill-founded reconstructions. This is true mainly for the deciphering of the initial

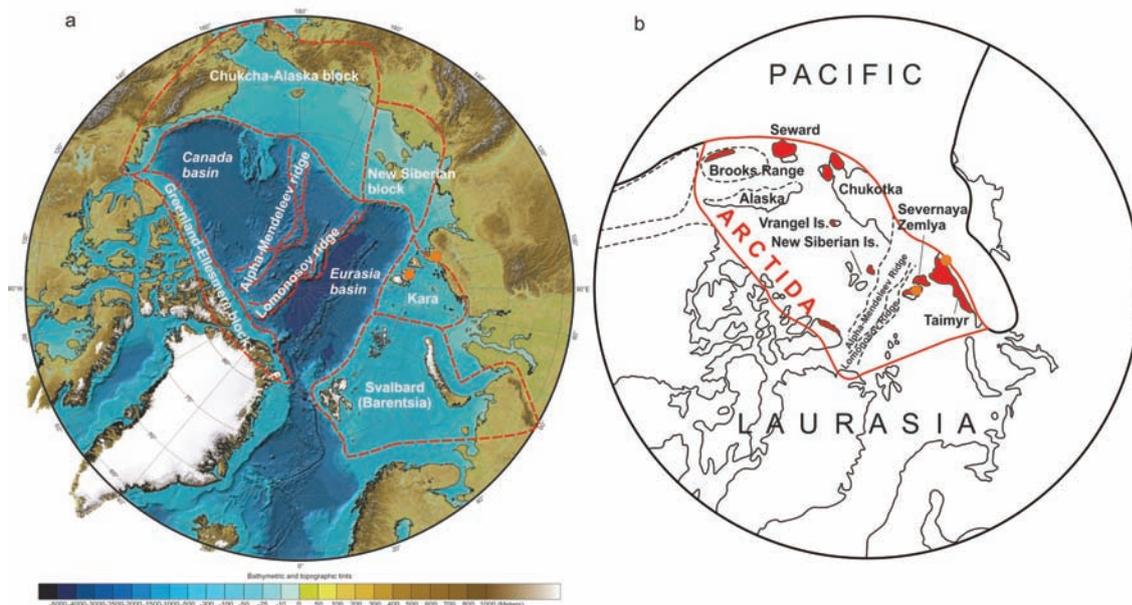


Fig. 1. (a) The main blocks, microcontinents, plates, and basins of the Arctic on the International Bathymetric Chart of the Arctic Ocean and (b) a reconstruction for the Early Jurassic, showing the Precambrian Arctic blocks (in red), amalgamated into the Arctida continent, which is attached to Laurasia (Zonenshain and Natapov, 1987; Zonenshain et al., 1990). The approximate location of the field study area within the Taimyr folded area is shown by orange dots.

structure of Arctida, the reasons and mechanisms of its breakup, the drift trajectories of the continental blocks that composed it. The very existence of oceanic basins that supposedly separated the paleocontinents is uncertain. All these are largely debatable topics, especially the early stages of the Arctic Region tectonic evolution – the Late Precambrian and the Early Paleozoic, which are the subject of this paper. In this study we have attempted to integrate the available geologic and geophysical material for the early evolution stages of the Arctic Ocean in the form of a series of paleotectonic reconstructions, as well as to create a new development model for the structures of Arctida.

The determination of the relative positions of the blocks composing Arctida could be done with paleomagnetic data. However, such data are very sparse for the Late Precambrian and the Paleozoic. For the entire Arctic Region the IAGA Global Paleomagnetic Database counts no more than 30 paleomagnetic determinations. Nearly all of the available data represent the Late Paleozoic and Early Mesozoic of the Barentsia and Greenland-Ellesmere regions. There are no data for the New Siberian Islands and the territories of Chukotka and Northern Alaska, which represent most of the classic

Arctida area. Reliable paleomagnetic determinations for the Neoproterozoic-Paleozoic time interval are available only for fragments of a 960 Ma island arc from Central Taimyr (Vernikovsky et al., 2011) and for which the paleomagnetic pole is comparable to the approximately synchronous pole of Siberia from (Pavlov et al., 2002). There are other reliable data for the Kara microcontinent: this includes three paleomagnetic poles for 500, 450 and 420 Ma (Metelkin et al., 2000; 2005). It is these data that are placed at the core of our paleotectonic reconstructions along with the extensive paleomagnetic database for the Laurentia, Baltica, Siberia and Gondwana cratons (Pechersky and Didenko, 1995; Torsvik et al., 1996; Smethurst et al., 1998; McElhinny and MacFadden, 2000; Wingate and Giddings, 2000; Pavlov et al., 2002; Torsvik and Van der Voo, 2002; Meert and Torsvik, 2003; Metelkin et al., 2007, 2012; Li et al., 2008). The paleogeographic position of the cratons is corrected (within confidence limits for paleopoles) in accordance with the general model and available global reconstructions, including structures of the Arctic sector (Scotese, 1997; Lawver et al., 2002, 2011; Golonka et al., 2003, 2006; Kurenkov et al., 2005; Cocks and Torsvik, 2002, 2007).

THE OLDEST ISLAND ARC COMPLEX OF CENTRAL TAIMYR

The Central-Taimyr accretionary belt is located between two large continental blocks – the Siberian craton on the south and the Kara microcontinent on the north (Fig. 2a). It is composed of paleo-island arc fragments, granite-metamorphic terranes, passive continental margin terranes of mainly carbonate composition and ophiolites, which were amalgamated and accreted to the Siberian craton in the Late Neoproterozoic and then unconformably overlain by a Vendian (Ediacaran) – Early Carboniferous cover (Uflyand et al., 1991; Vernikovskiy, 1996; Khain et al., 1997; Vernikovskiy and Vernikovskaya, 2001; Pease et al., 2001). In this model a significant role is played by ophiolites and island arcs, whose zircons U–Pb age has been established in the interval of 755–730 Ma from plagiogranites, gabbros and volcanogenic rocks (Vernikovskiy et al., 1994; 2004). However, no paleomagnetic data have been obtained for the 755–730 Ma rocks. Investigations carried out in the North-Eastern Taimyr in recent years allowed us to identify an older (960 Ma) paleo-island arc complex in the Central Taimyr accretionary belt and to establish its location at the time of formation by using paleomagnetic data.

The studied area of the Three Sisters Lake (Fig. 2b) is the junction zone of Zhdanov formation rocks (Zabiyaka et al., 1986) and mainly volcanogenic rocks previously included in the Borzov (Bezzubtsev et al., 1986) or Laptev (Zabiyaka et al., 1986) formations. Zhdanov formation rocks are mainly terrigenous (greenish-grey and grey sandstones, siltstones, black and dark-grey phyllites with separate layers of carbonate rocks, andesite-basalts, acid effusive rocks and their tuffs), metamorphosed in greenschist facies conditions. Borzov/Laptev formation rocks (metamorphosed basalts, andesites, dacites, plagioryhodacites) are host to plagiogranites and plagiogranite porphyry. Both formations are intruded by slightly metamorphosed gabbro-dolerite sills and dikes with thicknesses ranging from tens of centimeters to hundreds of meters which compose a wide dike belt with a total length of over 100 km. The Zhdanov and Borzov/Laptev formations are overlain by the coarse-grained terrigenous deposits of the Oktyabrsk formation. In the studied region the rocks that compose the island arc are tectonically

composited with sedimentary and volcanogenic-sedimentary deposits that we consider to have formed in an adjoining back-arc basin.

In the study area andesites, dacites, and plagioryhodacites are the dominant rock types in the paleo-island arc complex. These rocks range from dark-grey with a lilac hue to bottle-green, sometimes with 2–4 mm phenocrysts of plagioclase, quartz and less frequently of subordinate potassium feldspar. Andesites are distinguished by their fine-grained matrix textures and by the presence of suites of ore minerals. Hypabyssal rocks are represented by metamorphosed plagiogranite porphyry with medium-grained matrix texture and consisting mainly of quartz and plagioclase. Plagiogranites also contain hornblende and clinopyroxene. Secondary minerals are albite, chlorite, biotite, carbonate, and epidote. These rocks are often schistose and highly fractured and veined. Diabases and gabbrodolerites are fine- and medium-grained and intensely amphibolized.

The studied acid-intermediate volcanic and intrusive rocks are attributed to the tholeiitic magmatic series. They have weakly or moderately fractionated REE spectra ($(La/Yb)_N = 3.3–11.5$) with small negative Eu anomalies ($Eu/Eu^* = 0.7–0.9$), the total REE concentration is 330–781 ppm. On the spider diagrams the rocks are enriched in La, Ce and also Th and U and depleted in Sr, Ti, P, Ta, and Nb. For the island arc metabasites the total REE concentration varies in a wider interval from 233–375 to 1290 ppm. They can have small Eu anomalies ($Eu/Eu^* = 0.9–1.1$), whereas the $(La/Yb)_N$ ratio values vary widely from 1–3.5 to 36.6. The REE spectra are flat, typical of MORB and close to those of island arc basalts. The established Rare Earth and other trace elements distribution types for the entire complex are similar to those of the igneous rocks of Neoproterozoic island arc of other Taimyr regions (Vernikovskiy et al., 1994, 2004).

We performed U–Pb isotopic analysis utilizing a multicollector Finnigan MAT-261 mass spectrometer and the Sm, Nd, Rb, and Sr analysis – on a 7-collector Triton T1 mass spectrometer at the Institute of Precambrian Geology and Geochronology of the RAS, St. Petersburg (Russia).

The accessory zircons from a plagioryhodacite and a plagiogranite are semitransparent and

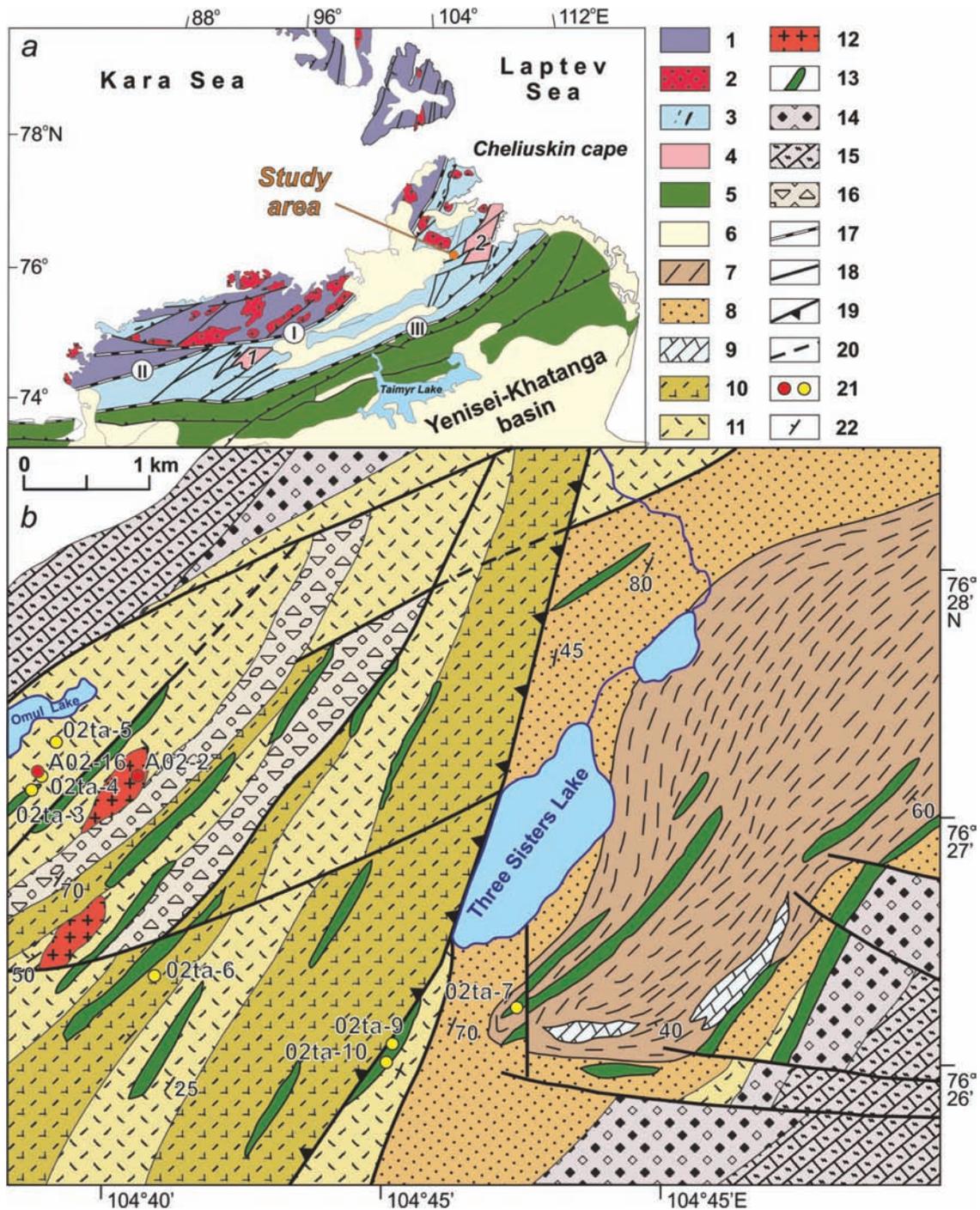


Fig. 2. (a) The Three Sisters Lake study area location on the tectonic scheme of the Taimyr folded area; and (b) a geologic map of the Three Sisters Lake study area composed using the data of Bezzubtsev et al., (1986); Zabiya et al., (1986); and Vernikovskiy (1996). 1–6 – Tectonic elements and geodynamic complexes on the tectonic scheme shown on the regional map (a): 1 – Kara microcontinent (NP–PZ); 2 – collisional granitoids (300–264 Ma, after Vernikovskiy et al., (1995; 1998)); 3 – Central Taimyr accretionary belt (NP) including 4 – Mamont-Shrenk (1) and Faddey (2) cratonic terranes; 5 – South Taimyr folded belt (PZ–MZ); 6 – overlapping sedimentary complex. 7–16 – Neoproterozoic rocks shown on the geological map (b): 7–9 – Zhdanov formation including: 7 – black phyllites and siltstones; 8 – sandstones and siltstones with subordinate interbeds of phyllites; 9 – lenses of limestones and dolomites; 10–11 – Borzov/Laptev formation including: 10 – andesites, dacites, subordinate basalts and andesite-basalts; 11 – plagioryhodacites; 12 – intrusions of plagiogranites; 13 – gabbro-dolerite sill; 14–16 – overlapping strata of Oktyabrsk formation including: 14 – quartz and polymict conglomerates; 15 – oligomict and quartz sandstones and gritstones; 16 – polymict and quartz conglomerates, breccias. 17–21 – faults and other symbols shown in both maps (a) and (b): 17 – sutures: I – Main Taimyr, II – Diabasovy, III – Pyasina-Faddey; 18 – normal faults, reverse faults, strike slip faults, 19 – thrusts; 20 – inferred faults; 21 – sampling sites for geochronological (red) and paleomagnetic (yellow) investigations; 22 – strata bedding.

transparent subidiomorphic pink crystals of prismatic and short-prismatic shape. The morphological particularities of the zircon grains indicate their magmatic origin. The isotopic composition points of the studied zircons from a plagioryhodacite (sample A02-16) are approximated by a regression line, where the upper intersection with the concordia corresponds to the age 966 ± 5 Ma and the lower intersection corresponds to 279 ± 30 Ma, with $MSWD = 0.84$ (Vernikovskiy et al., 2011). At the same time the isotopic composition points for the zircon residue after acid treatment with longer exposition is located on the concordia, and its concordia age is 961 ± 3 Ma ($MSWD = 0.72$, probability = 0.4) and can be accepted as the most precise crystallization time estimate for the studied zircons.

The isotopic composition points for 20 untreated zircon grains from a plagiogranite (sample A02-2) and for two residues after acid treatment form a discordia whose upper intersection with the concordia corresponds to the age of 989 ± 41 Ma, and the lower one – 508 ± 410 Ma, $MSWD = 0.05$. The mean age value, calculated from the $^{207}\text{Pb}/^{206}\text{Pb}$ ratio of the three fractions of the studied zircon grains correspond to 969 ± 17 Ma and is close to the age value obtained from the upper intersection with the discordia. This age estimate may be used as the most precise one (Vernikovskiy et al., 2011).

Sm–Nd isotopic data for island arc acid intrusive and volcanic rocks of the Three Sisters Lake region yield a Mesoproterozoic model age: $T_{\text{Nd}}(\text{DM})$ varies from 1170 to 1219 Ma. These data as well as Rb–Sr isotopic investigations indicate a predominance of a mantle component in the magmatic sources of these rocks: $\epsilon\text{Nd}_{(967-961)} = 5.1-5.2$ and $(^{87}\text{Sr}/^{86}\text{Sr}_0) = 0.70258-0.70391$ (Vernikovskiy et al., 2011).

The paleomagnetic analysis was performed on the apparatus of the Paleomagnetic Center in the Laboratory of Geodynamics and Paleomagnetism of the IPGG SB RAS (Novosibirsk). The hardware system comprises new generation measurers including a 2G Enterprises Superconductive Magnetometer (USA) with built-in AF-demagnetizer and an HSM superconductive spinner-magnetometer (Germany), as well as the well-known JR-4 and JR-6 spin-magnetometers (Czech Republic) and other instruments, placed in a shielded room. The investigation includes a detailed stepwise

thermal demagnetization (*T*-demagnetization) and/or alternating field demagnetization (*AF*-demagnetization) of all studied samples until their complete demagnetization.

The experimental results were processed with specialized software products that use standard techniques for component analysis (Butler, 1992); and various modifications of the fold test (McFadden, 1990; Watson and Enkin, 1993, Enkin, 2003) and reversal test (McFadden and McElhinny, 1990) for dating the magnetization components. The sample collection includes volcanic as well as intrusive rocks of the paleo-island arc complex (Fig. 2b). One site (02ta-4) corresponds to an outcrop of plagioryhodacites (sample A02-16), which has been dated by U-Pb method. The studied rocks are characterized by relatively low values of natural remnant magnetization, NRM (tens of mA/m, thousands for one outcrop) and a high magnetic susceptibility - about 10^{-3} SI units. For the analysis of the NRM components *T*-, and *AF*-demagnetization were used. Typical orthogonal plots are given in Fig. 3. Most of the samples are characterized by two often unidirectional components – a titanomagnetite component with a blocking temperature T_B of about 400°C and a magnetite component with $T_B \sim 580^\circ\text{C}$. Distinctive particularities in the NRM vector behavior during the demagnetization of rocks from various outcrops are mainly due to the input of the titanomagnetite and magnetite components. In some samples the component of characteristic remnant magnetization (*ChRM*; shown as dashed lines in vertical plane projections (open circles), Fig. 3) is exactly registered in a high temperature interval $400-580^\circ\text{C}$, and the almost complete demagnetization of others is reached with the heating to 400°C or lower. In the last case (lower right in Fig. 3) the *AF*-demagnetization is more informative. The value of the median destructive field (MDF) is no more than 20–30 mT, and the complete demagnetization is reached by the impact of the alternating magnetic field no more than 100 mT. The established average *ChRM* directions are given in Table 1 (In situ and Tilt corrected). The primary nature of the *ChRM* can be substantiated by positive results of the reversals and fold tests. The upper five of the studied sample groups have a normal polarity, the mean direction in stratigraphic coordinates: $D = 319.2$, $I = 13.7$,

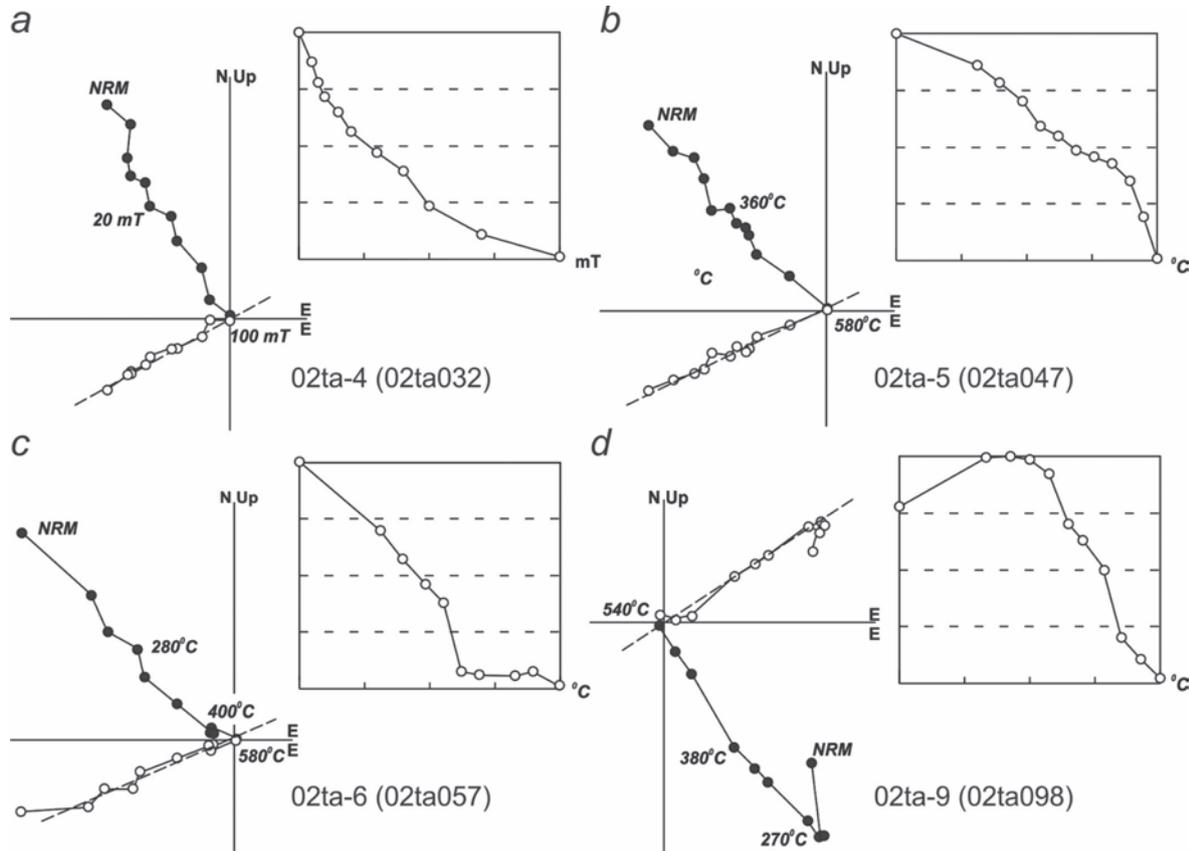


Fig. 3. Typical orthogonal plots in tilt-corrected coordinates and corresponding *NRM* vs. *T(AF)* graphs based on the results of *T*- and *AF*-demagnetization: (a) dacite from site 02ta-4 (sample number 02ta032); (b) rhyolite from site 02ta-5 (sample number 02ta047); (c) andesite from site 02ta-6 (sample number 02ta057); gabbro-dolerite from site 02ta-9 (sample number 02ta098). Solid dots represent projections of vector endpoint on the horizontal plane, and the open ones – on the vertical plane, the dashed line shows the stable *ChRM* component.

Table 1. Paleomagnetic directions and coordinates of virtual geomagnetic poles of the studied 960-Ma volcanogenic formation from the Three Sisters Lake region

Site numbers, rock type	<i>n/N</i>	In situ		Tilt corrected		<i>k</i>	α_{95}	VGPole			PL
		<i>D</i> (°)	<i>I</i> (°)	<i>D</i> (°)	<i>I</i> (°)			<i>PLat</i>	<i>PLong</i>	<i>dp/dm</i>	
02ta-3, gabbro-dolerite	10/11	70.0	86.1	323.9	16.3	374.1	2.5	19.1	322.7	1.3/2.6	8.3±1.9
02ta-4, dacite	8/10	152.5	88.0	319.5	16.9	118.9	5.1	18.8	327.3	2.7/5.3	8.6±3.8
02ta-5, rhyolite	10/10	327.5	89.4	320.1	14.4	75.4	5.6	17.6	326.5	2.9/5.7	7.3±4.1
02ta-6, andesite	7/10	209.4	80.5	310.7	18.1	92.7	6.3	17.9	336.5	3.4/6.5	9.3±4.7
02ta-7, gabbro-dolerite	9/12	300.4	-82.0	321.7	2.3	38.5	8.4	11.7	324.1	4.2/8.4	1.2±5.9
02ta-9, gabbro-dolerite	8/10	85.8	-69.1	138.9	-17.9	89.1	5.9	19.2	328.2	3.2/6.1	9.2±4.4
02ta-10, gabbro-dolerite	10/10	94.6	-74.0	144.8	-17.6	72.8	5.7	20.0	322.0	3.1/5.9	9.0±4.3
Mean		264.2	81.2	320.0	14.8	127.3	5.4	17.8	326.8	$A_{95}=4.0$	7.5±4.0

Note: *n/N* – ratio of the number of samples, used in the statistics, to the total number of studied samples; *D* – declination in degrees; *I* – inclination in degrees; *k* – precision parameter, α_{95} – 95% confidence limit, *VGPole* – the virtual geomagnetic pole coordinates (the inverted positions of the poles are given); *PLat* – latitude, *PLong* – longitude, *dp/dm* – semiaxes of the confidence circle of paleomagnetic pole; the mean pole is calculated as the average from the *VGPole* batch where A_{95} – 95% confidence limit; PL – paleolatitude for the reconstructed block in northern hemisphere.

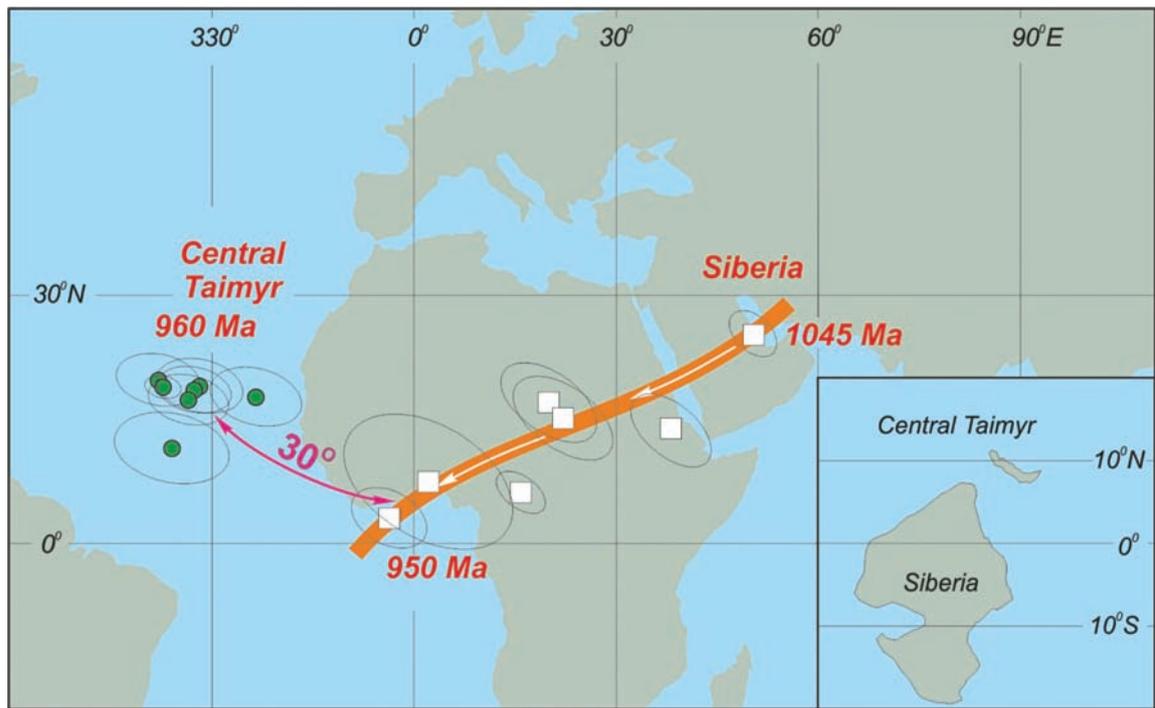


Fig. 4. The relative positions of the calculated Central Taimyr 960 Ma paleomagnetic poles (Table 1, circles) and APWP of Siberia for the time period 1,045–950 Ma by (Pavlov et al., 2002) and a paleogeographic reconstruction (inset) of the Central Taimyr margin and Siberia at 960 Ma.

$k = 100.8$, $\alpha_{95} = 7.7$. A reverse polarity has been established for two outcrops (sample numbers 02ta-9 and 02ta-10, table 1) of gabbro-dolerite sill, with the mean direction being: $D = 141.9$, $I = -17.8$, $k = 414.8$, $\alpha_{95} = 12.3$. The angle between the means of the normal and reverse polarity is $\gamma = 4.8^\circ$ with $\gamma_C = 9.4^\circ$ as the critical value. The precision parameter (k) is significantly higher in the stratigraphic coordinates (the ratio $k_s/k_g = 43.8$ is higher than the critical value – 4.16 for $n = 7$ at the 99% confidence level), the optimal concentration of magnetic directions (when k is maximum) is found at $109 \pm 4\%$ untilting. The correlation test (McFadden, 1990) is positive: the test parameter (distribution function) ξ_1 in stratigraphic coordinates – 3.166 exceeds the critical value at 95% confidence level – 3.086, at the same time in geographic (in situ) coordinates ξ_1 is 2.552, which is lower than the critical value. The main stage of deformations of the island arc complexes of Central Taimyr corresponds to the ~600 Ma boundary (Vernikovskiy, Vernikovskaya, 2001), therefore we can safely assume that the 960 Ma age of the established *ChRM* is pre-Ediacaran (630-542 Ma; Walker and Geissman, 2009). In all probability the

ChRM corresponds to the time of formation of these rocks at 960 Ma or in the Early Neoproterozoic.

The mean paleomagnetic pole for the Central Taimyr rocks (Table 1; PLat=17.8, Plong=326.8, A95=4.0) is close to synchronous poles for Uya series sedimentary rocks in the Uchuro-Maya region in the south-east of the Siberian craton, hosting basic intrusions (Pavlov et al., 2002). The age of those intrusions is substantiated by results of Sm–Nd, 942 ± 19 Ma (Pavlov et al., 2002) and U–Pb, 947 ± 7 , 1005 ± 4 Ma (Rainbird et al., 1998) dating. The angular divergence in the poles position, considering the confidence interval, is equal to $31.7^\circ \pm 4.3^\circ$ in paleolongitude and $-8.7^\circ \pm 3.7^\circ$ in paleolatitude (Fig. 4). Consequently, the island arc whose relicts are preserved in the modern structure of the Three Sisters Lake region was located in some distance away from the Taimyr margin of Siberia at the time of its formation. Judging by the difference in synchronous paleomagnetic latitudes, the Central Taimyr island arc could have been separated from the Siberian continent by a back-arc basin 550–1,380 km wide (Fig. 4). During the back-arc basin's closure the arc must have been rotated for $\sim 30^\circ$

clockwise. From these observations, we reach the following conclusions:

1. 960 m.y. ago the paleo-island arc of Central Taimyr was located in the subequatorial zone, near the northern margin of Siberia, and had a sublatitudinal strike. The sizes of the back-arc basin that existed between the arc and the continent at this time could reach 556–1,378 km incorporating the estimated error in paleomagnetic determinations.
2. The established age of the island arc in the Central Taimyr indicates that the transformation of the passive continental margin regime into an active one in the north of Siberia took place as early as the beginning of the Neoproterozoic (1 Ga). It is of fundamental significance for paleotectonic reconstructions of Siberia's position within the framework of Rodinia because it does not allow the joining of the Taimyr margin with the Canadian margin of Laurentia as it is assumed in alternative reconstructions (Dalziel, 1991; Hoffman, 1991).

3. The accretion of the island arc to the craton incorporated mutual rotation around a vertical axis, which implies the existence of a significant strike-slip component in the kinematics of the accretion process in the north of Siberia in the Late Precambrian.

THE KARA MICROCONTINENT DURING THE PALEOZOIC

The Kara microcontinent (or Kara plate) is one of the largest fragments of the ancient Arctida paleocontinent (Fig.1). Therefore the problems related to the reconstruction of its formation, kinematics and interactions with other continental blocks are very important for the understanding of the entire Arctic region. The Kara microcontinent's Precambrian basement is heterogeneous, which is consistent with refraction velocities ranging from 5.7 to 7.1 km/s (Bogolepov et al., 1991). According to individual seismic soundings, the crust thickness in the Kara microcontinent may exceed 40 km, with a 14–16-km-thick lower crust. The sedimentary

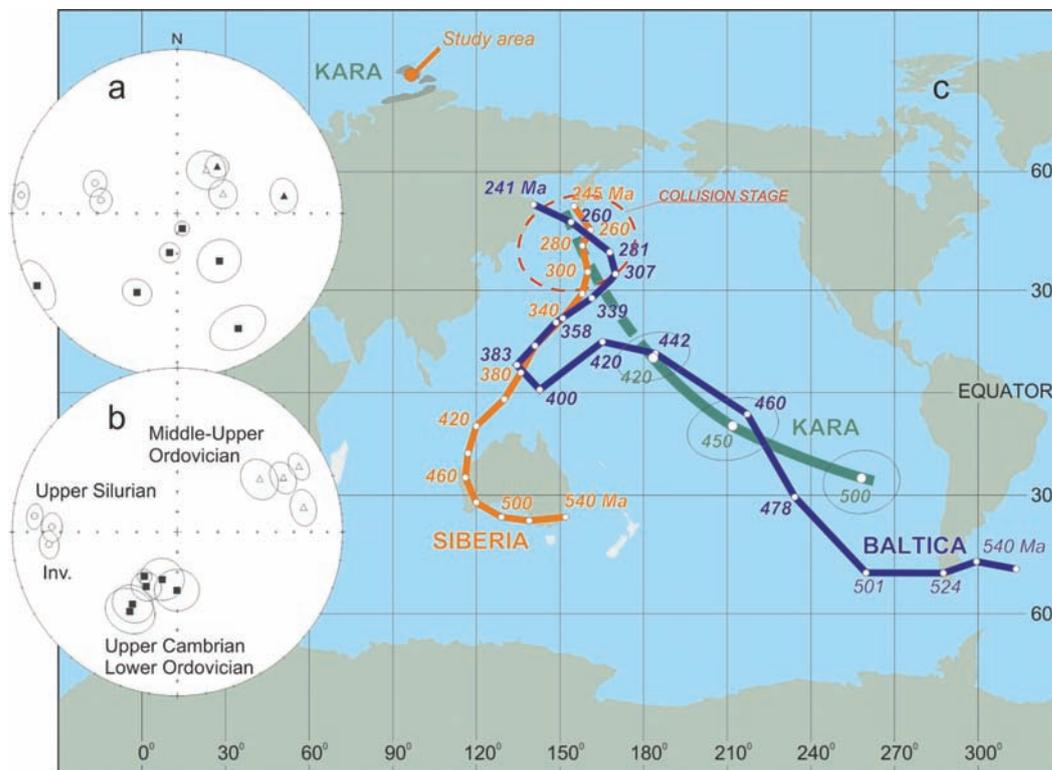


Fig. 5. (a) Paleomagnetic site mean directions in situ; and (b) tilt corrected; and (c) APWP for the Kara microcontinent and its comparison with APWP for Siberia by (Pechersky and Didenko, 1995) and APWP for Baltica by (Torsvik et al., 1996). Modified after (Metelkin et al., 2005).

section includes two units. The lower unit is up to 14 km thick and is apparently composed of Late Cambrian–Ordovician and Silurian to Permian carbonate, evaporate, and terrigenous deposits (Kaban'kov, Sobolevskaya., 1981; Kaban'kov et al., 1982). The upper unit, as thin as 2 km, consists of Triassic–Jurassic sequences.

Analysis of geostructural, paleomagnetic, geochronological and biostratigraphic data showed that the Kara microcontinent was tectonically isolated from neighboring continents in the Early Paleozoic (Fig.5) and collided with Siberia at ~300 Ma or in the Carboniferous (Vernikovskiy et al., 1995; 1998; 2004; Metelkin et al., 2000; 2005; 2012, Shipilov, Vernikovskiy, 2010). The 300 Ma collision is thought to have closed an oceanic basin that once separated Kara from Siberia and the Central Taimyr island arc that collided with the Siberian continent at ~600 Ma. The Northern Taimyr forms the collision belt between the Kara microcontinent on the north and the Central Taimyr/Siberian amalgamation on the south. The absence of Middle-Late Paleozoic ophiolite and island arc complexes in the Northern Taimyr is therefore curious.

We propose a reconstruction of the Early Paleozoic history of the Kara microcontinent as part of the amalgamation of the Arctida paleocontinent. Our reconstruction describes the mechanism of the 300 Ma collision of Kara with Siberia and the subsequent collision-caused deformation processes in the amalgamation of the greater Pangea supercontinent that was culminated in Permian time or ~280 Ma (Fig. 9) (Metelkin et al., 2011; 2012, Vernikovskiy et al., 2011).

Our paleotectonic analysis is based on paleomagnetic and geochronological data. The results indicate that the collision between Siberia and the Kara microcontinent was an oblique event. The orogen that was formed can be characterized as a transform orogen (Metelkin et al., 2005). During the Early Paleozoic and prior to the collision Kara moved northward on a system of large transform faults from the sub-tropic zone of the southern hemisphere to the subequatorial latitudes of the northern hemisphere while at the same time rotating counter clockwise, whereas the Siberian plate underwent a clockwise rotation (e.g., Fig. 8, 450 Ma).

The oppositely directed rotation of the interacting

Kara and Siberian continental masses led to their oblique convergence and “soft” collision. In the Late Silurian – Devonian, when there still was a “lens” of oceanic crust between the Siberian and Kara plates (Fig. 6, 430–400 Ma; Fig. 8, 420 Ma), the margins of the converging continents escaped significant shortening while sliding along the transforms and maintaining intact margins (Fig. 6, 430–400 Ma). It is possible that the oceanic crust was partially subducted beneath the Siberian plate; however the strike-slip processes were dominating. As a result the supra-subductional geologic complexes are lacking. The continent-microcontinent collision took place in the Late Carboniferous and culminated in the Permian (Fig. 6, panel for Carboniferous-Permian). The Carboniferous-Permian (300–260 Ma) collision re-deformed the Central Taimyr island-arc complex that was originally deformed when it accreted to the Siberian continent at 600 Ma.

In the course of oblique collision there was a thickening of the crust, accompanied by folding which migrated to the south-west (in modern geographic coordinates), regional metamorphism, and the formation of collisional granites (Vernikovskiy et al., 1995; Pease, 2001). As a result of compression in the frontal part of the Kara tectonic domain there was a gradual exhumation of the deeper parts of the crust of the deformed plate.

The transform faults which controlled the collision of the Kara and Siberia continental masses gradually evolved into thrust faults as shortening progressed. The oblique collision may have evolved into a more orthogonal-directed collision because of far-field interactions with the nearby continental masses of Laurentia, Baltica, Alaska-Chukotka, and Svalbard (Fig. 9). The most important among these thrusts is the Main Taimyr fault zone (Fig. 6), which can be regarded as the main suture of the Late Paleozoic Taimyr orogen. The progressing compression and the associated crustal thickening led to the “collapse” – fast thrusting and imbrication of the crust and post-collisional granitoid magmatism (Vernikovskiy et al., 1998) on the Main Taimyr thrust which separates the Central and North Taimyr zones.

This geodynamic paleoreconstruction for the Kara microcontinent shows the need and significance of such studies for the entire Arctic.

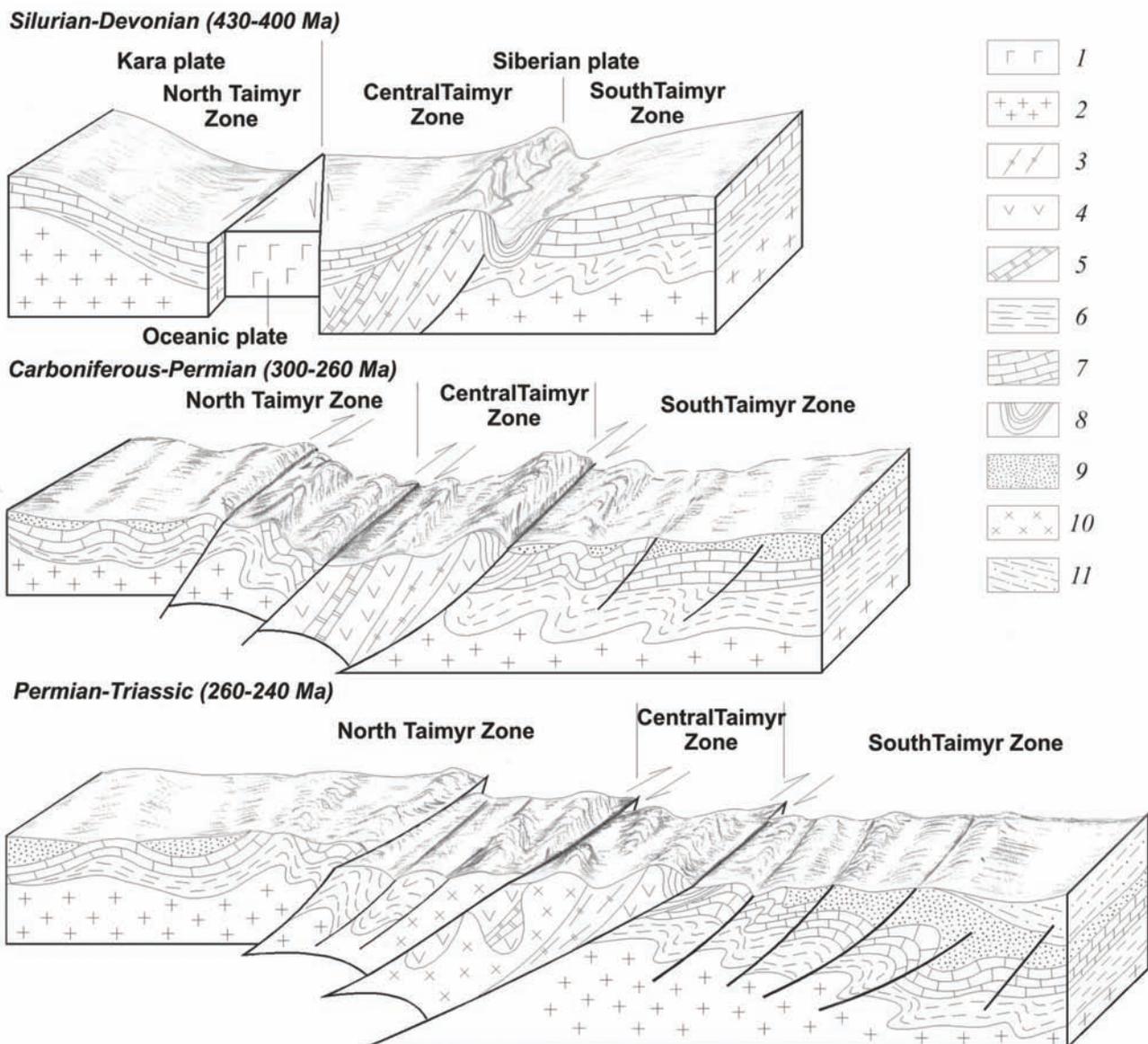


Fig. 6. A model for the structure transformation of the Siberian Taimyr margin during the interaction with the Kara microcontinent. 1 – oceanic complexes; 2 – Early Precambrian complexes of the Kara microcontinent and Siberian craton crystalline basement; 3–5 – Late Precambrian complexes of the Central Taimyr accretion zone: 3 – gneissic from cratonic terranes, 4 – volcanogenic-sedimentary from island arc terranes with ophiolites, 5 – carbonate shelf of passive continental margin terranes; 6 – Neoproterozoic-Cambrian flyschoid deposits of the Kara and Siberian continental margins; 7 – Paleozoic mainly carbonate shelf deposits (Ordovician-Silurian on the Kara microcontinent and Ordovician-Early Carboniferous on the Taimyr Siberian margin); 8 – Ediacaran-Early Carboniferous hemipelagic argillaceous-carbonate and black-schists deposits of the Pyasina-Faddey abyssal trough; 9 – Late Paleozoic mainly terrigenous deposits (Devonian and Carboniferous-Permian for Kara and Late Carboniferous-Permian for Southern Taimyr); 10 – Late Paleozoic (300–260 Ma) collisional granitoids; 11 – Triassic sandy-agrillaceous deposits, including the trap complex in the front of the Late Paleozoic orogen (basal horizons of the Mesozoic-Cenozoic Yenisey-Khatanga basin).

SPECULATIVE PALEOTECTONIC RECONSTRUCTIONS FOR THE ARCTIDA PALEOCONTINENT AND THE GREATER ARCTIC DURING LATE NEOPROTEROZOIC – PERMIAN TIME

Cryogenian (~750 Ma)

The Cryogenian is marked by the breakup stage of Rodinia – the supercontinent that formed around 1 Ga. According to current interpretations, the breakup of Rodinia began as soon as ~950 Ma and continued for a very long time until the Ediacaran (630–542 Ma) (Li et al., 2008). We join with Li et al., (2008) in believing that most of the classic Arctida blocks were composited into a continuous belt from fragments originating in diverse settings including the present-day northern margin of Laurentia, the former (750–650 Ma) southern margin of Siberia, and the present-day north-eastern margin of Baltica (Fig. 7, 750 Ma).

The Svalbard plate has a Grenvillian (1.3–1.0 Ga) basement, which has been confirmed by the identification of Grenvillian complexes on Spitsbergen (Gee et al., 1995) and on Novaya Zemlya (Korago et al., 2004). This allows us to infer the formation of Svalbard from collisional events during the establishment of Rodinia. On the basis of paleomagnetic data, Baltica is usually positioned in paleoreconstructions in such a way that in modern geographic coordinates the Grenvillian Sveconorwegian structures serve as the northern “ending” of the Grenvillian structures of Laurentia’s eastern margin. The Meso-Neoproterozoic fold belts of Amazonia are oriented in a linear fashion along Laurentia’s Grenvillian margin (Cawood and Pisarevsky, 2006). In this context it is logical to suppose that the Svalbard orogen structures form the northern (present-day) extension of the Grenville belt that marks the collisions between Laurentia, Baltica, and Amazonia that formed northern Rodinia.

Paleoproterozoic(?) crystalline complexes of the Kara microcontinent basement are known on the Severnaya Zemlya archipelago (Proskurnin, 1999; Proskurnin and Shul’ga, 2000) and in the northern part of the Taimyr Peninsula (Vernikovskiy and Vernikovskaya, 2001). The sedimentary cover on the Kara microcontinent is floored by Late Neoproterozoic flyschoid deposits which are overlain by a Paleozoic (Ordovician to Early

Carboniferous) sequence composed of carbonates, evaporates and terrigenous formations that indicate an epicontinental shelf regime. The structure of the gravity, magnetic, and other geophysical fields for the Kara microcontinent differ significantly from adjacent plates or blocks. The Kara microcontinent thus appears to form an independent block with a distinct internal structure.

Despite the distinctive differences between the geologic and geophysical structures of the Kara microcontinent and the Svalbard plate, the emplacement history and evolution of their modern margin (St. Anna Trough and North Siberian Sill) display a characteristic dextral strike-slip component, which was probably inherited from the Neoproterozoic-Paleozoic transform boundary between Svalbard and Kara (Shipilov and Vernikovskiy, 2010). From these observations we speculate that in the Meso-Neoproterozoic (Cryogenian) structure of Arctida (during the formation of Rodinia) the Kara microcontinent was located between the Greenland-Ellesmere block and the Svalbard block, from the latter possibly separated by a strike-slip fault system (Fig. 7, 750 Ma).

In our reconstruction the Early Precambrian structures of Arctida’s Greenland-Ellesmere block correspond to their current position near the Canadian margin of Laurentia. Our reconstruction infers that the Alaska-Chukotka block was located in close proximity to the Greenland-Ellesmere block and they did not change their positions as the northern (present-day) margin of Laurentia throughout the period 750–255 Ma. The detachment of the Alaska-Chukotka tectonic element from Laurentia occurred in the Jurassic (202–146 Ma), as part of the opening of the Canada basin (Grantz et al., 1998, Lawyer et al., 2002, Alvey et al., 2008). The Alaska-Chukotka tectonic element later collided with the Verkhoyan-Kolyma Siberian plate along the South Anyui (Novosibirsk-Chukotka) suture (Sokolov et al., 2002; 2009).

Unlike earlier models, our reconstruction does not include the New Siberian Islands and the Laptev Sea continental shelf (New Siberian block) in the structure of Meso-Neoproterozoic Arctida. The Neoproterozoic-Paleozoic evolution of the New Siberian block took place in a passive continental margin setting (Kuzmichev, 2009). The Paleozoic

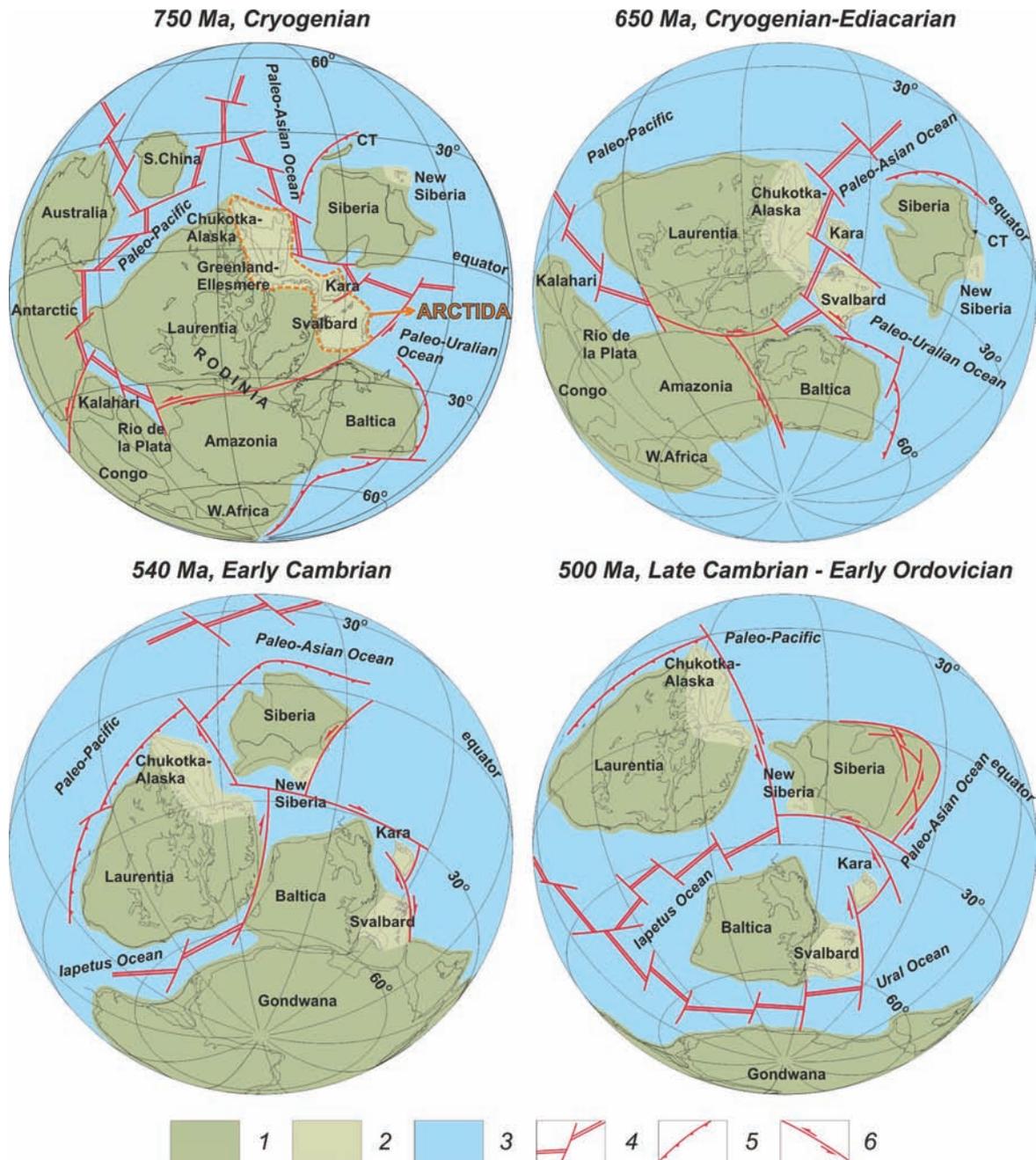


Fig. 7. Plate tectonic reconstructions for the evolution of Arctida and its dispersed fragments (Chukotka-Alaska, Kara, and Svalbard) from the Neoproterozoic to the Early Ordovician. 1 – continental masses; 2 – continental blocks of Arctida; 3 – oceanic basins; 4 – inferred position of spreading zones; 5 – active continental margins; 6 – general strike of the transform/strike-slip zones with indicated strike-slip kinematics.

geological complexes that exist on the New Siberian Islands are amazingly similar to the deposits of the Cis-Verkhoyan and South-Taimyr margins of Siberia. The lack of a pronounced tectonic suture in the Laptev shelf allows us to infer the genetic unity of the Paleozoic complexes of the New Siberian block and the north-eastern Siberian margin. Thus in our model in the Cryogenian the New Siberian block

was located far from the other Arctida blocks and during the Paleozoic it evolved as a part of the north-eastern (in geographic coordinates) Siberian margin.

According to our model, strike-slip displacements took a major role in the process of Rodinia's breakup and basically conditioned the tectonic dispersal of the supercontinent (Metelkin et al., 2007, 2012; Vernikovskiy et al., 2009). The

accepted position of the Siberian craton is based on paleomagnetic data for the south of Siberia (Metelkin et al., 2007, 2012) and is supplemented by data for the Taimyr margin (Vernikovskiy et al., 2011). The latter indicate that Central Taimyr island arcs were situated to the north from the Arctic margin of Siberia since 960 Ma. Consequently, at the time of a unified Rodinia (> 950 Ma) and later its breakup the Siberian northern margin should have been facing a paleo-ocean (Fig. 7, 750 Ma). The paleogeographic position of the Arctica subcontinent was to the southwest relative to Siberia and straddling the equator (Fig. 7, 750 Ma). The position of Kara, Svalbard and Alaska–Chukotka within Arctica is debatable and taken from (Li et al., 2008).

Cryogenian-Ediacaran (~650 Ma)

By the beginning of Ediacaran the Arctica subcontinent on the northern margin of Laurentia had moved south of the equator to subtropical latitudes (Fig. 7, 650 Ma). Some Arctica blocks were probably involved at this time in the breakup and dispersal of Rodinia, including the detachment of Baltica from Rodinia. Many examples show that the Rodinia breakup was accompanied by the shredding of the Rodinia margin into independent terranes such as the Kara microcontinent and the Svalbard plate. At the base of the Paleozoic sedimentary cover of those plates Late Precambrian riftogenic troughs and basins are present, which are clearly revealed by seismic measurements (Shipilov and Vernikovskiy, 2010).

At the same time on the eastern periphery of Baltica (Timan-Ural margin) the evolution of an active subduction zone can be inferred (Kuznetsov et al., 2007). Oblique subduction on one side of the Svalbard plate and extension on the other caused a transform regime of its displacement and interaction with the Kara plate.

Early Cambrian (~540 Ma)

Traces of the Cadomian orogenic event on the territory of Barentsia (Puchkov, 2003; Kuznetsov et al., 2007) in our opinion relate directly to the evolution of the Arctica structures. We believe this event to be a result of the collision between the Timan margins of Baltica (present-day northeast margin) with the Svalbard plate. From this time Barentsia was

joined to the East-European paleocontinent (Fig. 7, 540 Ma). The collision was structurally manifested in the formation of the divergent Timan-Pechora orogen. Its existence is confirmed by a deep cut-out of the Late Precambrian complexes in the basement of the Timan-Pechora sedimentary basin and by an outstanding unconformity in the base of the Paleozoic sedimentary cover (Kuznetsov et al., 2007). The 540 Ma collision was accompanied by the emplacement of I-type granitoid plutons, characterized by isotopic dates from 695 to 515 Ma (Kuznetsov et al., 2007). Kara continued to experience a mainly transform displacement relative to Svalbard. The transform/strike-slip regime characterized the entire northeastern Siberian margin (in geographic coordinates) and its displacements relative to distant Laurentia and Baltica. On the boundary between Laurentia and Baltica the Iapetus Ocean began to open (Fig. 7, 540 Ma).

Late Cambrian – Early Ordovician (~500 Ma)

By the Cambrian-Ordovician boundary (488 Ma) an active spreading regime widened the Iapetus oceanic basin (Fig. 7, 500 Ma). The breakup of the continental crust along the eastern (in present-day coordinates) Baltica margin and the formation of the Ural oceanic basin began at this time (Puchkov, 2003). Thus Baltica on almost all its periphery (except the north) was surrounded by young oceanic spreading centers whose growth dynamic set up a counter-clockwise rotation of the plate, which is confirmed by paleomagnetic data (Torsvik et al., 1991; Cocks and Torsvik, 2002). The northern Baltica margin including Svalbard was separated from Siberia by large-scale strike-slip faults, which caused a gradual drift of the Kara block towards Siberia (Metelkin et al., 2005).

Late Ordovician (~450 Ma)

The Iapetus oceanic basin began to close at the end of the Middle Ordovician. Active subduction occurred widely on all of the margins of the continents that surrounded the Iapetus Ocean. Baltica began its movement across Iapetus toward Laurentia. The Svalbard-Baltica margin and Kara located on its periphery were drawn significantly closer to the Taimyr margin of Siberia by mainly multidirectional rotation of these continental masses.

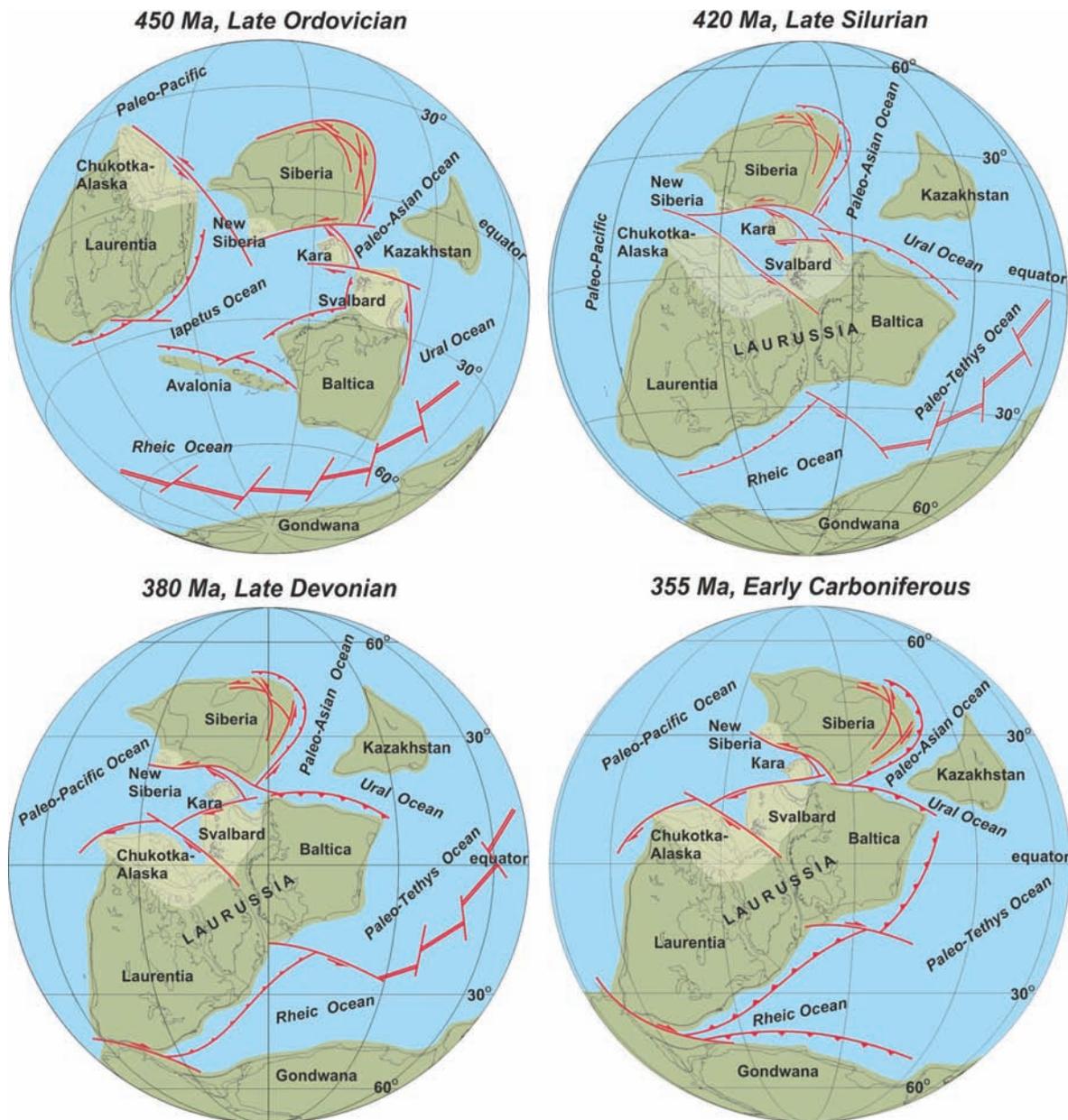


Fig. 8. Plate tectonic reconstructions for the evolution of Arctida's dispersed fragments from the Late Ordovician to the Early Carboniferous. See legend keys on Fig. 7.

This entire system continued its general drift towards the equator (Fig. 8, 450 Ma and 420 Ma).

Late Silurian – Late Devonian (~420–380 Ma)

During this time the collision between Laurentia and Baltica (Laurussia) took place (Golonka et al., 2003). Along with the formation of the Scandinavian orogen the Caledonian orogeny also affected Svalbard and the north-eastern Greenland margin, later spreading along the Greenland-Ellesmere area

of Laurentia. Thus, by the end of the Silurian the Ellesmere-Alaskan margin of Laurentia we infer the existence of an active subduction zone where the relicts of the Iapetus Ocean were consumed. The Kara microcontinent already was approaching the Taimyr margin of Siberia (Metelkin et al., 2005). The early stages of the Kara-Siberia collision occurred along a transform fault. The inferred transform fault collision mechanism does not exclude the existence of oceanic crust fragments between the Siberian

continent and the Kara microcontinent. Apparently there also existed a narrow space of oceanic crust between Svalbard and Kara. The Ural margin of Baltica and the south-western Siberian margin were characterized by intense subduction magmatism, which indicates the closure of the Ural and Paleo-Asian oceanic basins that were separated by the Kazakhstan plate (Fig. 8, 420 Ma). The collisions of the Siberian and Baltic plates took place along strike-slip faults within their modern arctic margins. As a result it was already the mid-Paleozoic (~380 Ma) when the component Arctic blocks of Arctida were reassembled into their Cryogenian (~750 Ma) configuration. By this time the Arctida blocks were located near the equator (Fig. 8, 380 Ma). By the end of the Devonian the Arctida assemblage formed a continental “bridge” between Siberia and Laurussia (Laurentia/Baltica). According to our reconstructions and available paleomagnetic data for the Early Paleozoic of the Kara microcontinent (Metelkin et al., 2005) we are inclined to believe that the Siberian margin in the Silurian-Devonian did not have any common boundaries with Laurussia. On the west, an embayment of the Paleo-Pacific Ocean separated the Siberian and Laurussia continental blocks during subsequent Paleozoic evolution. The Siberia and Laurussia plates were closest to each other by the end of the Silurian. The transform regime was dominating along all continental margins of Arctida at ~380 Ma. Strike-slip faults accommodated the sliding of Siberia and Kara to the east along the north-western (in paleogeographic coordinates) margin of Laurussia. This displacement widened the embayment facing the Paleo-Pacific Ocean into a wide marginal sea basin lapping the margins of Alaska-Chukotka, Svalbard, Kara, and New Siberia–Cis-Verkhoyan. It is probable that the inferred strike-slip displacements were driven by seafloor spreading Paleo-Pacific Ocean. To the east, subduction and the closing of the Paleo-Pacific and Ural Oceans added to the retreat of Siberia (and the Arctic blocks sutured to its margin) away from the Alaska-Chukotka margin of Laurentia.

Early Carboniferous (~355 and 330 Ma) and Late Carboniferous (~305 Ma)

The Carboniferous period witnessed the closing of the oceanic basins that divided the continental

masses of Laurussia (Baltica and Laurentia), Siberia, and the Kazakhstan composite terrane. These collisions culminated with the formation of Laurasia – the supercontinent that along with Gondwana formed Pangea at the Carboniferous-Permian boundary (Fig. 9, 280 Ma) (Zonenshain et al., 1990; Golonka, 2002).

At the beginning of the Carboniferous (355 Ma) the main blocks of Arctida (e.g., the Alaska-Chukotka, Svalbard, Kara, and New Siberian blocks) and the related continental margins of Laurentia, Baltica and Siberia occupied the space between the equator and 30° N. This entire paleo-shelf was tectonically stable and underwent a slow “opening” caused by the eastward retreat of Siberia. The main cause for this retreat probably was seafloor spreading in the Paleo-Pacific Ocean. The closing of the Paleo-Asian and Ural Oceans as well as the progressive narrowing of the Rheic and Paleo-Tethys Oceans was essentially complete by ~305 Ma. These collisions and the interactions with the Paleo-Pacific Ocean on the western side of the Laurasian continental agglomerate contributed to the transform-fault regime of the paleo-shelf described above and to the clockwise rotation of the system (Fig. 8, 355 Ma and Fig. 9, 330 Ma).

By the Late Early Carboniferous (330 Ma) all the continents continued drifting northwards, moving closer to each other. for the final amalgamation of continental masses into a unified supercontinent began in Late Carboniferous time (Fig. 9, 305 Ma). Subduction of the Ural Ocean at the northeast Baltica margin was completed (Puchkov, 2003). The Paleo-Asian Ocean collapsed in a regime of oblique subduction (Dobretsov, 2003; Windley et al., 2007). At the end of the Early Carboniferous (Fig. 9, 330 Ma), collision tectonics began at the Taimyr margin of Siberia (Vernikovskiy et al., 1995; Vernikovskiy, 1996). At Taimyr, the collision proceeded as a soft interaction between sialic masses in oblique impact conditions with them rotating relatively to each other (Metelkin et al., 2005, 2012). Geochronological data indicates that as early as in the Late Carboniferous (305 Ma) syn-collisional calc-alkaline granitoids began to intrude Taimyr (Vernikovskiy et al., 1995; Pease, 2001). Paleomagnetic data, described above, forms the chief evidence for the inferred strike-slip component of Taimyr deformation. Thus, large-

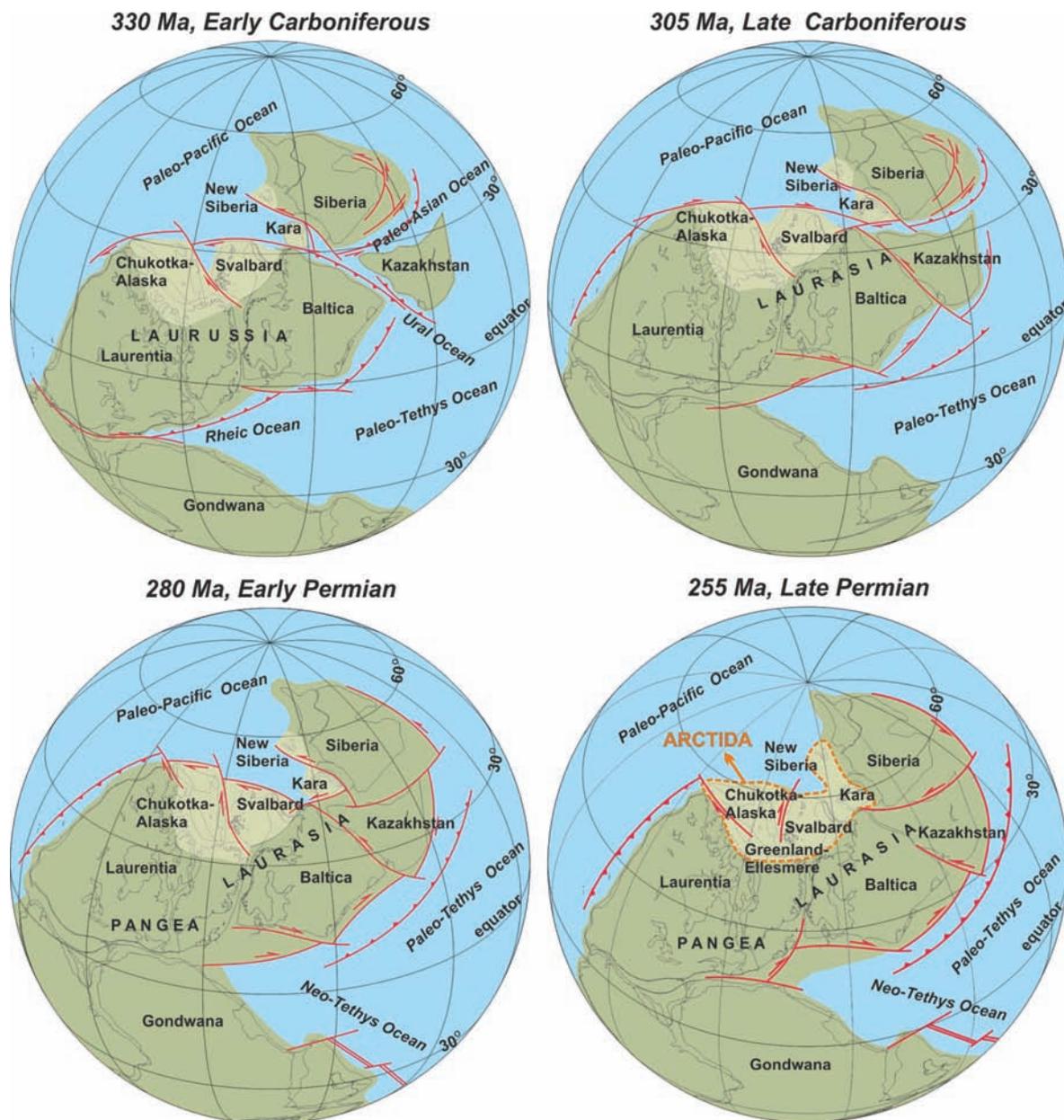


Fig. 9. Plate tectonic reconstructions for the evolution of Arctida and its dispersed fragments from the Early Carboniferous to the Late Permian. See legend keys on Fig. 7.

scale strike-slip fault zones, along which Kara “slid” during the entire Paleozoic, in the end led to the collision between the Kara microcontinent and the Siberian continent and subsequent formation of the fold-and-thrust structure of the Taimyr – Severnaya Zemlya folded area (located in Fig. 1b).

Early to Late Permian (280–255 Ma)

At the beginning of the Late Carboniferous (Fig. 9, 305 Ma) the main continental collisions involved in the formation of Pangea had already started (Zonenshain et al., 1990; Golonka, 2002; Metcalfe,

2002; Dobretsov, 2003). The Carboniferous-Permian boundary (280 Ma) is the time when the Laurasia and Gondwana blocks united in a single supercontinent – Pangea (Fig. 9, 280 Ma). The deformations caused by the collision and orogenic events continued within Laurasia, related mostly to strike-slip displacements along old sutures. Available paleomagnetic data indicate that the intraplate strike-slip displacements between rigid tectonic units of Eurasia (the Siberian and East European cratons) continued until the Cenozoic (Metelkin et al., 2010, 2012). By the end of the Permian (Fig. 9, 255 Ma)

the mainly transform-fault-driven amalgamation of the Kara – New Siberian and Svalbard – Novaya Zemlya continental margins into a single shelf structure was accomplished (Shipilov, 2003; 2008). The collisions caused the curved structure of the Pay-Khoy – Novaya Zemlya area (Korago et al., 1992; Scott et al., 2010).

Thus the Permian-Triassic boundary can be considered as the time of the second formation of Arctida or “Arctida-II”. Arctida-II is located in Pangea’s northern edge near the 60th parallel, occupying the moderate and sub-polar regions of the Northern Hemisphere (Fig. 9, 255 Ma). Subsequently, in the Mesozoic as a result of the opening of the Amerasian basin, the large Alaska-Chukotka block was rifted away from the Greenland-Ellesmere margin (Grantz et al., 1998). Its collision with the Cis-Verkhoyan Siberian margin in the Cretaceous along the South Anyui suture (Drachev et al., 1998; Sokolov et al., 2002; 2009; Kuzmichev, 2009) established the main structural features of the current arctic shelves of the Eurasian and North-American continents (Natalin, 1999; Khain et al., 2009).

CONCLUSION

We infer the existence at two different times of two Arctic subcontinents comprised of essentially the same crustal fragments. The first subcontinent, “Arctida-I” broke apart and the fragments were dispersed through independent plate movement paths before being reassembled as the second subcontinent, “Arctida-II.”

Arctida-I was an amalgamation of Mesoproterozoic terranes that “welded” together elements of Laurentia, Siberia and Baltica within the Rodinia supercontinent at 1 Ga. The Rodinia disintegration caused the breakup of Arctida-I into independent tectonic fragments which experienced highly diverse displacement paths over the next 720 million years (1,000 to 280 Ma). The evolution of Neoproterozoic and Paleozoic oceanic basins between these tectonic fragments led to their reorganization into a new configuration in Arctida-II – a Late Paleozoic subcontinent which again “welded” together the continental masses of Laurentia, Siberia and Baltica within Pangea. The breakup of Arctida-II in the Mesozoic and the

Cenozoic with the formation of the north Atlantic basin and the Amerasian and Eurasian basins of the Arctic Ocean led to a significant redistribution of the continental masses, especially in the north-eastern part of the modern Arctic and to the formation of the modern shelves of the Eurasian and North-American continents.

Our paleotectonic reconstructions will of course be improved after further investigations. For this purpose complex geostructural, geochronological, paleontological and especially paleomagnetic data will be of paramount importance.

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