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Tectonics of the Ural Paleozoides in Comparison with the Tien Shan

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Abstract—The main differences and similarities between the tectonic features of the Urals and the Tien Shan are considered. In the Neoproterozoic and Early and Middle Paleozoic, the Ural and Turkestan oceanic basins were parts of one oceanic domain, with several distinct regions in which tectonic events took different courses. The Baltic continental margin of the Ural paleoocean was active, whereas the Tarim–Alay margin of the Turkestan ocean, similar in position, was passive. The opposite continental margin in the Urals is known beginning from the Devonian as the Kazakh–Kyrgyz paleocontinent. In the Tien Shan, a similar margin developed until the Late Ordovician as the Syr Darya block with the ancient continental crust. In the Silurian, this block became a part of the Kazakh–Kyrgyz paleocontinent. The internal structures of the Ural and Turkestan paleooceans were different. The East Ural microcontinent occurred in the Ural paleoocean during the Early and Middle Paleozoic. No microcontinents are established in the Turkestan oceanic basin. Volcanic arcs in the Ural paleoocean were formed in the Vendian (Ediacarian), at the Ordovician–Silurian boundary, and in the Devonian largely along the Baltic margin at different distances from its edge. In the Turkestan paleoocean, a volcanic arc probably existed in the Ordovician at its Syr Darya margin, i.e., on the other side of the ocean in comparison with the Urals. The subduction of the Turkestan oceanic crust developed with interruptions always in the same direction. The evolution of subduction in the Urals was more complicated. The island arc–continent collision occurred here in the Late Devonian–Early Carboniferous; the continent–continent collision took place in the Moscovian simultaneously with the same process in the Tien Shan. The deepwater flysch basins induced by collision appeared at the Baltic margin in the Famennian and Visean, whereas in the Bashkirian and Moscovian they appeared at the Alay–Tarim margin. In the Devonian and Early Carboniferous, the Ural and Turkestan paleooceans had a common active margin along the Kazakh–Kyrgyz paleocontinent. The sudduction of the oceanic crust beneath this paleocontinent in both the Urals and the Tien Shan started, recommenced after interruptions, and finally ceased synchronously. In the South Ural segment, the Early Carboniferous subduction developed beneath both Baltica and the Kazakh–Kyrgyz paleocontinent, whereas in the Tien Shan, it occurred only beneath the latter paleocontinent. A divergent nappe–fold orogen was formed in the Urals as a result of collision of the Kazakh–Kyrgyz paleocontinent with the Baltic and Alay–Tarim paleocontinents, whereas a unilateral nappe–fold orogen arose in the Tien Shan. The growth of the high divergent orogen brought about the appearance of the Ural Foredeep filled with molasse beginning from the Kungurian. In the Tien Shan, a similar foredeep was not developed; a granitic axis similar to the main granitic axis in the Urals was not formed in the Tien Shan either.

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INTRODUCTION

The relationship between the Paleozoic fold system of the Urals and the Tien Shan has attracted the attention of many researchers. Over the vast territory between the Urals and the Tien Shan, the Paleozoic and older rocks are inaccessible for observation (Fig. 1). The efforts of researchers have been focused on comparison of the tectonic zones pertaining to both regions and on the interpretation of the geophysical fields in the territories overlapped by younger sedimentary cover. Stratigraphic sections of the Urals and the South Tien Shan have been the main objects of comparison. On this basis, many geologists have looked for extension of the Ural tectonic zones in the Tien Shan [7, 8, 70]. Some authors have arrived at the conclusion that no persistent links existed between these provinces [9, 35, 36].

In this paper, we make an attempt to compare the tectonic evolution of the Urals and the Tien Shan in the Neoproterozoic and Paleozoic.

TECTONIC HISTORY OF THE URAL PALEOZOIDES

The Ural Foldbelt is composed of rocks belonging to the East European paleocontinent (Baltica), the adjacent paleooceanic domain, and the marginal part of the Kazakh–Kyrgyz continental massif. The history of the tectonic evolution of the Urals is divided into the pre-

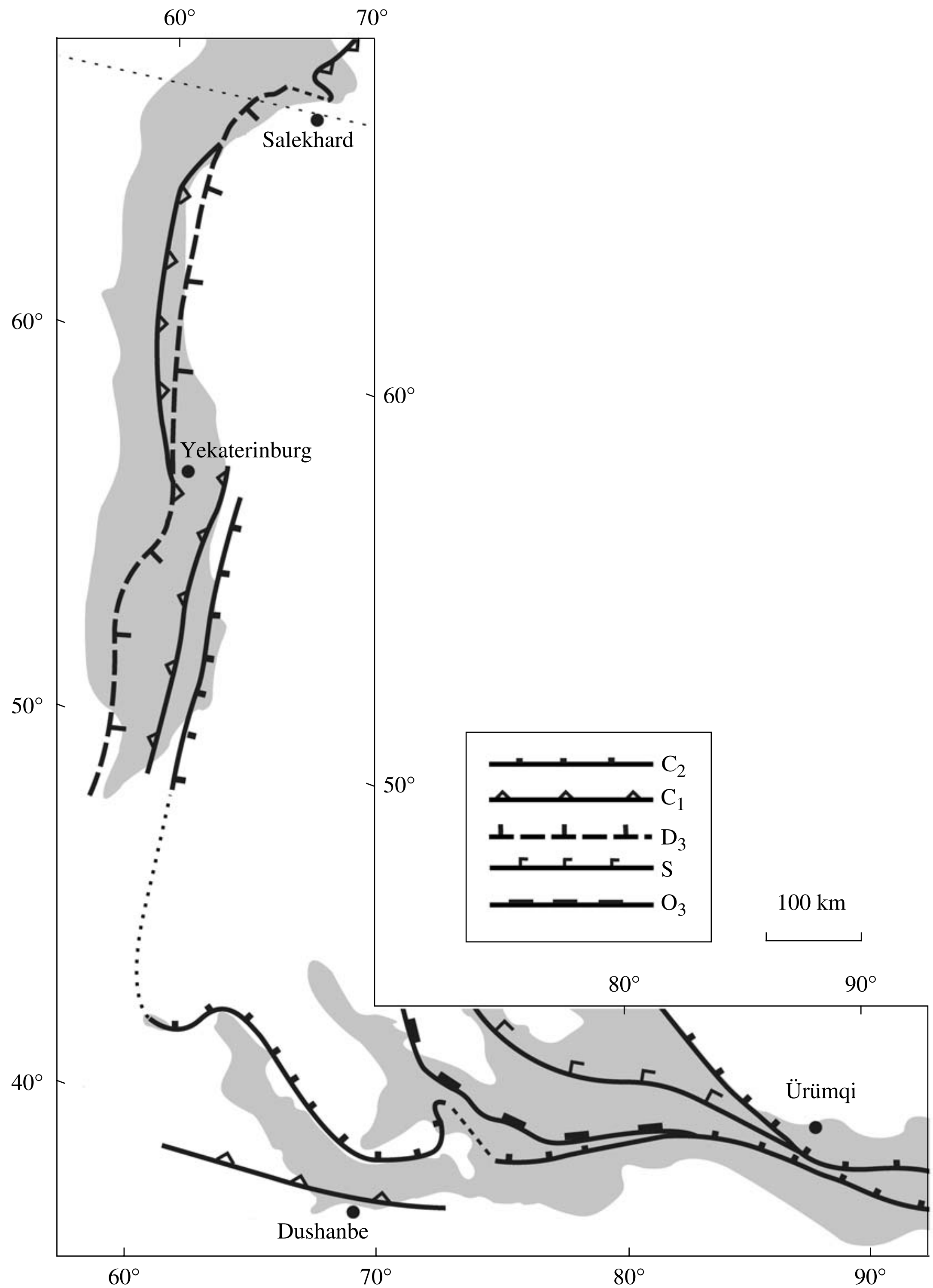


Fig. 1. Paleozoides of the Urals and the Tien Shan. The territory of the exposed Paleozoic and older rocks is toned. The sutures of the Paleozoic oceanic basins are denoted by lines; ticks indicate the age of suture and polarity of subduction of oceanic crust during closure of oceanic basins.

Ordovician and the Ordovician–Permian stages. The rocks formed during the first stage are termed as Protouralides and the younger rocks as Uralides. Both will be considered first for the western slope of the Urals and then for the eastern slope, to the east of the Main Ural Fault.

Protouralides

The Ural oceanic domain existed as early as in the Neoproterozoic [49, 78]. The relationship between the Baltic continental margin and the oceanic domain during the pre-Ordovician stage was different than in the Paleozoic (after the Cambrian–Ordovician boundary). As in the Uralides, the oceanic domain of the Protouralides consisted of various basins with oceanic crust, island arcs, and microcontinents. Later on, as a result of tectonic accretion, some of these structural elements were incorporated into the paleocontinental sector of the Uralides on the western slope of the Urals.

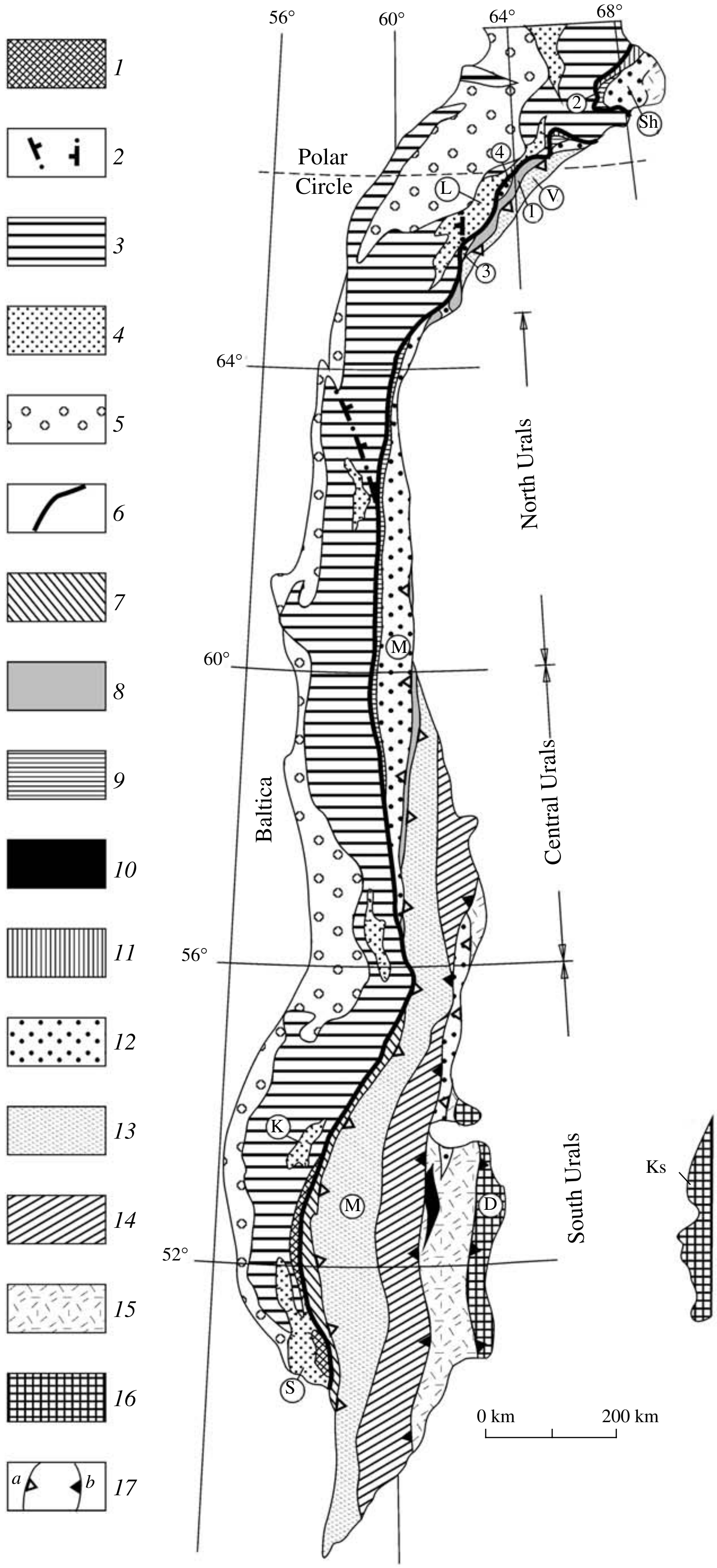
In the Riphean (Mesoproterozoic and Early Neoproterozoic), terrigenous and carbonate rocks with subordinate rift-related volcanics were deposited at the Baltic margin. The extended Late Riphean shelf of the Subpolar region spread toward the Timan. The passive evolution of the continental margin in the Polar and Subpolar Urals ended in the terminal Late Riphean, having given way to the Andean-type marginal volcanic–plutonic belt above the subduction zone plunging beneath Baltica [37]. As follows from the widespread suprasubduction complexes (Fig. 2), this zone was gently dipping. The volcanic rocks comprise calc-alkaline and subalkaline basalt, andesite, dacite, and rhyolite (the Man'ya, Sablya Mount, and Molyudvozh formations) [3, 59]. The Pb/Pb age of the intermediate rocks is 695 ± 19 Ma; the Rb–Sr age of the felsic rocks is 586 ± 21 and 535 ± 10 Ma. The Pb/Pb and U–Pb ages of the granitic rocks, diorites, and gabbros associated with the volcanics range from 632 ± 7 to 515 ± 8 Ma [59]. The volcanic molasse (Laptopai Formation) was deposited within the marginal belt in the Vendian (Ediacarian) and likely in the Early Cambrian [3].

A complexly built island arc and a microcontinental block (Paleoproterozoic Kharbei metamorphic complex [39] and the Middle–Upper Riphean (Ectasian and Cryogenian) Nyarovei volcanosedimentary group [60]) existed at that time in the marginal part of the Polar Ural oceanic basin. The island-arc complexes include the Bedamel basalt–andesite–dacite formation and the Lyadgei basaltic andesite–rhyolite formation (the U–Pb age of the rhyolite is 555–547 Ma [69]). The U–Pb age of plagiogranite from ophiolitic melange at the base of this formation is 670 ± 5 Ma [67]. In the Late Vendian (Edicarian) and Early Cambrian, the intrabasinal arc and the Polar Ural microcontinent (Fig. 3, notation PU) collided with Baltica and blocked up the subduction zone plunging beneath this microcontinent. Their attachment was accompanied by orogeny and emplacement of syncollision granites [65], increasing the initial area of the continental margin substantially. As a result, a new continent–ocean boundary was formed and its configuration changed. The Main Ural Fault marked this boundary. In the late Middle Paleozoic, the rock complexes of the paleooceanic sector of the Uralides were thrust along this fault over the enlarged margin of Baltica (Fig. 2).

The passive, rift-type continental margin existed in the Neoproterozoic in the south of the North Urals and in the Central Urals. The mainly subalkaline igneous rocks occurring here were derived from sources that formed at different depths above a mantle diapir [19, 30]. The sedimentary section was characterized by the Lower Vendian marine and glacial rocks deposited in graben-like depressions on the shelf and continental slope [27].

As in the Polar Urals, the South Ural margin of Baltica bore a convergent character. The convergent geodynamic setting originated here later than in the polar segment and was manifested in another way. The island-arc system, which arose in the south along the paleocontinental margin approximately at the Riphean–Vendian (Cryogenian–Edicarian) boundary was related to the steep subduction zone dipping, as in the north, toward Baltica [53, 55]. The intermediate and felsic volcanics along with younger alkali basalt (Lushnikovka Complex) occurred in a relatively small ensialic island arc in the Uraltau block (Fig. 3, notation U). The U–Pb age of the subvolcanic quartz diorite that crystallized at the final stage of the formation of the older group is 590 ± 4 Ma [55]. The rocks of the Lushnikovka Complex are cut through by a plagiogranitic pluton; the intrusive rocks occur as pebbles in the Tremadocian beds. To the north of the Lushnikovka Complex, small granitic bodies are dated at 543 ± 4.6 Ma (U–Pb, SHRIMP) [73]. To the east (in present-day coordinates), packets of imbricate tectonic sheets occur, which are interpreted as a forearc accretionary wedge (East Ebeta and Maksyutovo complexes) neighboring upon the remnants of volcanic arc. Fragments of the ophiolitic section of the Protoural oceanic basin and sedimentary cover of the microcontinent are numerous in these tectonic packets [53].

The Uraltau island arc was separated from the margin of Baltica by a backarc basin with the oceanic crust. Accretion at the end of the Cadomian tectonic epoch led to the termination of subduction and the disappearance of the backarc basin but did not change drastically the structural grain of the continental margin, which was enlarged at the expense of island-arc complexes, disintegrated and metamorphosed to various degrees, that were distributed uniformly along the strike of the continental margin. The findings of Cr-spinel grains in the Tremadocian sandstone from the Sakmara Allochton [17] are indirect evidence for melanocratic basement of the pre-Ordovician backarc basin. In addition, the serpenitite melange and olistostrome of this allochthon contain orthoamphibolite blocks spatially associated with tectonic lenses of crystalline schists, which probably formed in the Late Vendian (Edicarian) as products of metamorphism of arkose, graywacke, clayey siderite rock, and evaporites [43, 44]. These metasedimentary rocks may be regarded as fragments of the cover in the former backarc basin.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In the adjacent areas of the South Ural shelf of Baltica, the sedimentation in the Late Riphean and Early Vendian (Cryogenian–Edicarian) was accompanied by eruption of subalkali basalts (Arsha Formation). The tillite-like conglomerate (diamictite) close in composition to the coeval rocks of the Central Urals also testifies to the Early Vendian rifting at the shelf. In the Late Vendian (Edicarian), the sedimentation conditions changed. As a result of the Cadomian accretion, an eastern provenance appears, and the terrigenous polymictic sequence traditionally classified as molasse was deposited [35].

The island arcs and microcontinental blocks were accreted to the ancient margin of Baltica along almost its entire extent, giving rise to local orogenic uplifting and metamorphism. Glaucophane fragments were found in the Tremadocian sandstone of the Sakmara Allochthon [17]; the early generation of eclogites exposed near the Maksyutovo high-pressure complex in the Uraltau Zone has a U–Pb age of 547 ± 40 Ma [23]. Metamorphism of the Beloretsk high-temperature eclogite-bearing complex in the north of the South Urals is dated at 550 ± 5 Ma with the Ar/Ar method [76]. The glaucophane-schist complex at the Central and South Urals boundary is somewhat younger (535–539 ± 7 Ma, Rb–Sr method [32]). The greenschist metamorphism is coeval; its isogrades trend here in the northwestern Timan direction rather than in the meridional Ural direction [42].

On the eastern slope of the Urals, the rocks of the Protoural ophiolitic association occur immediately to the east of the Main Ural Fault. The Sm–Nd age of the ultramafic rocks of the Khadata (Syumkeu) massif in the Polar Urals (Fig. 2, notation 2) is 604 ± 39 Ma [10]. In the Voikar–Syn'ya massif, the U–Pb age of upper mantle activity in the ophiolitic complex is 585 ± 6 Ma [47]. In the Central Urals, the Silurian island-arc volcanic rocks contain xenogenic zircon grains dated at 990 Ma with the Pb/Pb method [38]; they were probably captured from the upper mantle mafic material as a result of subduction of ancient oceanic crust. In the north of the South Urals, metabasalt and plagiogranite with oceanic geochemical signatures occur as particular sheets in the Il'menogorsky and Sysert metamorphic complexes near the Main Ural Fault; their U–Pb ages are 643 ± 46 and 576 ± 65 Ma, respectively [22]. The Sm–Nd age of the peridotite in the fault-line Mindyak massif situated at 54° N is 882 ± 83 Ma, while the Re–Os age of the associated gabbro is 804 ± 37 Ma [79]. To the south, a metaterrigenous sequence with remains of the Late Vendian flora overlaps with scouring the altered ultramafic rocks of the ophiolitic association [16].

Uralides

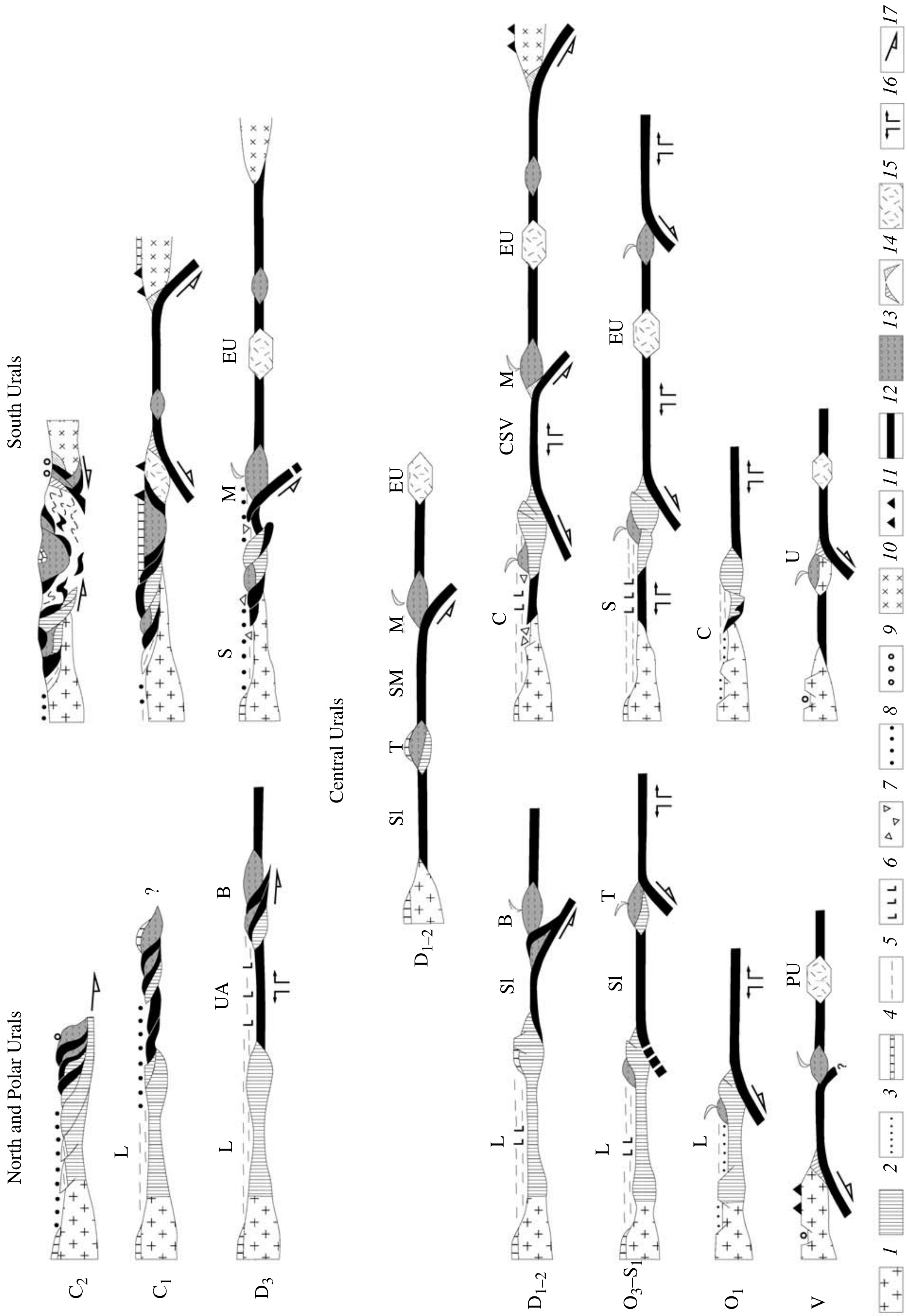
In the terminal Cambrian, a new stage of the tectonic evolution of the Urals started with structural rearrangement (mainly in the north), complication of the continent–ocean interface, and destruction of the continental margin formed during the Cadomian epoch.

The paleocontinental and paleooceanic sectors of the Uralides corresponding to the western and eastern slopes of the Urals are separated by the Main Ural Fault. Farther to the east, the marginal jut of the Kazakh–Kyrgyz continental massif crops out in the widest southern part of the region. The Late Paleozoic Ural Orogen is a result of collision between Baltica and this continent formed in the Silurian.

Paleocontinental sector. Marginal basins arose at the Paleozoic margin of Baltica. The fragments of their sequences make up a chain of allochthons along the western slope of the Urals. The sections of the South Ural and Polar Ural basins in the Sakmara and Lemva allochthons (Figs. 2, 3, notations S and L) are the most representative. The basins started to develop at the Cambrian–Ordovician boundary as rift basins [40, 64], which were filled with shallow-water graben facies locally associated with bimodal volcanics (Kidryasovo and Kuagach formations (Figs. 2, 3, notation S); Pogurei, Kokpel, and Manitanyrd formations (Figs. 2, 3, notation L)). The expansion of the sedimentation area in the Arenigian began in both the south and the north with deposition of variegated silty and clayey sediments (the Kuragan Formation in the Sakmara Zone and the Grubeyu Formation in the Lemva Zone).

The initial basins were separated by marginal uplifts from the oceanic domain, which continued to evolve. The uplifts were composed of Protouralide complexes [54, 56, 64], which accreted to Baltica in the preceding tectonic epoch. From the Middle Ordovician up to the Late Devonian, both basins were filled with deepwater, largely cherty sediments (the Novokursky, Sakmara, and Kyzykflot formations in the Sakmara Zone; the

Fig. 2. The major paleotectonic elements and tectonic zones of the Urals. (*1, 2*) Protouralides of the western slope: (*1*) Vendian–Early Cambrian island arc and accretionary wedge of the Uraltau Zone, (*2*) boundary of the Late Riphean–Early Cambrian (?) marginal volcanic–plutonic belt; (*3–5*) paleocontinental sector of the Uralides: (*3*) Protouralides and Uralides (shelf and flysch complexes, unspecified), (*4*) allochthons with the Early and Middle Paleozoic marginal-sea complexes, (*5*) Permian molasse of the Ural Foredeep; (*6*) Main Ural Fault; (*7–15*) paleooceanic sector of the Uralides: (*7–11*) oceanic sutures and ophiolitic allochthons of the (*7*) Cis-Sakmara–Voznesenka, (*8*) Serov–Mauk, (*9*) Salatim, (*10*) Transural, and (*11*) Ural–Arctic basins; (*12, 13*) island-arc systems: (*12*) Silurian and (*13*) Devonian; (*14*) East Ural Zone of continental microterranes; (*15*) Transural Zone of melange; (*16*) Kazakhstanides (notations in figure): D, Denisovka Zone; Ks, a fragment of the Kokshetau Block with ancient sialic crust; (*17*) frontal zones of the Middle Paleozoic island-arc systems (*a*) and marginal volcanic–plutonic belts (*b*). Notations in circles: allochthons: S, Sakmara; K, Kraka, L, Lemva; zones: M, Magnitogorsk; T, Tagil; V Voikar; Sh, Shchuch'ya; mafic and ultramafic massifs and blocks: 1, Voikar Syn'ya; 2, Khadata; 3, Khulga; 4, Khord'yu.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Kachamyl'k, Kharota, Pagina formations and Chernogorsky Group in the Lemva Zone), and a considerable amount of basalts erupted in their central parts (Sugraly Complex in the Sakmara Zone and the Grubeshor Group, or Lagorta Complex in the Lemva Zone). The ocean-type crust [14] with a complete section of ophiolites was formed in the Sakmara Basin as early as in the beginning of the Middle Ordovician, whereas such a crust did not exist in the Lemva Basin, where gabbro–ultramafic complex of the ophiolitic association is absent and the Middle Ordovician basalts rest on the Lower Ordovician terrigenous sequence [64]. Another difference is the asynchronous development of the basins in the backarc regime, where they were rimmed in the east (in present-day coordinates) by island-arc volcanics, which were located along the western edge of marginal uplifts of the Protouralides [54]. Subduction occurred on opposite sides of these uplifts and was directed beneath Baltica. The volcanic arc conjugated with the Sakmata Basin that existed during the Middle Ordovician–Middle Devonian (the Guberlya, Baulus, Blyava, Kosistek formations and their analogues). In the Polar Urals, a similar arc did not shift and functioned in the Arenigian–Ashgillian (Igyadei Complex). The geodynamic setting in the Lemva Basin did not change from the Arenigian to the Tournaisian inclusive, when the deposition of bathyal clayey and cherty sediments ended (Fig. 3). In the Sakmara Basin, the tectonic stacking started as early as in the Early Devonian [40]. This process brought about the obduction of ophiolites on the outer Uraltau marginal uplift and the complication of its internal structure with the formation of the high-pressure Maksyutovo Complex in the beginning of Late Devonian. The rise of intrabasinal uplifts led to their destruction and deposition of mixtite–olistostrome units and sequences [41, 48]. In the Late Devonian, the Sakmara Basin was closed, and only a small deepwater trough was left at the rise of the continental slope of Baltica.

The sequences of the Paleozoic shelf of Baltica consist of shallow-water carbonate and terrigenous–carbonate sedimentary rocks deposited from the Middle Ordovician (the Early Ordovician in the extreme north) and up to the formation of collision orogen. During this time interval, the sedimentation shifted inland, from the Devonian–Carboniferous boundary in the South Urals and from the mid-Early Carboniferous in the Polar Urals. The transverse zoning of the shelf domain changed with time. In the Polar Urals, the following series of structural–facies zones may be outlined for the stratigraphic interval of the Silurian to the Middle Devonian: carbonate platform cover–relatively deepwater clayey chert, fine platy limestone, and marlstone of inner shelf (depression zone)–barrier reef–chert and shale, pelagic loope limestone, calcarenite, and calcitutite of the outer shelf and continental slope–the Lemva deepwater basin–shallow-water terrigenous–shelf-type carbonate section (Paipudyn Formation) of the marginal uplift that separated the Lemva Basin from oceanic domain [64].

Paleooceanic sector. The destruction of the Baltic margin in the Paleozoic was combined with the generation of a new oceanic crust multiply formed due to spreading in the adjacent domain. The upper members of ophiolitic associations—complex of parallel dolerite dikes and comagmatic mafic lava—are indicators of this process. The oldest Paleozoic dike complex was established in the northeastern Voikar–Syn'ya massif (Fig. 2, notation 1). The dikes were intruded in the terminal Cambrian 490 ± 7 Ma ago [68]. The Ordovician oceanic basalts are widespread in the South Urals [3]. In the Cis-Sakmara–Voznesenka suture zone of the paleoocean, basalts with various geochemical signatures were formed from the Arenigian to the Emsian (the Polyakovka and Dergaish formations, the Aratau Complex). Diverse cherty and less abundant terrigenous rocks were deposited from the middle Llanvirnian to the Frasnian (the Sakmara, Mazovo, Turata, and Mukasovo formations) [41, 50].

The basalt–sediment interface, sliding from the Llandeilian to the Emsian, was most likely related to the long-working dispersed spreading in the western South Ural oceanic domain.

In the Central and North Urals, the oceanic domain increased due to the opening of the Salatim marginal basin (Fig. 3, notation Sl) at the Middle–Late Ordovician boundary [57]; likely, this was one event. After the eruption of mafic lavas from the Late Ordovician to the Early Devonian (Lochkovian), finely intercalating carbonaceous siltstone, claystone, and cherty shale were deposited [31]. In the Polar Urals, the Ural–Arctic Basin (Fig. 3, notation UA) that opened in the early Late Devonian evolved in line with the same scenario. The Frasnian sequence of tholeiitic basalts and dolerites was built up here by tephroturbidite, shale, carbonaceous cherty shale, phtanite, and limestone with Famennian and Tournaisian conodonts [64]. The origination of the Ural–Arctic Basin led to the expansion of the deepwater sedimentation in the paleocontinental sector due to the subsidence of marginal uplift of the

Fig. 3. Tectonic evolution of the Paleozoides in the Urals. (*1, 2*) Baltic paleocontinent: (*1*) pre-Vendian crust, (*2*) margin accreted in the Late Cadomian epoch; (*3–9*) marginal continental complexes: (*3*) rift-related, (*4*) shelf, (*5*) slope and bathyal basinal, (*6*) basaltic, (*7*) olistostrome, (*8*) graywacke and polymictic flysch, (*9*) molasse; (*10*) Kazakh–Kyrgyz continental massif; (*11*) marginal volcanic belt; (*12*) oceanic crust; (*13*) island arc, (*14*) accretionary wedge; (*15*) microcontinent; (*16*) spreading zone; (*17*) direction of subduction and underthrusting; (V) Vendian (Ediacarian). Notations in figure. Marginal continental basins: L, Lemva; S, Sakmara; basins of the Ural paleoocean: CSV, Cis-Sakmara–Voznesenka, Sl, Salatim; SM, Serov–Mauk; UA, Ural–Arctic; island arcs: V, Voikar; M, Magnitogorsk, T, Tagil; U, Uraltau; microcontinents: EU, East Ural; PU, Polar Ural.

Protouralides, where the Lemva-type bathyal sediments were deposited. The relics of younger paleooceanic crust have been retained in the east of the South Ural (in the Transural Zone (Fig. 2)), where they are composed of pillow lavas and hyaloclastites with xenoliths of the upper Visean–Serpukhovian limestone and associated gabbroic rocks. These rocks are compared with the complexes of present-day intraoceanic rises with a rather thick crust [61].

The East Ural microcontinent. Fragments of the ancient continental crust in the form of gneissic and granite-gneissic complexes occur in the paleooceanic sector of the South and Central Urals, largely in the East Ural Zone. As follows from the zircon geochronology, these complexes are Paleo- and Mezoproterozoic in age [29]. In the Neoproterozoic (Late Riphean–Vendian), they probably made up a continuous microcontinent covered by Vendian quartz and polymictic clastic sediments. The terrigenous–carbonate cover was formed in the Arenigian–Late Devonian; the Silurian–Upper Devonian cherty shales occurred sporadically [35]. The locally developed basalts belonging to various petrochemical series, including subalkali basalts, that erupted in the Early and Middle Ordovician [46], mark the activation of magmatism in the adjacent oceanic domain. In the Middle Ordovician, the distance from the East Ural microcontinent to the nearest edge of Baltica was more than 750 km along a paleomeridian [5].

Island arcs were formed in the Middle Paleozoic in the marginal portion of the Ural ocean that adjoined Baltica. The oldest Tagil arc in the Central and North Urals (Figs. 2, 3, notation T) started to form in the Late Ordovician. This arc underwent all stages of evolution from initial to mature with final subalkaline magmatism in the Pridolian–Lochkovian [58]. At the late stage, the volcanic rocks were locally replaced with limestone bioherms and reefs. In the Early and Middle Devonian, the extinct arc was built up by carbonate islands with bauxite deposits. The Tagil paleoarc was magmatically active during approximately 35 Ma and located above the subduction zone plunging toward the paleocontinent, relatively close (500–1000 km) to the outer edge of the Baltic shelf [33, 51]. Such a position ruled out its collision with Baltica.

The composite dunite–clinopyroxenite–gabbro massifs of the Platinum belt and the related plagiogranite which occurs at the base of the back portion of the Tagil island-arc system mark the completion of the evolution of these belts at the Silurian–Devonian boundary. The U–Pb age of the plagiogranite of one of the central plutons is 415 ± 10 and 416.6 ± 1.6 Ma; dates of 419 ± 12 Ma (Sm–Nd method) and 428 ± 7 Ma (U–Pb method) have been published for gabbro from the other two plutons [13]. In addition, some clinopyroxenites have been dated at 441 ± 27 Ma [34]. The gabbroic rocks from a group of closely spaced plutons in the south of the Platinum belt contain zircons within a chronological interval from 422 ± 11 to 462 ± 15 Ma and xenogenic zircon crystals that yielded Proterozoic dates from 1200 to 2200 Ma [21, 66]. In the opinion of the authors who published these data, the older dates indicate that a block of ancient continental crust was a source of Proterozoic zircons. If this was the case, such a block probably was detached from the Baltic margin in the Late Ordovician or somewhat earlier as a result of opening of the Salatim Basin. Before this opening, Vendian rocks of the subcontinental nature existed at the Baltica margin. These rocks were subsequently incorporated into the complexly built central plutons of the Platinum belt. In particular, these are ultramafic rocks and olivine gabbro dated at 551 ± 32 and 561 ± 28 Ma [26, 34]. In the South Ural, the Tagil arc pinches out without indications of a continental block therein [51].

In the Polar Urals, fragments of this arc are exposed in the northern Voikar Zone and more prominently in the Shchuch'ya Zone, composed of Silurian and calc-alkaline volcanics and Lower–Middle Devonian bauxite-bearing limestone; high-Sr gabbro of the Maslovo Complex are noted [12]. As in the north of the Central Urals, the sections are reduced here; in particular, the final subalkaline volcanic complex is not developed [18]. To the south of the Polar Circle, the volcanic rocks belonging to the Tagil arc are not exposed. Only inliers of its ensimatic granulite–metabasic basement with deep-seated analogues of gabbro from Platinum belt (Khord'yu and Khulga blocks) occur here [12, 45]. In the Khulga block (Fig. 2, notation 3), the U–Pb age of the protolith is 578 ± 11 Ma [77]. When a perioceanic arc appeared in the Silurian, the synchronous suprasubduction volcanism in the outer zone of the marginal continental Lemva Basin ceased (Fig. 3).

In addition to the Tagil arc, another arc, intraoceanic in nature, originated at the Ordovician–Silurian boundary and functioned up to the Early Devonian. The fragments of this arc extend along the boundary of the East Ural and Transural zones of the South Ural. In age, composition, and general trend of evolution, the section of this arc is similar to the Tagil section [72]. The subduction zone probably had the same polarity.

The end of the Early Devonian was characterized by great structural rearrangement of the Ural active margin: the overall Ural island arc system related to subduction in the direction opposite to Baltica originated between Baltica and the East Ural microcontinent. This system is most representative in the south, i.e., in the Magnitogorsk Zone (Figs. 2, 3, notation M). The Magnitogorsk arc started to evolve on the oceanic crust and was active during 45 Ma from the Emsian to the Famennian inclusive. The volcanic axis shifted eastward with time in the same direction as the subduction zone plunged, having a variable dip angle [37]. The igneous series evolved from tholeiitic, with boninites at the base in the western frontal part, to calc-alkaline and then to subalkaline and alkaline in the Famennian. This general sequence was locally complicated by backarc and intra-arc spreading (Aktogai dike complex, Emsian–Givetian Mugodzhary and Kurkuduk formations, and Eifelian Karamalytash Formation). The age of the island-arc intrusive rocks determined with the U–Pb and Pb/Pb methods varies from 393 ± 6 Ma (tonalite) to 368 ± 7 or 352 ± 7 Ma (gabbro, diorite, granodiorite, and granite) [66]. Large reefs are not characteristic of the Magnitogorsk arc. Its continuation in the Central Urals is distinguished by reduced magmatic activity in the Emsian–Frasnian. In the eastern, extinct Tagil arc, on the opposite side of the Serov–Mauk interarc basin with oceanic crust, a calc-alkaline volcanic–plutonic association and an upper association of elevated alkalinity are known [58, 71]; their geodynamic setting is a matter of debate. In the North Urals, the Devonian arc is buried beneath the cover of the West Siberian Plate. The flank of this arc crops out in the Voikar Zone of the Polar Urals. In this segment, the volcanic activity proceeded from the Late Silurian (?) to the Middle Devonian and is completed by shoshonite-like rocks [64]. The specific feature of the Voikar segment is the suprasubduction granodiorite–tonalite batholith dated with the Rb–Sr method at 400 ± 10, 399 ± 24, and 385 ± 4 Ma [2]. To the north, in the Shchuch'ya Zone, small coeval granitoid plutons cut through the Silurian–Middle Devonian Tagil island-arc complexes [2]. Outcrops of Middle–Late Devonian calc-alkaline and subalkaline volcanic flows and subvolcanic intrusions (Yenzor and Tal'bei complexes) are known here as well [69]. All this testifies to the degeneration of the Voikar arc in the northern direction and its juxtaposition with the Silurian arc of the Shchuch'ya Zone.

Arc–continent collision. Different segments of the Devonian island-arc systems arose at a variable distance from the edge of Baltica but everywhere above the subduction zone that plunged away from the paleocontinent. This difference predetermined the different duration of their convergence and asynchronous collision, as well as the dissimilar character of attachment (accretion) of the suprasubduction complexes to the continental plate. Apparently, the Magnitogorsk arc in the South Ural was the nearest to the paleocontinent [6]. Its convergence with Baltica began in the Middle Devonian after the termination of spreading in the Cis-Sakmara–Voznesenka basin (Fig. 3, notation CSV). Collision with the subsided continental margin occurred in the Famennian. The polymictic graywacke flysch of the Zilair Group and replacing olistostromes appeared in front of the arc and at the Baltic margin on the place of the closed Sakmara Basin. The clastic material accumulated owing to the erosion of the island arc, salients of the forearc accretionary wedge consisting of fragments of sections pertaining to the Cis-Sakmara–Voznesenka oceanic basin, and the marginal Ulutau uplift with high-pressure metamorphic rocks [35, 48]. The collision resulted in blockage of the subduction zone and cessation of related volcanic activity. The westward obduction of forearc ophiolites dismembered into a series of sheets and the formation of nappe packets composed of rocks belonging to the Sakmara basin started at the Devonian–Carboniferous boundary. The Baltic margin grew owing to the attachment of the extinct Devonian island arc and the East Ural microcontinent (Fig. 3, notation EU) behind it. In the middle Tournaisian, subduction was resumed along the oceanic margin of this microcontinent, beneath which the jumped subduction zone began to plunge [35, 52]. At the same time, the suprasubduction magmatism became marginal continental in its geodynamic setting.

At the boundary of the South and Middle Urals, the Magnitogorsk arc collided first with the southern end of the Tagil arc. As a result, the imbricate sections of the dividing basin were thrust over the Tagil complexes as early as in the Frasnian Age [51]. The volcanic activity of the Magnitogorsk arc waned in this region by the Famennian. A new stage of tectonic stacking along the marginal jut of Baltica ended in the mid-Early Carboniferous. The collision boundary with island-arc complexes—melange of the Main Ural Fault Zone—was sealed by deformed gabbroic rocks and granodiorite dated at 334 ± 4–5 Ma with the Pb/Pb method for zircon; the cutting through massive granite is 327 ± 4 Ma in age [28].

Over most part of the Central and Northern Urals, the accretion of island-arc systems developed without obduction (a mild scenario). The Devonian arc, having sharply deviated eastward, was the most distant from the Silurian Tagil arc and correspondingly from the Baltic paleocontinent. The subduction beneath this arc, i.e., toward the paleoocean, gave rise to disintegration and consumption of the Serov–Mauk interarc basin (Fig. 3, notation SM), which transformed into the tectonic suture (Fig. 2). Afterward, the subduction zone migrated westward, beyond the inactive Tagil arc, and Baltica began to approach the double island-arc assembly with the Tagil complexes at its front. The onset of such convergence is confirmed indirectly by the Frasnian olistostrome at the boundary between Baltica and the Salatim basin [31]. As a result of accretion, a narrow shear zone affected by greenschist and blueschist metamorphism was formed on the place of the Salatim marginal basin. This metamorphism is dated at 370 ± 35 Ma with the Sm–Nd method [32].

In the Polar Urals, the arc collided with the continental margin according to another scenario. The Voikar terminal segment of the Devonian island-arc system, in contrast to its southern part, was formed somewhat earlier and much closer to the Tagil arc. The latter was involved into the Devonian subduction zone and incorporated into the basement of the Voikar arc (Fig. 3, notation V) as early as at the Early Devonian stage of its formation [45]. A large sheetlike intrusive body of moderately silicic granitoids was probably formed at the base of the Devonian section. With onset of collision at the Middle–Late Devonian boundary, the island arc was involved into intense erosion. After almost complete consumption as a result of ongoing subduction of the Salatim basin, large-scale obduction took place, and the allochthonous mafic–ultramafic complex, which previously had been a basement of interarc trough at the extension of the Serov–Mauk basin (Figs. 2, 3), appeared at the front of obduction. The largest Voikar–Syn'ya ophiolitic nappe, with fragments of the basement of the Tagil arc soldered to its bottom (the Khulga and Khord'yu blocks (Fig. 2, notations 3 and 4)), thin lenticular sheets of the Salatim Shear Zone, and melange were thrust over the margin of Baltica. The complexes which underlay the ophiolites underwent high-pressure metamorphism of various grades [45]. At the boundary between the Polar and North Urals, the high-pressure metamorphic rocks (Nerkayu Complex) are dated at 352 ± 3.6 Ma (early Tournaisian) with the Ar/Ar method [15].

The Voikar arc degenerated in the northern direction. In the Shchuch'ya Zone it was poorly active and juxtaposed with a flank of the Tagil arc. The Khadata ophiolitic allochthon (Fig. 2, notation 2) exposed to the west is a composite structural element consisting of fragments of mafic–ultramafic basement of the Tagil arc and the bottom of the adjacent basin. In this district and somewhat to the south, the tectonic stacking was interrupted by the formation of the Ural–Arctic oceanic basin in the Frasnian–Tournaisian at the junction of the margin of Baltica and the Middle Paleozoic accretionary system (Fig. 3). A new stage of compression along the edge of Baltica and the final collision with complexes belonging to the paleooceanic sector started here in the Visean. As a result, the relatively small Ural–Arctic basin was crushed, and graywacke flysch (the Rai-Iz Formation) was deposited in front of the overthrust ophiolitic masses [64].

In general, the considered part of the Ordovician and Silurian Urals resembles the Melanesian region of conjugation of the Australian and Pacific plates, where various marginal basins were opened at different times and island arcs distinct in polarity arose repeatedly. A relatively extended and very tortuous garland of island arcs is situated above the subduction zones plunging toward the ocean.

Marginal volcanic–plutonic belt. In the Early Carboniferous, the Magnitogorsk island-arc system together with the East Ural microcontinent were attached to Baltica, and a westward verging subduction zone originated along a new accretionary boundary (Fig. 3). Calc-alkaline, alkaline, and bimodal subalkaline volcanic series of the marginal continental type were formed above this zone from the late Tournaisian to the late Visean [61]. Comagmatic gabbro–diorite–granite, granosyenite, and tonalite–plagiogranite intrusions were emplaced at the same time. Their Rb–Sr age varies from 346 ± 1 to 330 ± 4 Ma [35, 62]. Similar dates were obtained with the U–Pb method [66].

Early Carboniferous volcanic complexes are widespread in the east of the Magnitogorsk Zone and locally develop in the East Ural Zone. Almost all of these complexes are bimodal in composition; intermediate rocks are much less abundant. To the west of the volcanic area, terrigenous, locally coal-bearing and carbonate (in the upper part of the section) rocks were deposited in the shallow-water setting of the open shelf. A deep-water trough inherited from the Late Devonian was located to the west of the Uraltau Zone, where deposition of flysch alternated with formation of pelagic limestone and cherty rocks [35]. On the east side of the volcanic belt, an accretionary wedge existed. The retained fragments of this wedge comprise intercalating tectonic lenses of the Riphean–Middle Ordovician sedimentary rocks deposited at the margin of the microcontinent, Lower Carboniferous limestone (not abundant), various clastic rocks with carbonaceous and tuffaceous interlayers, and serpentinite and diverse schists as products of dynamometamorphism [62]. The remnants of the Silurian intraoceanic island arc have been attached to the accretionary wedge (Figs. 2, 3). The relics of oceanic – mafic volcanics, the Ordovician cherty anf fine clastic sediments, the Lower Silurian black shale, the Upper Devonian chert and cherty tuffite, the post-Famennian sandshale members with olistostromes [11, 62] – occur in the Transural Zone along with the aforementioned Late Visean–Serpukhovian aceanic mafic complex. Small outcrops of serpentinite are abundant.

Igneous rocks are replaced with terrigenous–carbonate sequences both across and along the strike of the volcanic–plutonic belt. Volcanic eruptions often occurred in grabens, so that the thickness of the adjacent sections turns out to be sharply different. The temporal limits of volcanic activity in various localities are variable as well. In general, the onset of volcanic activity becomes younger in the eastern direction. The termination of this process opened the way for deposition of subplatform limestone up to the Late Carboniferous [61]. It is believed that the Early Carboniferous volcanic complexes with mixed geochemical attributes of suprasubduction and intraplate igneous rocks were formed at the active margin resembling the Californian margin of North America. However, the Ural marginal volcanic–plutonic complex also resembles the Cenozoic volcanic belt of Kamchatka, where the igneous rocks become younger toward the Pacific Ocean and where the typical suprasubduction rocks often associate with the volcanics having intraplate geochemical signatures [1].

The Ural Margin of the Kazakh–Kyrgyz Paleocontinent

This large composite continental massif underwent long-term and intricate history of its evolution completed in the Devonian.

Its western boundary is traced from the South Urals (Fig. 2) beneath the Mesozoic–Cenozoic sedimentary cover of the West Siberian Plate up to intersection of 66° E and 60° N and further northward [24]. The subduction of the Middle Paleozoic Ural oceanic crust beneath the newly formed Kazakh–Kyrgyz continent gave rise to the formation of two volcanic belts of different ages along its margin. The igneous rocks of the Early–Middle Devonian continental belt occupy a large territory and pertain to the calc-alkaline bimodal series. In the South Ural, they are exposed extremely poorly and were studied mainly from cores of numerous boreholes in the Torghay Trough and at the margin of the West Siberian lowland. The rocks become more alkaline and enriched in potassium toward Kazakhstan [24]. The accretionary complex in the front of the area of Devonian volcanism consists of tectonic blocks and sheets of various sizes composed of serpentinites and fragments of pre-Devonian sequences; variegeted, mainly coarse-clastic fossiliferous Emsian–Eifelian sedimentary rocks with a tuffaceous admixture [25]; coeval polymictic sandstone and shale with limestone interbeds; and Middle Devonian reef limestone [35].

Another marginal continental belt was formed under subaqueous conditions in the Early Carboniferous. This belt was narrower that its Devonian counterpart and situated closer to the edge of the Kazakh–Kyrgyz paleocontinent above a probably steeper subduction zone. Volcanism migrated with time eastward. The belt is composed of two spatially separated calc-alkaline complexes. The western, Aleksandrovka Complex that built up the Devonian accretionary wedge consists of basalt, basaltic andesite, andesite, less abundant felsic rocks, and coarse tuffs; andesite is a predominant rock. Limestone interbeds contain middle and late Visean brachyopods. The eastern, Valer'yanovka Complex differs in having a greater thickness and consists of middle–upper Visean and Serpukhovian–Bashkirian sequences. Claystone, siltstone, tuffite, and calcareous sandstone lie at the base of the older sequence. Upsection, they give way to basaltic andesite with thin interbeds of calcareous tuffite and limestone. Further, the section is built on by basalt, basaltic andesite, andesite, tuff, tuffite, and tuffstone. This sequence is abruptly replaced eastward by a tuffaceous–terrigenous–carbonate association of sedimentary rocks. The younger sequence is composed of basalt, basaltic andesite, andesite (lavas and tuffs), red beds, calcareous sandstone, siltstone, and small bodies of organogenic clastic limestone. In general, leucocratic plagiophyric and plagioclase–pyroxene basalts and basaltic andesites are predominant among the igneous rocks of the Valer'yanovka Comples; basic and internediate tuffs are also abundant. Gabbro, diorite, and granodiorite intrusions are related to volcanic centers. The geochemistry of the igneous rocks in the Carboniferous volcanic belt at the margin of the Kazakh–Kyrgyz paleocontinent is close to that of the reference suprasubduction volcanics of the active continental margins [61].

Collision of the Baltic and Kazakh–Kyrgyz Paleocontinents

The shortening of the eastern part of the Ural paleoocean started with the Early–Middle Devonian subduction of the oceanic crust beneath the Kazakh–Kyrgyz paleocontinent; no Devonian spreading centers are known here. Nevertheless, in the Middle Devonian, the width of the oceanic domain between the western frontal complex of the Magnitogorsk island-arc system and the accretionary wedge at the margin of the eastern paleocontinent was 2800 ± 450 km along the paleomeridian [6]. Subduction beneath the young continent ceased in the Late Devonian (Fig. 3). The consumption of oceanic crust resumed in the late Tournaisian, when a subduction zone started to operate along the new, eastern boundary of Baltica and reinforced in the middle Visean as a result of recommenced subduction on the other side of the Ural Paleoocean. At that time, the active Andean-type margin arose again along the Kazakh–Kyrgyz paleoocean, whereas the paleoocean was transformed into a residual basin bounded by converging continental masses.

At the end of the Visean, the subduction zone plunging beneath Baltica was blocked by terranes of microcontinents and island-arcs. In the terminal Bashkirian, subduction was completed in the east as well, because the oceanic crust of the residual Ural basin, except for recently arisen areas of increased thickness, disappeared beneath the Kazakh–Kyrgyz paleocontinent. As a result, the Early Carboniferous accretionary margin of Baltica collided with this paleocontinent. The Transural Zone of melange and shearing was formed along the collision suture. The belt of dislocation high-pressure metamorphism with glaucophane schists extends along this suture [20].

After the eventual closure of the Ural paleoocean in the Moscovian, the bilateral orogen started to form. This process was accompanied by extrusion and differently oriented thrusting of various lithotectonic complexes. Continental near-shore marine molasses were deposited contemporaneously in the eastern intermontane basins. In the west, a packet of deformed tectonic nappes, including such large allochthons as the Sakmara and Kraka (Fig. 2, notations S and K), propagated for a considerable distance over the Devonian margin of Baltica. Beyond the growing orogen, the foredeep inherited from the previous epoch gradually shifted toward the platform shelf. As a result, the thin deepwater cherty sedimentary rocks conformably rest upon the shallow-water carbonate facies and are, in turn, overlapped by polymictic flysch with olistostromes [35].

At the final, Permian stage of collision, likely developing in a transpressional setting [62], the orogen continued its growth and the erosion of the uplift increased. The region of emergence and erosion enlarged at the expense of the western zones, so that the Ural Foredeep migrated as before toward the platform. In the Late Permian, flysch in the foredeep was replaced by molasse. The maximum thickness of the crust beneath the orogen was reached in the East Ural Zone, where most of the continental blocks were gathered. The main granitic axis consisting of large multiphase plutons was formed here as a result of palingenesis [35]. Some of the granitic plutons were located to the west of the main axis. The generation of granites was accompanied by regional amphibolite-facies metamorphism. Two peaks of magmatic activity—290–280 Ma (the major peak) and 260–250 Ma—have been dated by various geochronological methods [66].

In the north of the Urals, collisional orogeny developed according to the same scenario. The exposed western limb of the orogen (the eastern limb is buried beneath the Mesozoic and Cenozoic sedimentary cover) has principally the same structure and geological history as in the south. In the Polar Urals, the onset of these processes was complicated by the northward pinch-out of the Ural–Arctic oceanic basin in the Frasnian–Tournaisian. After its closure, graywacke flysch was deposited in the Visean in front of the obducted complexes of the paleooceanic sector. This flysch forced out bathyal, condensed sediments that occurred in the west (Fig. 3). The graywacke sequence is built on by polymictic flysch that prograded toward the Baltic shelf in the Visean–Artinskian following the displacement of the front of tectonic stacking initiated by the growing and widening orogenic uplift. The temporal shift of terrigenous sedimentation caused sliding of the lower boundary of the flysch complex and its westward rejuvenation. In the Shchuch'ya Zone, the marine molasse appeared in the Late Carboniferous after deposition of shallow-water terrigenous–carbonate sediments in the Early Carboniferous [18, 45]. In the Ural Foredeep, flysch was replaced with coal-bearing molasse, which accumulated from the end of the Artinskian and during the Kungurian and the entire Late Permian [64].

COMPARISON OF TECTONIC PROCESSES IN THE URALS AND THE TIEN SHAN

Paleozoic Tectonic History of the Tien Shan

The tectonic history of the Tien Shan was comprehensively considered in [4, 74, 75]. The principal features of the tectonic evolution of this region are discussed below.

The Alay–Tarim and the Kazakh–Kyrgyz paleocontinental massifs of the Tien Shan have different geological records. These massifs are divided by a suture of the Turkestan oceanic basin, which existed from the Neoproterozoic to the Late Carboniferous (Figs. 4–6).

The Alay–Tarim paleocontinent was separated from Paleogondwana. In the Sinian, Cryogenian, and Edicarian, volcanic, carbonate, and clastic rocks with tilloid diamictites at the Edicarian (Vendian) level accumulated in the territory of the future Tien Shan. In the Early Paleozoic, carbonate sediments typical of the shelf of passive continental margins were deposited. Pelagic silicites and terrigenous turbidites were formed on the low-angle continental slope and rise in the Ordovician and Silurian. In the Devonian and Early Carboniferous, carbonate shelf sediments occupied a vast territory. On the continental slope, terrigenous sediments gave way to condensed cherty sediments, whose bottom was sliding up the Devonian section inland of the Alay–Tarim paleocontinent and along its slope from the west eastward (in present-day coordinates). The formation of turbitides recommenced in the Late Carboniferous. The domain of deepwater flysch sedimentation gradually spread over the territory of the former shelf.

The Kazakh–Kyrgyz paleocontinent was formed in the Ordovician and Silurian as a result of amalgamation of three sialic blocks: Syr Darya, Ysyk-Kol, and Borohoro. In the Neoproterozoic and Early Paleozoic, they were separated by the Terskey and Ili oceanic basins (Fig. 4, notations *Ter*, *I*), which opened after the breakup of the ancient continental plate. The Terskey oceanic basin probably existed as early as in the Cryogenian (Early Sinian in terms of Chinese geologists) and closed in the Middle–Late Ordovician. As a result, the composite Syr Darya–Ysyk-Kol microcontinent was created (Fig. 6). The Ili oceanic basin appeared in the Late Edicarian or Early Cambrian and closed in the Silurian. The vast Kazakh–Kyrgyz continental block was formed after attachment of the Borohoro block to the Late Ordovician Syr Darya–Ysyk-Kol microcontinent.

The Turkestan oceanic basin underwent long evolution (Figs. 5, 6). Proterozoic and Early Cambrian (Pb/Pb dates) ultramafic and mafic rocks, as well as oceanic basalts intercalated by sedimentary rocks with Cambrian, Ordovician, Silurian, and Devonian fauna are known in the West Tien Shan. Pelagic cherty rocks that accumulated up to the end of the Serpukhovian rest upon these basalts. In the Early Cambrian, the Turkestan oceanic basin separated the Indian–Australian and Pacific–Atlantic biogeographic provinces of trilobite fauna. This basin was the main biogeographic barrier in the wide Cambrian ocean. An ensimatic island arc that existed in the Turkestan paleoocean in the Ordovician separated a backarc basin (Fig. 6). The spreading of oceanic crust proceeded in the paleoocean up to the Late Devonian. Dispersed spreading of various duration and magmatic history probably was predominant in particular portions of the basin.

In the Silurian and the Early–Middle Devonian, the crust of the Turkestan paleoocean was plunging beneath the Kazakh–Kyrgyz paleocontinent, where suprasubduction volcanic rocks formed on land. Subduction ceased in the Givetian. After the break that lasted for 50–60 Ma, the consumption of oceanic crust beneath the Kazakh–Kyrgyz paleocontinent was resumed. The Carboniferous suprasubduction magmatic belt originated at the margin of this paleocontinent in the Visean. The magmatic activity developed under shallow-water marine and subaerial conditions. In the Moscovian, the development of the accretionary

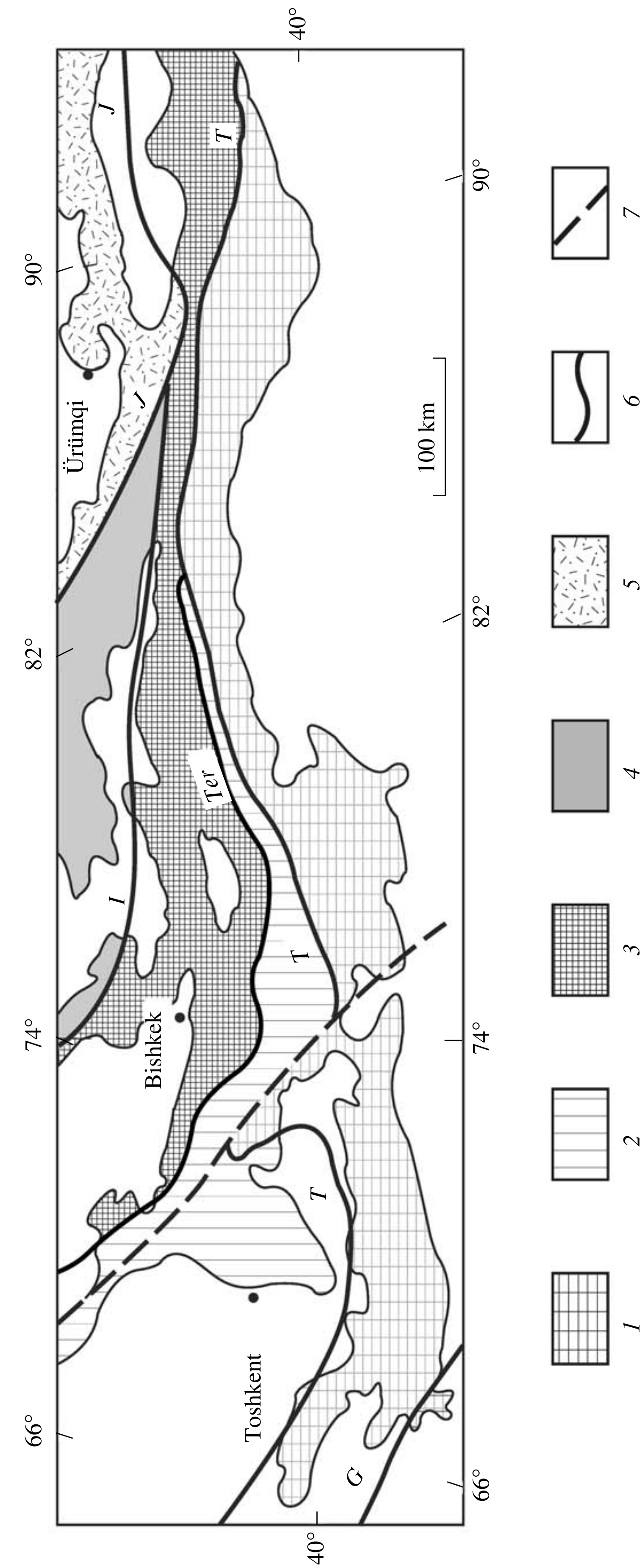

Fig. 4. The major paleotectonic elements of the Tien Shan. *(1)* Alay–Tarim paleocontinent; *(2–4)* blocks of the ancient continental crust in the Kazakh–Kyrgyz paleocontinents: *(2)* Syr Darya, *(3)* Ysyk-Kol, *(4)* Borohoro; *(5)* Carboniferous Bogdoshan island-arc system; *(6)* oceanic sutures (notations in figure): *G*, Gyssar; *J*, Junggar; *I*, Ili; *T*, Turkestan; *Ter*, Terskey; *(7)* Late Paleozoic Talas–Fergana Strike-Slip Fault.

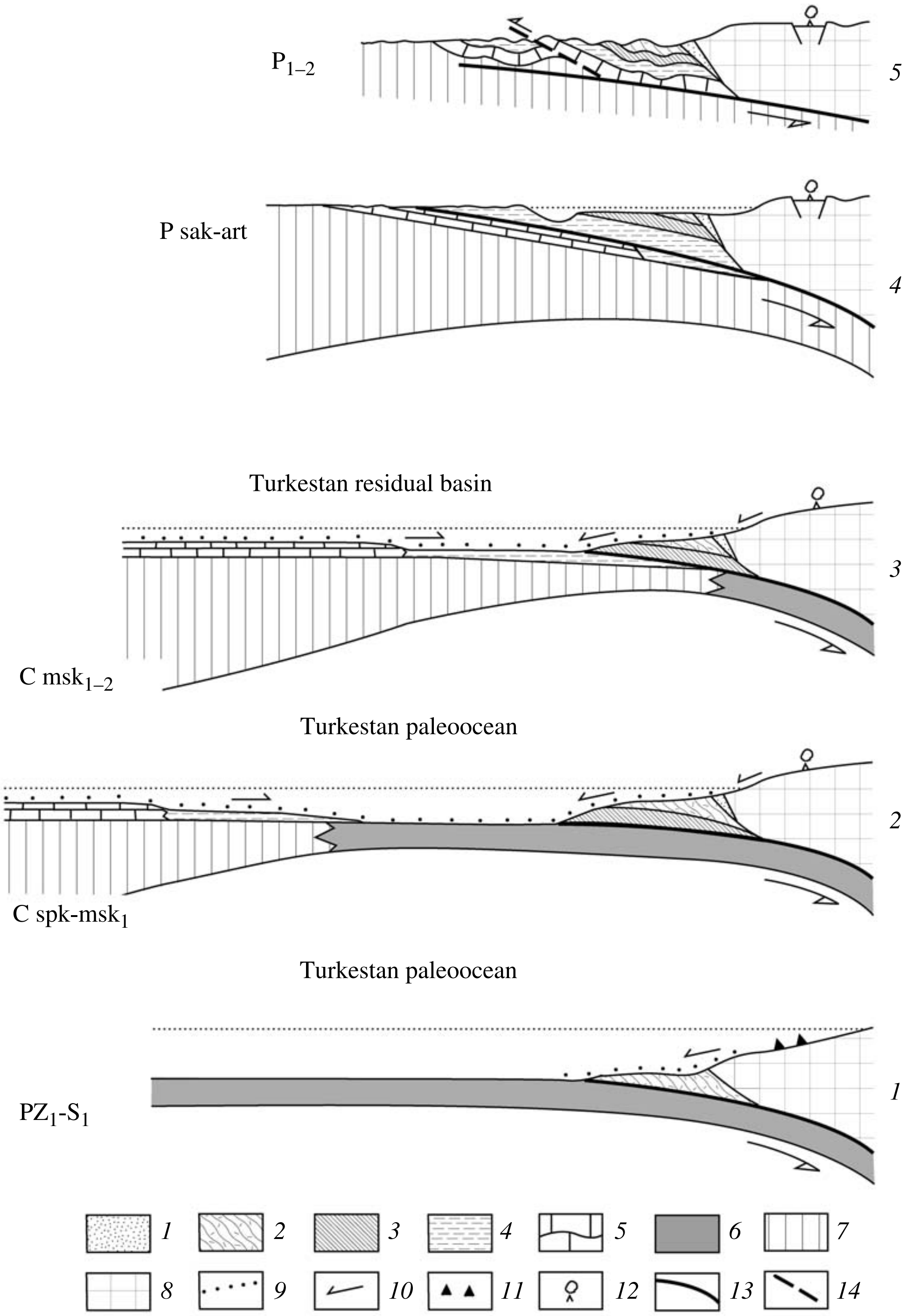

Fig. 5. Oceanic and continental subduction in the Tien Shan. (*1*) Terrigenous rocks metamorphosed in the Early Paleozoic; (*2, 3*) rocks of oceanic crust and oceanic island arcs: (*2*) metamorphosed in the Early Paleozoic, (*3*) unmetamorphosed; (*4, 5*) sedimentary rocks deposited on (*4*) slope and (*5*) shelf of the Alay–Tarim paleocontinent; (*6*) oceanic crust; (*7, 8*) continental crust: (*7*) Alay–Tarim and (*8*) Kazakh–Kyrgyz; (*9*) sediments coeval with thrusting; (*10*) direction of sediment transportation; (*11, 12*) volcanic activity: (*11*) submarine and (*12*) subaerial; (*13*) zones of subduction and underthrusting; (*14*) thrust fault.

wedge completed at the boundary of paleoocean with the Kazakh–Kyrgyz paleocontinent.

The collision of the Alay–Tarim and the Kazakh–Kyrgyz paleocontinents occurred in the Moscovian, when the Turkestan oceanic crust was completely subducted. The width of the oceanic plate plunged over 25 Ma of subduction could have reached 2500 km. Of the Turkestan paleoocean, only its marginal portion in

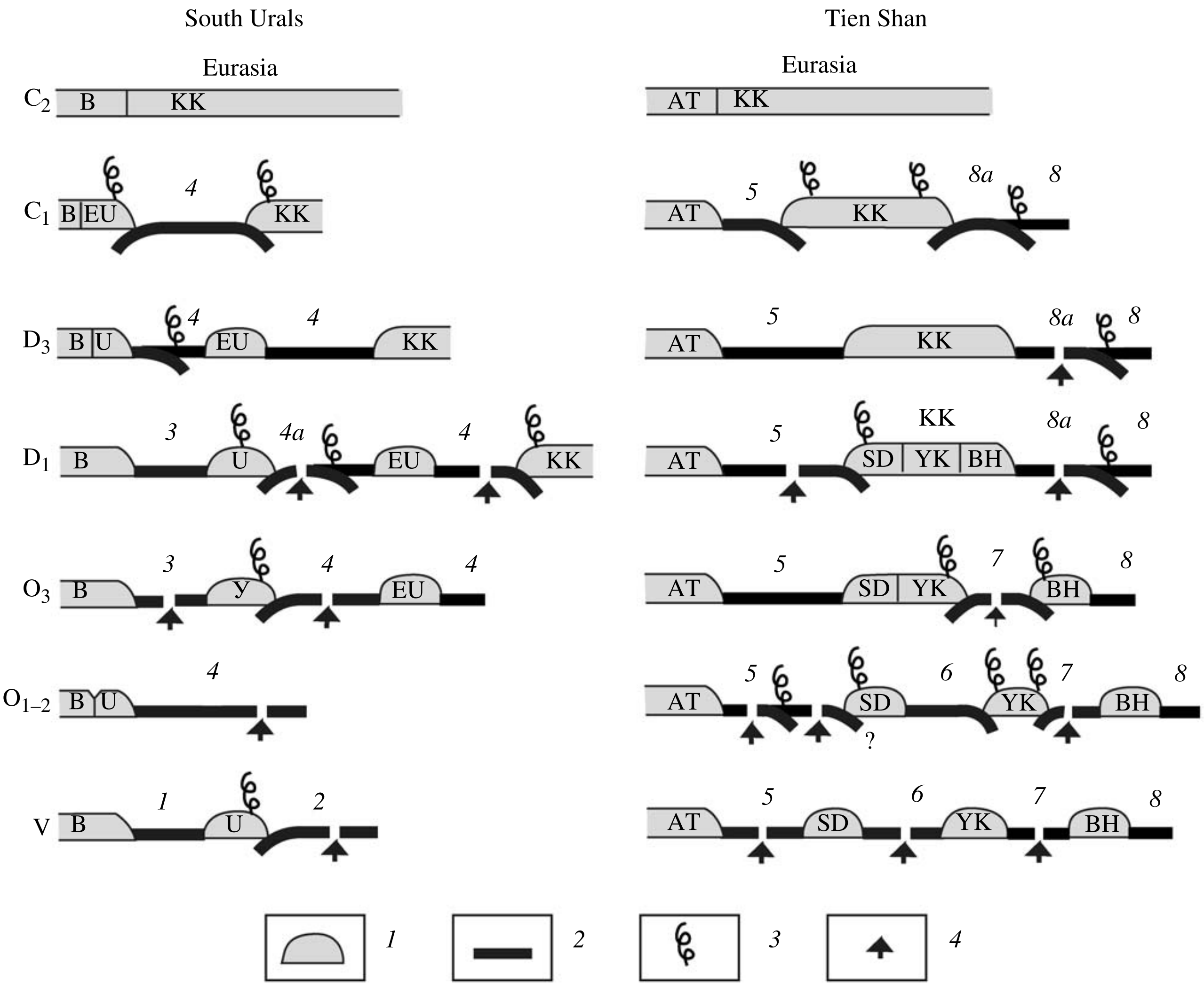

Fig. 6. Geodynamic sections across the South Urals and the Tien Shan. (*1*) Continental crust: continents, microcontinents, and island arcs with continental basement; (*2*) oceanic crust; (*3*) volcanic island arcs and marginal continental volcanic belts; (*4*) spreading zone; (V) Vendian (Edicarian). Notations in figure. Oceanic basins: *1*, Uraltau backarc basin; *2*, Protoural; *3*, Sakmara; *4*, Ural (*4a*, Cis-Sakmara–Voznesenka); *5*, Turkestan; *6*, Terskey; *7*, Ili; *8*, Paleoasizn (*8a*, Junggar–Balqash); paleocontinents: AT, Alay–Tarim; B, Baltic; KK, Kazakh–Kyrgyz; blocks with sialic crust: BH, Borohoro; EU, East Utal; YK, Ysysk-Kol; SD, Syr Darya; U, Uraltau.

the form of a marine basin with continental crust remained. This residual marine basin with flysch sedimentation existed until the Sakmarian or the Late Permian at the northern margin of Tarim.

In the Moscovian, the subduction of the oceanic crust was replaced with continental subduction, i.e., underthrusting of the Alay–Tarim continental margin beneath the accretionary wedge and the Kazakh–Kygyz paleocontinent (Fig. 5). The continental subduction, which lasted until the Late Permian, resulted in multifold shortening of the passive margin, whose initial width was more than 500 km. The shortening was compensated by the formation of a multilayer assembly of nappes that thrust over the Alay–Tarim paleocontinent. The nappes composed of the rocks pertaining to the Turkestan oceanic crust originated in the accretionary wedge at the margin of the Kazakh–Kyrgyz paleocontinent before its collision with the Alay–Tarim paleocontinent (Fig. 5, notations *1, 2*). After collision, the accretionary wedge was thrust along the bottom of the residual marine basin over the lower Moscovian rocks in the lower portion of the Alay Tarim continental slope (Fig. 5, notation *3*). Later on, the tectonic delamination of the sedimentary cover started at the margin of the Alay–Tarim paleocontinent. The detached sedimentary complex of the former slope together with overlying ophiolitic nappes was displaced inland of the paleocontinent (Fig 5, notation *4*). In the Permian, the assembly of nappes was deformed along with the autochthon into folds and thrust faults (Fig 5, notation *5*). The ongoing lateral compression led to squeezing of the arisen foldbelt and longitudinal tectonic flow. The Permian deformation was accompanied by orogeny, emplacement of collision and postcollision granitic olutons and alkaline intrusions likely related to strike-slip displacements along the foldbelt.

Comparison of Tectonic History of the Urals and the Tien Shan

Let us consider the similarities and differences between the Urals and the Tien Shan. The Ural and Turkestan paleooceans existed as early as in the Neoproterozoic and closed almost simultaneously in the Moscovian. The dispersed spreading was predominant in both paleooceans. In the Vendian (Edicarian) and Early and Middle Paleozoic, the Baltic continental margin of the Ural paleoocean was active, whereas the Alay–Tarim margin of the Turkestan paleoocean, which is comparable with the Baltic margin in its position, was passive. The opposite continental margin in the Urals is unknown before the Devonian. In the Tien Shan, a similar margin evolved on the Syr Darya block with ancient continental crust up to the Late Ordovician and became a part of the Kazakh–Kyrgyz paleocontinent in the Silurian (Fig. 6). This continental margin was active in the Silurian, Early Devonian–Eifelian, and Serpukhovian–Late Carboniferous.

The destruction resulted in the formation of marginal continental blocks developed in different manners. On the Baltic paleocontinent, a chain of rift basins originated near the Ural paleoocean at the Cambrian–Ordovician boundary and existed until the Late Devonian. The oceanic crust appeared in rift basins in the Middle Ordovician only in the South Ural segment. At the Alay–Tarim continental margin, local manifestations of rift-related magmatism, most intense in the Early Devonian, are known.

The internal structures of the Ural and Turkestan paleooceans were different. The East Ural microcontinent occurred in the Ural paleoocean in the Early and Middle Paleozoic. In the Turkestan paleoocean, such microcontinents are not established. The volcanic arcs in the Ural paleoocean appeared largely along the Baltic margin at different distances from it in the Vendian (Ediacarian), at the Ordovician–Silurian boundary, and in the Devonian. In the Turkestan paleoocean, a volcanic arc existed in the Ordovician near its Syr Darya margin, i.e., on another side of the ocean with respect to the Urals. The subduction of the Turkestan oceanic crust was interrupted but always recommenced in the same direction. The situation in the Urals was more complicated (Fig. 6).

In the Late Devonian and Early Carboniferous, Baltica collided with the Devonian island-arc system and in the Moscovian, with the Kazakh–Kyrgyz paleocontinent. In the Tien Shan, only continent–continent collision is documented. The deepwater flysch basins initiated by collision appeared on the Baltic margin in the Famennian and Visean; at the Alay–Tarim margin, they appeared in the Bashkirian and Moscovian. The flysch basins existed in the Urals up to the Kungurian, whereas in the Tien Shan, they were present up to the Sakmarian and locally up to the Late Permian.

In the Devonian and Carboniferous, the Ural and Turkestan paleooceans had a joint active margin along the Kazakh–Kyrgyz paleocontinent. The subduction of the oceanic crust beneath this paleocontinent in the Urals and Tien Shan started, terminated, recommenced, and ceased again synchronously. In the South Ural segment, the subduction in the Early Carboniferous developed beneath Baltica and the Kazakh–Kyrgyz paleocontinent, i.e., in opposite directions, whereas in the Tien Shan they developed only beneath this paleocontinent. As a result, the collision of the Kazakh–Kyrgyz paleocontinent with Baltica and the Alay–Tarim microcontinent gave rise to the formation of the divergent Ural Orogen and the unilateral fold–nappe belt in the Tien Shan.

The consumption of the Turkestan oceanic crust in the Tien Shan was followed by plunging of the relatively homogeneous passive margin of the Alay Tarim microcontinent into the subduction zone. In the Urals, collision was predated by the formation of the extensive active margin of Baltica. This margin consisted of an agglomeration of tectonic blocks differing in structure, size, and thickness. On colliding with the Kazakh–Kyrgyz paleocontinent, these blocks were overridden onto one another. Such a heterogeneous accretionary margin could not have been pulled into the Kazakh subduction zone after the oceanic crust. The direction of tectonic stacking toward the Kazakhstanides was predetermined by the subduction zone blocked in the early Serpukhovian and plunging in the opposite direction toward Baltica.

The growth of the high divergent orogen caused the development of the Ural Foredeep, which was filled with molasse since the Kungurian. No counterparts of this foredeep are known in the Tien Shan. The formation of the main granitic axis in the Urals was related to palingenesis of buried sequences into the zone with thick continental crust and numerous continental microblocks. No such granitic axis formed in the Tien Shan. The final collisional deformation in both the Urals and the Tien Shan developed in the transpressional geodynamic setting.

The comparison of the tectonic history of the Urals and Tien Shan has shown that during the Neoproterozoic and Early and Middle Paleozoic, the Ural and Turkestan oceanic basins was parts of the same oceanic domain. The tectonic events in its particular regions proceeded in different styles. In contrast to the passive Alay–Tarim margin, the Baltic margin was active. The data presented above indicate that the Alay–Tarim and Baltic continents were autonomous within the Early–Middle Paleozoic ocean. During the Early Carboniferous, a suprasubduction volcanic belt arose at the margin of the Kazakh–Kyrgyz continent. The magnetic anomalies make it possible to trace this belt from the eastern zone of the South Urals to the West Tien Shan. The formation of this belt accompanied the closure of the Ural–Turkestan oceanic domain, which was completed by amalgamation of the Urals, Tien Shan, and Kazakhstan in the Moscovian.

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