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ORIGINAL PAPER

Structural stages and evolution of the Urals

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Abstract Five main structural and historical stages are established in the territory of the Urals: 1) Archean-Paleoproterozoic, a time of formation of the Volgo-Uralia subcontinent and its amalgamation with the other blocks of the craton of Baltica; 2) Riphean-Vendian (Meso- and Neoproterozoic), a stage that was finished with formation of Timanides; 3) Paleozoic-Early Mesozoic stage, corresponding to the development of the Uralides; 4) Mid-Jurassic-to Miocene platform stage; 5) Pliocene-Quaternary neo-orogenic stage. In this paper stratigraphic data are discussed, schemes of the structural zonation are presented, and the problems of the structural geology and geodynamics of sedimentary and magmatic complexes are discussed in a chronological order. Ideologically, the paper is based on plate and plume tectonics, in their modern versions.

Abbreviations

MUF Main Uralian Fault
MGA Main Granitic axis
LIP Large Igneous Province
BMA Bashkirian megaanticlinorium

Introduction

When we say the Urals and Uralides, we mean first and foremost a linear Paleozoic orogen, ca. 2,000 km long, composing the western part of the huge Uralo-Mongolian foldbelt. But when we say Ural Mountains, we mean a moderately-high mountain range and accompanying linear hills, formed as the result of a recent intraplate deformation of a peneplain and a rather weak orogeny. In addition, the earlier Precambrian structural evolution and history of the region is quite different and has nothing to do with the Paleozoic structure of the Uralides.

The rocks occurring in the Urals range from the Archean to the Quaternary. They compose five major structural levels and epochs referred to in this paper as stages (Fig. 1): (1) highly metamorphosed Archean-Paleoproterozoic crystalline basement, formed in a process of assembling of the Craton of Baltica; (2) Meso- to Neoproterozoic rift, continental margin and island arc-oceanic complexes, deformed as a result of subduction, followed by orogeny that culminated with the formation of Timanide foldbelt along the periphery of Craton of Baltica; (3) Paleozoic–Lower Jurassic rift, continental margin, island arc-oceanic and orogenic complexes of the Uralides; (4) Middle Jurassic–Paleogene–Miocene platform cover, formed during a peneplain formation and preserved mostly at the periphery of the Urals; and (5) Pliocene–Quaternary (neo-orogenic) complexes, formed as the result of a far-field influence of the Alpine-Himalayan orogenesis. Every latter stage is shorter than the previous one.

A detailed description of these stages, accompanied by a discussion of most daunting problems of the stratigraphy, tectonics, geodynamics and metallogeny of the Urals is provided in a recent book, published in Russian (Puchkov 2010a) and awarded of A.D.Archangelsky academic Premium in 2011.

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The stages of the Urals

Archean-Paleoproterozoic stage

This stage is represented by exclusively para- and orthometamorphic rocks which have undergone regional granulite and

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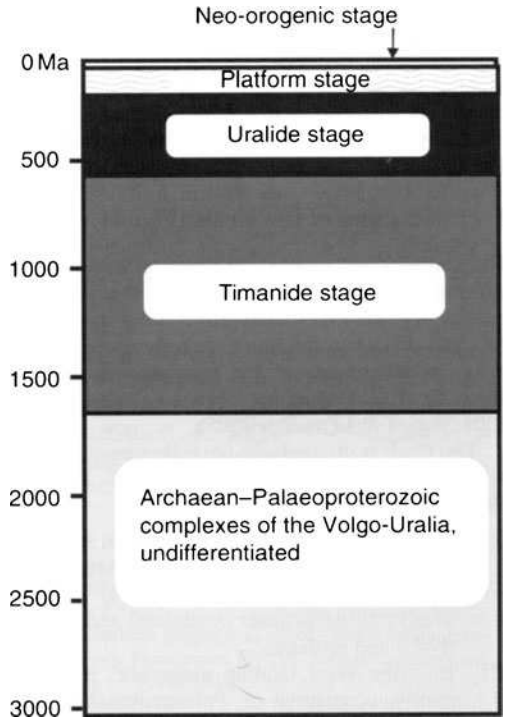


Fig. 1 Structural stages of the Urals (Puchkov 2005)

then amphibolite facies metamorphism, and more locally, in shear zones, a retrograde low metamorphism. The only place in the Urals where these basement rocks are exposed at surface is the Taratash uplift zone in the Southern Urals. The oldest U-Pb age of zircons, acquired from granulites of this block, is about 2.9 Ga. The youngest age for the ampibolite metamorphism and granite formation is close to 1.8 Ga. (Sindern et al. 2006; Krasnobaev et al. 2011). The block is rather small-sized, but is linked to the crystalline basement of the East European platform. According to seismic profiles URSEIS-95 and SB ESRU, discussed in the "Continent–continent collision and orogeny. The main Bashkirian-Permian phase" of the paper, the basement is traced continuously, gently dipping, far under the Urals and its signature is lost under the middle of the Tagilo-Magnitogorsk zone. In the Polar Regions of the Urals, Archean rocks are not recognized and may not exist, but Paleoproterozoic rocks–are suspected in the most metamorphosed cores of some thermal domes, based on rare U-Pb ages. Proterozoic crystalline basement as a whole can be traced under the Polar Urals according to seismic data.

Meso-Neoproterozoic stage and the Timanides

The Timanides got their name from the Timan Range in the Northwest of Russia. But the Timanian orogen is a much larger structure (Fig. 2), that can be traced along strike at least for 3,000 km from the southernmost Urals to Varanger-fjord in Norway along the margin of Craton of Baltica. Across its strike, it is traced from the Timan Range under the cover of the Timan-Pechora basin to the basement of the Old Kimmerian foldbelt of Novaya Zemlya. The complexes of Timanides are uplifted in anticlinal structures of Uralides (Central Uralian zone) and Caledonides (Svalbard).

The term Timanides was coined by N.S.Shatsky in 1937 (Shatsky 1964). Later on, he suggested the term Baikalides, with ages between Cryogenian and the Early Paleozoic, which is, however, a too general term to be used in our case since the structure of Timanides corresponds to a collisional orogen formed mostly between 600 and 550 Ma. According to modern reconstructions, supported by geophysical data, Timanides make a wide fan-shaped (in plan view) structure open to the north. It can be subdivided into two major structural zones: Externides and Internides (Fig. 2).

The Externides compose the outer part of the Timanian orogen, adjacent to the Archean-Paleoproterozoic craton (continent) of Baltica. The crystalline basement of the craton continues underneath the shelf and bathyal facies sediments of the former continental margin. It is exposed at the surface in the Taratash uplift zone of the Bashkirian megaanticlinorium (BMA). The important structure of the Externides is the foredeep filled with Upper Vendian (Late Ediacaran) molasse sediments formed as a result of erosion of the Timanide orogen, and a marginal anticlinorium of Timan Range. The major boundaries in the Externides are the line dividing shallow shelf and bathyal facies sediments of the margin and the outer boundary of regional metamorphism. The Internides are divided from the Externides by a hypothetical suture zone and include a series of anticlinoria and synclinoria, with some ophiolite sutures in the latter and thermal domes with Paleoproterozoic cores in the former.

Externides

Meso- and Neoproterozoic complexes The Meso- and Neoproterozoic deposits of Externides are exposed most completely in the western parts slope of the Southern Urals, where one of the most important Meso- and Neoproterosoic sections of weakly altered sedimentary and volcanic rocks is developed and accepted as the stratotype for the Riphean eratheme. This is the major Precambrian stratigraphic unit in Russia (Table 1). The Meso-Neoproterozoic section in the Bashkirian megaanticlinorium (BMA) of the Southern Urals is well correlated with platform aulacogen (graben) sections of the adjacent East European platform, and belongs to the eastern flank of a great sedimentary basin, probably grading eastwards into a continental margin (poorly exposed). The Uralian part of the basin was folded in Vendian and Late Paleozoic time and belongs to Externides of Timanian orogen (Fig. 2). Owing to the uniqueness of this section we include a more detailed description in the following.

The Riphean of BMA is divided into four systems (Burzyanian, Yurmatinian, Karatavian and Arshinian (the

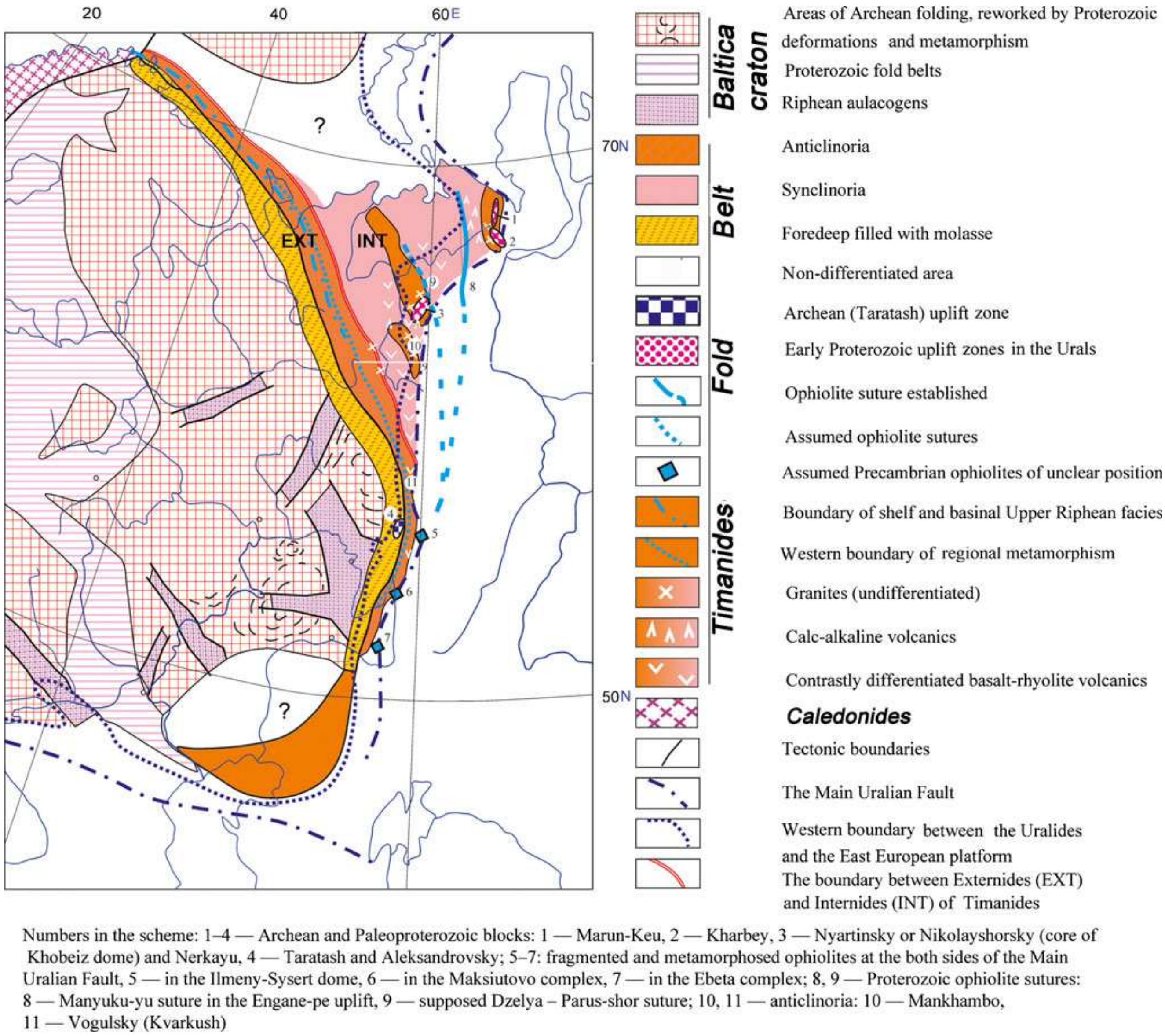


Fig. 2 The Tectonic scheme of Timanides (Puchkov 2010a, modified)

latter was recently added to the Riphean as a result of new isotope data) (Figs. 3 and 4).

The Lower Riphean (Burzyanian) consists of three Formations (Keller et al. 1983; Kozlov et al. 1989). The lower, Ai Formation is composed of basal polymictic conglomerates, overlying the crystalline Taratash complex and containing its fragments, coarse-grained polymictic sandstones, arkose sandstones and shales, often carbonaceous, with rare layers of sandy dolomites. The characteristic member of the Formation is a thick (500–1,000 m) volcanosedimentary unit, called the Navysh subformation, represented by metabasalt flows and variously grained polymictic sandstones, quartz and arkose sandstones and siltstones, gravelstones and conglomerates. The total thickness of the Ai Formation varies from 1,700 to 2,500- m. In the lower part of the volcanosedimentary unit, ca. 200 m above the base of the Formation, the age of the basalts was determined as 1,752±11 Ma (Puchkov et al. 2012b; Krasnobaev et al. 2013). Taking into account the presence of sedimentary rocks below the dated volcanics, we accept the age of the lower boundary of the Riphean at ~1,760 Ma, admitting ca. 10 Ma that were needed to accumulate 100–150 m of sediments (Fig. 3, Table 1).

The middle, Satka Formation is composed of predominant dolomites and limestones, often with stromatolites, with layers of shales and carbonaceous shales, often limy, with Lower Riphean microfossils. Its thickness is 2,000–2,400 m.

The upper, Bakal formation is composed of carbonaceous shales, dolomites and subordinate limestones with layers of

Table 1 Correlation between the International (Gradstein et al. 2004) and Uralian (suggested by Puchkov, this paper) stratigraphic charts

International scale					Uralian scale		
Eonotheme, eon	Eratheme, era	System, period	Age limit, Ma		Eratheme, Era		System, period
Proterozoic			542				
	Neoproterozoic	Ediacaran	630	600	Upper (Late) Proterozoic		Vendian
		Cryogenian	850	750		Riphean	Arshinian
		Tonian	1000	1030			Karatavian
	Mesoproterozoic	Stenian	1200				
		Ecstasian	1400	1400			Yurmatinian
		Calymmian	1600				Burzyanian
	Paleoproterozoic	Staterian	1800	1760	Lower (Early Proterozoic)		
		Orosirian	2050				
		Rhiacian	2300				
		Siderian	2500				

quartz siltstones and sandstones. The carbonates contain stromatolites, the shales–microfossils. Its thickness is 1,200–1,400 m. The formation completes the section of the Lower Riphean that has a total thickness between 4,900 and 6,300 m (Fig. 3).

Such a stratigraphy is characteristic only for the northern part of BMA. In the southern part of it, the Lower Riphean is divided into the terrigenous-carbonate Bolsheinzer, Suran and Yusha Formations, analogous to the Ai, Satka and Bakal Formations of the northern part.

The Bolsheinzer Formation (2,100 m) is composed of quartz and (to a lesser extent) of feldspar-quartz siltstones, with subordinate dolomites, limestones, shales, and carbonaceous shales; rare layers of small-pebble conglomerates are also present. The lower boundary of the formation is not exposed: volcanic rocks are absent.

The Suran Formation (1,000–2,000 m) is represented by dolomites and dolomitized limestones at the base and top of the formation, while the middle part is composed of shales, carbonaceous shales, marls, feldspar-quartz siltstones and rare sandstones. The upper and lower boundaries are transitional.

The Yusha Formation (900–1,100 m) is predominantly terrigenous, and mostly contains quartz and feldspar-quartz sandstones and siltstones, shales, carbonaceous shales; layers of dolomites and limestones are rare. The formation completes the Lower Riphean section of the southern part of the megaanticlinorium; here the total thickness of the Lower Riphean is 4,400–6,000 m.

The Middle Riphean (Yurmatinian) includes the type sections of the Middle Riphean. The Yurmatinian series are developed on both sides of the South Uralian watershed and is represented by quartz and feldspar-quartz sandstones and siltstones, dolomites, limestones, shales and carbonaceous shales. In the eastern sections, the thick volcanogenic and volcanosedimentary Mashak Formation lies at the base of the series with an unconformity, while in the west it is absent; the younger conglomerates of Zigalga Formation overlie the Burzyan series with a pronounced erosional contact. The Mashak Formation can be subdivided into 3 subformations.

The Lower and Middle subformations of Mashak Formation (from bottom to top) are composed of: basal conglomerates with quartzite pebble, rhyodacites, metabasalt flows, carbonaceous shales, polymictic siltstones and sandstones, conglomerates with quartzite pebble, quartzite-sandstones, tuff siltstones, and finally tuff sandstones. The total thickness of the Lower subformation is 800–1,000 m, and the Middle subformation–900 m. The rhyodacites were reliably dated at 1,380–1,385 Ma by U-Pb methods (IDTIMS, CA IDTIMS, SHRIMP) in three laboratories around the world using zircons obtained from several places in the Bashkirian meganticlinorium (Puchkov et al. 2012a, b). So we accept the lower boundary of the Yurmatinian at ca. 1,400 Ma, taking into account the presence of 300-m thick sediments and volcanics between the base of the Mashak Formation and the dated horizon (Table 1). A geochemical study of the volcanic rocks indicates an interplate, rifting geodynamic setting, and continental reconstructions suggest that the area where Mashak Formation formed was part of a LIP (Large Igneous Province) dated at ca. 1,380 Ma (Ernst et al. 2006; Puchkov et al. 2012a). The Upper subformation is represented predominantly by polymictic siltstones, shales, carbonaceous shales; a single layer of limy dolomites is present.

In the western limb of BMA, Mashak formation is not developed, and the basal unit of the Yurmatinian series is the Zigalga formation, which unconformably overlies the formations of the Burzyan series (Bakal and Satka Formations). In the eastern slope of BMA, it overlies the Mashak formation with a transitional contact. The formation is divided into three

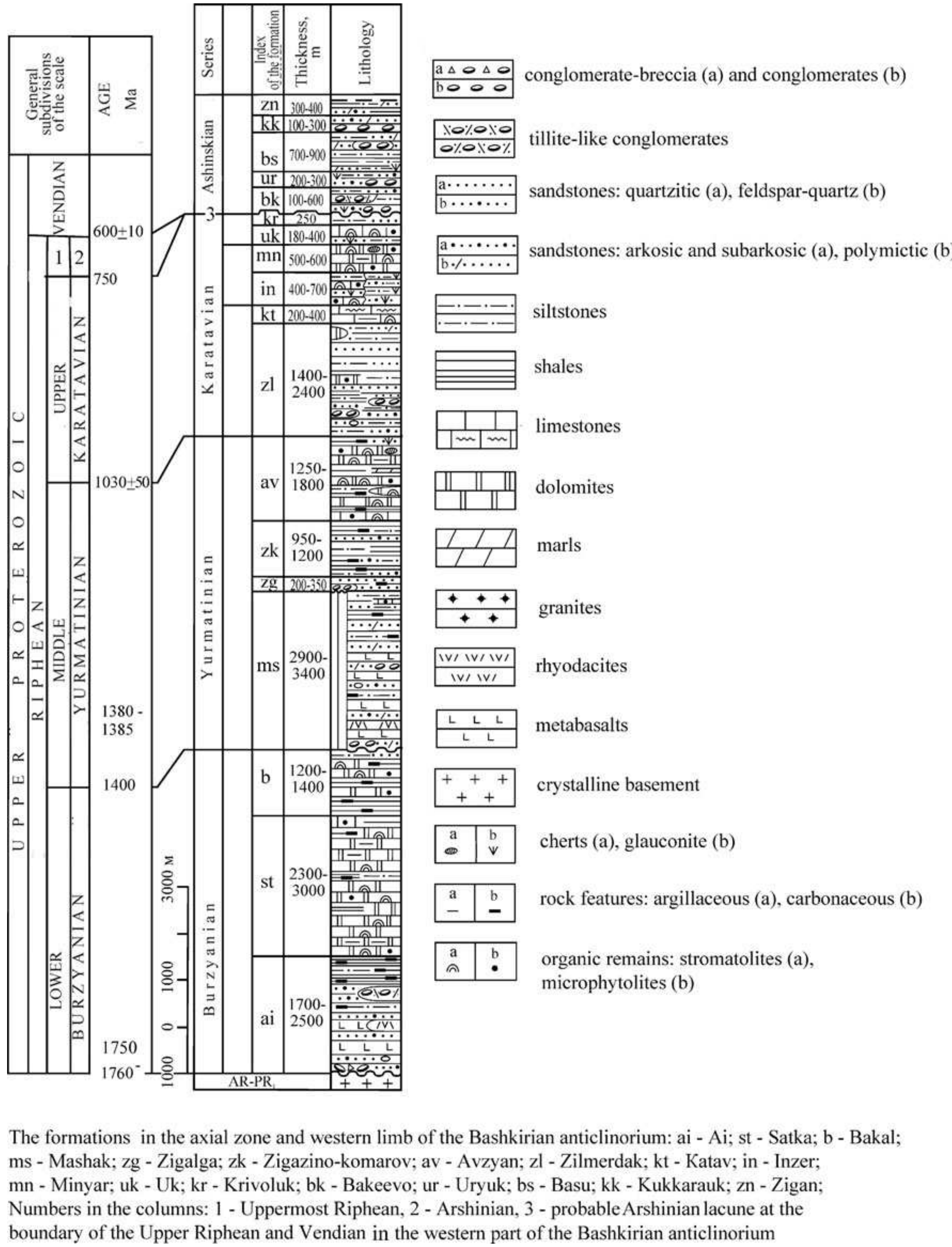


Fig. 3 Stratigraphic column of the Meso-to-Neoproterozoic deposits in the Southern Urals (Bashkirian anticlinorium). After VI Kozlov 2006, strongly simplified and changed after incorporating the latest data

subformations; the Lower and the Upper ones sandy and the Middle–siltstone-shaly, often carbonaceous. The total thickness of the formation is 250 to 400 m (Fig. 3). The predominant lithology of the Zigalga formation is a monomineral light-coloured quartz sandstone, often quartzite-like. It is thick-layered and often forms characteristic big-blocky surfaces on the tops of ridges. The sandstones also exhibit a cross-bedding and ripple marks.

Above that, only sedimentary rocks accumulated: terrigenous (shales, often carbonaceous, alternating with quartz and

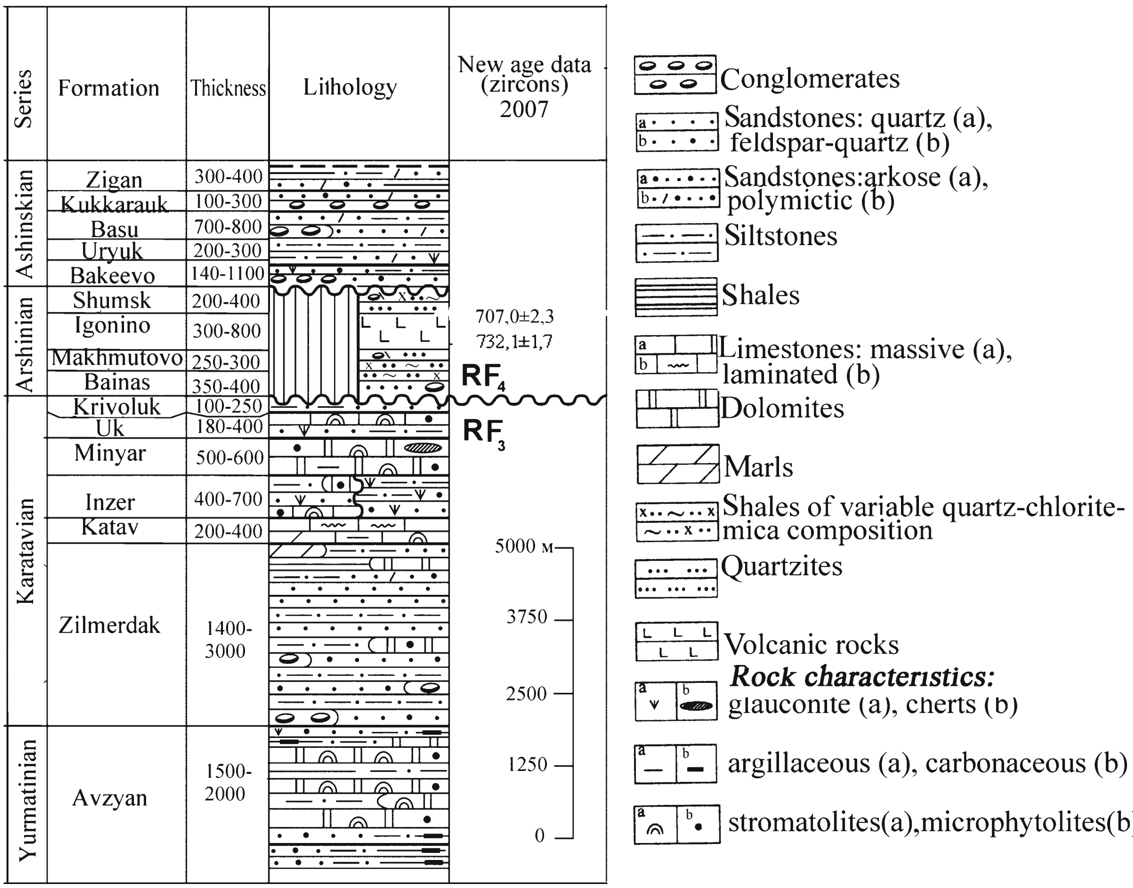


Fig. 4 Stratigraphic column of the Upper Riphean–Vendian of the eastern limb of the Bashkirian meganticlinorium (Kozlov et al. 2011, modified)

feldspar-quartz siltstones and sandstones of the Zigazino-Komarov Formation, 1,000–1,200 m thick) and carbonate-terrigenous (dolomite and limestones, often with Middle Riphean stromatolites and microphytolites), sandstones of quartz and feldspar-quartz composition, shales and carbonaceous shales of Avzyan Formation, 1,500–2,000 m thick. The general thickness of the Yurmatinian series in the eastern sections reaches 5,500–6,700 m

The Upper Riphean (Karatavian) is represented by carbonate-terrigenous sediments of the Karatavian series, which vary in lithology and colour and is the most widespread among the Upper Proterozoic sections of BMA.

The basal unit of the Upper Riphean is the Zilmerdak Formation. It overlies the Yuruzan series conformably, and is represented by four subformations, each of a distinct appearance, composed mostly of terrigenous sediments and representing three cycles of sedimentation (Biryan, Nugush, Lemeza). The lower, basal Biryan subformation consists of feldspar-quartz, arkosic to subarkosic sandstones, siltstones with layers and lenses of shales, conglomerates and rarely, dolomites. This subformation has a variable thickness averaging about 800 m.

The next unit, Nugush subformation is represented mostly by siltstones and shales with interlayers and packages of sandstones. These sediments strongly differ from the Biryan sediments because of their less coarse grained character and grey colour. The thickness of the Nugush subformation varies from 200 to 300 m.

The next, Lemeza cycle is represented by two subformations: Lemeza and Bederysh. The Lemeza subformation is composed predominately of quartz sandstones with rare layers of siltstones and shales. The sandstones are monomineralic (quartzite-like) and light-colored. It is one of the best marker horizons in the Karatavian of the Southern Urals. The thickness is 100–250 m. The contact with the underlying Nugush subformation is transitional.

The Bederysh subformation is represented by quartz-felspar, quartz and polymictic sandstones and siltstones, shales and rare dolomites. Its thickness is 250–400 m, and total thickness of the Zilmerdak formation is from 2,000 to 3,000 m.

Higher up is the Katav Formation, composed of limestones, marly limestones and marls. Owing to this composition, the occurrence of specific stromatolites and characteristic pink and greenish colours, the formation is easily recognized and serves as one of the best marker horizons in the Riphean of the Southern Urals. It has gradual transitions with the underlying Zilmerdak formation; its thickness is 200–400 m. It is overlain by the Inzer Formation, mostly terrigenous in composition, represented by quartz-felspar and quartz sandstones, often with glauconite and shales; less common are limestones and dolomites, with Upper Riphean stromatolites and microphytolites. The Inzer Formation shows gradual transitions with the Katav Formation. The thickness is 400–700 m.

Next is the Minyar Formation which is composed of dolomites and dolomitized limestones with Upper Riphean stromatolites and microphytolites. In the upper part of the section there are layers and lenses of cherts. It has gradual transitions with the underlying Inzer formation. The thickness is 500–600 m.

The overlying Uk Formation is represented by limestones with stromatolites and microphytolites, sandstones and siltstones of glauconite-quartz, quartz, feldspar-quartz composition, with shale, often containing microfossils (Uk microbiota). The thickness of the formation is 180–400 m.

The uppermost Krivoluk Formation is developed only in the easternmost sections of BMA. It overlies the Uk formation and is composed of polymictic, quartz and feldspar-quartz sandstones and siltstones with layers of shale and glauconite shale. It has gradual contacts with the underlying Uk formation. The thickness of the formation is 100–250 m and it marks the top of the Karatau series; the total thickness of the latter varies between 3,380 and 5,350 m.

The Uppermost Riphean (Arshinian) series was previously treated as a Formation and correlated with the Lower Vendian Bakeevo Formation of the western sections of the Southern Urals. In the General Precambrian Scale of Russia the lower boundary of the Vendian is decreed at 600±10 Ma (Amendments to the Stratigraphic Code of Russia 2000). But according to the latest data (Puchkov et al. 2012b), zircons, obtained in eight samples of volcanics in the middle of the series (Igonino Fm), indicate at their polychronous character with two peaks at 707,0±2,3 and 732,1±1,7 Ma, respectively what let us suppose a long history of volcanism and sedimentation during the Igonino time. Therefore, we need to introduce a new stratigraphic unit–the Final (Uppermost) Riphean (Arshinian series), indexed as RF_4. The lower boundary of this unit is accepted at ca. 750 Ma, taking into account the thick sedimentary succession between the site of the dated sample and the base of the Arshinian (Fig. 4, Table 1). In the stratotypic sections the Arshinian series includes the Bainas, Makhmutovo, Igonino and Shumsk Formations (Fig. 4, Kozlov et al. 2011).

The Bainas Formation is composed of sericite-chlorite-quartz slates, often limy, with carbonate–quartz and dolomite layers, and at the base of the formation there are conglomerates and quartz sandstones. It overlies the Karatavian with an erosional contact. The thickness is 350–400 m. The Makhmutovo Formation is represented by quartz and feldspar-quartz sandstones, tillite-like conglomerates, sericite-quartz schists and quartzites. It is conformable with the Bainas Formation and its thickness is 250–300 m. Taking into account recent dating of the Igonino Formation, the tillite-like conglomerates can be correlated with Kaigas–the earliest Cryogenian glaciation (cf. Chumakov 2011). The Igonino Formation is composed of metabasalts, their tuffs and tuff breccias, with layers of terrigenous rocks. It has a gradational contact with the underlying sediments. Its thickness is 300–800 m. The 200–400 m thick Shumsk Formation is represented by sericite-chlorite-quartz schists, quartzite sandstones and tillite-like conglomerates (= Rapitan or Sturtian glaciation?). The contact with the underlying formation is transitional. The total thickness of the Arshinian deposits varies between 1,100 and 1,900 -m, depending on the thickness of volcanic rocks, which is very variable. Contacts with the younger Vendian deposits do not exist: the Arshinian series is developed only in the eastern limb of BMA, while the type section of the Vendian is restricted to its western limb.

The Vendian (Asha series) in its type section is divided into five formations: Bakeevo, Uryuk, Basu, Kukkarauk and Zigan, with a total thickness of 1,500 m. In the stratigraphic schemes of the Urals (All-Russian Stratigraphic Committee 1993) the series is attributed to the Upper Vendian, while the Arshinian is attributed to the Lower Vendian. However, the lower boundary of the Vendian was decreed at 600±10 Ma (Amendments to the Stratigraphic Code of Russia 2000, p.21). Even if we accept an older age of this boundary–say, 630 Ma, like the Ediacaran of the international scale, or ~650 Ma as supposed by Semikhatov (2008), Chumakov (2011), the Arshinian, according to the abovementioned age data, must be correlated with the Cryogenian, and in no case can be Vendian.

The Bakeevo Formation in the western sections usually overlies the Uk formation with a stratigraphic unconformity. It is represented by feldspar-quartz sandstones with glauconite, layers of shales, locally with layers of tillite-like conglomerates and hematite ores. The thickness of the Formation is up to 140 m. The Bakeevo formation locally grades into the Tolparovo and Suirovo formations–a thick series of mostly terrigenous sediments with ordinary and tillite-like conglomerates, filling a deep trough, 18 km wide and 1,000 m deep (probably, on our opinion, a buried fjord, Gorozhanin 1988). The glaciation can be correlated with the Gaskiers or Marinoan glaciation (cf. Chumakov 2011).

The Uryuk Formation is represented predominantly by felspar-quartz and arkosic sandstones with layers of gravelstones and conglomerates, up to 300 m thick.

Higher up section, the character of the section changes. The Basu and Zigan Formations, divided by polymictic conglomerates of the Kukkarauk Formation, acquire a distinct polymictic character themselves. These three upper units have a total thickness of 1,500 m and belong to molasse of Timanides and have specific mineralogical and isotopic characteristics, discriminating them from the underlying Proterozoic sediments (Willner et al. 2001, 2003). The tuffs in the lower part of the Zigan formation are dated by the U-Pb method (zircons) at 548,2±7,6 Ma (Grazhdankin et al. 2011).

The age of the upper part of Vendian section of the Southern Urals is a matter of debate: occurrences of Cambrian fauna had been reported recently in the Kukkarauk Formation (Kuznetsov and Shatsillo 2011). But in fact, these occurences are very small (1 mm or hundreds of microns) and consist of shell splinters, reported as fragments of Obolidae (oral communication of L.E.Popov). Taking into account that the first fragments of a skeletal fauna appear in the Nemakit-Daldyn horizon at ca. 550 Ma and the Vendian/Cambrian boundary is established at 542 Ma, one may suppose that the fragments belong to some representatives of the Nemakit-Daldyn biota in the uppermost Vendian. One cannot speak about Cambrian fauna with certainty until the paleontological conclusions are fully detailed in a publication

In general, the above-described Riphean and Vendian (ca. Meso- and Neoproterozoic) section, which is more than 15 km thick, is one of the best in the world, and now has good isotopic age markers, which is of a great importance for characterizing the stratigraphy of the Proterozoic. In the International Stratigraphic Chart (Gradstein et al. 2004), approved by the International Commission on Stratigraphy, the Mesoproterozoic (1,000 to 1,600 Ma), is divided into Systems/Periods of equal length: Stenian (1,000–1,200 Ma), Ecstasian (1,200–1,400 Ma) and Calymmian (1,400–1,600 Ma). Such a division does not follow the general principles of stratigraphy, whereby boundaries should be defined on the basis of real stratigraphic sections and events to create a "natural" stratigraphic scale (e.g. Bleeker 2004). The described section has a favourable characteristics for becoming one of the global standards.

The sections of the Proterozoic of other parts of Externides–in the Middle Urals, Timan, Kanin peninsula and northeastern coast of Scandinavia belong mostly to the Upper Riphean and Vendian; their bottom parts are not exposed. In Timan and other areas to the north-west, a facial transition from shallow-marine to deep-water sediments is detected, indicating the probable relics of a passive margin, grading into the paleooceanic area of Internides. Neither Arshinian, nor Vendian are exposed here.

Metamorphism of the Riphean and Vendian rocks of Externides is very low in their outer, marginal parts (it is mostly sub-greenschist) and increases, approaching the border of Internides, acquiring a maximal intensity in the Beloretsk metamorphic dome at the east of BMA, where isogrades of omphacite and garnet are traced in the core of the dome. The age of the metamorphism is mostly Vendian (Glasmacher et al. 2001; Alexeiev et al. 2006). Further to the north, sub-greenschist to greenschist metamorphism is also predominant, with subordinate development of epidote-amphibolite, amphibolite and (very locally) glaucophane-schist metamorphism (Getsen 1987; Beckholmen and Glodny 2004).

The Timanides are bordered by a foredeep, filled with the Upper Vendian molasse and traced along the whole orogen.

Internides

They have a sharp boundary with the Externides on aeromagnetic maps: the anomalies in the Internides become more intense and linear (Fig. 5). One can suppose a suture zone here, though there is no direct proof for it. The Internides are exposed in the Central Uralian zone of the Cis-Polar and Polar Urals, and to a lesser extent–in the uplift zones of Pay-Khoy and Novaya Zemlya. The combination of the map of magnetic anomalies and data on geology of exposed areas permits to decipher the structure of the part of Timanides, submerged

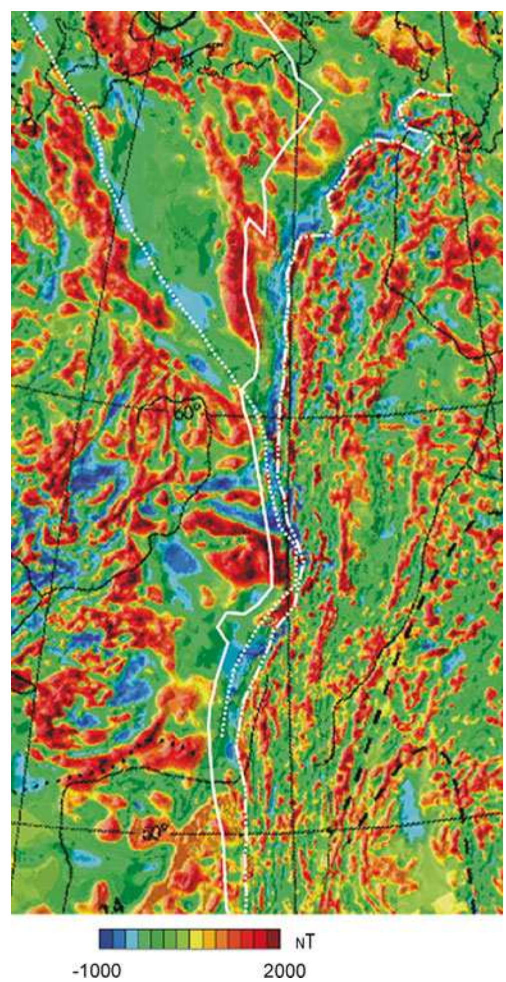


Fig. 5 Magnetic anomalies of the Uralo-Timanian mobile belt. *Solid line*—the western boundary of the Uralide folding, *dotted and dashed line*—Main Uralian Fault; *dotted line*—western boundary of Timanide folding

under the Paleozoic platform cover in the Timano-Pechora basin. In general, the positive anomalies correspond to synclinoria, with thick volcanic formations, and the negative ones to anticlinoria, with thermal domes and granites (cf. Figs. 2 and 5). This correlation is supported by deep boreholes, drilled in the Timano-Pechora basin that reach crystalline basement.

The main feature of Internides is the presence of ophiolites and supra-subductional complexes (Belyakova et al. 2008; Scarrow et al. 2001; Corfu et al. 2010; Puchkov 2005). Their primary distribution could have been wide, just because of a wide presence of these supra-subductional volcanics, so we suggested (Puchkov 2005) that they are associated with so-called Pechora Ocean which existed during the Upper Riphean. But now the Internides are composed mostly of sialic blocks, separated by narrow suture zones with poorly preserved Neoproterozoic ophiolites (e.g. Scarrow et al. 2001). In the cores of these dome-like blocks there are metamorphic complexes with a complicated history, consistent with their interpretation as retrograde metamorphosed fragments of Archean?-Paleoproterozoic basement (Pystina and Pystin 2002). Such uplifts as Khobeiz, Kharbey, Marunkeu (Fig. 2) may be interpreted as microcontinents, accreted to the craton of Baltica by Vendian time, during the Timanian collision and orogeny and bounded by sutures and crypto-sutures.

Paleozoic-Early Mesozoic structural stage and the Uralides

Structures of the Uralides

Uralides are not Variscides

The term "Uralides" was introduced very long ago, as far back as E. Suess' works, who coined many other useful terms. But for some reason, Uralides did not appear in the Stille's canon; probably because the comparative timing of orogenies in European Variscides and Uralides was not known, and the Uralides were treated as a Hercynian (Variscan) orogen for a long time, including in some very recent publications (Atlas of geological maps of Central Asia 2008). But in fact, orogeny in Western and Central Europe came to an end before the beginning of the Permian, while in the Uralides the orogeny lasted during the entire Permian time, and even resumed for a short time at the end of the Lower Jurassic (Puchkov 2003, 2009b). The orogenic processes in the Taymyr and Tyan-Shan behaved almost in the same way, which makes Uralides an inter-regional structure (Puchkov 2009b)

The stage under discussion started after a short transitional Cambrian stage of almost total non–sedimentation and erosion. There is a possibility that the Upper Vendian grades into the Lowermost Cambrian, but this remains to be proven. The Lower Cambrian reefal limestones are present only in two small areas of the Southern Urals: the Northern part of the Sakmara allochthon and a section in the Sanarka river (Western and East-Uralian zones, correspondingly). But in both cases, the blocks of limestones with Cambrian algae and archaeocyathids represent olistoliths in olistostromes of a younger age (Ordovician and Devonian). The latter contain also an olistolith of basalt and cherty siltstone with Upper Cambrian conodonts (Puchkov 2000). The presence of Upper Cambrian terrigenous sediments as a basal part of a thick Ordovician succession of terrigenous rocks is suggested, but is still a matter of discussion, taking into consideration the uncertainty of Cambrian/Ordovician boundary and stratigraphic significance of Inarticulata brachiopods, found in the Southern and Polar Urals.

Geodynamically, the Paleozoic-Early Mesozoic structural development of the Urals can be interpreted as a full Wilson cycle, from epicontinental rifting in the Ordovician till collisions in the Late Paleozoic and Mesozoic. As a result, the Uralian orogen was formed–a linear zonal bi-vergent folded structure with a significant participation of continental margin, oceanic and suprasubductional formations. The zones of the Urals are the most prominent and unobliterated features of its structure, and therefore serve as a reference frame for the Urals as a whole, from the Archean to the Quaternary.

Structural zones of the Uralides

The Urals consists of north–south striking structural zones, giving it a general appearance of an approximately linear foldbelt, in contrast to the more strongly mosaic and/or more oroclinal chains of the European Variscides, Alps, Himalayas and Kazakhstanides (Franke 2000; Khain 2001; Agard and Lemoine 2005). The Urals is divided into six structural zones (Fig. 6; Puchkov 1997, 2000, 2009a, b), which are (from west to east) as follows:

- A– the Preuralian foredeep, which inherited the western part of a larger and long-lived orogenic basin and overlies the older basin of a passive margin of the Uralian palaeocean. It is filled mostly by Permian preflysch (deep-water condensed sediments), flysch and molasse.
- B– the West Uralian megazone, predominantly consisting of Paleozoic shelf and deep-water passive margin sediments. This zone was affected by intense fold-and-thrust deformation, and includes klippe, containing easterly-derived ophiolites and arc volcanics (Puchkov 2010a).
- C– the Central Uralian megazone, where the Precambrian (predominantly Meso- and Neoproterozoic) crystalline basement of the Urals is exhumed. This basement is traced by geophysical data under the A, B and C megazones. Basically, these megazones were formed as a result of deformation of the continental margin of Baltica, although

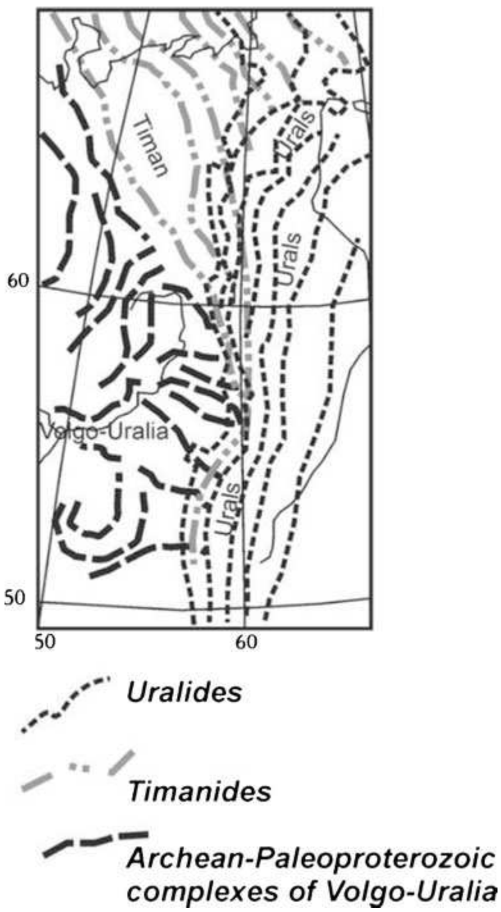


Fig. 6 Relationships between strikes of the Uralides, Timanides and Archean-Paleoproterozoic complexes

some allochthons, and partly the Ural-Tau antiform (UTA) were derived from more eastern oceanic structures.

D– the Tagilo–Magnitogorskian megazone, bordered to the west by serpentinitic mélange within the Main Uralian Fault (MUF) zone and to the east by the East Magnitogorsk Fault (EMF) and Serov-Mauk Fault zones (SMF). This megazone predominantly consists of Ordovician–Lower Carboniferous complexes of oceanic crust and ensimatic island arc, including the Platinum-bearing Belt (PBB) of layered basic-ultramafic massifs, overlain by shelf carbonate and rift-related volcanic rocks.

E– the East Uralian zone, bordered to the west by the East Magnitogorskian mélange zone (EMF) and to the east by the Kartaly (Troitsk) Fault (KRF) (Fig. 7). This zone comprises Proterozoic gneisses and schists overlain by weakly metamorphosed Ordovician to Devonian sedimentary clastic and carbonate strata and by tectonically emplaced sheets of Paleozoic (Ordovician–Lower Carboniferous) oceanic

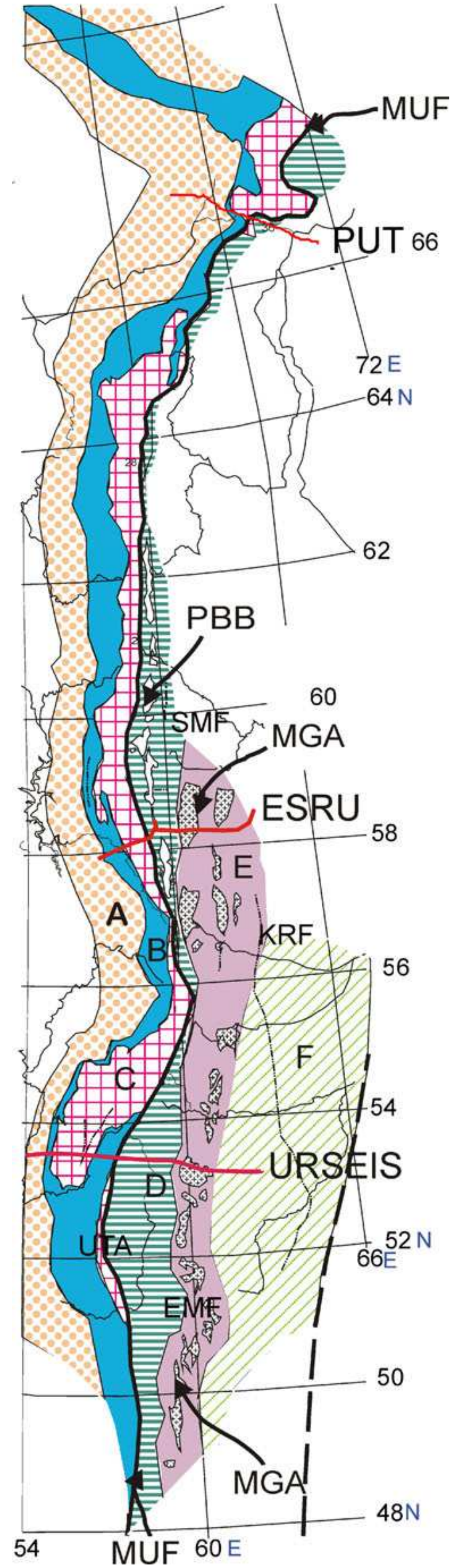


Fig. 7 Structural zones of the Uralides. Explanations of letter indexes and abbreviations–are in the text. *Red lines* show the most important seismic profiles across the Urals (URSEIS-95, ESRU-SB 93-95, and PUT–Polar Urals Transect, completed by 2010)

and island arc complexes. The East Uralian Zone is intruded by voluminous Late Paleozoic granite bodies of the Main Granitic Axis (MGA) of the Urals.

F– the Transuralian zone, the easternmost zone of the Urals has probably an accretionary nature. It contains pre-Carboniferous complexes which preserve a variety of tectonic settings, including Proterozoic(?) blocks of gneisses, crystalline schists and weakly metamorphosed sediments, Ordovician rift (coarse terrigenous and volcanic) and oceanic (ophiolite) deposits, Silurian island-arc complexes and Devonian deep-water deposits overlain unconformably by the Lower Carboniferous suprasubductional volcanogenic strata.

Zones D–F, together with MUF, are usually interpreted to preserve some vestiges of the palaeoceanic component of the Urals, relics of the Palaeouralian ocean (Peyve et al. 1977; Puchkov 2000; Puchkov 2010a). Although ophiolites with MORB signatures are poorly preserved in most orogens, the ophiolites in the eastern zones of the Uralian orogen are abundant, which is a rather anomalous feature, explained probably by a comparatively low degree of orogenic shortening.

All megazones are either exposed or are near the Earth's surface only in the Southern Urals. To the north, the easternmost zones are covered by the Mesozoic and Cenozoic strata of the West Siberian basin, and in the Northern and Polar areas only the Preuralian foredeep, West Uralian, Central Uralian and western part of the Tagil- Magnitogorskian megazone are exposed, while the thickness of the Mesozoic and Cenozoic cover becomes considerable immediately to the east.

The Wilson cycle of the Uralides

The development of the Uralides can be described as a succession of stages of a standard Wilson cycle:

(1) Rifting of Baltica continental crust, composed of a cratonic crystalline basement and the Neoproterozoic Timanide foldbelt (Cambrian?-Middle Ordovician);
(2) Formation of a continental passive margin, an oceanic basin and microcontinent(s) (Middle- Late Ordovician);
(3) Subduction of the oceanic crust, accompanied by arc generation (Late Ordovician- Early Devonian);
(4) Arc-continent collision (Late Devonian-Early Carboniferous), diachronous along-strike of the Uralides;
(5) Subduction of the last portions of oceanic crust and its complete disappearance (Lower Carboniferous-Bashkirian);
(6) Continent–continent (Laurussia–Kazakhstania and then Laurussia–Siberia) collisions (Bashkirian-Late Permian).
(7) The Uralo-Siberian superplume activity (rifting and flood basalt magmatism in the Early-Middle Triassic)
(8) the posthumous Middle Jurassic (Old Cimmerian) fold-thrust-wrench fault deformations.

Structural relationships between the Uralides and Timanides

The Uralides tectonic system obviously overprints the Timanides. In the South the strikes of Timanides and Uralides coincide more or less, but in the north they have truncating relationships, which is best seen on the map of magnetic anomalies which are easily correlated with relict structures of Timanides, reconstructed in the Central Uralian zone (Figs. 2 and 5). The greatest discrepancy between Timanides and underlying Archean-Paleoproterozoic structures is displayed most evidently in the southern area of their development (Fig. 6). This basic fact was also reflected in a presence of transversal structures in the western zones of Uralides explained as older structures showing through the younger.

An absence of a structural inheritance of general strikes of the foldbelts is best explained by the fact that the strikes of epicontinental rifts (as precursors of oceans) do not necessarily depend on strikes of structures of a crystalline basement of a continent. One can see this fact very well on opposite sides of the Middle Atlantic. Therefore we can suppose that the Riphean and Early Paleozoic rifts determined the future positions of the Timanian and Uralian orogens at the boundaries between weak oceanic and strong, undestroyed continental lithospheric blocks.

The development of the Uralides, deduced from study of corresponding igneous and sedimentary complexes

The structures preceding strongest orogenic deformations cannot be reconstructed only with usual methods of a structural analysis. The analysis of index igneous and sedimentary formations corresponding to certain geodynamic situations is much more helpful.

Epicontinental rift complexes of the Uralides

The main manifestation of rifting in the Early Paleozoic is the presence of rift facies at the western slope of the Urals, mostly in the Central Uralian zone. A detailed description of the Uralian Early Paleozoic rift facies in the western slope of the Urals is given in Puchkov (2002, and references therein). Typically, the rift facies consists of uppermost Cambrian (?)–Tremadocian to Middle Ordovician coarse terrigenous sediments (conglomerates, sandstones, siltstones of highly variable thickness, combined with interlayered subalkaline flows and tuffs). They overlie unconformably the crystalline basement and are overlain either by shelf or deepwater facies, reflecting the development of the eastern passive continental margin of Baltica. Rift facies of a comparable type are also present in the ensialic, microcontinental East Uralian zone, but their age is Middle Ordovician (Fig. 8, Kliuzhina 1985; Snachev et al. 2006), which suggests no direct ties between the Baltica (East European continent) and the microcontinent. The transition between rifting and ocean floor spreading took place in the Arenigian and led to the formation of the Uralian palaeocean. Figure 8 shows a localization of relic rift facies in the Urals, overlain on the Uralian zones. All the occurrences occupy rather restricted areas (always on a continental crust), and volcanic rocks in them do not form a LIP, so the continental margin, originated in a consequence of rifting process, belong to an avolcanic type.

Alternatively, some researchers (Scarrow et al. 2001; Samygin and Ruzhentsev 2003) maintain that the Palaeouralian Ocean was inherited from the Proterozoic time. However, the truncation of the Timanides against the younger Uralides is clear evidence against that idea. In addition, the lower age limit of the ophiolites attributed to the Palaeouralian ocean, as determined by more than 25- years of conodont studies is Arenig–Llandeilian, which can be correlated with Floian, Daipingian and Darriwilian of the modern global scale (Bergström et al. 2009) and is clearly younger than the Ediacaran–Tremadocian age that would be expected if there was uninterrupted ocean development from Proterozoic time onwards (Puchkov 2005; Borozdina et al. 2004, 2010; Borozdina 2006; Smirnov et al. 2006).

The ophiolites of the Urals as relics of a crust of the Palaeouralian ocean

Many detailed studies describe the Uralian ophiolites (e.g. Savelieva 1987; Savelieva and Nesbitt 1996; Melcher et al. 1999; Spadea et al. 2003; Savelieva et al. 2006a, b). Such huge ophiolite massifs as Syum-Keu, Ray-is, Voykar, Kraka, Khabarny, Kempirsay are known worldwide and well studied, but there are tens of smaller massifs also deserving attention (the most important among them are Shevchenko, Akkarga, Mindyak, Nurali, Kliuchevsk, Alapayevsk, Olysya-Musyur). In addition, there are many zones of serpentinitic mélanges, tracing major suture zones and thrust surfaces of allochthons (Kuvandyk, Sakmaro-Voznesensk, East Magnitogorsk, Serov-Mauk and others. As a whole, the Urals gives the impression of being a unique, most significant ophiolite belt (Fig. 9).

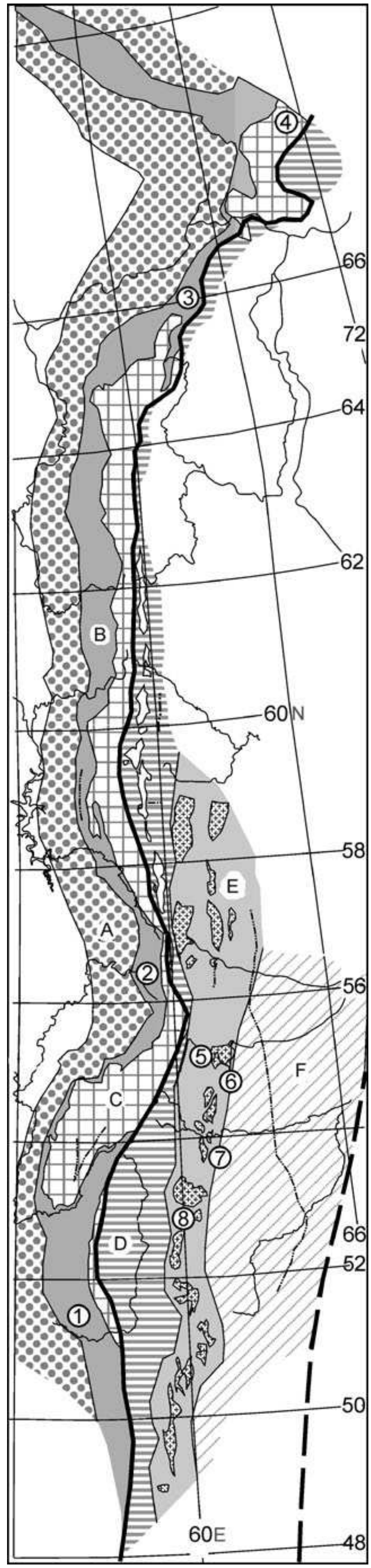
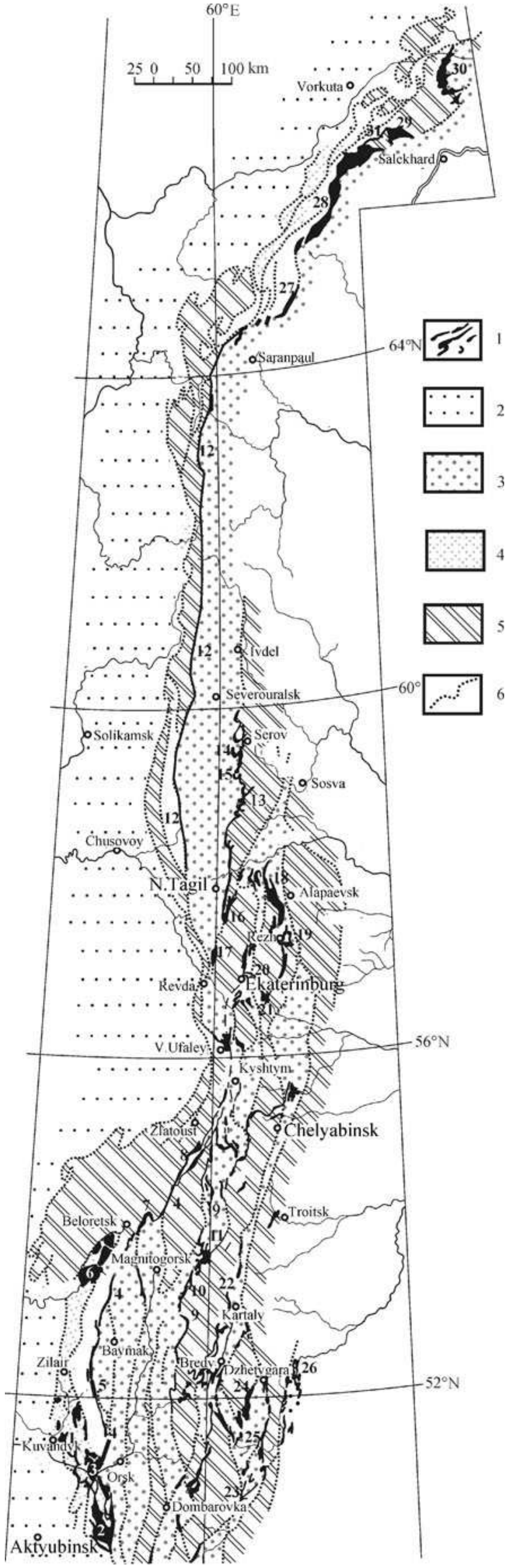


Fig. 8 Position of Ordovician rift complexes with respect to the structural zones of the Urals (see Fig. 7, according to Puchkov 2005, modified). *Numbers in circles*—areas of development of Ordovician rift complexes with volcanism of a matching petrochemistry (see the explanations in the text)



◀**Fig. 9** Localization of Paleozoic ophiolites in the Urals: traces of the Paleouralian ocean. Based on a scheme in Savelieva et al. (2006a) and earlier publications, strongly modified

The 'ideal' section of the Uralian ophiolites (from top to bottom) consists of:

(1) Tholeiitic basalts (mostly pillow lavas) with layers of pelagic sediments (typically jaspers containing relics of half-dissolved radiolarians). The age of these basalts, constrained by many occurrences of conodonts, is never older than Arenigian–Llandeilian (as discussed above). In the Tagil zone, ophiolitic complexes are overlain by Upper Ordovician island-arc formations, whereas in the Magnitogorsk zone, the condensed oceanic sediments overlie Ordovician–Llandoverian basalts, and persist until the onset of island arc magmatism in the Early Devonian;
(2) Dyke-in-dyke sheeted complexes;
(3) Alpine-type gabbro;
(4) Banded dunite-wehrlite-clinopyroxenite complexes, interpreted as a fossil MOHO boundary; and
(5) Peridotite complexes, represented by lherzolites, harzburgites and dunites in different proportions and combinations.

However, continuous sections where all components are present in a normal succession are rare, if present at all. Usually they are recognized after some reconstruction because of the presence of subsequent strong deformations of accretionary and orogenic nature. On the other hand, not all the ophiolitic complexes display this simple sequence. For example, Ishkinino, Ivanovka and Dergamysh Ni–Co-rich pyrite deposits in the mélange of MUF zone of the Southern Urals (Sakmaro-Voznesensk mélange zone) are attributed to ocean floor black smokers which overlie and partly penetrate peridotitic host-rock; they are devoid of several 'standard' members of the ophiolite section: peridotites are often overlain immediately by basalts and/or pyritic sandstones.

Although the paleontological data (conodonts in cherts among basalts) support the idea of a Lower Paleozoic age for the Uralian ophiolites, the isotopic data, especially the age determinations of zircons in peridotites, gabbro and even in basalts are more ambiguous: as often as not they give Proterozoic ages (Gurskaya and Smelova 2003; Savelieva et al. 2006b; Tessalina 2005; Batanova et al. 2007; Krasnobaev et al. 2008). Two contrasting explanations have been proposed. According to Tessalina (2005), the ultramafic complexes are Neoproterozoic ophiolites and represent relics of the oceanic crust developed during Timanide history. Alternatively, Puchkov (2006b) suggested that the Neoproterozoic dates in the Paleozoic ophiolites reflect a relict signature preserved in the mantle part of the younger ophiolites (peridotites and partly ex-eclogitic mantle gabbro), notwithstanding the overprinting during subsequent processes of ophiolite formation. The additional alternative is a contamination of ophiolites by older zircons during lithosphere recycling. The possibility of input and preservation of ancient zircons in mantle and their contamination of younger MOR and island arc volcanics have been underlined in many publications (Bea et al. 2001; Bortnikov et al. 2005; Puchkov et al. 2006). Some other isotopic ratios may be also relic, because the mantle is much more ancient than the last Wilson cycle that it experienced.

Sedimentary and igneous complexes of the continental passive margin

Following rifting, and parallel to the onset of ocean floor spreading, a passive margin of the continent forms. The process was described in detail by Puchkov (2002). Typically, the succession starts with rift facies–Uppermost Cambrian–Lower Ordovician coarse-grained terrigenous deposits, in some cases accompanied by minor volcanic rocks (see the "Epicontinental rift complexes of the Uralides"). The margin can be classified as a non-volcanic type (in a sense of Melankholina 2008). For many modern passive margins it is not unusual that initial rifts are filled with evaporites. For many years, we had no evidence for such phenomenon in the Urals. But recently Ordovician salts were discovered by drillhole 1-Vostochno-Lemvinskaya in the Polar Urals (Gudelman et al. 2009).

Two facies were established after the rifting phase in the passive margin structure–an internal (shelf) and an outer (continental slope, grading to continental rise) (Fig. 10, Table 2). Usually, shelf sediments are developed as a continuous, para-autochthonous sedimentary cover of an elevated thickness, represented by shallow-water carbonates (limestones, dolomitic limestones, dolomites) and terrigenous sediments with west-derived oligomictic, quartz sandstones. Regressions are marked by barrier reefs at the outer margin of the shelf zone, while the transgressions favour formation of deepwater, starved basins with condensed facies of marls and oil shales (called 'domanik' in the Russian literature), surrounded by reefs and bioherms.

The outer, continental slope and rise (bathyal and abyssal) sections are developed mostly in separate allochthons, preserved from erosion (Sakmara, Kraka, Bardym, Upper Pechora, Lemva, Baydarata and Kara allochthons). Their sections consist of thick westerly-derived quartz sandstones and siltstones of flysch-like type and thin, condensed units of shales, cherts and minor limestones–depending on the abundance of a terrigenous provenance. Fauna are mostly pelagic: radiolarians, conodonts, graptolites, foraminifers and rare goniatites. The absence of limestones indicates at a transition to abyssal conditions which are not favourable for deposition of carbonates. These facies change upwards to polymictic, flysch deposits, accompanied by olistostromes, signifying a sharp change of provenance that is connected with the start of orogenesis (Puchkov 2002; Willner et al. 2002, 2004). This change happens in the Southern Urals by the beginning of the Famennian, but in the Polar Urals it takes place only in the Early Visean.

Notwithstanding the allochthonous character of the deep-water facies, the amplitude of their thrusting was moderate, and they follow the Main Uralian Fault, controlling the ophiolites and marking the continent/ocean ancient boundary. Figure 10 shows, that the boundary had promontories (Ufimian and Bolshesemelian), divided by recesses, that influenced the pattern of the later folded structures.

Apart from the early volcanic phase connected with rifting, two volcanic stages were recorded in the passive margin (Puchkov 2012). One, Ordovician/Silurian, discovered quite recently, is developed in the BMA of the Southern Urals and probably connected with one of the stages of development of the Vishnevegorsk alkaline complex; it is represented by trachybasaltic dykes. Another, represented by dolerite dykes, has a Devonian age. Both can be attributed to mantle plume activity (at least, Devonian volcanism was very abundant in the East European platform).

Subduction complexes

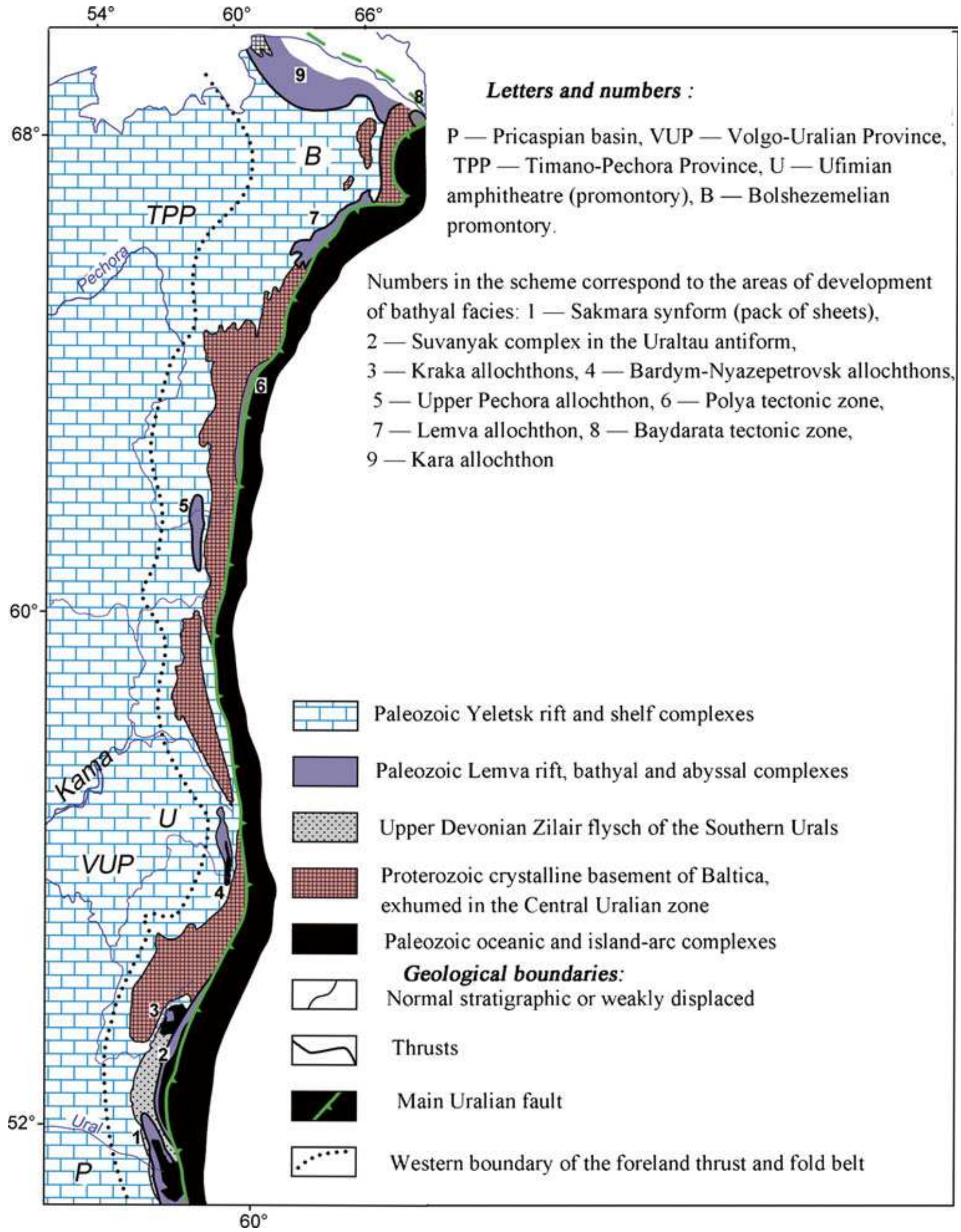
Not only ophiolites, but also subductional complexes (mostly island arc), are exceptionally well developed. Four main subduction zones of different age and localization can be reconstructed in the history of the Paleozoic Urals: Sakmaro-Vosnesensk, Tagil, Magnitogorsk and Valerianovka (Fig. 11):

The Ordovician arc and continental margin subduction complexes in the Sakmara allochthon and MUF mélange of the Southern Urals referred to as hypothetical Guberlya arc (Zonenshain et al. 1990) were reconsidered recently on the base of an extensive study of stratigraphy and geochemistry of highly tectonized Ordovician volcanic rocks (Ryazantsev 2012). Some time is still needed to evaluate these results, but the existence of subduction between the Late Early Odovician and Late Ordovician in this zone seems very probable.

The Late Ordovician–Early Devonian Tagil arc complexes are developed in the Middle, Northern, Cis-Polar and Polar parts of the Tagilo–Magnitogorskian zone (Fig. 11-I). The best island-arc sections, well constrained by conodonts, are preserved in the Middle Urals. However, the lowermost Paleozoic volcanics, including dyke-in-dyke complexes are reported to have rather contradictory petrochemical characteristics (Afar-type basalts, MORB type, along with calc-alkaline basalts), probably due to the imbricated tectonic structure of MUF zone (Smirnov et al. 2006; Petrov 2007; Puchkov 2010a).

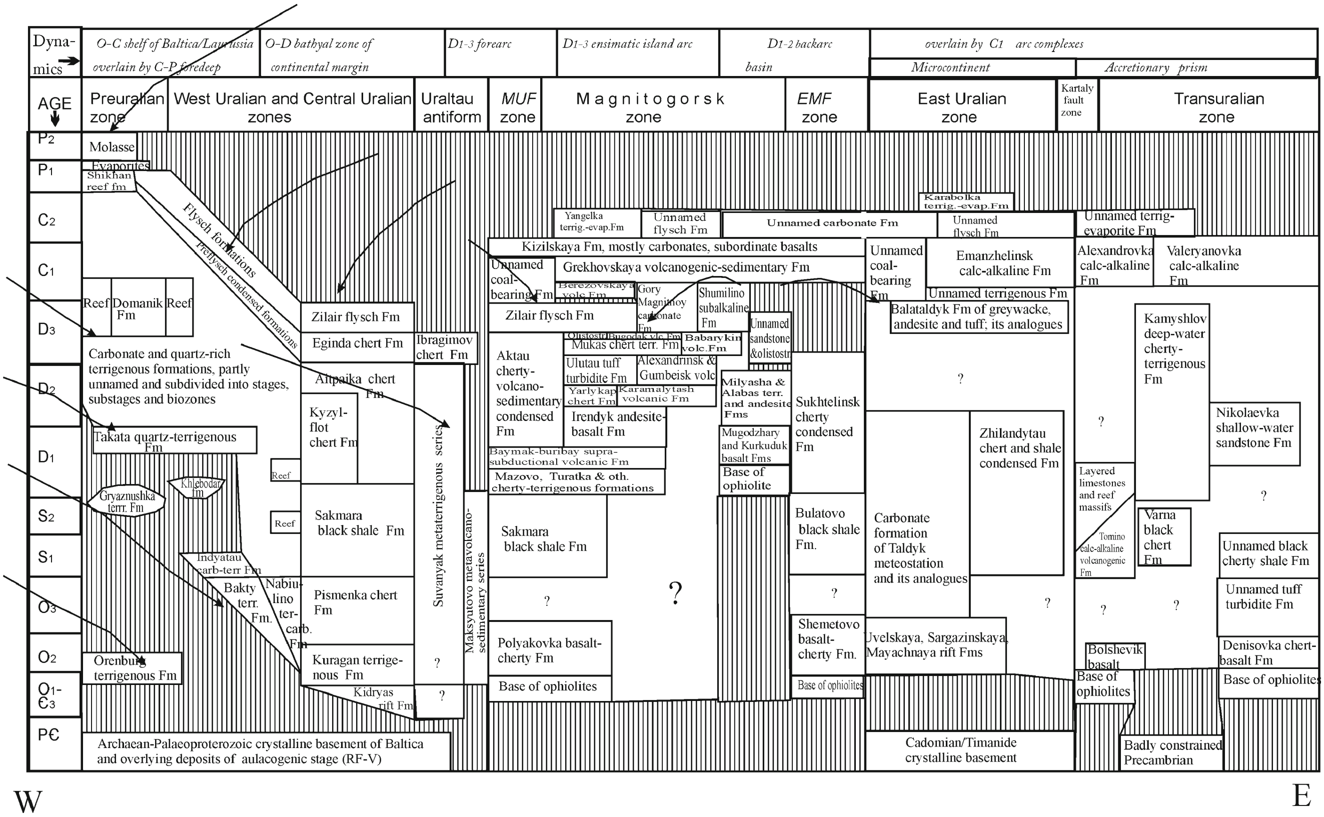
According to stratigraphic and petrochemical studies (Narkissova 2005; Borozdina 2006), the oldest ensimatic

Fig. 10 Major structural elements and complexes of Baltica Paleozoic passive margin that was involved in the Urals (Puchkov 2002, modified)



island-arc succession is predominately represented by the basaltic (O_3), basalt-plagiorhyolitic (O_3) and basalt–andesite–plagiodacite (S_1) volcanic associations. The latter two have calc-alkaline affinities (Narkissova 2005). The overlying Silurian association (S_1ll_2–S_2w_1) is represented by flysch consisting of interbedded black cherty siltstones, tuffites and tuffaceous sandstones (arc slope deposits), overlain by andesites, dacites, basalts and very abundant tuffs. In the Wenlockian, volcanic rocks are developed simultaneously with reefal limestones (All-Russian Stratigraphic Committee 1993). The biohermal deposits persisted until the Pridolian as an unstable, narrow carbonate shelf on the perimeter of the island arc. These Silurian facies are substituted laterally by a volcanic association (S_1ll_3–S_2ld_1), represented by basalts, andesites and tuffs with rare layers of cherty siltstones. After the Late Ludlovian, the marine conditions partly changed to continental conditions: the Pridolian association is represented by predominant coarse-grained red-coloured polymictic terrigenous-volcanogenic deposits with fragments of the older rocks, such as volcanics, with sublakalic and alkalic basalts being predominant. The island arc succession is terminated by shoshonitic mafic to intermediate volcanic rocks (Narkissova 2005) and minor flysch-like volcaniclastic deposits (S_2pr–D_1lh). The volcanism of the Tagil arc evolved from a uniformly tholeiitic affinity to a differentiated calc-alkaline and then to subalkalic shoshonitic affinity, suggesting deeper levels of partial melting in mantle with

Table 2 Tectonostratigraphic chart of the Paleozoic formations of the Southern Urals. Arrows show the provenance of terrigenous material. Puchkov (2009a, b), with minor changes



time. Geochemically, the volcanics are thought to be typical of ensimatic island arcs (Narkissova 2005; Borozdina 2006), though there are different opinions (Fershtater 2012).

The specific feature of the southern part of the Tagil subduction zone is the so-called Platinum-bearing Belt (PBB on Fig. 7), attributed to Uralo-Alaskan type concentric-zonal massifs. The belt consists of dunites, clinopyroxenites, gabbro and plagiogranites; mafic rocks comprise up to 80 % of the belt. Disseminated platinum is hosted by dunites, and industrial deposits are represented mostly by Quaternary or reworked Meso-Cenozoic placers, though platinum was produced also from some small mines. The geodynamic setting of the belt is controversial, but the most probable is a supra-subduction zone setting (Ivanov and Shmelev 1996; Ivanov et al. 2006; Fershtater 2012). The age of the belt, determined by several methods is 420–430 Ma, and the similarity of its trace and rare earth element patterns with island-arc tholeiites supports the supra-subduction zone model. However, the origin of the belt can not be explained as a result of a single process, because it demonstrates many anomalous (Precambrian) age determinations and consists of several rock associations, of which the most important are two: (1) the earlier, dunite-clinopyroxenite-gabbro and (2) the later, gabbro, intruding the first. In the author's opinion (Puchkov 2010a), the first complex was formed as a result of rift magmatism and mantle exhumation, accompanying the formation of the subduction zone in the Ordovician; at that, the initial, differentiating melt had a picrobasalt composition. The second complex, represented mostly by bi-pyroxene gabbro-norites, was formed as a result of partial melting of the supra-subduction mantle wedge in the Silurian.

Ordovician–Early Devonian island-arc volcanism was followed by the development of a relatively stable carbonate shelf which caps the western part of the island arc complexes and contains bauxite deposits formed owing to a favourable combination of a tropic climate, intense karst development and weathering of volcanic rocks. After that time the Tagil arc lost its individuality and by the Emsian, it became a "dead" terrane that accreted to the new-formed Magnitogorsk island arc, (see below). The location of the subduction zone changed and the region became characterized by presence of two subzones: the western, Petropavlovsk and the eastern, Turyinsk sub-zones. In the Petropavlovsk sub-zone, Lower to Middle Devonian shallow-water limestones and bauxite were followed in the Late Devonian by deepwater cherty shales and polymictic terrigenous sediments. In the eastern, Turyinsk sub-zone, sedimentary strata (shallow-water limestones, shales and cherty shales) are interlayered with andesites, basalts, tuffs and volcanogenic sandstones (All-Russian Stratigraphic Committee 1993). Yazeva and Bochkarev

(1993) pointed out that these thick (up to 4–5 km) Devonian volcanic complexes occur with comagmatic intrusions.

It is important to know when the arc was sutured against the continent of Laurussia. Although the Tagil arc lost its identity as an active structure in the Early Devonian, it did not collide with the continent until at least the end of the Devonian or beginning of Carboniferous. The evidences for this is as follows. In the western slope of the Middle Urals, against the southern end of the Tagil zone, there is the Bardym allochthon, composed of two units. The upper unit is represented by oceanic to island–arc volcano-sedimentary complexes of Ordovician to Lower Devonian age and a Pd–Pt-enriched gabbro-ultramafic Suroyam massif of the lowermost Devonian age (Zhilin and Puchkov 2009). All the complexes demonstrate a closest resemblance to those of the Tagil zone. The lower unit comprises the Ordovician to Frasnian terrigenous and cherty bathyal sediments, demonstrating calm, uninterrupted conditions of sedimentation at a deeper part of the continental margin during the entire duration of subduction zone to the east of it (Puchkov 2002). Further to the North, in the Salatim zone, close to the Main Uralian Fault, a more fragmentary section of Ordovician to Frasnian bathyal sediments is reconstructed, also revealing no sedimentary record of collision in the Lower or Middle Devonian (Petrov 2007). Such indications are also not present in the Upper Pechora and Lemva bathyal allochthons of the Cis-Polar and Polar Urals, refuting the possibility of the arc–continent collision earlier than Carboniferous (Puchkov 2002).

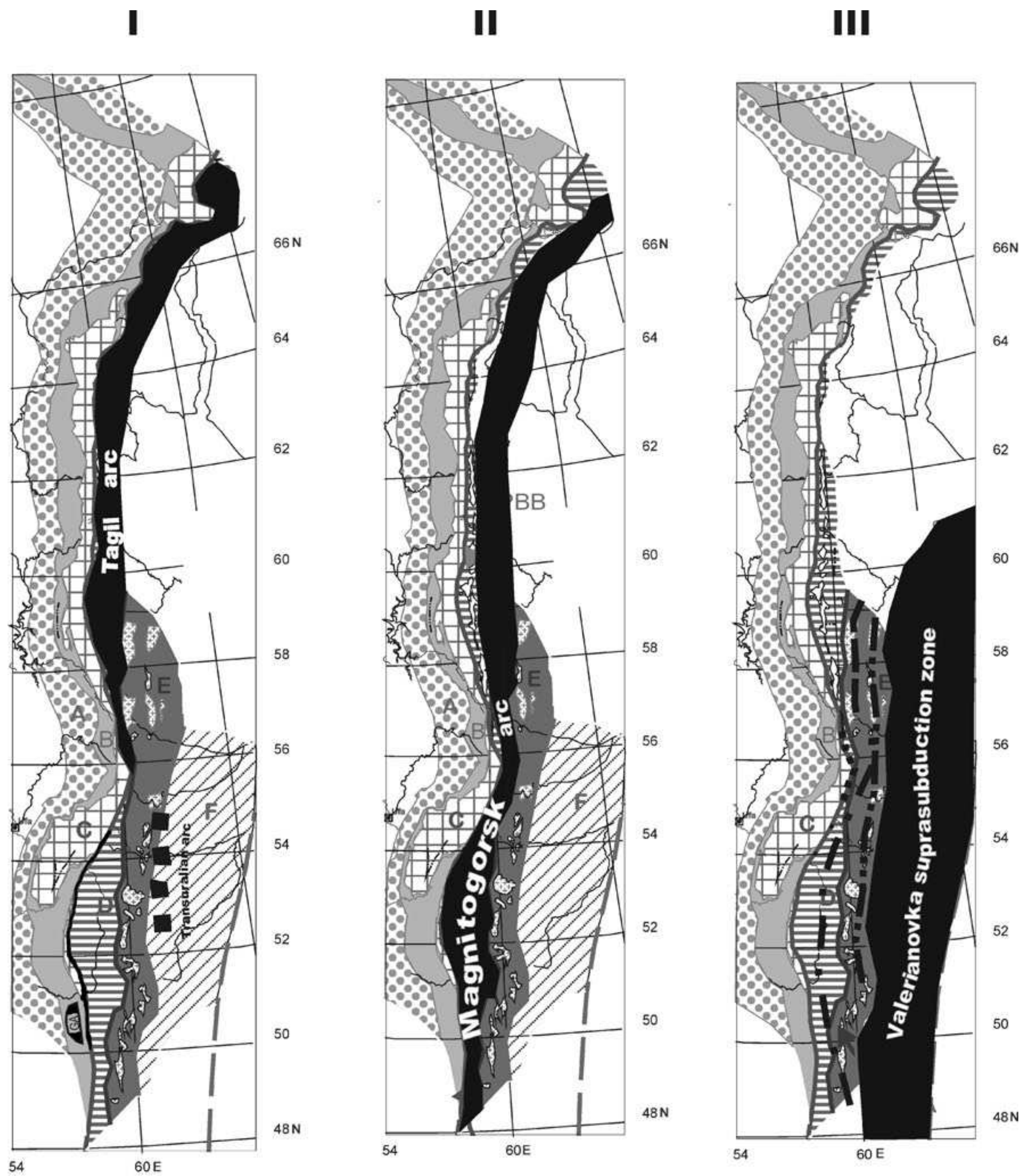
The Devonian Magnitigorsk arc (Fig. 11-II) started to develop almost simultaneously with the end of the cycle of active development of the Tagil arc. The location of the Turinsk zone of Tagil arc along the extension of the Magnitogorsk arc suggests that the Magnitogorsk subduction zone was partly inherited from the Tagil zone. The Middle–Upper Devonian calc-alkaline complexes can be traced northward to the Polar Urals. But in the Southern Urals, island arc development was quite new; it was preceded by a long period of deposition of deepwater oceanic cherts and carbonaceous cherty shales accompanied by MOR basalts in the Llandoverian (Table 2). Most of the Silurian is represented by 300 m of distal, condensed cherty shales. They are considered to represent the sedimentary cover of the ophiolites. Non-volcanic sections of the Lower Devonian (Lochkovian–Lowermost Emsian) are represented by either deep-water terrigenous chert, argillaceous cherty sediments of the Masovo, Turatka, Ishkinino and other formations or bioherm limestones, often reworked into olistostromes and coarse-grained sediments (Artiushkova and Maslov 2003). The appearance of olistostromes may herald the start of subduction (at its most early, non-volcanic stage). The first volcanic series appeared in Emsian.

Volcanic complexes of the Magnitogorsk arc in the Southern Urals are well described (Brown et al. 2001, 2006; Kosarev et al. 2005, 2006, and references therein). The volcanic succession is represented by a characteristic alternation of tholeiitic basalts, bimodal basalt-rhyolite and successively differentiated basalt-andesite-dacite-rhyolite series, grading upwards to subalkaline shoshonitic series; this is a similar pattern to that observed in the older, Tagil arc.

The succession (Table 2, Magnitogorsk zone) starts with a tholeiite-boninite complex, overlain by a bimodal rhyolite-basalt unit. In turn, it is overlain by a basalt-andesite-dacite-rhyolite series. The whole succession is usually labelled as the Baimak-Buribay Formation of the Emsian age. Next upsection, the Irendyk Formation, of Uppermost Emsian–Lower Eifelian age is represented by an andesite-basalt series, with a combination of tuffaceous successions and lava flows. It changes upwards into the contrastly differentiated rhyolite-basalt Karamalytash Formation (Upper Eifelian), that forms separate volcanic structures and grades into a contemporaneous condensed Jarlykapovo Jasper Formation. After a deposition of a widely developed Bugulygyr marker jasper horizon, volcanic activity resumed and differentiated basalt-andesite-dacite-rhyolite volcanics of Givetian–Lower Frasnian age are formed. These consist of the Ulutau Formation with thick volcaniclastic turbidites of an arc slope setting, and Alexandrinsk and Gumbeisk Formations belonging to the volcanic range of the arc. Upper Frasnian volcanism is represented by a basalt-andesite type substituted laterally by cherty flysch, condensed cherts and olistostrome. In the Famennian–Tournaisian, a Shumilino shoshonite-absarokite formation and its analogues are locally developed. It is facially substituted by contemporaneous shallow-water limestones of the Magnitogorsk Formation, corresponding to a narrow shelf of the dying island arc. Simultaneously, an accretionary prism of the arc abutts the margin of the continent in the west, growing as a non-volcanic cordilliera, and a large-volume greywacke flysch of the Zilair Formation (Famennian-Lower Tournaisian) starts to form. The arc-continent collision starts and this means the end of active arc development and subduction.

In the Early Carboniferous, slab breakoff probably took place, so the character of volcanism changed again. The volcanics of Berezovskaya and Grekhovskaya formations of the Tournaisian-Visean age, described in detail by Salikhov (1997), are dominated by tholeiitic basalt in the west and by more widely developed subalkaline bimodal basalt-rhyolite in the east, and are combined with limestones, becoming increasingly abundant upsection. They are accompanied by a chain of coeval (335–315 Ma, Bea et al. 2002) bimodal gabbro-granitoid intrusions (Magnitogorsk-type plutons). Both volcanic and intrusive bimodal series, according to their field relationships, mineralogy and geochemistry, suggest an extensional or passive within-plate non-arc origin (Fershtater et al. 2006; Kosarev et al. 2006).

There are many parallels with the development of the Tagil arc, including the alkaline trend towards the upper part



Areas of suprasubductional volcanism

Suprasubductional granodiorite–tonalite intrusions at the Devonian/Tournaisian boundary

Mid-Vizean Turgoyak–Sukhtelinsky complex of granitoids

Serpukhovian Verkhisetsk group of tonalite–granodiorite untrusions

The axial zone of volcano-plutonic Tournaisian–Lower Bashkirian bi-modal rift magmatic complex

◀ **Fig. 11** The major subduction zones in the Paleozoic history of the Urals: *I*–Ordovician to Pragian, *II*–Emsian to Famennian, *III*–Late Famennian–Early Bashkirian (Puchkov 2009a, b, slightly modified). GA–position of Ordovician Guberlya subduction indicators on the scheme I

of the succession. Trace element abundances for contemporaneous volcanic rocks indicate at an eastward dip of the subduction zone. To the west and east of the main volcanic body of the arc, mostly in mélanges of the MUF and EMF zones and associated allochthons, condensed cherty-terrigenous series contemporaneous with the arc were developed, corresponding to the forearc and backarc basins (Table 2). Therefore the island arcs of the Urals demonstrate a unique state of preservation, unusual for the Paleozoic, and even for Mesozoic orogens elsewhere. The explanation was a specific (comparatively weak) development of Uralide orogenic compression between rigid Laurussia (former Baltica) and a softer Kazakhstania continents.

The arc-continent collision started with the end of the active stage of the Magnitogorsk arc in the Southern Urals when it collided with the passive margin of the Laurussia continent (former Baltica) (Brown and Puchkov 2004; Brown et al. 2006; Puchkov 2009a, 2010a and references therein).

One of the specific features accompanying this collision process was development of a linear belt of high-pressure-low temperature (HP-LT) metamorphic complexes of eclogite–glaucophane–schist and glaucophane–schist types, with some spots of granulitic garnet pyroxenite occurrences (Fig. 12).

The complexes are traced as a unique 2,000-km-long belt along the MUF, which of course is not coincidental (Udovkina 1971; Dobretsov 1974; Lennykh 1977; Karsten and Puchkov 1990; Petrov 2007; Puchkov 2010b, and many references therein). Owing to good stratigraphic control and an extensive isotope dating done mostly in the last decade, the information on the exhumation age of these metamorphic complexes gives good support to geological conclusions concerning the timing of the arc–continent collision in the Urals.

As it was stated above, the trace element concentrations for volcanic rocks of the Magnitogorsk arc indicate a dip of the subduction zone from Laurussia. Generally speaking, it is a necessary condition for a continent and island arc to collide, at the very moment when the strip of remaining oceanic lithosphere between them is subducted (Puchkov 2010b). This happened for the Magnitogorsk island arc and Laurussia continent by the end of Frasnian (Fig. 13).

The model at the Fig. 13 shows that the collision was accompanied by the following events: (1) scraping up of the deep-water sediments of the continental passive margin by the rigid edge (backstop) of the arc and formation of an accretionary prism, (2) jamming of the subduction zone followed by a jump in the location of the subduction zone, (3) slab break-off and opening of a slab gap permitting the deeper, more fertile and hotter mantle to produce subalkaline, non–subduction volcanics (including K–Na subalkaline hawaiite–mugearite–trachyrhyolite formation after Bochkarev and Yazeva 2000), (4) uplift of the buoyant continental part of the slab, exhumation of UHP(?) and HP/LT metamorphic complexes and their erosion, (5) formation of the accretionary cordillera of Uraltau antiform (UTA in the Fig. 7), comparable to an accretionary avolcanic arc of Indonesia, and two flysch basins flanking both sides of it: forearc and foredeep basins; (6) formation of the suture zone of the Main Uralian Fault, which divides the accretionary prism overlying the continent, from remnants of the island arc, which itself became an accreted part of the continent.

The collision did not take place along the entire length of the passive margin of the Laurussia continent. To the north of the Ufimian promontory of the continent, the margin of Laurussia shows no evidence of the Late Devonian arc–continent collision. Two remarkable Late Devonian events in the Southern Urals can be regarded (along with direct structural data) as important indicators of the transition from subduction to collision. The first is the Zilair greywacke flysch Formation of the eastern provenance (uppermost Frasnian to Famennian) which overlies Frasnian deep-water and Famennian shallow-water deposits of the continental margin of Laurussia (Puchkov 2002, Fig. 13). The second is a culmination of HP/LT metamorphism at 372–378 Ma against the background of a general range of Sm-Nd, U-Pb, Ar-Ar, Rb-Sr determinations between 415 and 360 Ma (reviewed by Brown et al. 2006 and Puchkov 2009a, 2010a) that provides additional evidence for the end of subduction, onset of collision and exumation of HP metamorphics (Fig. 14, left). The diagram shows, that age determinations, made by different methods, not necessarily date the age of the exhumation process, making a rather continuous succession. The oldest Sm-Nd and U-Pb dates (410–415 Ma) correspond to an age of a pre-subductional situation, and characterize the garnet pyroxenites (meta-gabbro) corresponding to a deep-mantle high-pressure granulitic metamorphism. These rocks were exhumed along with the HT-LP rocks and preserved pre-subductional mantle ages.

In the Northern and Polar Urals, the age of the oldest known Paleozoic easterly-derived polymictic sandstones and conglomerates on the continental margin of Laurussia to the west of the Main Uralian Fault, is the Lower Visean (Puchkov 2002 and references therein). It must be noted however, that the Early Visean (345 Ma) is the upper limit of the event, because in the Lemva zone the easternmost sections are concealed under the large-scale thrust of the Voykar–Synya ultramafic massif. As for the age determinations of the Paleozoic metamorphism in the Cis-Polar and Polar Urals, they range between 370 and 348 Ma based on the same methods as in the Southern Urals (Fig. 14, right). These age determinations support the idea of a later

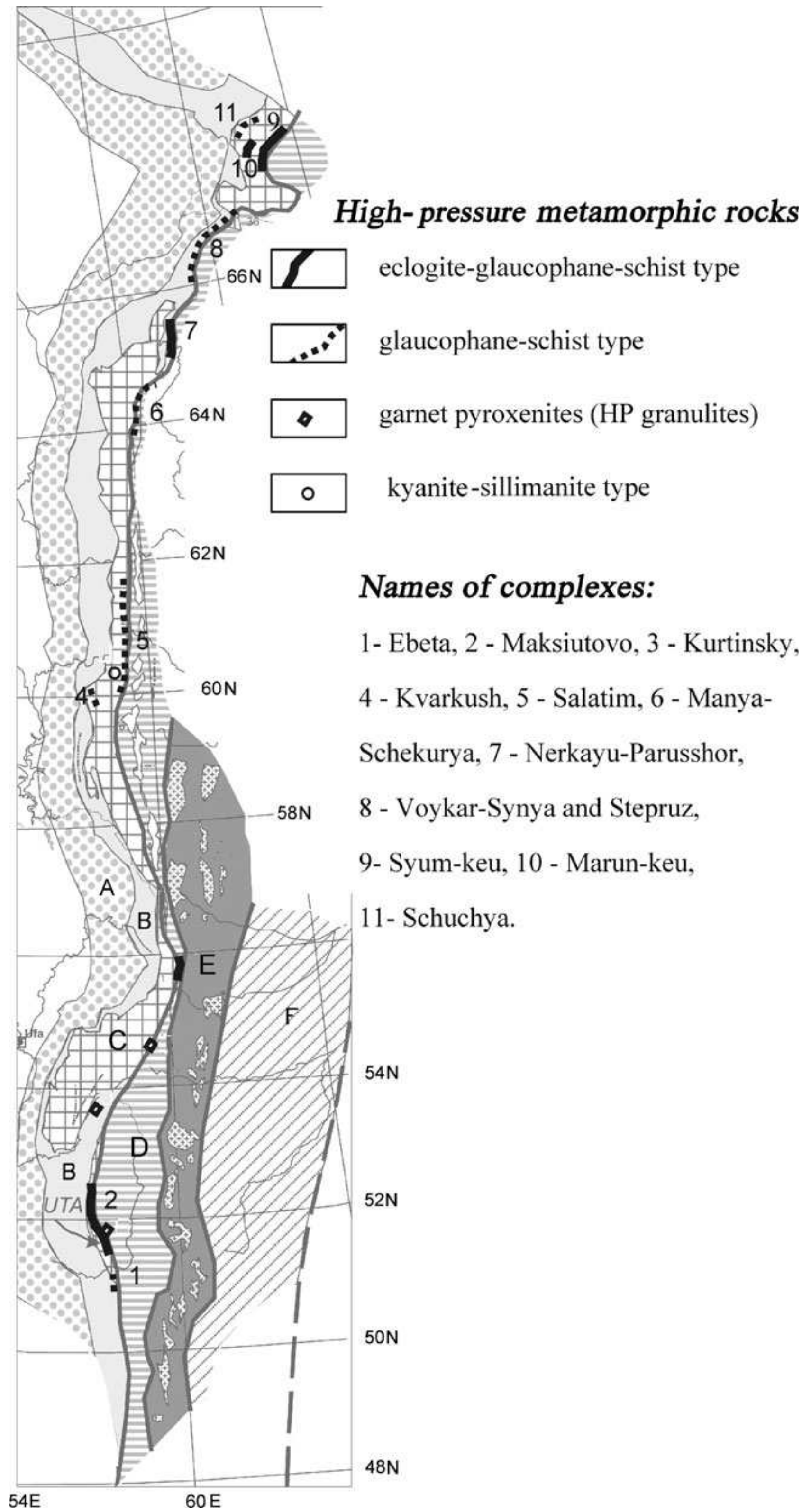


Fig. 12 The All-Uralian high-pressure/low-temperature (HP-LT) metamorphic belt and local high-pressure metamorphic rocks accompanying the Tagil-Magnitogorskian island-arc zone in the walls of the Main Uralian Fault (Puchkov 2009a, modified). Based on Fig. 7

exhumation of the HP-LT metamorphic rocks in the northern parts of the Urals, compared with the Southern Urals.

Therefore, the collision of the Magnitogorsk arc with Laurussia may have occurred in two discrete stages. First, in the Southern Urals by the Famennian (at which time the Magnitogorsk arc was accreted), a triangular–shaped gap was left between the arc and the continent, similar to the modern Bengal Bay, Northwest Australian Bay or South

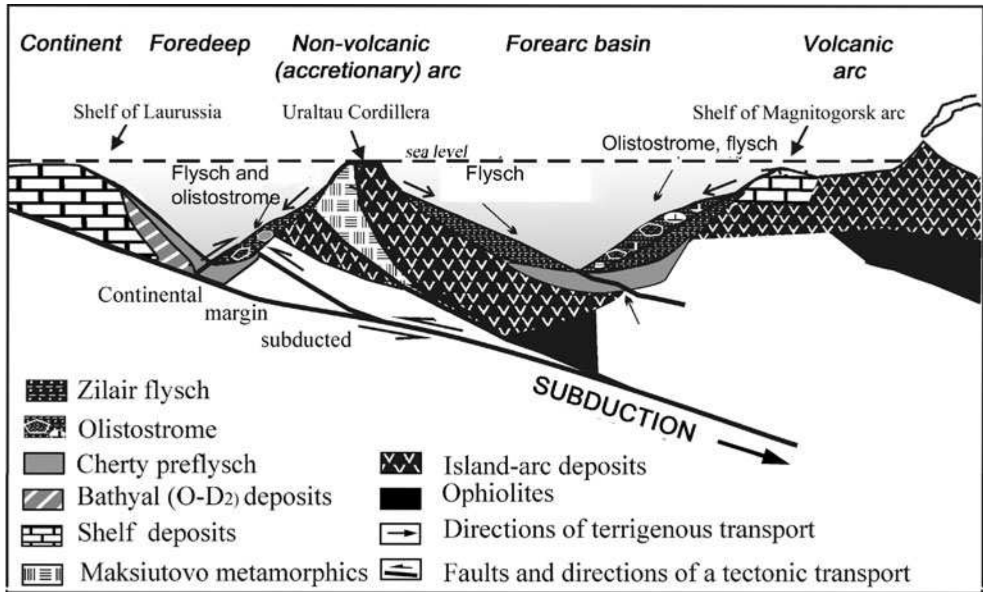


Fig. 13 Reconstructed paleogeological section across the Magnitogorsk arc for the Mid-Famennian time (Puchkov 2002, 2009b, modified)

China sea. Second, in the Early Carboniferous, the northern half of the arc bent to the west and moored to the continental margin (Fig. 15).

The Early Carboniferous-Bashkirian subduction stage followed the above-described Early Carboniferous (Tournaisian–Visean) stage and began in the Serpukhovian, as indicated by the Verkhisetsk chain of granitoids (320 Ma) (Fig. 11-III), related by Fershtater et al. (2006) to another east-dipping subduction zone. By the middle of the Lower Carboniferous, the suture zone was established along the whole length of the MUF (Puchkov 2000, 2002). The chain of 335–330 Ma (mid-Visean) massifs (Turgoyak–Syrostan group of granites) of the Middle Urals intrude the suture zone (Fershtater et al. 2006) and therefore post-date the Magnitogorsk subduction, providing an upper age limit for MUF. This conclusion is supported by the age of the Ufaley granite intrusion (concordant U–Pb, 316±1 Ma, Early Bashkirian), which seals the Main Uralian Fault in the northern part of the Ufimian promontory (Hetzel and Romer 1999).

Meanwhile in the eastern Urals, from the latest Devonian until the Mid-Bashkirian, formation of an ensialic subduction zone occurred. The Main Granitic Axis of the Urals (Fig. 7) developed first as a chain of suprasubductional tonalite–granodiorite massifs at the time of the boundary between Famennian and Tournaisian (ca. 360 Ma), when the southern part of the Magnitogorsk subduction zone ceased to operate (Bea et al. 2002; Fershtater et al. 2006). Simultaneously, immediately to the east, a wide NNE-trending band of calc-alkaline and partly within-plate volcanic rocks and associated plutonic complexes ranging up to mid-Bashkirian in age were formed, suggesting a close affinity with the massifs of the Main Granitic Axis (All-

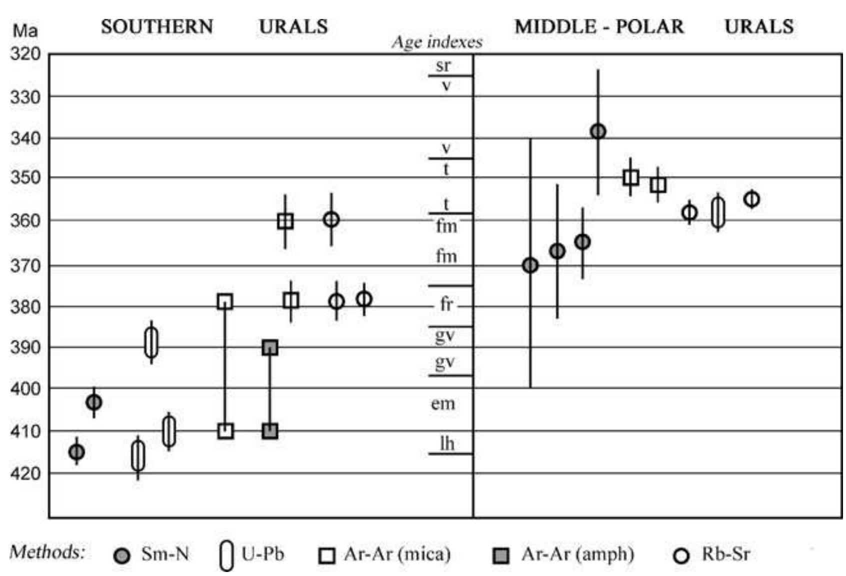


Fig. 14 A comparison of isotopic ages of HP metamorphic rocks from the southern and northern parts of the Urals (Puchkov 2010a). sr–Serpukhovian,v–Vizean, t–Tournaisian, fm–Famennian, fr–Frasnian, gv–Givetian, em–Emsian, lh–Lochkovian

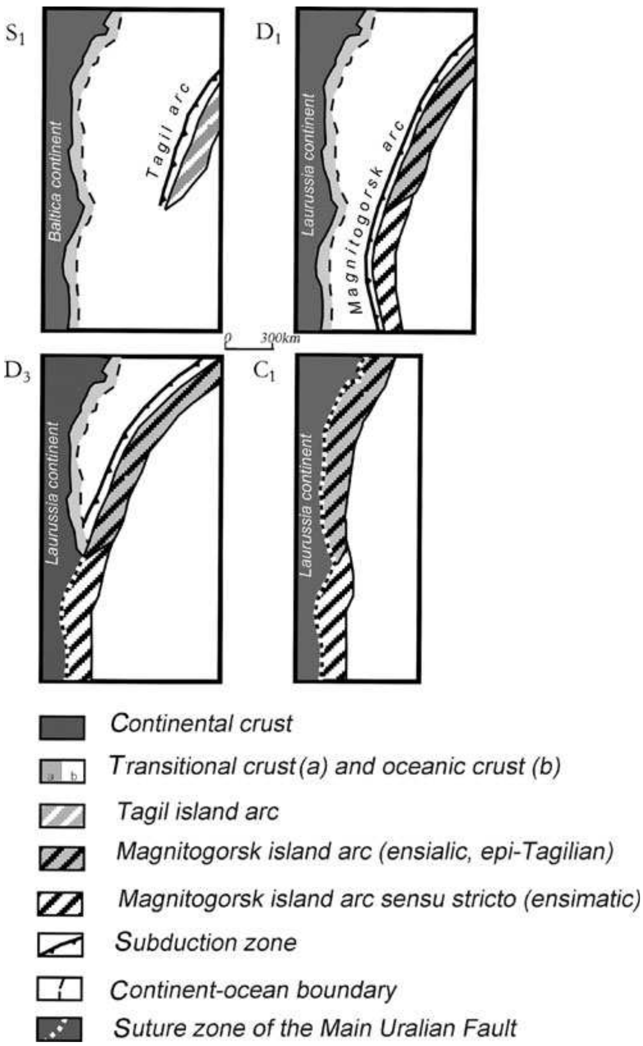


Fig. 15 A model for a two-stage Upper Devonian–Lower Carboniferous arc–continent collision in the Urals (Puchkov 2009a, slightly modified). S_1–plate tectonic situation for the Early Sllurian. Tagil arc at some distance from Baltica continent. D_1–situation in the Early Devonian (Emsian). Magnitogorsk arc, incorporating inactive Tagil arc as a terrane, is divided from Laurussia continent by an oceanic space. D_3–situation in the Late Devonian. Magnitogorsk arc collides with Laurussia continent in the area of the future Southern and (partly) Middle Urals. C_1–situation for the Early Carboniferous (Visean) time. Magnitogorsk arc collides with Laurussia continent in tne northern and polar areas of the future Urals

Russian Stratigraphic Committee 1993; Tevelev et al. 2005; Fershtater et al. 2006). Poor exposure of bedrock in the eastern areas of the Urals has hindered development of unbiased subduction models for this period of time, and there is a considerable difference of opinions as to the number of subduction zones and their polarities. There is also an uncertainty as to the exact geodynamic type of subduction. It was shown (Tevelev et al. 2005) that the Carboniferous volcanics of the eastern zones of the Urals show contradictory geochemical features of both supra-subductional and intraplate type. This can indicate a subduction of a MOR, as exemplified by several places on the eastern margin of the Pacific ocean. It is evident that calc-alkaline volcanic complexes in the Urals abruptly stopped forming by the mid-Bashkirian, marking the end of a wide-scale subduction of an oceanic crust (its disappearance) and transition to a continent–continent-type collision.

Continent–continent collision and orogeny. The main Bashkirian-Permian phase

The collision between Laurussia and Kazakhstania that resulted in mountain building in the Urals has been described recently by Brown et al. (2006) and Puchkov (2010a) and is briefly summarized here.

The collisional processes between Laurussia and Kazakhstania began by the Mid-Bashkirian, first as formation of linear uplifts and basins, documented in the Southern Urals (Puchkov 2000 and references therein). In the Late Bashkirian and Moscovian, widespread marine flysch sediments were deposited in troughs separated by slowly subsiding shelf zones and intensely eroded uplifted crustal blocks. As uplift continued, the basins contracted and inverted. By the Kasimovian time, the territory east of the MUF was dominated by erosion and subaerial deposits, locally with terrigenous sediments and evaporites.

To the west of MUF, a deep-water foredeep trough was being filled by easterly-derived flysch, prograding to the west (Puchkov 2000). The foredeep was divided by basement–controlled transverse uplifts into a series of semi-insulated basins, which were favourable conditions for a formation of thick evaporites of Kungurian time. Later on, the salt and gypsum were mobilized to form crest-like, elongated (horizontal stress-influenced) diapirs. A westerly prograding foreland west-vergent fold-and-thrust belt developed along the eastern margin of the foredeep, which deformed and uplifted flysch and underlying sediments of its eastern limb. Behind this thrust front, a series of synform-hosted nappes is preserved, represented by bathyal, ophiolite and island arc formations (Sakmara, Kraka, Bardym, and other allochthons, Fig. 16).

At the eastern limb of the orogen, east-vergent thrusts and folds were formed. Therefore, the orogen acquired a typical well-preserved bi-vergent structure, that was well demonstrated by deep seismic profiles, URSEIS-95 and ESRU-SB-93-95 in the Southern and Middle Urals (Brown et al. 2006; Kashubin et al. 2006; Rybalka et al. 2006; Puchkov 2009a, b, 2010a and references therein, Fig. 17). Recently the results of the third, Polar Uralian profile were released though still not published. All three profiles support the idea of the bi-vergent character of the orogen; all the profiles show a typical and constant asymmetry. The structures of the western limb have a main detachment at the top of the

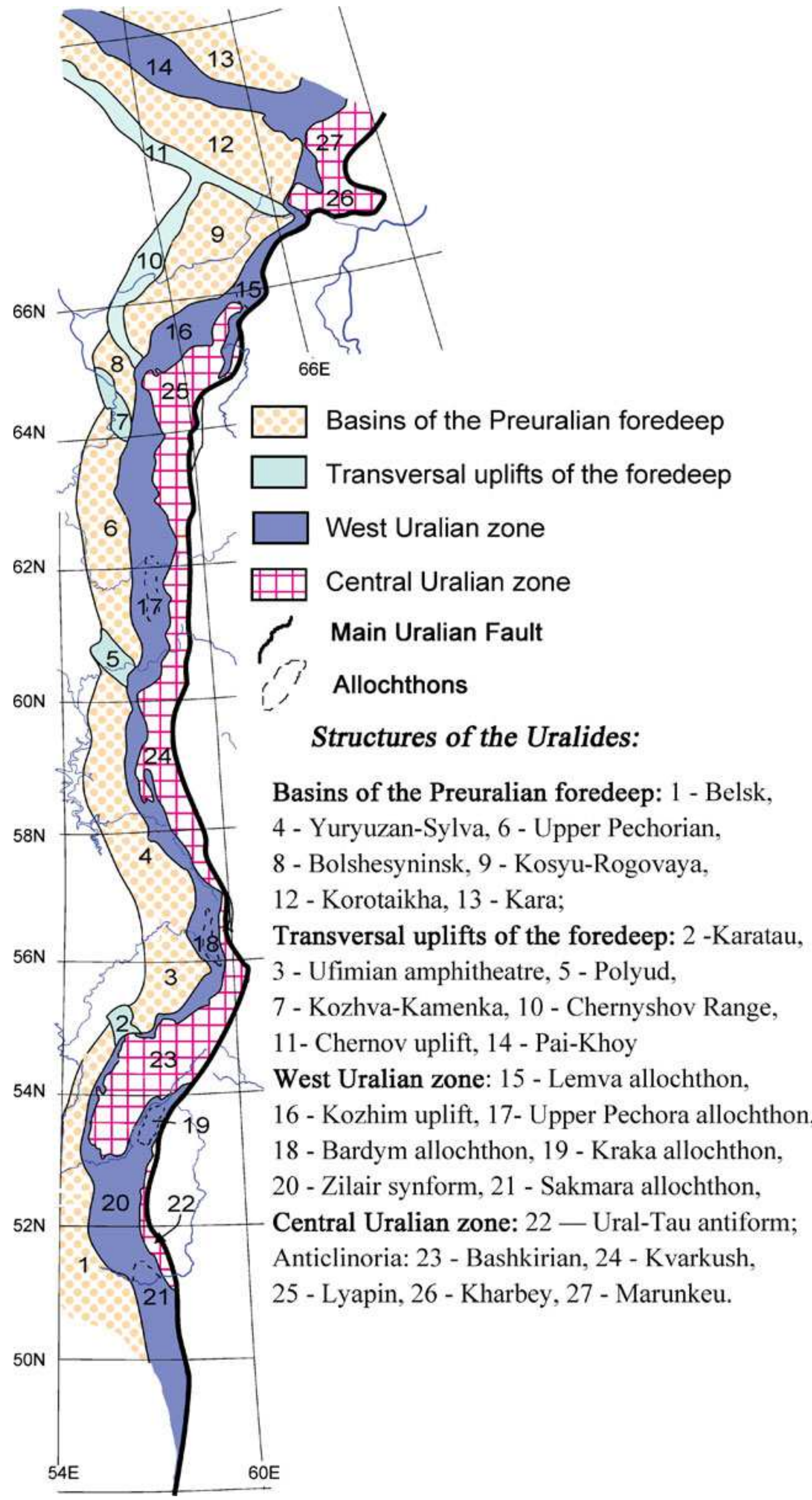


Fig. 16 Structures of the paleocontinental sector of the Urals (Puchkov 2010a slightly modified)

crystalline basement, resulting in "thick-skinned tectonics"; in the Middle and Polar Urals the fold and thrust structures grade to the west into the "thin-skinned" tectonics with a detachment at the base of the Paleozoic. Conversely, at the eastern limb of the orogen the main detachment is situated at the MOHO boundary, so the thrust tectonics here may be called "super-thick-skinned". Such an asymmetry can be explained by the comparative weakness of the young Kazakhstanian continent, devoid of a solid and thick Precambrian lithosphere.

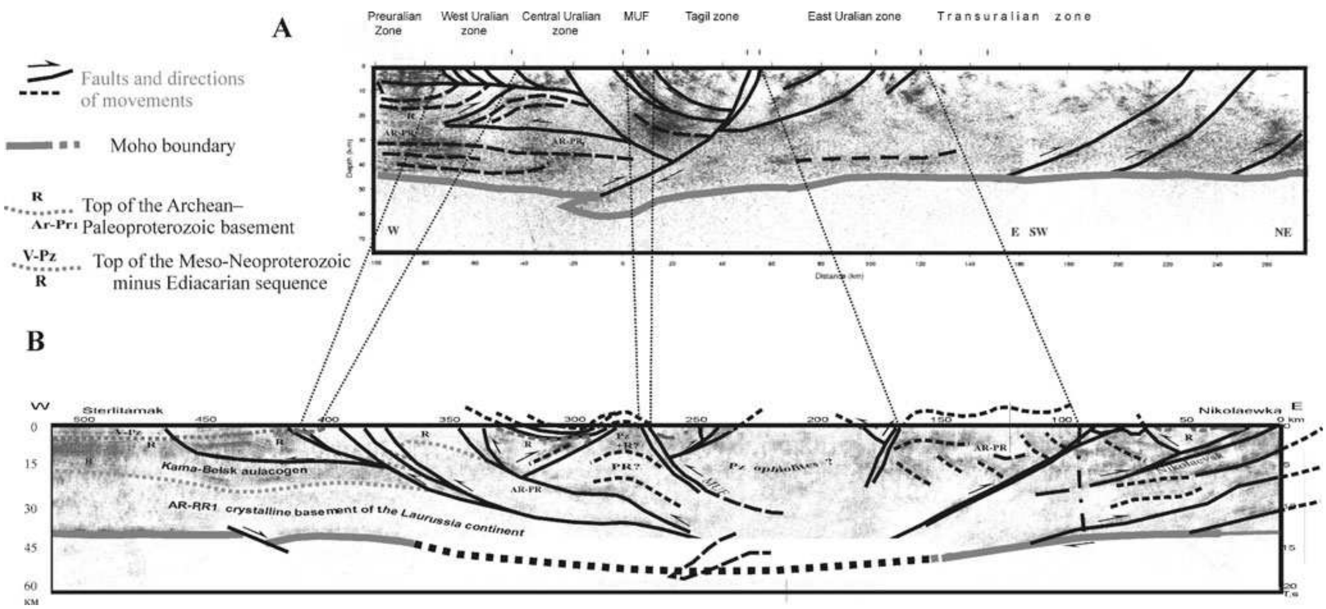


Fig. 17 Seismic profiles URSEIS-95 and ESRU-SB 93-96 with a structural correlation. The *lines* of the profiles are shown on the Fig. 7. (Puchkov 2010a and references therein)

In the west, the fold-and-thrust belt of the foreland (Externides) is much better studied than the eastern one. A series of complicated nappe structures is described, consisting of a regular succession of tectonic sheets, of which the higher is the older (Puchkov 2010a). The formation of these thrust packages is explained by a theory which likens the fold and thrust structures of the foreland to a wedge of a deformed earth before a bulldozer, when each new thrust originates under the wedge along a detachment surface (Davis et al. 1983). The exclusions from this rule belong either to the most early deformation, connected with the end of subduction and exhumation of HP-LT metamorphics, or to the latest stage when the wedge stops.

Tracing the structures from the western margin of the orogen toward its interior, we first meet so-called "thin-skin" structures, with a shallow, near-horizontal detachment surface; further on, they change abruptly, through a ramp, to "thick-skin" structures, where shear zones are strongly inclined inside the orogen. This succession of structures is completed by sutures—zones where the material is flattened, laminated and squeezed onto the outer zones as tectonic sheets. Such zones include the MUF, East Magnitogorsk and Serov-Mauk mélange zones which have the greatest transverse shortening. The Internides (zone D, Fig. 7) are contrastly characterized by simple non-squeezed plicative and disjunctive structures. Their good preservation is a specific feature of the Uralides.

In Gzhelian–Sakmarian time, thrusting and crustal thickening created a hot crustal root in the Southern Urals, which resulted in generation of 305–290 Ma syn-collisional "wet" granites of the Main Granitic Axis (MGA) (Fig. 7, zone E), followed by 10–15 km of erosion (Fershtater 2012; Fershtater et al. 2006). Crustal thickness at that time may have reached 65 km (the modern crustal thickness of the East Uralian zone is up to 50 km). According to drilling data, the MGA can be traced to the northern part of the orogen, where it is concealed under the Mesozoic–Cenozoic sedimentary cover of the West Siberian basin (Ivanov et al. 2005b); the ideas for large Precambrian microcontinental blocks (like the Uvat-Khanty-Mansy) existing in the Uralian basement of the West Siberian basin do not find any support in the recent research.

Syn-collisional granite magmatism in the Southern–Middle Urals migrated northward, and is thought to be a manifestation of the oblique, diachronous character of collision (305–290 Ma for the southern Uralian granites, 265 Ma for the Kisegach massif, and 250–255 Ma for Murzinka and Adui massifs of the Middle Urals (Fershtater 2012; Fershtater et al. 2006). Tuff layers in deep-water sections of Gzhelian to Lower Kungurian preflysch and distal flysch of Preuralian foredeep may represent remote volcanic equivalents of this magmatic activity (Davydov et al. 2002).

Proceeding with L. Kober's (1933) terminology zone E can be called the Metamorphides and zone F–attributed to the internal zone of the eastern Externides, with its "super-thick-skinned" tectonics. The outer zone of the eastern Externides is poorly exposed and must be sought in the outer part of Kazakhstanides, where the Upper Devonian–Lower Carboniferous sedimentary cover is still preserved, like in the Bigger Karatau, where "thin-skinned" tectonics can be demonstrated.

A change from thrust-dominated tectonics to sinistral transpression occurred in the Southern Urals in the Late Paleozoic

(Ivanov 1998 and references therein; Znamensky 2009), explaining the K-rich concentric-zoned post-tectonic ca. 283 Ma rift-related granite-monzonite massifs at the northern end of Magnitogorsk synclinorium (Fershtater et al. 2006) and the occurrence of ca. 301–310 Ma lamproite dykes in the Southern Urals (Pribavkin et al. 2006). The diachroneity of the magmatism is underlined by the presence of the Late Permian marine sediments with Tethyan fauna in the southernmost Urals that is coeval with granite magmatism in the Middle Urals (e.g. Chuvashov et al. 1984).

LIP formation and rifting in the Triassic

At the Permian–Triassic boundary, the waning manifestations of orogenesis were overprinted by the formation of the vast Uralo-Siberian LIP (large igneous province), extending from Taymyr in the north to the Kuznetsk and Turgay basins in the south and from the Tunguska basin in the east to the Urals in the west, therefore not controlled by previous tectonic structures. In the Urals, the western boundary of basalt eruptions goes across the orogenic structures, cutting them at a small angle. Volcanism started locally in Central Siberia with alkaline basalts and minor rhyolites. But very soon vast flows of flood basalts were emplaced, hundreds of metres thick, alternating with coarse-grained terrigenous sediments, filling rifts and wide basins. The area of the Triassic flood basalt volcanism changed at its periphery to manifestations of basalt-rhyolite volcanism and small granitoid intrusions. Ar–Ar data (Ivanov et al. 2005a) suggest that the bulk of the volcanism initiated in Siberia at the Permian–Triassic boundary but probably continued for 22–26 Ma, with several short surges. Recent Ar–Ar dates for plagioclase from basalts in the Polar Urals (249.5+0.7 Ma) and in the east of the Southern Urals (243.3+0.6 Ma, Reichow et al. 2009) support the simultaneous beginning of the LIP formation over a vast region followed by a more protracted period of reduced magmatism.

In contrast with eastern Siberia, where the volcanic flows overly the Permian sediments almost without unconformities, the Early Triassic history of the Urals is dominated by considerable uplift, erosion and formation of thick coarse-grained alluvial to proluvial sediments that resemble orogenic molasse but are attributed to the effects of the Uralo–Siberian distributed rifting, accompanied by a general uplift and LIP magmatism. Examples include the huge Triassic Koltogorsk–Urengoy graben of Western Siberia, the newly-identified Severososvinsky graben in the subsurface of the Cis-Polar Urals (Ivanov et al. 2004), and basins of the eastern parts of the LIP (Kurenkov et al. 2002; Ryabov and Grib 2005).

Old Kimmerian deformation in the Jurassic

A short phase of orogeny occurred at the end of the Early Jurassic, and its effects differ along the strike of the Uralides. The Triassic deposits of the Southern Urals are affected by this orogeny only in the Trans-Uralian zone (Chelyabinsk and other graben-like depressions), where Upper Triassic and older deposits are deformed by thrusting (Rasulov 1982), followed by uplift and peneplanation during the Middle and Upper Jurassic, and deposition of Upper Cretaceous marine strata. In the Northern Urals, three 'grabens' (Mostovskoi, Volchansky, Bogoslovsk-Veselovsky) (Tuzhikova 1973) containing Upper Triassic coal-bearing sediments were complexly deformed. In the Polar Urals, the Triassic deposits of the foredeep and the Chernyshov and Chernov range are all deformed, and are unconformably overlain by Middle Jurassic strata. However, in the nearby Severososvinsky graben to the east, Triassic and Jurassic deposits are not deformed (Ivanov et al. 2004; Puchkov 2010a), illustrating the localized nature of Kimmerian orogenic events. The Pay-Khoy and Novozemelsky ranges were formed as a whole in the Kimmerian phase (Korago et al. 1989; Yudin 1994). Kimmerian orogenesis is attributed to large-scale intra-Pangaean displacements, accompanied by block rotation, possibly reflecting lateral escape of Kazakhstania between Laurussia and Siberia and a collision of the latter two in the basement of the northwestern Siberia. According to palaeomagnetic data (Kazansky et al. 2004), Siberia rotated 30^o clockwise between the Triassic and the Late Cretaceous. The model of such rotation presupposes large-amplitude movements along sinistral strike-slip faults in the Urals, Altay, China and Mongolia (Kazansky et al. 2004; Bragin 2005; Metelkin 2010) (Fig. 18).

The Middle Jurassic-Miocene platform stage of development

The mountains of the Uralide orogen were eroded soon after the Old Kimmerian alpine-type deformations. After the second half of the Middle Jurassic, the sea started to come periodically very close to the Urals territory, from the southwest, and east, though most of the area experienced either slow erosion and weathering or formation of continental coal-bearing sediments, including mature quartz sandstones. During the peak of the vast Late Cretaceous (Santonian–Maastrichtian) transgression, the sea covered the whole Uralian foredeep, the southern part of the Uraltau antiform and Zilair synform including the Sakmara allochthon, and a major part of the Transuralian megazone. It is evident that by that time the Uralian foldbelt was no longer active. The surface of its axial part was probably above sea level, though not very high, taking into account the quartz composition of Cretaceous sandstones and presence of underlying weathering crusts and bauxites. The position of a restored shoreline at the peak of the next (and last) transgression in the mid-Eocenian was approximately the same

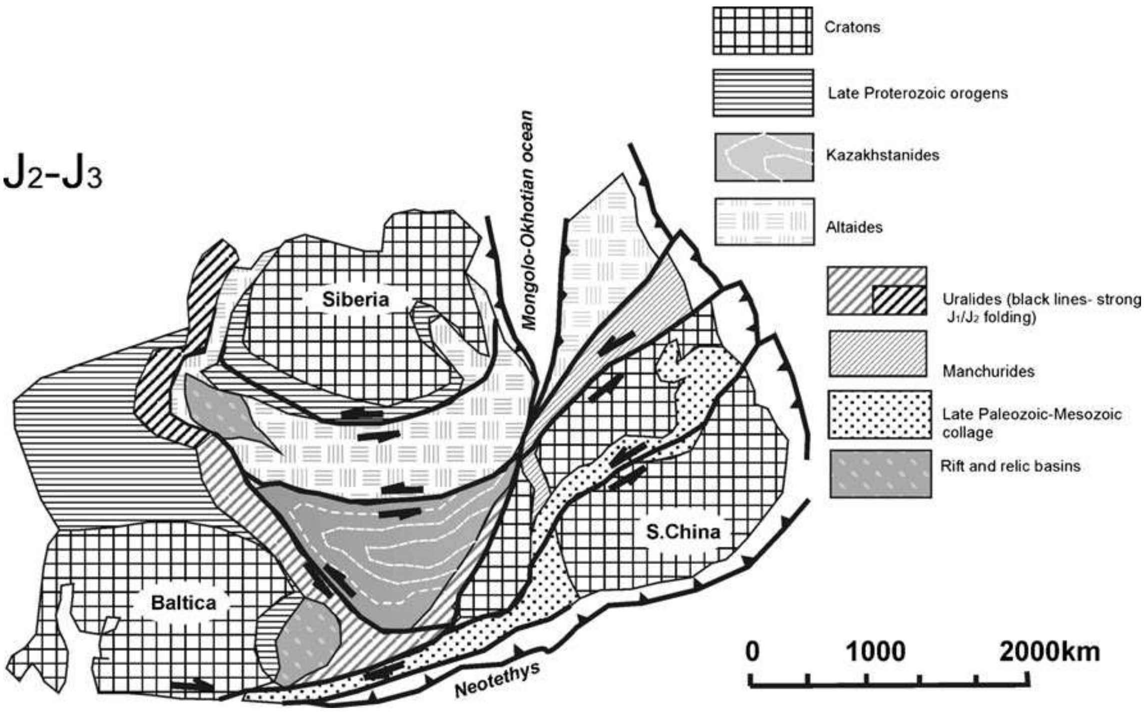


Fig. 18 Wrench-fault deformations between blocks of Eurasia with a clockwise rotation of Siberia (after Metelkin 2010; Bragin 2005, modified)

as in the Cretaceous, which would be explained by the tectonic stability of the area at that time. The northeastern direction of buried Late Jurassic and Early Cretaceous riverbeds in the eastern slope of the Southern-Middle Urals indicate a non–Uralian (northwestern) strike of the watershed during this time. The comparison of fauna in the eastern and southwestern seas around the Urals suggests a presence of marine straits, connecting them. Before the Oligocene, the seas left the territory of the Urals, and Oligocene and younger sediments are continental. Much later, the modern Ural Mountains started to be formed. The probable time limits of the neo-orogenic stage of the Urals are discussed below.

The neo-orogenic stage

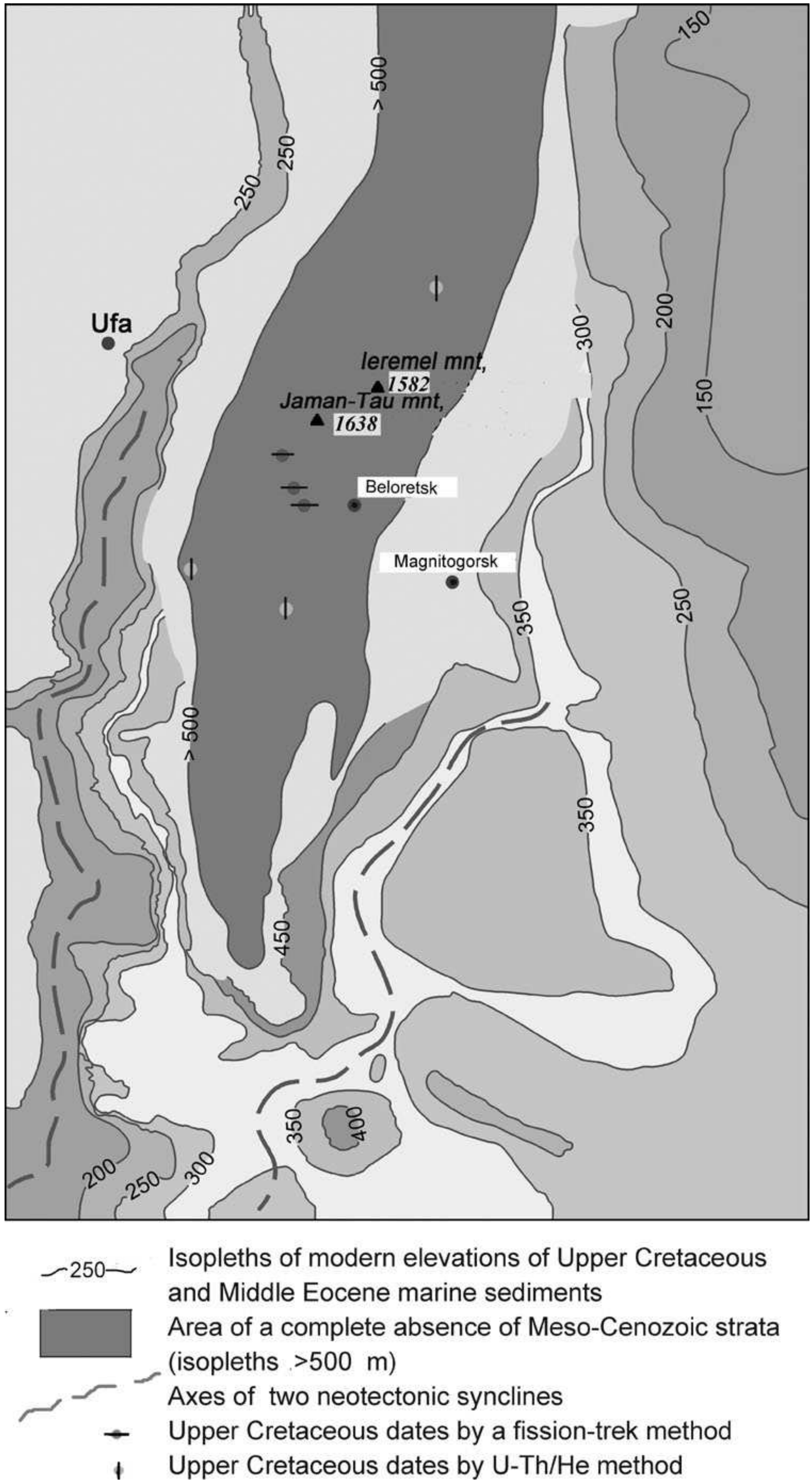
New information concerning the time of transition from the platform to the neo-orogenic stage was provided by recent fission-track data. These analyses gave an information concerning the low temperature history of rocks in the Ural Mountains. It was shown that cooling through the 110 °C temperature isotherm occurred mostly in the Jurassic (Seward et al. 1997, 2002), though some preliminary data show a possibility of Cretaceous normal faulting in the area to the south of the Jaman-Tau mountain (Glasmacher et al. 1999, 2002).

A new method of (U–Th)/He chronometry using apatite was suggested (Reiners 2002). The low closure temperature of apatite (U–Th)/He chronometry (~70 °C, as opposed to annealing temperatures of 110 °C for fission tracks in apatite) makes this method more promising for the study of the late neotectonic history of mountains, particularly the Urals. Studies using this method have been attempted in co-operation with our Spanish colleagues. In several places of the mountainous area of BMA, the exhumation was dated as Late Cretaceous.

Taken as a whole, the data means that the rocks which are now exposed in the surface were in the Jurassic (and locally in the Cretaceous) at a depth of about 2.5–3 km. It was still unanswered when in the later history these 2.5–3 km were eroded. To find the answer we need a geological approach.

The modern altitudes of the Upper Cretaceous and Mid-Eocene marine deposits in the Southern Urals taken mostly from 1:200,000-scale geological maps, and not corrected for the high stand of the Late Cretaceous and Mid-Eocenian transgressions, give the minimum value for the amplitude of the Cenozoic (post-mid-Eocenian) vertical deformations of some parts of the territory. These amplitudes are below 200 m in the Trans-Uralian zone, up to 500 m in the southern part of the Uraltau antiform, about 400 m in the Sakmara allochthon, 300–400 m in the Belaya River valley on the western slope of the Ural Mountains and at 170–260 m in the Russian Plain along the Belaya River valley (Fig. 19). The pre-neotectonic Cretaceous/Eocene denudation levels of those territories of the Urals where the transgressions did not reach, had still higher altitudes (>500 m), though it is very difficult to say how high they were.

Fig. 19 Deformations of a near-bottom surface of shallow-water Upper Cretaceous and Middle Eocene seas (isopleths show their modern absolute heights, with corrections for the thickness of the sediments)



Anyway, the Eocene-Cretaceous transgressive-erosional reference surfaces were considerably deformed during the neotectonic stage, but again it is hard to tell if it was a simple arch-like uplift or a more complex deformation, though a couple of synclines can be indicated (Fig. 19).

There is not much data available for active high-amplitude faultings during the neotectonic stage, though Bachmanov et al. (2001) give evidence for the Oligocene-Quaternary faults having an amplitude of up to 100–200 m. Palaeomagnetic studies of the Late Pleistocene deposits in the Yuruzan-Ai and Magnitogorsk depressions have shown the presence of local young folds and fault dislocations (Minibaev and Sulutdinov 2001). The direct studies of folds and flexures in the Pliocene and Pleistocene deposits in the Southern Urals also speak in favour of the localized character of some deformations accompanying the general uplift of the territory (Danukalova et al. 2002; Kopp and Yegorov 2002; Puchkov 2010a). In the northern territories of the Urals, Kopp (2007) has shown that where Cenozoic sediments are present, the participation of thrust and strike-slip dislocations in the neo-orogenic movements are quite evident. Tevelev (2002) also attributed great importance to strike-slip movements in the Urals. Systematic monitoring of weakest earthquakes has shown their concentration along some old faults in the Southern Urals, giving evidence of their rejuvenation (Kazansev et al. 1995).

The relief of the Urals is a combination of ridges and mountain massifs with relics of rather smooth denudation surfaces (e.g. the North Kraka Mountains with a denudation level of about 1,000 m) and lower plateaus (e.g. the Zilair plateau, 500–700 m high). The relief around the highest mountains (Jaman-Tau, Mashak, Ieremel and others) is a combination of narrow ridges with relics of plains at altitudes of up to 1,300 m and U-shaped valleys clearly suggesting their intramountain glacial origin in the Pleistocene (Kolokolov and Lvov 1945; Astakhov 1984; Levina et al. 2001), though good descriptions of moraines in the Southern Urals have still not been published. Further north, especially in the Cis-Polar and Polar Urals, the participation of mountain glaciers in the modeling of the modern relief is absolutely evident.

No fauna have been found to date the denudation levels in the higher part of the Urals. According to Borisevich (1992), evidently following the general ideas of some previous researchers (e.g. Sigov 1969), the ranges and massifs dominating over the level of peneplained watersheds of the Ural Mountains have a relic nature and originated as a result of erosion of the Middle Triassic peneplain. However, this point of view conflicts with the fission-track data showing that at least 3 km-thick mass of rocks was eroded in the Bashkirian anticlinorium since the Jurassic and Cretaceous. In this case, even the Jurassic and Lower Cretaceous peneplains had no chance to survive. A more realistic guess could be a Cenozoic age for it. The existence of Late Cretaceous and Eocene marine straits connecting the eastern and western seas surrounding the Urals land (Papulov 1974; Amon 2001; Benyamovsky et al. 2006a, b) suggests again that the modern Urals as a single continuous mountain chain was formed only in the Late Cenozoic, though some hills and even hill chains of the Uralian strike could already have existed locally in the Cretaceous.

The lithology and facies of the sediments of the neotectonic stage bear their own information on the character and development of the mountains (e.g., Astakhov 1984; Stefanovsky and Shub 1997; All-Russian Stratigraphic Committee 1997). The Oligocene in the Urals is preserved in the Fore-Uralian, Orsk-Tanalyk and Transuralian zones and is represented by quartz sandstones, siltstones and clays of alluvial and lacustrine nature. The thickness is up to 50 m. The Miocene in the Fore-Uralian zone is represented by a Priuralian series developed in karst depressions situated above the Kungurian evaporites and consists of quartz sands, silts, clays, sometimes conglomerates, with a total thickness of up to 300–350 m, with coal seams (Yakhimovich and Andrianova 1959). In higher places of the zone, the Miocene is represented by lacustrine and alluvial terrigenous sediments. In the Late Miocene, after the accumulation of the Priuralian series, a period of intense erosion started, though it is hard to explain it by an uplift of the territory. Conversely, it may have been due to a Messinian-like event of a great depression of the Caspian Sea (lake) level, which was a base level for the rivers of Paleo-Volga basin. The valleys of the Belaya and Ural Rivers were eroded down to 100–200 m below the modern sea level, while the depth of the Caspian Sea level was at ~550 m below the ocean level (Milanovsky 1963; Sydnev 1984; Rozhdestvensky and Zinyakhina 1997; Leonov et al. 1998). Meanwhile, due to the subsequent uplift of the Urals, the upper reaches of these rivers do not show so anomalously downcut valleys (Varlamov 1960). In the central part of the Urals, the Miocene occurs sporadically; in the mountain area of Beloretsk it is represented by the coal-bearing sediments of the Priuralian series and having the same features as in the Fore-Urals (Kozlov 1976); in other places, lacustrine and alluvial sediments are predominant. In the Transuralian megazone, erosion was prevailing, and only locally deluvial and eluvial-proluvial red-coloured sediments accumulated.

The Pliocene in the Fore-Uralian zone was a time of filling of the downcut river valleys. The initial accumulation was with coarse-grained alluvial-lacustrine sediments, which changed upwards to less coarse, silty-clayish Akchagyl sediments with marine fauna, up to 120 m thick. At that time, due to a rise of the Caspian Lake above the normal level, an ingression of brackish waters far up river took place (Sydnev 1984, 1985). In the Central Urals, the Pliocene is represented sporadically in erosional depressions by red-coloured alluvial, deluvial-lacustrine sediments (clay, silt, gravels and pebbles). Since that time, the sediments were clearly polymictic which is evidence for an acceleration of the erosion. In the Trans-uralian zone, the

Pliocene is represented by thin differently coloured clays, some sands, gravels and pebbles. Most of the river terraces are dated as Quaternary (Stefanovsky 1997). An Early Pliocene and Miocene age of some terraces has been put forward from time to time (e.g. Kazakov 2003), but these proposals are not supported by any geological data. No cave sediments older than the Neopleistocene have been recorded so far; therefore the observed cave formation, a process closely dependent on mountain growth, could have started only recently. That is why V. Puchkov (in Danukalova et al. 2002) has expressed the opinion that the mountain growth was strongly accelerating through the neotectonic stage towards modern times.

Modern tectonic activity of the Urals is shown in many ways. First of all, the intense modern movements of the earth's surface has been proven by repeated topographic leveling. The velocities of the surface uplift are up to 5 mm/year (Kononenko et al. 1990), which means that either the vertical movements of the Urals' surface were strongly oscillating (which is not proven by any independent method), or that the Ural Mountains have grown mostly during the Late Pliocene–Quaternary. Recently, the first data on a GPS-monitoring of horizontal movements of the earth's surface in the Middle Urals were published (Utkin et al. 2010). It was shown that the Urals moves with a velocity of 2 cm/year to the east which is well reconciled with previous data showing that the Eurasian plate is rotating clockwise. Furthermore, minor differences in velocities and directions for individual points permit differential movement of tectonic blocks of the area. This new GPS data for active differential movement is consistent with the fact that the Urals are known for their seismicity. Some strong and even destructive earthquakes were recorded during the historical period (Utkin 2001). They are concentrated mostly around the protruding rigid Ufimian salient of the East European Platform, which is acting as an indenter. The measured maximum stress directions in the Middle Urals are oriented perpendicular or slightly oblique to the structural grain of the Paleozoic basement. Data on the current stress patterns in the Urals was used in an attempt to explain the formation of the modern Ural Mountains (Mikhailov et al. 2002). When the presence of the cold, rigid Magnitogorsk block is taken into account, it was shown that under sufficient stress the territory immediately to the west of this block ought to be deformed; the fault (destructive zone) responsible for this deformation originates deep in the crust under the Central and Western Uralian zones, where maximum deformation and uplift are concentrated.

The questions of the modern history of the Southern Urals have not been addressed in full for a long enough time, since the last works of Sydnev (1985) and Rozhdestvensky(1997). Meanwhile, a well substantiated stratigraphic scheme for the Late Pliocene and Quaternary sediments has been presented (Danukalova in Puchkov and Danukalova 2009; Danukalova 2010), and provides a good opportunity for a correlation of European and Siberian schemes for the same time span.

The improved stratigraphy, combined with a set of new data (geological, geophysical, fission track etc.) provides an opportunity to revise some ideas that were accepted as absolutely true by previous generations of researchers. For example, it had previously been taken as granted that globally the neotectonic epoch is a time span between the Oligocene and the present (e.g. Trifonov 1999), and it was thought that the Ural Mountains were growing during the whole of the Late Oligocene, Neogene and Quaternary (Rozhdestvensky 1971, 1997; Rozhdestvensky and Zinyakhina 1997; Kazakov 2003). It was also thought that the morphology of the Uralian relief preserves some surfaces as old as the Triassic. However, these ideas conflict with our new data. The fission-track and geological data show that the relief of the Southern Urals stabilized by the end of the Cretaceous/beginning of the Cenozoic, and there was no chance of preservation of erosional features as old as Triassic and Jurassic. On the other hand, the idea that the Ural Mountains were growing during the whole Late Oligocene and Neogene is contradicted by various data. Among these the most important are: (1). The Miocene Priuralian series contains considerable amounts of quartz sandstone and sand, and therefore are witness to non-orogenic conditions of weathering, erosion and accumulation. The same type of sediments is recorded even deep in the mountains near the town of Beloretsk. As for the polymictic sediments, they appear only by the Late Pliocene (Verbitskaya 1964); (2). The deep Neogene erosion of the rivers is better explained not by the uplift of the Urals but by a fall of the Caspian Sea (lake) level; (3). There are no well-documented river terraces of Miocene–Early Pliocene age; (4). No cave deposits older than the Neopleistocene have been detected so far; (5). The velocity of the modern uplift of the Urals surface, determined by topographic leveling, is 10 times more than needed to make the Ural Mountains if it had been constant since the Oligocene. Therefore, we think that the modern Urals were formed as a result of mostly Late Pliocene–Quaternary uplifts. The data on the fractures, flexures and faults in the young deposits speak in favor of a combination of a general dome-like and local thrust and fold deformations experienced by the Urals in the neo-orogenic epoch.

A comparison of the Urals with larger neo-orogens of the same type, such as the Tien-Shan and Altay mountains supports the idea of their intraplate origin. It is the intraplate stress, originated as a far-field influence of the Alpine-Himalayan orogeny, which is probably responsible for the modern deformation, reactivating some 'weak' zones and subsequently uplifting the Urals.

Discussion and concluding remarks

As follows from the previous review of the Uralian geology, the structure and history of the Urals is utterly complicated and encompasses the time spell of 3,5 billion years, which corresponds to the most part of the Earth's history. Nevertheless, the thorough analysis permits to establish in the Urals five clearly outlined and distinct stages, organized quite differently and demonstrating a certain evolutionary trend.

The earliest and longest (unfortunately not subdivided into shorter periods) stage is typical for the Archean- Early Proterozoic (Paleoproterozoic) Earth's history: a wide development of relics of granulite metamorphic facies, thoroughly reworked by a later amphibolitic metamorphism and granitization. The magnetic anomalies in the adjacent area of the Volgo-Uralian province are partly linear and partly curvilinear, ovoid-like, reflecting a specific ancient tectonic style of this stage, corresponding to the crystalline basement of the platform, traced under the western zones of the Urals.

The structures of Timanides, corresponding to the Riphean-Vendian (Meso- and Neoproterozoic) time of development, demonstrate quite a different style: they are linear and make a fan-like pattern in plan. Timanides have a foredeep–probably one of the earliest recorded in the Earth's history. The sedimentary succession of the Externides, exposed in the Timan Range and (at its best) in the BMA, corresponds to the continental margin of Baltica. It is very weakly metamorphosed, has a great thickness, shows no conspicuous lacunes and contains rift-type volcanics at several levels. A thorough work on geochronology of these volcanics with application of new methods permitted us recently to refine the data on age of stratigraphic boundaries in this standard section and suggest it for a further discussion concerning an upgrade of the International Stratigraphic Scale of the Meso- and Neoproterozoic. The Internides, exposed in the Northern parts of the Central Uralian zone, are more metamorphosed, reveal presence of oceanic (ophiolite) formations (although poorly developed), suprasubductional volcanics of calc-alkaline affinity and granites. The tectonic features include a characteristic presence of thermal domes. At that, HP-LT metamorphism was weakly developed, though present in the north of the Kvarkush anticlinorium. It can be also attributed to a general trend of evolution: such metamorphism is developed at its best only in the Phanerozoic fold belts. The structure of Timanides suggests its uncomplete preservation, connected with the Ordovician rifting and Late Paleozoic orogenic deformations which reworked their crust.

The Paleozoic Wilson cycle, and the Uralides as its result, is again quite different from the previous cycle and the Timanides.

Extremely good preservation of oceanic and island-arc complexes and a low degree of shortening in the foreland belt are unprecedented in Paleozoic orogens and give the Uralides a real individuality. Many other features of the Urals are also rare, such as its well-preserved bi-vergent structure, island arc-related platinum-bearing belt, model arc-continent collision, diachroneity of collisions, a combination of orogenic and rift-related magmatism in a single stage of transpressive deformation of the lithosphere, an exemplary development of an eclogite- glaucophane-schist metamorphism and preservation of a heavy, relatively 'cold', isostatically equilibrated root (Puchkov 2010a, b and references therein). On a plate scale, however, the history of both the Timanides and Uralides follows the main stages of a Wilson cycle, finished by orogenic movements and modified somewhat by episodes of plume-related tectonics and magmatism in the Middle and Uppermost Riphean, Devonian and Triassic.

On the contrary, the latest stage of orogenesis, which created the modern Uralian mountains, was not connected with Wilson cycle. The neo-orogen was built on the Meso-Cenozoic peneplain and its origin resulted from the influence of strongest compressional forces, spread from the Alpine-Himalayan collisional belt at its latest stages of development. Probably a wide development of regenerated mountains like the neo-orogenic Urals, Tian-Shan, Altay and others was the next innovation in the tectonic style of the Earth and a witness of its further evolution.

Summing up, we stress upon that the geology of the Urals is a very dynamic field of knowledge that develops continuously and quickly. Even during the short period when this paper was being prepared, the author ought to make some changes and amendments in it, to take into account new data, acquired quite recently. On the other hand, the main concepts of the paper are now well developed and stable. They are based on plate and plume tectonics, in their modern versions.

Looking forward, we must admit that the more we know about the Urals, the more there is to discover in the future. Here we can indicate some problems which await further research. One of these unresolved problems is the place of Baltica continent and Timanides relative to the continents of the Gondwana group and Vendian-Early Cambrian orogens. Only recently, reliable paleomagnetic poles had been acquired for the Upper Vendian of the Urals and there is a hope for better continental reconstruction of this time. In the field of statigraphy, the most important is now a further refinement of the standard Riphean and Vendian sections of the Urals. Many questions arise in connection with new isotope analyses of high-pressure members of the metamorphic Maksiutovo complex. The timing and setting of the Vishnevogorsk carbonatite complex is also a potential object of a very hot discussion. A specific problem is a description of events accompanying the very beginning of a subduction: origin of ophiolite-like formations with dyke-in-dyke complexes and formation of intrusive massifs of the Ural-Alaskan type. Many types of sulfide deposits of the Urals (except VMS deposits) have poor age control and many questions, concerning the timing of metasomatism, metamorphism and metallogeny are still to be elucidated. Application of new methods for dating of sulfides will be of critical value. Another specific topic is dating of dike swarms of the Urals through the Project "Dolerite dikes of Russia" (www.largeigneousprovinces.org/projects). In the field of the study of neo-orogenic stage of the Urals, the most difficult problems include the dating of high planation surfaces of the modern Urals, a search for Quaternary moraines from a hypothesized mountain glaciation of the Southern Urals. This enumeration of current problems is just a fraction of actual questions in the Uralian geology.

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