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Devonian palaeogeography of the Southern Urals

Gunar A. MIZENS

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Devonian deep-water deposits of the Southern Urals are represented mainly by flysch and cherty units. The main sedimentary basins (marginal sea and back-arc basin) and their origin, evolution and principal depositional environments are characterised. The main sources of clastic material were the Uraltau microcontinent (especially in the Famennian, when a mountain range formed following collision with an island arc) and two island arcs: the Irendyk, at the end of the Early and at the beginning of the Mid Devonian; and the Magnitogorsk, in the Mid to Late Devonian. Comparison with transgressive-regressive cycles established in Devonian epicontinental basins worldwide indicates that these global sea level fluctuations were recorded also in the Southern Urals deep-water settings. This applies primarily to the Eifelian and Frasnian-Famennian transgressive-regressive cycles.

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Key words: Southern Urals, Devonian, sedimentary basin, palaeogeography, olistostrome, eustatic changes.

INTRODUCTION

The structure of the modern Ural Mountains largely reflects processes and events that took place in the Devonian. Devonian deposits there are also linked with important mineral resources such as copper, zinc, iron and so on. Increased knowledge of the geology of the Urals foldbelt, much new factual material and the emergence of new methods of study have stimulated revision of concepts regarding its geological history. In particular there has been considerable progress in establishing a reliable stratigraphy, mostly based on conodonts, allowing modification of palaeogeographic and palaeotectonic models, elucidation of the character of sedimentary basins and sources of clastics, thus constraining hypothesis of their evolution in time and space. This paper discusses these topics, based on geological materials from the area between latitudes 51–55°N and longitudes 57–60°E (Fig. 1).

GEOLOGICAL FRAMEWORK OF THE SOUTHERN URALS

The main structural elements of the southern part of the Urals are the West Uralian, Magnitogorsk, East Uralian and Transuralian megazones (Figs. 1 and 2), as well as the West Uralian Foredeep (Ivanov *et al*., 1986). On the western slope of the Urals (within the West Uralian megazone) three large allochthonous units (Sakmara, Kraka and Bardym) are located, which, presumably, were displaced from more easterly areas.

The West Uralian megazone comprises the Zilair Synclinorium (synform), the Bashkirian Anticlinorium and the Uraltau antiform, though some workers (e.g. Puchkov, 2000) refer the two last structures to an independent Central Uralian megazone. In the Zilair Synclinorium mainly Lower and Middle Palaeozoic cherts and muddy cherts are developed. They are overlain by flysch of the Famennian-Tournaisian Zilair Formation. The Bashkirian Anticlinorium is composed mostly of Riphean sedimentary rocks, and its limbs comprise Vendian and Palaeozoic strata, including Devonian deposits. The Uraltau antiform is represented by two units differing sharply in composition and degree of metamorphism: the Suvanyak unit in the west and the Maksutov unit in the east. Quartzitic sandstones and quartzites of the Suvanyak complex have been metamorphosed to greenschist facies. Metamorphism of the Maksutov complex (arkose sandstones, volcanics, cherts, rare carbonates) reaches the eclogite-glaucophane facies. The age of both complexes is mainly Palaeozoic (Late Ordovician–Mid Devonian), but the presence of a Precambrian core can not be excluded (Krasnobaev *et al*., 1996).

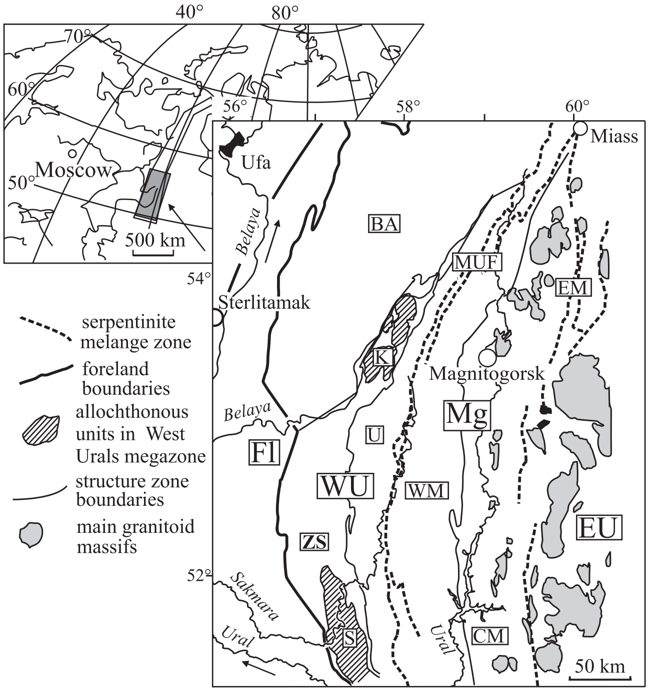

Fig. 1. Geological sketch showing the location of main structures of the Southern Urals

Megazones: WU — West Uralian, EU — East Uralian, Mg — Magnitogorsk, Fl — Foreland (West Uralian Foredeep); zones: BA — Bashkirian Anticlinorium, ZS — Zilair Synclinorium, U — Uraltau, MUF — Main Uralian Fault, WM — West Magnitogorsk, CM — Central Magnitogorsk, EM — East Magnitogorsk; allochthonous units: K — Kraka, S — Sakmara

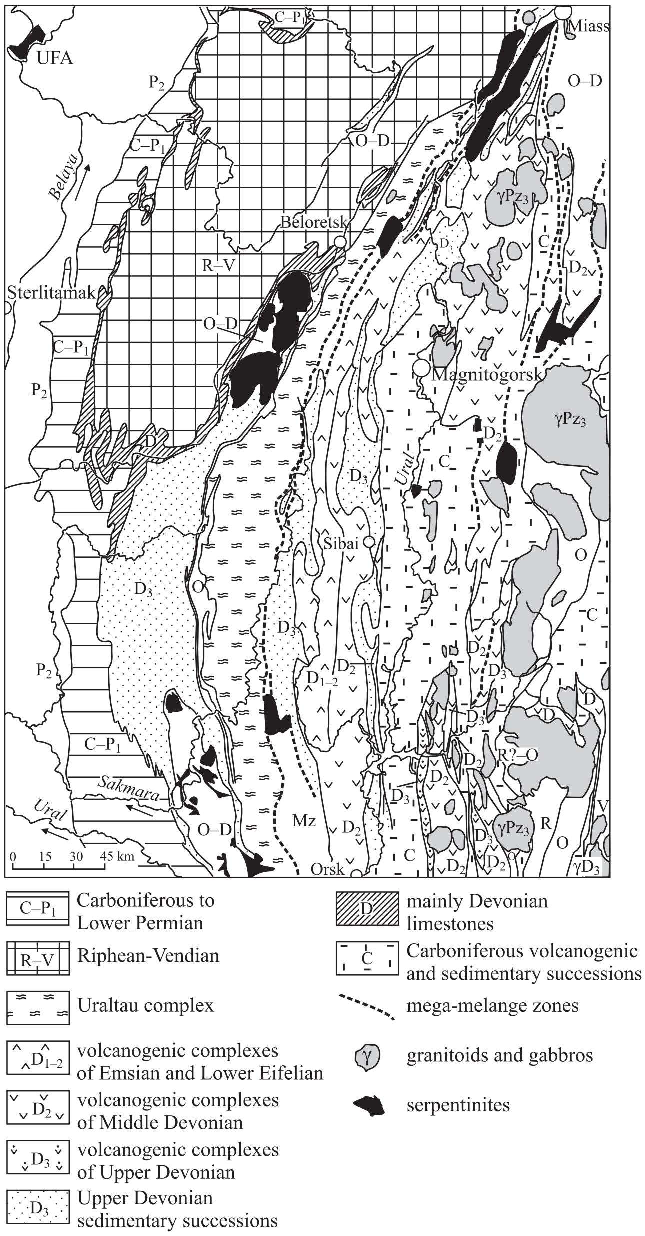

Fig. 2. Geological map of the western part of the Southern Urals based on the Urals map, series Geology of the USSR 1:500 000 (1979)

R — Riphean, V — Vendian, O — Ordovician, O–D — Ordovician to Devonian, P_2 — Upper Permian, Pz_3 — Upper Palaeozoic, Mz — Mesozoic

The Magnitogorsk megazone (together with its southern continuation, the West Mugodzhary zone) can be traced approximately from the latitude of the town of Miass northwards to where the Urals structures plunge to the south. It has a synformal character and includes the Voznesensk–Prisakmara, West, Central and East Magnitogorsk zones. The Voznesensk–Prisakmara structure represents the "Main Uralian Fault Zone" (MUF), a serpentinite mega-melange in which a significant role is played by Ordovician and, to a lesser extent, Silurian ophiolites. Upper Devonian, locally also Middle Devonian sedimentary sequences discordantly overlap the ophiolite complex. Nevertheless, Devonian strata are frequently incorporated in the melange, which probably represents evidence of repeated fault activity. The MUF zone dips towards the east at about 20–40° (Puchkov, 2000). The Magnitogorsk megazone is also bounded to the east by a belt of serpentinite mega-melange, which is usually distinguished as the East Magnitogorsk Fault Zone. In contrast to the MUF, this dislocation dips towards the west. Internal zones of megastructure, i.e., the West and Central Magnitogorsk zones, are composed mainly of sedimentary and volcanogenic strata of Devonian and Carboniferous age.

The East Uralian megazone is characterised by a widespread development of intrusive granitoids and gneissomigmatite complexes and therefore it is known as the "Main granite axis of Urals". Volcanogenic, volcanogenic-sedimentary and sedimentary successions of the Lower and Middle Palaeozoic are also widespread. A significant role is also played by Lower Palaeozoic ophiolites (the Denisovsk and Varna zones). Serpentinite melange belts often separate the diverse rock successions.

The Transuralian megazone includes the Valerianovsk, Borovsk and Ubagansk zones in the far east of the Urals, close to the Kazakhstanides boundary (Ivanov *et al*., 1986). Carboniferous, especially Lower Carboniferous volcanogenic, volcanogenic-sedimentary and sedimentary successions are widespread in the area. The presence of Middle and Upper Devonian deposits is also established, but these have been little studied.

The West Uralian Foredeep is built of clastic successions: flysch (Lower Carboniferous–Lower Permian) and molasse (Upper Permian–Triassic), in general onlapping shelf limestones of the East European (Russian) Platform margin (Mizens, 1997). It is a classical foredeep and its depositional fill is usually attributed to a number of subordinate depressions differing in structure and in thickness of sediment accumulations. In the south, these are the Aktyubinsk, Belsk, Sim and Yuruzan–Ai depressions.

DEVONIAN SUCCESSIONS

Devonian deposits are present within all of the above-mentioned regional units, but they are developed most completely in the territory of the West Uralian and Magnitogorsk megazones, as well as on the cratonic basement of the West Uralian Foredeep. The latter Devonian strata are considered to belong to the Russian Platform and in part they frame the Bashkirian Anticlinorium. In the East Uralian and Transuralian megazones only fragmentary Devonian successions are present, which are poorly studied, because of poor exposure.

WEST URALIAN MEGAZONE

The Lower and Middle Devonian stratigraphy of the Zilair Synclinorium (Table 1) is not entirely clear as regards stage boundaries, stratigraphical succession and lithology. The Emsian and Middle Devonian in the eastern part of synclinorium are represented by interbedded quartz sandstones, cherts and argillaceous slates, locally with conglomerate and limestone blocks (Puchkov, 1979, 2000). The cherts contain Emsian, Eifelian and Givetian conodonts, but few age determinations have been made. In the central part of the synclinorium, these strata are strongly thrusted and have not been studied. In the strongly tectonically disturbed Kraka and Sakmara zones, Lower and Middle Devonian deposits are exposed only as isolated fragments. Puchkov (2000) proposed that in the Kraka serpentinite massifs two types of Devonian sequence are present: (1) para-autochthonous (cherts with quartz sandstone interbeds) and (2) allochthonous (cherts associated with ultrabasites, volcanics and isolated limestone blocks). However, there is little convincing evidence to support this interpretation. Alternatively, Yakupov *et al*., (2002) noted that undoubted sedimentary contacts between cherts and volcanic rocks have not been found. The chert age everywhere corresponds to the Lower Devonian and Eifelian. Givetian conodonts have not yet been found in this lithology. This stage likely comprises other rock types (sandstones or volcanics ?).

The Sakmara zone is marked by a similar complexity of tectonic structure (Table 1). In general, one can distinguish up to five structural-facies complexes, the major part of which is presumably allochthonous (Ruzhentsev and Aristov, 2003). At a first approximation, there are the following complexes: the siliceous Sakmara complex (Middle Ordovician to Middle Devonian); the volcanogenic-siliceous Sugralinsk complex (Middle Ordovician to Upper Devonian); the siliceous-terrigenous Khersona complex (Lower Devonian to Lower Carboniferous); the siliceous-volcanogenic Guberlya-Kosistek complex (Middle Ordovician to Middle Devonian), and the siliceous Sarbayevo complex (Lower Devonian to Upper Devonian). Nevertheless, the presence of para-autochthonous strata, representing two types of successions, is possible (Ivanov and Puchkov, 1984; Puchkov and Ivanov, 1985; Maslov *et al*., 1993; Puchkov, 2000):

1. Argillaceous and cherty slates with interbeds of argillaceous limestones containing Pragian conodonts and cherts with conodonts of the upper Eifelian. These are overlain by argillaceous and carbonaceous-argillaceous slates of the Eginda Formation, which in the upper Frasnian is gradually replaced by flysch of the Zilair Formation, and in the west by the calcareous-argillaceous Kiya Formation.

2. Cherts and argillaceous slates of the Lower Devonian — Eifelian Kizilflot and the Lower Devonian — Givetian Aitpai formations occur above the Lower Palaeozoic Sakmara Formation. Within this succession, there are frequent subaqueous slump deposits: cloddy cherts, cherty-clastic breccias, conglomerates and sandstones, and frequently blocks of shallow-water biohermal limestones mainly of Pragian-Emsian age (also Givetian limestones within the Aitpai Formation).

Table 1

Stratigraphic subdivision and correlation of Devonian deposits of the Zilair Synclinorium

<table>
<tr><th></th><th></th><th>East zone</th><th colspan="2">Kraka zone</th><th colspan="2">Sakmara zone</th></tr>
<tr><td rowspan="2">Upper Devonian</td><td>Famennian</td><td colspan="5">alternation of polimictic sandstones and argillites, sometimes there are interbeds and packets of siliceous slates, interbeds of calcarenites are in the upper part (Zilair Formation) more than 3000 m</td></tr>
<tr><td>Frasnian</td><td colspan="3">cherts, argillaceous and siliceous slates (Ibragimovo Horizon) up to 200 m</td><td colspan="2">cherts, argillaceous slates (Eginda Formation) up to 250 m</td></tr>
<tr><td rowspan="2">Middle Devonian</td><td>Givetian</td><td rowspan="3">siliceous and argillaceous slates, quartz sandstones, sometimes limestone blocks and conglomerates more than 300 m</td><td colspan="2">?</td><td>?</td><td rowspan="5">cherts, argillaceous and siliceous slates, cherty breccias, often with limestone blocks up to 500 m</td></tr>
<tr><td>Eifelian</td><td rowspan="2">cherts, siliceous slates, quartz sandstones 100–200 m (?)</td><td rowspan="2">cherts, basaltoids, sometimes blocks of limestones ?</td><td rowspan="3">argillaceous and siliceous slates, interbeds of argillaceous limestones up to 200 m</td></tr>
<tr><td rowspan="3">Lower Devonian</td><td>Emsian</td></tr>
<tr><td>Pragian</td><td rowspan="2">?</td><td colspan="2" rowspan="2">?</td></tr>
<tr><td>Lochkovian</td><td>?</td></tr>
</table>

table includes data from: Ivanov and Puchkov (1984); Ivanov (1998); Puchkov (2000); Yakupov *et al*. (2002)

The Upper Devonian in the Zilair Synclinorium is represented by Frasnian cherts and Famennian-Tournaisian greywacke flysch of the Zilair Formation. Frasnian cherts (Ibragimovo Horizon) may be traced along the eastern limb of the synclinorium (Puchkov, 2000), while on the western flank the Frasnian is developed in cherty-argillaceous facies (Yakupov *et al*., 2002). At the end of Frasnian and at the beginning of Famennian, cherts were replaced gradually (through interbedding) by sandstones and mudstones of the Zilair Formation. In the eastern areas they were deposited in the Late *rhenana* to *linguiformis* conodont zones (Puchkov, 2000), while to the west somewhat later, in the Late *triangularis* Zone (Yakupov *et al*., 2002). The Zilair flysch typically forms a relatively monotonous and thick (over 3–4 km) succession. Rudaceous deposits are rarely found. It is possible, though, to distinguish calcareous olistoliths near the western (platform) synclinorium flank, containing conodonts of the *triangularis* and Early *crepida* zones, and rare boulder-rich units of small thickness (up to 10–15 m) in the middle part of the Zilair Formation.

In southern areas of the Zilair Synclinorium (Sakmara zone), the Devonian succession contains at least four levels of olistostromes (Ivanov and Puchkov 1984; Puchkov, 2000; Ruzhentsev and Aristov, 2003): Pragian-Emsian, Eifelian, Givetian-Frasnian and Famennian.

In the Silurian to Middle Devonian of the Uraltau territory shallow-water limestones were deposited locally. Displaced blocks and fragments of limestones occur in the Devonian argillaceous-siliceous and flysch strata along the western margin of Uraltau (e.g. Keller, 1949; Ozhiganov, 1964; Puchkov, 1979), as well as in the melange of the Kraka ultrabasite massifs (Kamaletdinov and Kazantseva, 1983). Blocks of Pragian-Givetian limestones are also common in olistostromes of the Sakmara zone (Ivanov and Puchkov, 1984). The debris could not be derived from a shelf zone of the East European continent as they often include polymictic clastic rocks with grains and pebbles of metamorphic and magmatic rocks, including serpentinites. The source area for the latter material could only have been the Uraltau massif. Middle Palaeozoic conodonts have also been found in the marbles of the Uraltau succession (Zakharov and Zakharova, 1998).

MAGNITOGORSK MEGAZONE

The geological structure of the Magnitogorsk megazone is more complex. Devonian deposits are here characterised by great diversity and lateral variability (Table 2). The complex tectonic structure of this area obscures correlation of the successions. However, during the last 20 years, the Devonian biostratigraphy of this megazone has become relatively clear owing to the application of conodont biostratigraphy (e.g. Maslov, 1980; Maslov *et al*., 1993, 1999; Artyushkova and Maslov, 1998; Maslov and Artyushkova, 2000, 2002).

Lower Devonian deposits are known practically only in the Voznesensk–Prisakmara zone as isolated fragments of successions, most often with tectonic boundaries. Admixtures of volcanogenic components are common in the sedimentary rocks. These are mainly tephra-rich and volcanomictic sandstones, cherty argillites and cherts, less often limestones and subalkaline volcanic rocks.

Eifelian successions are particulary variable due to changing proportions of volcanogenic, volcaniclastic, clastic and siliceous components, with olistostromes also being important volumetrically. The thickness of this stage is extremely variable: from 100–150 m in the condensed sections of the Aktau Formation, up to 3000–4000 m in the area of the volcanogenic Irendyk Formation. The volcanogenic rocks of the Irendyk island arc in the east are overlapped by rudaceous debrites and olistostromes of the Gadilev complex and jaspers of the Yarlykapov Formation, which contact across the synsedimentary growth fault with volcanic rocks developed during back-arc spreading (the Karamalytash Formation). Stratigraphically higher, relatively thin Bugulygyr jaspers occur over a considerable part of territory. According to Maslov and Artyushkova (2002) their age corresponds to the upper Eifelian (*kockelianus* Zone).

Table 2

Stratigraphic subdivision and correlation of Devonian deposits of the Magnitogorsk megazone

		Voznesensk–Prisakmara zone	West Magnitogorsk zone	Central and East Magnitogorsk zone	
Upper Devonian	Famennian	polimictic sandstones and argillites, in the eastern part tephroids (Zilair Formation) more than 2000 m; mixtites up to 800 m		limestones 40–200 m	subalkaline basalts, andesites, dacites, tuffs up to 2000 m
Upper Devonian	Frasnian	cherts, siliceous slates, in the north and east also packages of tuffs and tuffites (Mukasovo Formation) from 50–100 m up to 700–800 m			tuffs, tephroids, cherts up to 500 m
Middle Devonian	Givetian	cherts, argillaceous-siliceous slates, volcanomictic sandstones, mixtites up to 250 m	tephroids, tuffs, argillaceous-siliceous slates, mixtites (Ulutau Formation) 500–1800 m; mixtites	basalts, andesibasalts, tuffs, cherts up to 2000 m	
Middle Devonian	Eifelian	volcanomictic and tephrogenic sandstones, argillites, mixtites up to 400 m	jaspers 5–120 m; jaspers 20–100 m; basalts, rhyolites up to 2000 m; basalts, andesibasalts, tuffs, tephroids (Irendyk Formation) 200–3500 m	basalts, andesites, rhyolites, jaspers 1500– 2300 m; ?	
Lower Devonian	Emsian	basalts, rhyolites up to 2000 m; subalkaline vulcanites up to 1000 m	?		
Lower Devonian	Pragian	siliceous and argillaceous slates, volcanomictic and tephrogenic sandstones, gravelstones, sometimes limestone blocks 300–700 m		limestones, volcanomictic sandstones, argillites up to 200 m	
Lower Devonian	Lochkovian			?	

table includes data from: Artyushkova and Maslov (1998), Maslov and Artyushkova (2000, 2002), Mizens (2002*b*)

In the Givetian and early Frasnian, in the west Magnitogorsk zone, a thick (up to 2000 m) volcanogenic flysch of the Ulutau Formation was deposited, comprising volcanomictic rocks, tephroids and tuffs mainly of intermediate and acid composition. Sandstones dominate, but conglomeratic deposits are also common: conglomerates and debris-flow sediments that contain pebbles and blocks of limestones. The Magnitogorsk island arc situated to the east (in modern coordinates) was a sediment source, now preserved as volcanic lavas and tuffs of the Central Magnitogorsk zone. During this timespan, the massif of the Irendyk arc continued to form a morphological barrier in spite of the fact that it was already submerged below sea level. To the west of the barrier, in the Voznesensk–Prisakmara zone, thin cherts and argillaceous-siliceous sediments were deposited. This area was not reached by turbidite flows from the Magnitogorsk arc.

The middle-upper part of the Frasnian, corresponding to the *punctata-rhenana* zonal interval (Artyushkova and Maslov, 1998; Maslov *et al*., 1999; Maslov and Artyushkova, 2002), is represented by cherts (Fig. 3) and cherty-argillaceous slates of the Mukasovo Formation, to the north with interbeds of tephra-derived sandstones, and locally conglomerates. These facies have occupied a major part of the Magnitogorsk megazone, except for eastern areas where volcanic complexes are developed. The thickness of the sequence varies from 50–100 m in the southern part up to 700–800 m northwards. It is not quite clear whether the Irendyk barrier still existed at that time, as cherts were deposited both to the west and east of the Irendyk volcanic complex.

The Famennian succession displays a great similarity to coeval sequences from the western slope of Urals. Sandy-argillaceous Zilair-type deposits are also developed, but their characteristics are somewhat different. Using overall composition and provenances two formations may be distinguishedin the Magnitogorsk megazone: the western Prisakmara Formation and the eastern Bolshekizil Formation (Mizens, 2002*b*). The main difference between them is the presence (Prisakmara) or absence (Bolshekizil) of metamorphic rock clasts. The lower boundary of the Prisakmarian flysch corresponds to the upper Frasnian *linguiformis* Zone. Mountains formed at the site of the former Uraltau microcontinent acted as a source of the detrital sediments; EUROPROBE project participants came to a similar conclusion (Wilner *et al*., 2002). Gravity-mass deposition of

Fig. 3. Frasnian cherts of the Aktau settlement area

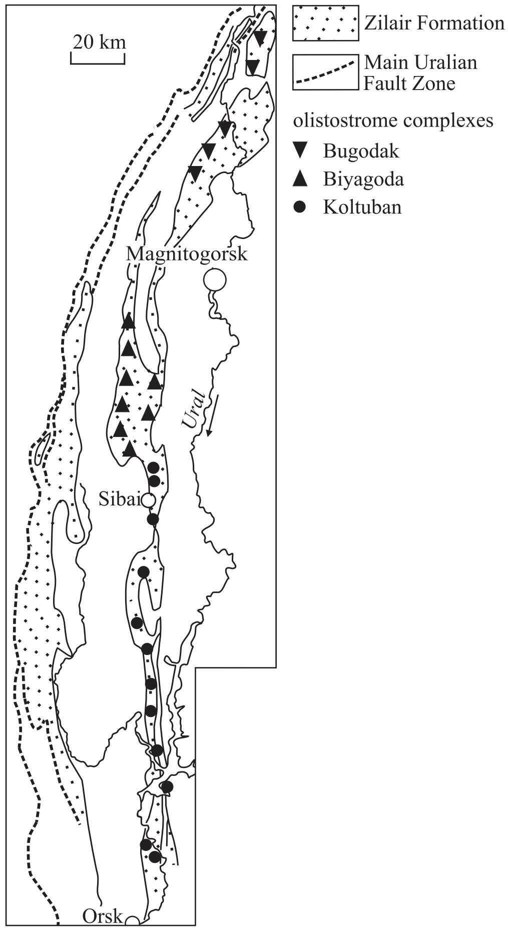

Fig. 4. Geological sketch showing the location of main mixtite complexes at the Frasnian/Famennian boundary

the Bolshekizil Formation, coupled with submarine sliding, started later, close to the middle *triangularis* Zone, and was related to a volcanic island arc.

Substantial rudaceous successions (Biyagoda olistostrome, Koltuban calcareous conglomerates, Bugodak complex *etc*.) have been traced in the lower part of the Bolshekizil Formation (Fig. 4). The thickest Biyagoda olistostrome (up to 700–800 m; Fig. 5), widespread in the drainage area of Great Kizil River, is represented by a complex irregular pile of blocks and clasts of basic and acid volcanics, limestones, cherts, sandstones and sandy-argillaceous layers. Some erratic blocks (i.e. olistoliths) reach sizes of 1–2 km. The olistostrome is marked by a poorly sorted gravel-sandy matrix. It is everywhere underlain by turbidites of the Prisakmara Formation (Mizens, 2002*b*). The "Koltuban limestones" (after Nalivkin, 1951), unsorted conglomerates of mainly calcareous composition, were deposited in the southern Magnitogorsk megazone. Polymictic conglomerates with scattered blocks of limestone replace them at the level of the town of Gai. The conglomerates can be observed as a discontinuous horizon up to 150 km in length and up to 100 m thick in places. This deposit represents a densely stacked boulder-pebble mass with irregular distributed blocks (up to hundreds of metres across), but in other places the coarse clasts are loosely scattered in a matrix. In the northern part of the Magnitogorsk megazone, the olistostromes correlate with the

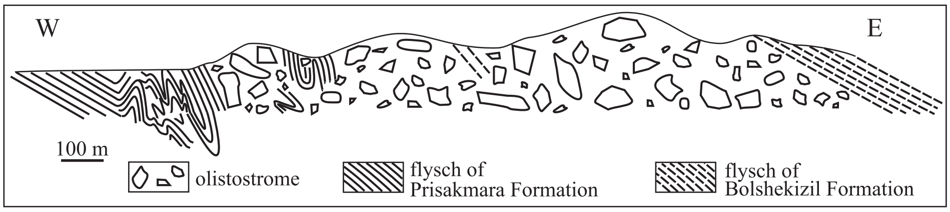

Fig. 5. Geological cross-section of the Biyagoda mixtite; Ikstimer River

upper part of the volcanogenic-sedimentary Bugodak complex described by Artyushkova and Maslov (1998). This succession also contains thick (up to 100–200 m) bodies of rudaceous rocks with blocks and clasts of basic volcanics and upper Givetian limestones.

The palaeontological data and interrelations with enclosing rocks prove that all the olistostromes were formed almost simultaneously in an early Famennian interval corresponding to the Middle and Late *triangularis* Zone, in places up to the Middle *crepida* Zone (Mizens, 2002*b*).

MAIN SEDIMENTARY BASINS, THEIR ORIGIN AND EVOLUTION

It is usually supposed (e.g. Ivanov *et al*., 1986; Zonenshain *et al*., 1990; Puchkov, 2000) that the Palaeozoic history of the Urals was connected with an ocean which had been formed in the Early Ordovician. But this palaeogeographic domain was not a single unit. In the territory of the present Southern Urals, there were probably at least three rigid blocks with a continental crust (microcontinents), which subdivided the Uralian ocean into several secondary basins. These (see Fig. 1) are the Uraltau (Central Uralian), East Uralian (East Mugodzharian) and Transuralian massifs.

WESTERN MARGINAL SEA

The Uraltau microcontinent was located near the margin of the East European continent (Fig. 6). The basin, which was separated by this continental block from the open ocean, probably had a suboceanic character, without fully developed or with only incipiently developed oceanic crust. In fact it was a deep marginal sea, resembling the modern Mozambique Strait between Africa and Madagascar. This strait existed from the Ordovician to the Permian, not considering its transformation into a foreland basin in the Carboniferous (Mizens, 1997, 2002*b*). Sequences of the Zilair Synclinorium represent an unquestionable sedimentary record of this marginal sea. The common viewpoint, that this area, together with the Uraltau block, represents a Palaeozoic bathyal slope and base of slope of the East European continent (e.g. Puchkov, 1979, 1993, 2000; Ivanov and Puchkov, 1984; Ivanov *et al*., 1986; Ivanov, 1998), is not confirmed here. Such an interpretation contradicts recent new data (Mizens, 2002*a*, *b*). In particular, the Uraltau massif in different times served as a source of clastics (including rudaceous debris) for the marginal basin. Flute casts suggest current directions in the basin from the south to the north; nevertheless, the origin of clastics is obviously connected with the Uraltau cordillera.

OCEANS, MICROCONTINENTS AND ISLAND ARCS

East Uralian and Transuralian massifs separated basins characterised by oceanic crust. Oceanic basin successions were less well preserved than those of the continental margins. Presumably they were mostly destroyed by subduction processes. Fragmentary sections can be encountered in serpentinite mega-melange zones, which are interpreted as the remains of accretionary prisms. The oceanic basin between the East Uralian and Transuralian microcontinents has only been reconstructed prior to the beginning of the Devonian (Yazeva and Bochkaryov, 1998; Puchkov, 2000), because pre-Devonian collision of the East Uralian microcontinent with an island arc

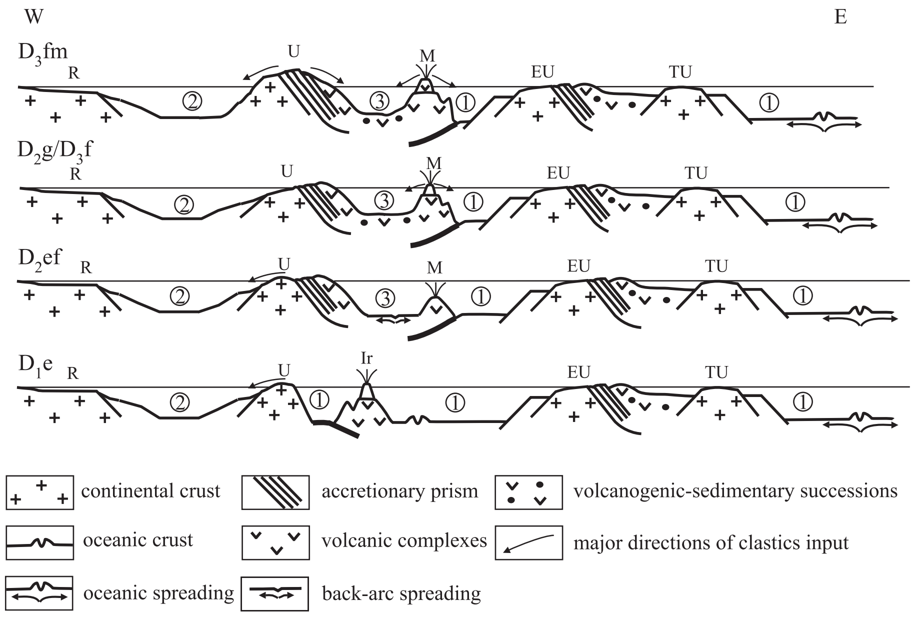

Fig. 6. Schematic model outlining development of the Devonian sedimentary basins in the Southern Urals

R — Russian (East European) platform; microcontinents: U — Uraltau, EU — East Uralian, TU — Transuralian; basins: 1 — ocean, 2 — marginal sea, 3 — back-arc; island arc: Ir — Irendyk, M — Magnitogorsk is supposed. An oceanic basin probably existed up to the Mid Carboniferous to the east of the Transuralian block (Puchkov, 2000). Microcontinents were mostly covered by shallow and sometimes deep seas.

Most researchers concluded that the ocean between the Uraltau continental block and the East Uralian massif existed up to the end of the Devonian (e.g. Ivanov *et al*., 1986; Zonenshain *et al*., 1990; Feist *et al*., 1997; Ivanov, 1998; Puchkov, 2000; Alvarez-Marron *et al*., 2000). Data from igneous rocks (Yazeva and Bochkaryov, 1998) indicate that a subduction zone developed in the area during the second half of the Early Devonian. The seismofocal plane dipped to the east (in modern coordinates). Above it, an island arc, represented by volcanic complexes of the Baimak–Buribai and Irendyk formations, had been formed. At the beginning of the Eifelian, the Uraltau microcontinent moved towards the island arc and they collided. The fore-arc accretionary prism, composed of ocean deduced from the floor rocks, was partly overthrusted on to the microcontinent, as deduced from a tectonic structure and lithological composition of the Maksutov complex of Uraltau (Puchkov, 2000; Mizens, 2002*b*).

Non-metamorphosed parts of the accretionary prism have been preserved in the Main Uralian Fault Zone and its surroundings. These are ultrabasites and volcanics of Ordovician and late Early Devonian age, and argillaceous-siliceous sequences of the Silurian and Lower Devonian. The succeeding microcontinent edge also came to be involved in subduction, beginning to be uplifted after slab break-off. This process led to mountain building by the end of the Frasnian. Traces of high-pressure metamorphism, typical of sialic rocks of the Maksutov complex, have been associated with the descent of continental crust into a subduction zone. In the Devonian, however, these rocks were not eroded (Mizens, 2002*b*). The accretionary prism had been thrust over the microcontinent by the Mid Devonian. This is shown by composition of conglomerates (debrites), intercalated with argillaceous-siliceous parts of the Middle Devonian Betrya Formation in the Zilair Synclinorium. Pebbles in these conglomerates are composed of limestones, crystalline schists, gneisses, quartzites, basic volcanics and serpentinites.

After the collision of the island arc with the microcontinent the subduction zone was displaced into the oceanic area to the east. The new Magnitogorsk island arc was formed on its basement and existed throughout the Givetian and Late Devonian (Brown *et al*., 2001; Mizens, 2002*a*, *b*). Magmatism of this arc consequently evolved (Yazeva and Bochkaryov, 1998) from tholeiites of a primitive island arc of Tonga type (found among Eifelian volcanics of the Karamalytash complex) to the volcanic formations of a young arc (the Givetian Ulutau complex) and a developing arc (Upper Frasnian Novovoronino complex = Koltuban complex), and was completed by the volcanic and plutonic rocks of a mature arc (the Late Devonian Verkhneuralsk complex). Thus, it is evident that the Magnitogorsk arc did not arise as a result of the evolution of the Irendyk arc, as has been thought (e.g. Ivanov *et al*., 1986; Alvarez-Marron *et al*., 2000; Brown *et al*., 2001), but developed independently. Seravkin (1986) confirmed the undoubted presence of primitive island arc complexes at two stratigraphical levels: Eifelian (= Emsian in recent terms) in the west and Givetian (= Eifelian) in the east.

The majority of authors think that the subduction zone, despite the Irendyk arc`s collision with a microcontinent (or continent) maintained its position and easterly dip up to the end of the Devonian (e.g. Ivanov *et al*., 1986; Chemenda *et al*., 1997; Ivanov, 1998; Yazeva and Bochkaryov, 1998; Alvarez-Marron *et al*., 2000; Brown *et al*., 2001). There is considerable evidence, however, that the polarity and location of the subduction zone changed in the Middle Devonian (e.g. Samygin *et al*., 1987; Zonenshain *et al*., 1990; Seravkin *et al*., 1992; Surin and Moseichuk, 1995; Mizens, 2002*a*, *b*). The most convincing data here comprise sources of clastics for the Zilair Formation, as well as that regarding migration of the Magnitogorsk volcanic arc. New palaeomagnetic data also support this interpretation (Burtman *et al*., 2000). Geochemical studies give inconsistent results (Yazeva and Bochkaryov, 1998), but on the whole they also suggest a westwards dipping subduction system.

The events described (collision of an island arc with a microcontinent, accretionary prism obduction on to a microcontinent) essentially did not influence the western marginal sea. However, the geotectonic situation in the eastern part (in modern coordinates) was changed radically. The oceanic basin between the Irendyk island arc and the Uraltau microcontinent disappeared in the Middle Devonian.

FORE-ARC BASIN

Between the Irendyk volcanic arc and the accretionary prism a fore-arc basin formed in the Early Devonian and continued its development into the early Eifelian (Brown and Spadea, 1999; Alvarez-Marron *et al*., 2000; Puchkov, 2000; Brown *et al*., 2001). In this area there is little direct evidence of volcanism, and the sedimentary succession is represented by volcanogenic sandstones, argillaceous and siliceous rocks with olistostromes, debrites and individual limestone blocks. Clastic sediments were mainly supplied from the flank of a volcanic arc (Iltaban, Mansurov, Ryskuzhin, Turat, Aktau and other formations), and to a lesser degree, probably also from the accretionary prism area (monomictic cherty debrites of the Mazov Formation).

In the eastern part of the Magnitogorsk megazone, the fore-arc basin (associated with the Mid–Late Devonian island arc) is also assumed to have been present, but the corresponding deposits (mainly turbidites) are poorly exposed in the East Magnitogorsk Fault Zone and are little studied.

BACK-ARC BASIN

The participants of the EUROPROBE Uralides project came to the conclusion that the area between Uraltau and the Magnitogorsk arc in the Middle and Late Devonian had the character of a suture fore-arc basin (e.g. Alvarez-Marron *et al*., 2000; Brown *et al*., 2001) as, according to their opinion, the subduction zone maintained an easterly dip during the entire Devonian. Contradictory evidence is represented by a huge complex of Zilair flysch, which, together with Mukasovo Formation cherts, was deposited on both sides of the Irendyk volcanic edifice. If the Uraltau cordillera served as the source area for this flysch neither an oceanic domain nor the subduction zone between the Irendyk and Uraltau in the Late Devonian could have already existed. Hence, they did not exist in the Late Devonian, because Middle and Upper Devonian magmatic and clastic complexes to the east of the Irendyk volcanic edifice are genetically connected (see above). The subduction zone and the ocean could have existed only to the east of the arc, but then a seismofocal plane had to plunge under the arc to the west. A regular rejuvenation of Magnitogorsk arc barrier zone igneous activity from east to west also testifies in favour of such a supposition. It is clearly visible on the geodynamic map of Yazeva and Bochkaryov (1998).

The situation described is evidently not a unique one. "Jumping" of the subduction zone associated with changes of polarity after collision of an island arc with a passive continental margin is quite a common phenomenon. The mechanism of such process has been considered by Konstantinovskaya (1999). Using experimental evidence and geological data from the western and southwestern margin of the Pacific she showed that a change of subduction polarity after the collision of an island arc with a continental passive margin is practically inevitable under the conditions of a common regime of compression. Earlier Mores and Twiss (1995) came to a similar conclusion after analysing possible types of collisions and their consequences.

If so, then between the Magnitogorsk arc and the Uraltau microcontinent a back-arc basin developed in the Middle and Late Devonian. The lower part of the succession (Eifelian stage) is composed of jaspers of the Yarlykapov Formation (thickness 5–50 m) and the volcanic Karamalytash Formation (up to 2000 m), whose geochemical features suggest origin in a back-arc spreading setting (Yazeva and Bochkaryov, 1998). Volcanic rocks with interbeds of chert (in the east) and jasper (in the west) were deposited simultaneously in the Mid Eifelian and contact with each other across synsedimentary growth faults. The Eifelian upper part is represented by jaspers of the Bugulygyr Formation. The Givetian and lower part of Frasnian are composed of the Ulutau Formation (resedimented tuffaceous rocks and tuffs), which indicate eruptive activity on the island arc.

In the Eifelian to early Frasnian the back-arc basin was divided by a longitudinal barrier (mainly submerged), built of a volcanic belt of the extinct Irendyk arc. To the west of this barrier the type of sedimentation was inherited from the Early Devonian–early Eifelian fore-arc basin (cherts, cherty-argillaceous deposits, volcanomictic turbidites and debrites of the Turat and Aktau formations). By the end of the Frasnian the barrier was probably completely submerged and no longer influenced sedimentation. The upper part of the Frasnian is hence represented throughout the whole basin by siliceous deposits of the Mukasovo Formation, while the Famennian stage comprises a flysch succession of the Zilair Formation.

SEA LEVEL CHANGES

Some features of sedimentation in the Devonian deep-water basins of the Urals seem to correspond to the transgressive and regressive events known from the epeiric depositional record of all continents (e.g. Johnson *et al*., 1985; Alekseev *et al*., 1996; Racki, 1997). This overall coincidence was partly noted earlier by Fokin and Nikishin (1999), but it was interpreted as recording tectonic processes (convergence or extension), especially in relation to the Urals rudaceous successions, correlated with hiatuses on the platforms. Horizons of sedimentary cherts, however, were usually explained in terms of tectonic quiescence. The study of depositional environments in the Urals deep-water basins and their interrelations with palaeotectonics and geodynamic settings nevertheless has shown that causal links between tectonics and sea level fluctuations are rare or absent (Mizens, 2002*a*, *b*). By contrast, the facies characteristics allow, in some cases, to infer an influence of global sea level changes. Such a genetic explanation seems applicable to two sequences of associated pelitic (mainly siliceous) and rudaceous deposits of Eifelian and, especially, Frasnian-Famennian age, which developed on the eastern slope of the Southern Urals and formed in a back-arc basin. Evidence of similar Givetian eustatic fluctuations has not yet been found.

EIFELIAN TRANSGRESSIVE-REGRESSIVE CYCLE

As shown above, upper Eifelian cherts extend across the back-arc basin, both eastwards and westwards of the residual Irendyk volcanic arc. Cherts overlap the volcanic and argillaceous-siliceous sequences in the eastern part (Karamalytash and Yarlykapov formations), as well as clastic successions westwards (Turat Formation and sandy-conglomeratic part of the Aktau Formation). Their age, according to Maslov and Artyushkova (2002), corresponds to the *kockelianus* Zone, that is correlated with the late Eifelian transgression (approximately the basal If cycle of Johnson *et al*.,1985), which has been noted from many epicontinental basins, including the East European (Russian) Platform (e.g. Weddige, 1977; Johnson *et al*., 1985; Tikhomirov, 1995; Racki, 1997; Fokin and Nikishin, 1999; Yakupov *et al*., 2002). It may be inferred that the Bugulygyr jaspers originated in highstand conditions when the main clastic source areas were drowned.

The sea had retreated from a considerable part of the East European Platform near the Eifelian-Givetian boundary, leading to the development of the largest Middle Devonian hiatus (Tikhomirov, 1995; Alekseev *et al*., 1996; Fokin and Nikishin, 1999). Especially intensive processes of erosion occurred in the eastern part of the shelf where they continued up to the Late *ensensis* Zone (Alekseev *et al*., 1996). On the eastern slope of Urals the regressive stage (= upper If eustatic fall of Johnson *et al*., 1985), is evidently reflected in a hiatus at the base of the Ulutau Formation, as is observed near the eastern foot of the Irendyk ridge. Possibly, at that time, olistostromes of the Gadilev Formation with olistoliths of biohermal limestones were also formed.

Across the Zilair Synclinorium, the Eifelian transgressive-regressive couplet is indicated by widespread Eifelian cherts and their disappearance (or at least decrease) near the Eifelian-Givetian boundary. This stratigraphical pattern is observed both in northern and southern areas of the synclinorium.

FRASNIAN-FAMENNIAN TRANSGRESSIVE-REGRESSIVE CYCLE

A widespread transgression, similar to the Eifelian one, occurred in shallow-water basins of the East European Platform at the beginning of the Late Devonian. In general, very high sea levels continued almost to the end of Frasnian (Johnson *et al*., 1985; Tikhomirov, 1995; Alekseev *et al*., 1996; Racki, 1997; Yunusov *et al*., 1997; Fokin and Nikishin, 1999). In the mid and late Frasnian (*punctata* to *rhenana* zones) cherts and siliceous argillites (Mukasovo Formation) were also deposited in a back-arc basin of the Urals Palaeozoic ocean. These monotonous sediments, devoid of coarser clastic material across most of the basin (excluding the zone with active volcanism), may indicate flooding of sediment-source areas. Thus, the marine nearshore zone moved up the basin slope and supply of terrigenous material into the deep-water part of the basin decreased. In the eastern part of the Zilair Synclinorium, interbeds of quartz sandstones, typical of the Givetian Stage, disappear in the Frasnian. This was accompanied by the development of a succession of cherts (Ibragimovo Horizon). The temporal coincidence of the Mukasovo Formation and Ibragimovo Horizon with the eustatic sea level rise recorded in epeiric domains (including the East European Platform), suggests common eustatic controls on the onset of siliceous sedimentation.

The thick pile (up to 700–800 m) of olistostromes and debrites, deposited above the cherts in the lower part of the Zilair flysch, probably indicates a sea level fall. Their formation, as shown above, occurred synchronously, mainly in the *triangularis* Zone and can be correlated with co-eval signatures of regression across all continents (e.g. Johnson *et al*., 1985; Southgate *et al*., 1993; Veimarn *et al*., 1996, 2002; Racki, 1997, 1998; Tsien and Fong, 1997; Schindler *et al*., 1998). A direct connection with collision processes and major overthrusts, inferred by some researchers (e.g. Salikhov, 1997; Ivanov, 1998), seems unlikely, especially because clasts in the olistostromes complexes do not include allogenic material from an island arc or its slope. Besides, neither the geotectonic character of the basin, nor the facies types, changed after this apparently regressive event.

It is here regarded that deposition of the Biyagoda olistostrome and of the debrites of the Bugodak succession is connected with the reworking of the tephra masses accumulated within the nearshore zone of volcanic islands during the transgression. This gravity collapse occurred in the zone of enhanced erosion as a result of the sea level fall. Tephra flows carried away fragments of the lava cover, and deeply eroded underlying Frasnian and upper Givetian limestones. In the southern part (at the latitude of Sibai and southwards) volcanic activity ended at the end of Frasnian. In this area, the regression was accompanied by subaerial exposure and brecciation of upper Frasnian reefs leading to the formation of the Koltuban clumpy conglomerates. Similar large-scale reef-margin collapses and breakup of carbonate platform margins, recorded in the deposition of slope megabreccias and debris flows, are known in the F-F interval in many parts of the world (e.g. Hladil *et al*., 1991; Holmes and Christie-Blick, 1993; Southgate *et al*., 1993; Veimarn *et al*., 1997; Racki, 1998).

It shoud be noted that the olistostromes and debrites described are composed of the material from an island arc and its shelf zone. The rudaceous mass occurs in the basal part of the Bolshekizil Formation. Significant debris flows and submarine slides did not occur on the slopes of the Uraltau cordillera, a feature probably related to the absence of clastic accumulations within the nearshore zone. During the Frasnian, the Uraltau rise area remained completely submerged. The uplift of the cordillera began only at the end of the Frasnian, in the *linguiformis* Zone. Therefore rudaceous strata are absent in the eastern part of Zilair Synclinorium in the earliest Famennian. Only westward, near the East European Platform margin, are limestone pebbles and olistoliths, containing conodonts of the *triangularis* Zone, seen in the flysh deposits.

Upper Devonian conglomerates with blocks and boulders of Frasnian limestones are also widespread in the southern continuation of the Magnitogorsk megazone, in the southern part of Western Mugodzhary (Kochetkova *et al*., 1987). According to Veimarn *et al*. (2002), this succession represents the Late *triangularis* to *crepida* zones.

FINAL REMARKS

The main factors determining conditions of sedimentation in deep-sea Devonian basins of the Southern Urals were geodynamic setting, tectonics and volcanism. There were also certain periods of a considerable influence of eustatic changes, whose origin, nevertheless, may be also connected to global geodynamic patterns (e.g. Racki, 1998). Stable bottom currents locally influenced the distribution of clastics, as is seen from the elongation of deep-water fans in submeridional direction (using modern coordinates). There is only indirect evidence of regional climate. Possibly, it was arid, as the clastics show no traces of intensive chemical weathering.

Analysis of sedimentary and volcanogenic complexes of the Magnitogorsk megazone shows that considerable structural changes took place several times during the Devonian. In the Early Devonian, in the oceanic basin near the Uraltau microcontinent, a subduction zone and island arc developed. By the first half of the Eifelian the arc had collided with microcontinents, the subduction zone was displaced to the east, and a new island arc formed which in turn collided with the East Uralian continental block around the Devonian-Carboniferous boundary. The formation of two fore-arc and back-arc basins in the area analysed was connected to these events. The establishment of cordillera in the place of the Uraltau microcontinent had a considerable influence for changing the character of sedimentation on the Southern Urals territory. Before the Famennian, only tephra-derived and volcanomictic clastics assocciated with volcanic islands were brought into the Magnitogorsk megazone basins, while in the Famennian clastic deposits with a complex composition including sialic material were predominant. However, contrary to prevailing views (e.g. Puchkov, 1993, 2000; Ivanov, 1998), structural changes did not take place in this territory around the Frasnian-Famennian boundary (Mizens, 2002*a*, *b*). Continental collision began later, as late as the Carboniferous. The back-arc basin which formed, together with the western marginal sea where deposition of the Zilair Formation deposition took place, was situated at the accretionary margin of the East European continent, which was the source of clastics detritus.

Comparative analysis with transgressive-regressive cycles, recognised from Devonian epicontinental sequences (see Johnson *et al*., 1985), suggests that these global sea level fluctuations may be recognised in the Southern Urals deep-water settings. This eustatic control concerns primarily Eifelian and Frasnian-Famennian transgressive-regressive cycles. An alternative mechanism, large-scale facies changes due to collision processes and large overthrust movements (see Ivanov *et al*., 1986; Ivanov, 1998; Salikhov, 1997 and others), seems to be of secondary significance, as there is an absence of allogenic material (from an island arc and its slope) in the olistostromes.

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