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Earth-Science Reviews 79 (2006) 261 – 287
EARTH-SCIENCE REVIEWS
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Arc–continent collision in the Southern Urals

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Received 12 December 2005; accepted 8 August 2006
Available online 17 October 2006

Abstract

The Southern Urals of Russia contain what is arguably one of the best-preserved examples of an arc–continent collision in any Paleozoic orogen. The arc–continent collision history recorded in the rocks of the Southern Urals began in the Early Devonian with the onset of intra-oceanic subduction and the formation of the Magnitogorsk Arc and ended with its collision with the margin of Baltica during the Late Devonian. The Baltica margin consisted of a basement that was composed predominantly of rocks of Archean and Proterozoic age that, by the time of arc–continent collision, was overlain by Cambrian, Ordovician, Silurian, and Devonian sediments interpreted to have been deposited in rift-related grabens on the continental slope and rise, and on the shallow marine platform. The Magnitogorsk Arc consists of Early to Late Devonian island arc volcanic rocks and overlying volcaniclastic sediments. Arc–continent collision led to the development of an accretionary complex that includes shallowly and deeply subducted continental margin rocks, ophiolite fragments, and sediments that were deposited in a foreland-basin setting. The geochemistry of the Magnitogorsk Arc volcanic rocks, the structure of the arc–continent collision accretionary complex and the forearc, the high-pressure rocks beneath and along the suture zone, the mafic and ultramafic ophiolitic material, and the syn-tectonic sediments show that the Paleozoic tectonic processes recorded in the Southern Urals can be favorably compared with those in currently active settings such as the west Pacific.

Keywords: arc–continent collision; Southern Urals; arc-chemistry; high-pressure rocks; forearc basin; foreland basin; ophiolite

1. Introduction

Arc–continent collision has been an important tectonic process throughout geological time (Page and Suppe, 1981; Leggett et al., 1982; Charlton, 1991; Silver et al., 1991; Dorsey, 1992; Cousineau and St-Julien, 1992; van Staal, 1994; Rusmore and Woodsworth, 1994; Abbott et al., 1994; Abers and McCaffrey, 1994; Brown et al., 1998; Draut et al., 2002; Dewey, 2005; Huang et al., 2006) and is thought to be the major process by which the continental crust has grown (Rudnick, 1995). Arc–continent collision zones are also an important source of the world's mineral wealth, hosting large deposits of massive sulfides, gold,

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doi:10.1016/j.earscirev.2006.08.003

and platinum-group elements. Understanding the geological processes that have taken place in the crust during arc–continent collision is therefore of importance for our understanding of how collisional orogens evolve, how the continental crust grows, and how its mineral wealth is formed and preserved. A spectacular example of the importance of this process is the widespread occurrence of arc–continent collision zones in the Paleozoic orogens of Pangea (Leggett et al., 1982; Cousineau and St-Julien, 1992; Sengör et al., 1993; van Staal, 1994; Rusmore and Woodsworth, 1994; Puchkov, 1997; Brown et al., 1998). Most of these orogens have since been fragmented and dispersed, and are now found in a number of tectonic plates worldwide. The anatomy of many of their arc–continent collisions is often obscured by the plate fragmentation and the extensive reworking that occurred as a result of subsequent deformation, metamorphism, and intrusion. The Uralide orogen of Russia is one of the few Pangean orogens that remains intact. The Southern Urals contain what is arguably one of the best preserved of the Paleozoic arc–continent collisions zones, and the outcropping geology allows it to be compared to arc–continent collisions that are active today (Puchkov, 1997; Brown et al., 1998; Brown and Spadea, 1999; Alvarez-Marron et al., 2000; Brown et al., 2001). In this paper we present a review of the major results that have come out of the last ten years of our research into arc–continent collision in the Uralides. We focus on the Southern Urals, presenting a broad spectrum of data that is used to construct a model for the arc–continent collision that occurred there. The Geologic Time Scale 2004 (Gradstein et al., 2004) is used for the Paleozoic throughout this paper.

2. Geological framework of the Uralides

Extending for nearly 2500 km from near the Aral Sea in the south to the islands of Novaya Zemlya in the Arctic Ocean (Fig. 1A), the Uralide orogen of Russia records the Paleozoic collision of at least two intra-oceanic arcs with the margin of Baltica and its subsequent continent–continent collision with the Kazakhstan and Siberian plates during the assembly of Pangea. From west to east the Uralides are divided into a number of tectonic zones; the undeformed foreland basin, the foreland thrust and fold belt, the Magnitogorsk–Tagil Zone, the East Uralian Zone, and the Trans-Uralian Zone (Fig. 1A). For descriptive purposes, the Uralides are divided geographically into the Southern, Middle, Northern, Cis-Polar and Polar Urals.

The foreland basin comprises undeformed Late Carboniferous to Early Triassic syn-tectonic sediments that were derived from the growing Uralide orogen to the east (Mizens, 1997; Chuvashov, 1998). The foreland thrust and fold belt comprises deformed Late Carboniferous to Early Triassic sediments of the foreland basin, Paleozoic platform margin and continental slope rocks, Archean, Mesoproterozoic and Neoproterozoic rocks of the East European Craton (part of Baltica) and, in the Southern Urals, the Middle through Late Devonian arc–continent collision accretionary complex (hereafter referred to as accretionary complex) related to the collision of the Magnitogorsk Arc with the Baltica continental margin (Kamaletdinov, 1974; Brown et al., 1997, 2004, 2006) (Fig. 1B). The Magnitogorsk–Tagil Zone consists of Silurian to Devonian intra-oceanic island arc volcanic rocks and overlying volcaniclastic sediments of the Magnitogorsk and Tagil island arcs (Seravkin et al., 1992; Maslov et al., 1993; Brown and Spadea, 1999; Brown et al., 2001; Spadea et al., 2002; Herrington et al., 2002). The Magnitogorsk–Tagil Zone is sutured to the former continental margin of Baltica along the Main Uralian Fault. The East Uralian Zone is composed predominantly of deformed and metamorphosed volcanic arc fragments with minor amounts of Precambrian and Paleozoic rocks thought to represent continental crust (Puchkov, 1997, 2000; Friberg et al., 2000). The East Uralian Zone was extensively intruded by Carboniferous and Permian granitoids (Fershtater et al., 1997; Bea et al., 1997, 2002), forming the “main granite axis” of the Uralides. The East Uralian Zone is juxtaposed against the Magnitogorsk–Tagil Zone along the East Magnitogorsk–Serov-Mauk strike-slip fault system. The Trans-Uralian Zone is composed of Carboniferous volcano-plutonic complexes (Puchkov, 1997, 2000). Ophiolitic material and high-pressure rocks have also been reported (Puchkov, 2000). The contact between the East Uralian and Trans-Uralian zones is only exposed in the Southern Urals, where it is a serpentinite mélange. Rocks that unequivocally belong to either the Kazakhstan or Siberia plates do not outcrop in the Uralides. The remainder of this paper will deal with the Magnitogorsk Zone and the Southern Urals accretionary complex (Figs. 2 and 3), which record the arc–continent collision that took place between the Devonian margin of Baltica and the Magnitogorsk Arc.

3. The Paleozoic continental margin of Baltica in the Southern Urals

Rifting during the Late Cambrian to Early Ordovician led to the formation of a paleo-Uralian ocean basin with the Baltica passive margin facing it. The margin was built on basement that was composed predominantly of strongly folded and metamorphosed rocks of Archean

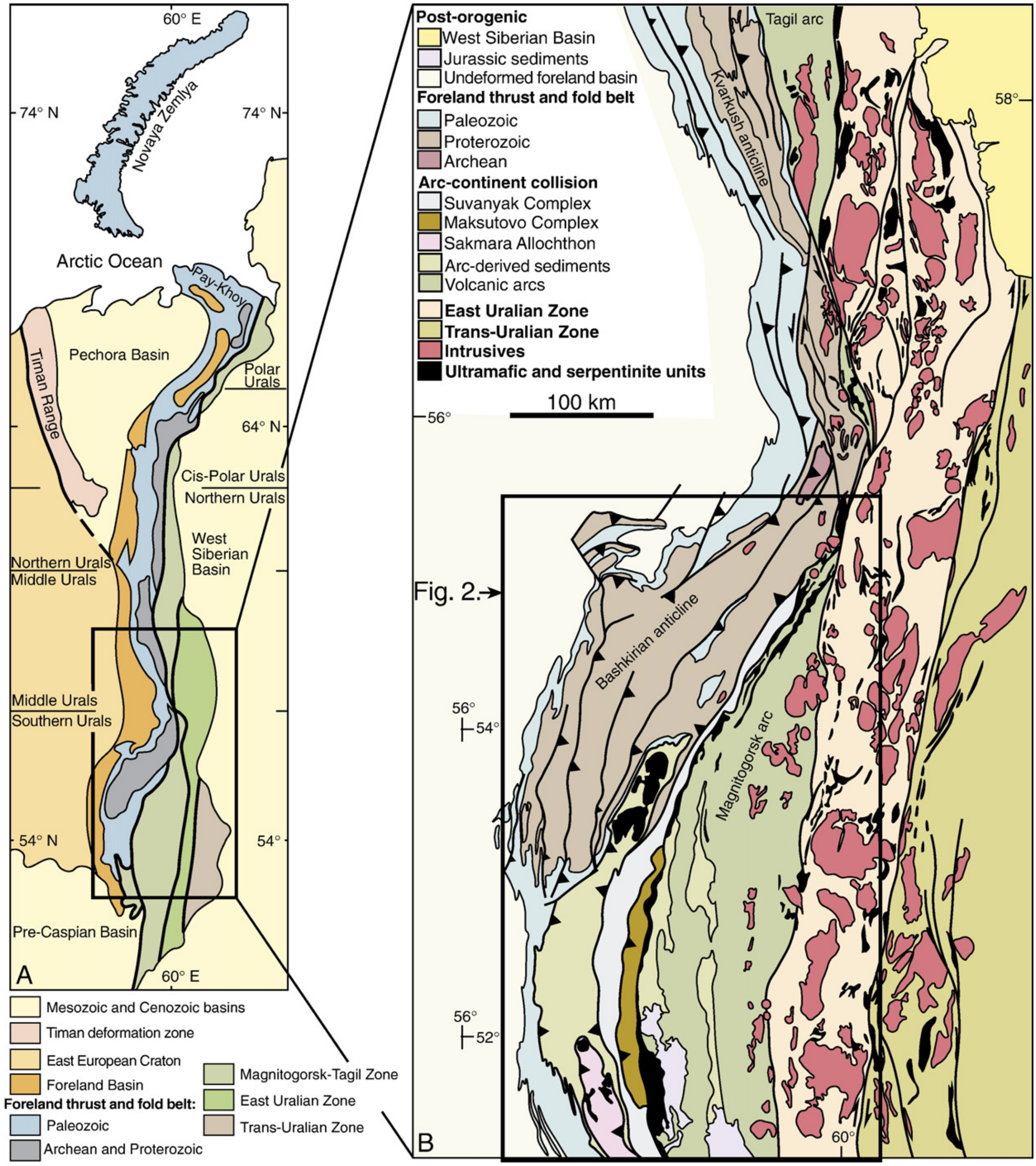

Fig. 1. A) Map showing the different zones of the Urals and its geographic divisions from north to south. The box indicates the area discussed in this paper. B) Geological map of the Southern and part of the Middle Urals. The legend shows the disposition of the various tectonic units discussed in this paper. The box indicates the location of Fig. 2.

age, and Mesoproterozoic (Riphean) sediments that filled aulacogens (Rusin, 1980; Maslov et al., 1997; Puchkov, 1997). Two major tectonostratigraphic units have been recognized in the basement: an Archean to Paleoproterozoic craton that was consolidated before 1.6 Ga and is overlain by Mesoproterozoic (Riphean) sediments, and a fold belt along the periphery of the craton that formed as a result of the Neoproterozoic (Late

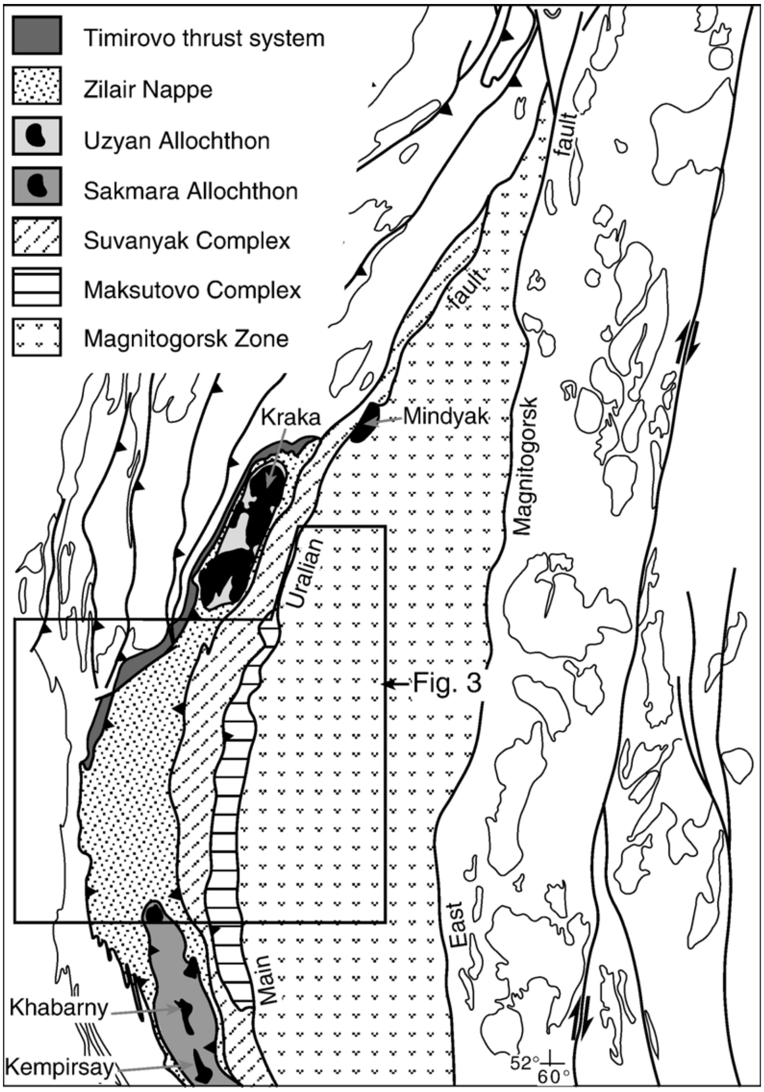

Fig. 2. Map of the Southern Urals showing the main tectonic units related to the arc–continent collision. The location of Fig. 3 is shown.

Vendian) Timanide Orogeny (Puchkov, 1997; Giese et al., 1999).

The Paleozoic continental margin sediments that entered into the collision with the Magnitogorsk Arc in the Middle Devonian consisted of a thin veneer of Ordovician, Silurian, and Devonian sediments above the Precambrian basement (Fig. 4) (Tyazheva, 1961; Einor et al., 1984). These rocks include sediments that are interpreted to have been deposited in rift-related grabens, sediments deposited on the continental slope and rise, and on the shallow marine platform (Puchkov, 2002). A well studied representative of the rift-stage facies in the Southern Urals is found in the Sakmara Allochthon (Fig. 1B) (Ruzhentsev, 1976; Khvorova et al., 1978) where it is composed of Upper Cambrian(?) to Ordovician (Tremadoc) coarse-grained, poorly sorted conglomerate, sandstone and siltstone with a highly variable thickness, and interlayers of trachybasalt, rhyolite tuffs, minor limestone, and jasper (Puchkov, 2002). These sediments are moderately folded and practically unmetamorphosed (Khvorova et al., 1978). Sediments that can be attributed to the continental slope and rise of the margin consist of polydeformed and metamorphosed quartzite, shale and phyllite that are currently found as separate, thrust-bound, units within the arc–continent collision accretionary complex (Puchkov, 2002). These rocks are described in more detail in Section 5. The shallow water platform margin consisted of up to 500 m of Ordovician

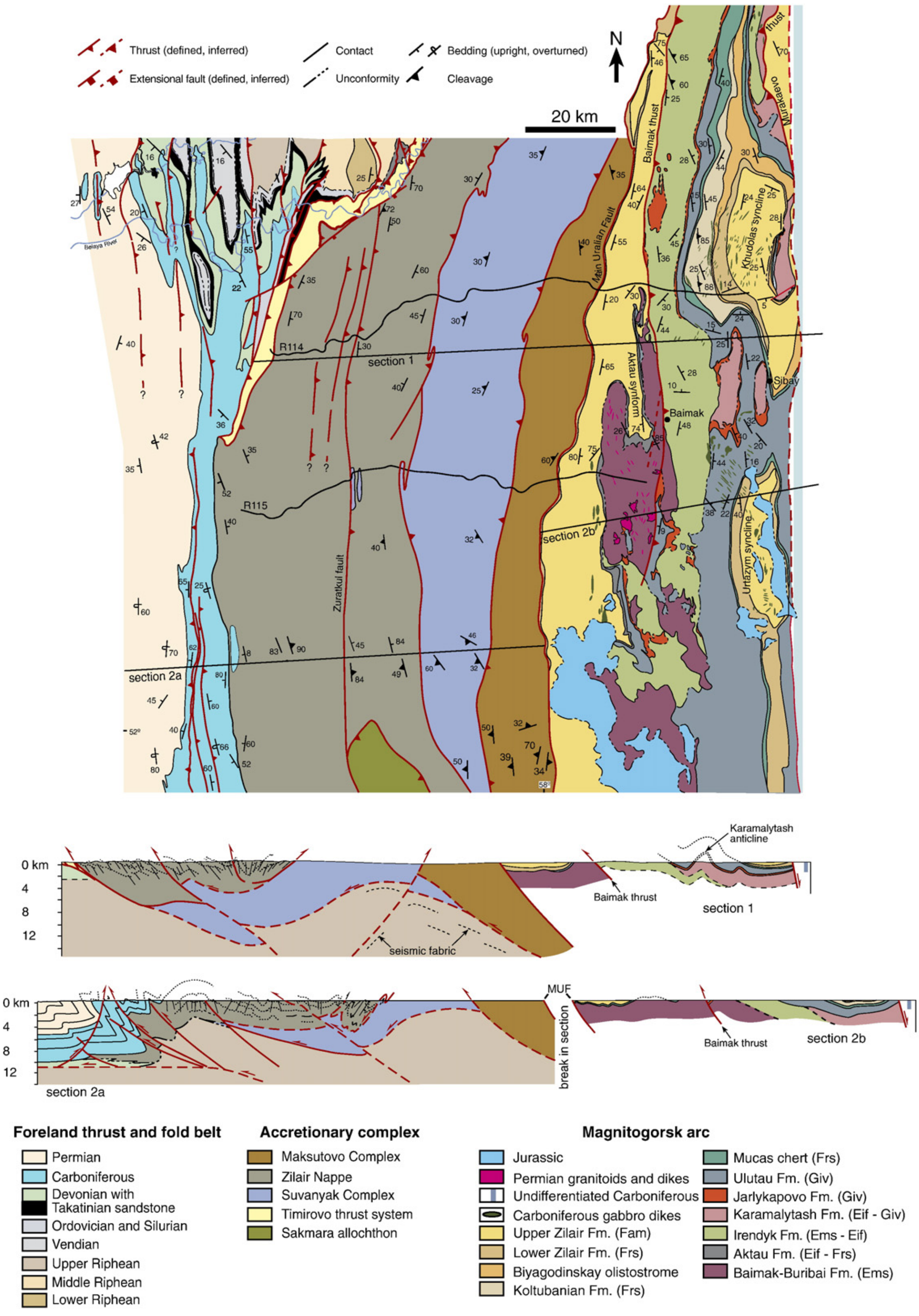

Fig. 3. A detailed geological map and cross-sections of part of the Magnitogorsk forearc and the accretionary complex. The location of the cross-sections are indicated on the map, as are the locations of seismic profiles R114 and R115 discussed in Section 5.

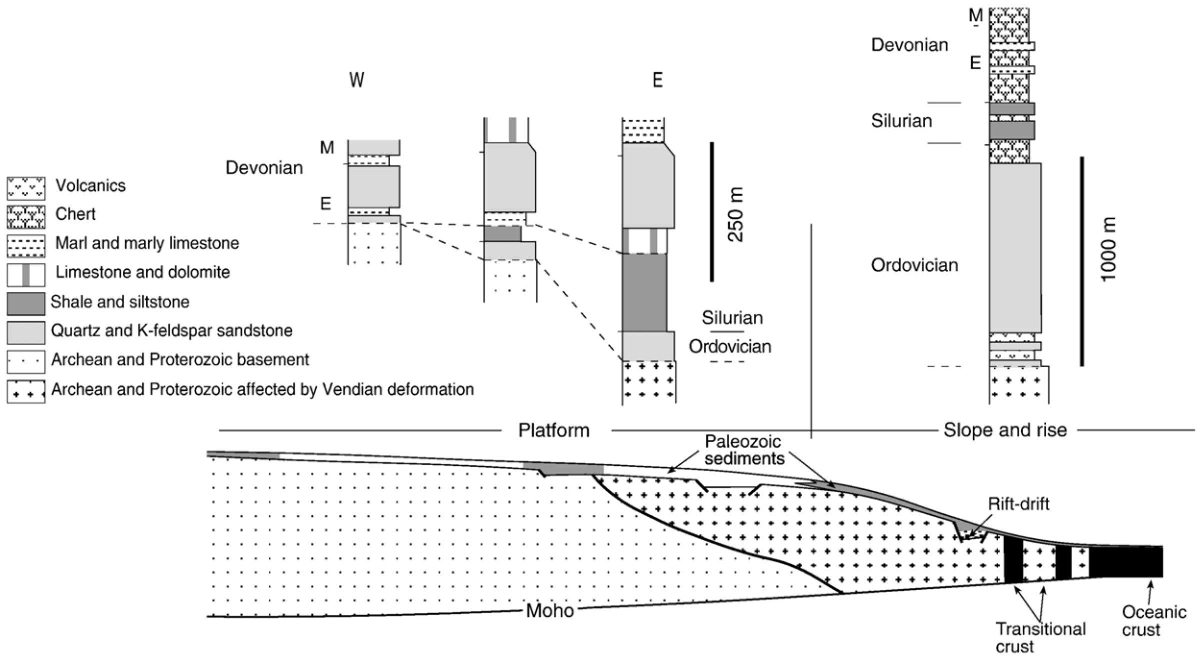

Fig. 4. An idealized section of the Devonian margin of Baltica that entered into the subduction zone beneath the Magnitogorsk Arc. The Paleozoic platform stratigraphy that was present at this time is represented by three schematic stratigraphic columns outcropping along the Belaya River, at the southern end of the Bashkirian Anticlinorium (see Brown et al., 1997). A schematic, composite rift-drift, slope, and rise facies stratigraphic column is interpreted from weakly deformed rocks found in the Sakmara Allochthon (after Puchkov, 2002).

and Silurian sandstone, conglomerate, siltstone, and limestone that pinch out westward (Einor et al., 1984). The Ordovician sequence is not present everywhere. Where it is present it displays rapid changes in thickness, suggesting deposition in grabens on a rifted basement. The Devonian consists of up to 1000 m of transgressive, westward thinning limestone, sandstone, interbedded limestone and sandstone, and marly limestone. A 100–800 m thick unit of Lower Devonian reef limestone disappears westward, and the Emsian sandstone of the Takatinian Formation unconformably overlie the Precambrian basement and the Silurian sediments (Maslov and Abramova, 1983; Bogoyavlenskaya et al., 1983; Chibrikova and Olli, 1991).

4. The Magnitogorsk Arc

4.1. The volcanic rocks

The oldest part of the Magnitogorsk Arc exposed in the Southern Urals is represented by the Emsian-age Baimak–Buribai Formation (Fig. 3). The Baimak–Buribai Formation is composed of picrite, high-Mg basalt and basaltic andesite and andesite lava flows intercalated with pyroclastics, hyaloclastites, agglomerates, volcaniclastics, thin beds of tuffaceous chert, and, locally, boninitic lavas and dikes (Kuz'min and Kabanova, 1991; Spadea et al., 1998). The boninites are petrographically distinct from the other arc rocks, consisting mostly of glassy, aphyric and sparsely olivine and pyroxene phyric rocks, containing chromite, and without plagioclase. They are chemically primitive (Mg# 81–68) and very low in TiO_2 (Fig. 5), are strongly depleted in high field-strength elements (HFSE) with respect to normal mid-ocean ridge basalt (N-MORB), and have a flat rare earth element (REE) pattern (Fig. 6A and B). The picrites have Mg# 81.4, very low TiO_2, and are the most HFSE-depleted rocks among the high-Mg lavas. The high-Mg basaltic andesite and andesite, are aphyric, or porphyritic, the latter being characterized by the presence of plagioclase phenocrysts. They are lower in MgO (Mg# 73–62) and slightly higher in TiO_2 than the boninites, are depleted in HFSE, and have a flat REE to slightly light rare earth element (LREE)-depleted pattern. The island arc tholeiites (IAT), basaltic andesites and andesites are aphyric, or porphyritic rocks with pyroxene (± plagioclase) phenocrysts. They have Mg# in the 68–42 range, low TiO_2 (0.4–0.8 wt.%), have low and variable HFSE MORB-normalized values, and a flat or slightly LREE-enriched pattern. The calc-alkaline (CA) basalts, basaltic andesites and andesites are mostly porphyritic with variable amounts of pyroxene, plagio-

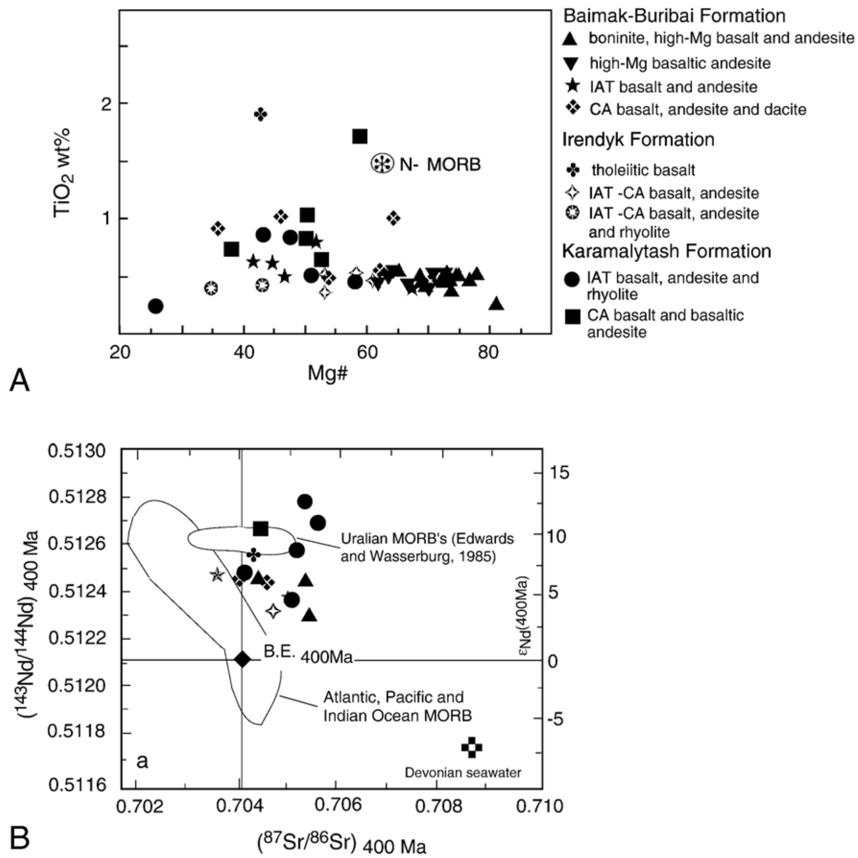

Fig. 5. A) TiO_2 versus Mg# for Magnitogorsk Arc Devonian extrusive rocks (after Spadea et al., 2002). B) Initial $^{143}Nd/^{144}Nd$ versus $^{87}S/^{86}Sr$ (back calculated to 400 Ma) for representative Magnitogorsk Zone Devonian extrusive rocks (after Spadea et al., 2002). The fields of Atlantic, Pacific and Indian MORB back calculated to 400 Ma, and Uralian MORB. The isotopic composition of Devonian seawater is from Burke et al. (1982) for Sr, and from Shaw and Wasserburg (1985) for Nd.

clase and olivine phenocrysts. They have Mg# in the 71–36 range, have higher TiO_2 contents than the other petrologic groups (0.5–1 wt.%), have depleted to slightly enriched N-MORB-normalized patterns, and have a relatively high REE content with a LREE-enriched pattern. Spadea et al. (1998, 2002) and Brown and Spadea (1999) interpret the Baimak–Buribai Formation to have been erupted in a suprasubduction-zone setting at an early stage of intra-oceanic convergence.

From the latter part of the Emsian through the lowermost Eifelian, volcanism evolved from the high-Mg basalts of the Baimak–Buribai Formation to island arc calc-alkaline suites of the uppermost Baimak–Buribai and Irendyk formations (Fig. 3). The Irendyk Formation consists of lava flows intercalated with pyroclastics, hyaloclastites, agglomerates, volcaniclastics, and, locally, thin beds of tuffaceous chert. Irendyk tholeiitic basalt has MORB-type characteristics with a high Fe_2O_3, with Mg# 43 (Fig. 5), a N-MORB-type HFSE pattern that is about 20 times chondritic, is LREE-depleted, and REE-enriched (Fig. 6C and D). IAT basalts and basaltic andesite are moderately fractionated (Mg# 61–54), typically low in TiO_2, are HFSE depleted (except P) with respect to MORB, and have a flat REE pattern that is 10 times less than chondritic. IAT andesites and dacites are low Mg# (44–35), low TiO_2 (around 0.4 wt.%), with MORB-type incompatible HFSE patterns and marked Ti depletion, and have REE contents that are more than 10 times chondritic and a marked negative Eu anomaly (Spadea et al., 2002).

The uppermost Eifelian Karamalytash Formation outcrops in the central and eastern part of the Magnitogorsk Zone (Fig. 3). It is composed of aphyric and plagioclase or clinopyroxene phyric basalt, and rhyolite with minor basaltic andesite and quartz andesite. In the central part of the zone, the IAT basaltic andesite of the Karamalytash Formation has a Mg# between 58 and 44, TiO_2 in the 0.4–0.8 wt.% range (Fig. 5), and other HFSE are slightly to strongly depleted with respect to MORB. Zr and Ti negative anomalies were identified in two

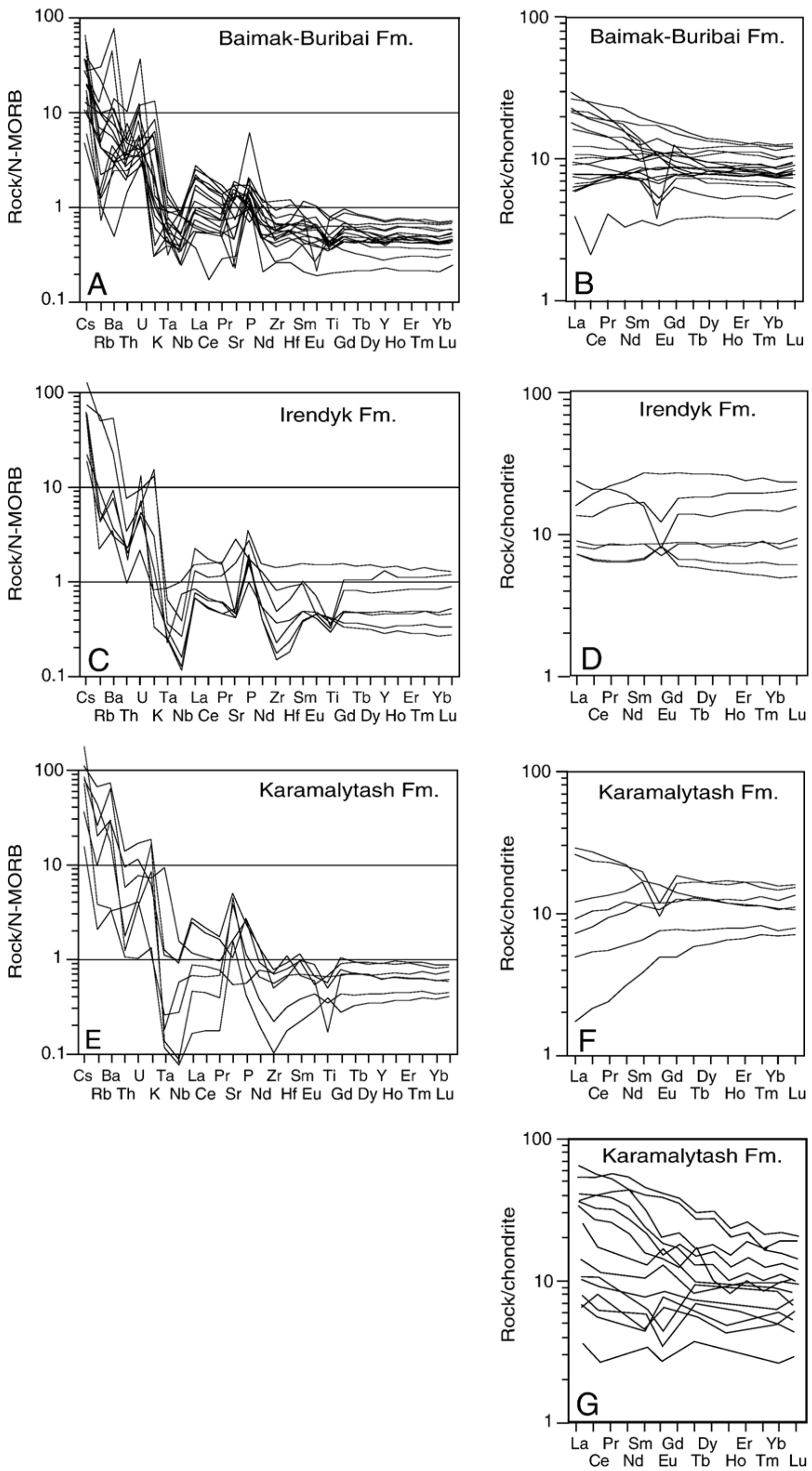

Fig. 6. Mid-ocean ridge basalt (N-MORB)-normalized incompatible element diagrams and chondrite-normalized rare earth elements diagrams for the Baimak–Buribai Fm. (A and B), the Irendyk Fm. (C and D) and the Karamalytash Fm. (E and F) (redrafted from Spadea et al., 2002). G) Chondrite-normalized rare earth elements diagram for the Karamalytash Fm. in the eastern area of the Magnitogorsk Arc (redrafted from Herrington et al., 2002).

samples. The REE patterns are flat to variably LREE-depleted (Fig. 6E and F). Calc-alkaline basalt and basaltic andesite have a Mg# between 52 and 38, is low in TiO_2 (0.63–0.72 wt.%), is characterized by low N-MORB-normalized HSFE patterns, and are LREE-enriched. In the eastern part of the Magnitogorsk Zone, the lower part of the Karamalytash Formation is characterized by high-Mg basalts with primitive REE patterns (Bochkarev and Surin, 1996). These basalts have an arc tholeiite signature, although they are poor in Ti (Frolova and Burikova, 1997). Rhyolites display two different affinities, one with a fractionated tholeiite signature and a second with a calc-alkaline signature (Gusev et al., 2000). Tholeiites have low Zr/Y ratios (5–7), indicative of an arc setting. The calc-alkaline rocks have LREE enrichment patterns with strong positive Eu anomalies.

From the Eifelian onwards volcanism shifted entirely to the eastern part of the Magnitogorsk Arc (Puchkov, 1997; Gusev et al., 2000; Herrington et al., 2002), to what we here call the East Magnitogorsk volcanic front. This area has not been as well studied as the west and central part of the zone. Below we summarize the existing data. In the northeast part of the Magnitogorsk Arc, the Givetian Kurosan Formation is composed of a continuous series of basalt (predominantly) to rhyodacite that is enriched in K and LREE (Fig. 6G). Elevated Sr-isotope values are suggestive of a contaminated magma source (Spadea et al., 2002). The Givetian to lowermost Frasnian Ulutau Formation consists mainly of andesite and dacite lava flows, near-vent breccias and sills intruded into tuffaceous and volcaniclastic sandstone and agglomerate (Yazeva et al., 1989). It comprises a series of largely calc-alkaline basaltic andesite through to rhyolite, with basaltic andesite and andesite forming around 40% of the sequence (Gusev et al., 2000). The Ulutau Formation is low-K calc-alkaline. Volcanism continued sporadically across the Magnitogorsk Arc until the earliest Bashkirian epoch of the Late Carboniferous, although they do not have a subduction zone geochemical signature. The East Magnitogorsk volcanic front was the primary source for the Late Devonian suture forearc sediment that is discussed below.

Preliminary lead-isotope data from the various volcanogenic massive sulfide (VMS) deposits associated with the Magnitogorsk Arc volcanic rocks indicate that there is broad spread between the mantle and upper crust curves, with a range of data of 0.15 for $^{207}Pb/^{204}Pb$ and 1.3 for $^{206}Pb/^{204}Pb$, straddling the Stacey–Kramers curve (S–K in Fig. 7) (Herrington et al., 2002). The VMS deposits in volcanic rocks found in the Main Uralian Fault plot closest to the mantle curve, suggesting they have a primitive mantle source. There is a clear difference in lead-isotope values for the VMS deposits found in the Baimak–Buribai Formation and those in the Karamalytash Formation and younger volcanic rocks. Those in the Baimak–Buribai Formation have lead that is less radiogenic, which is indicative of a source older than the age of the volcanics and implies a

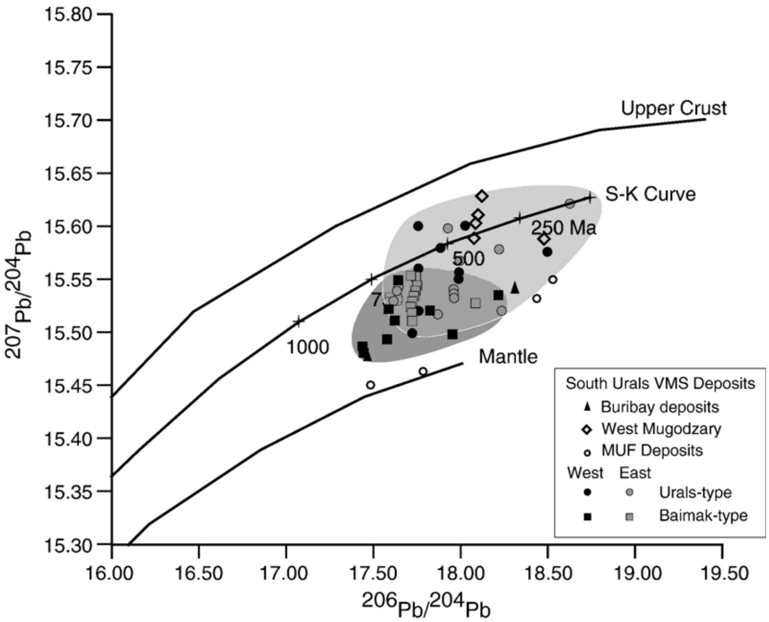

Fig. 7. Plot of $^{207/204}Pb$ versus $^{206/204}Pb$ for galena from a number of volcanogenic hosted massive sulfide deposits from the Magnitogorsk volcanic rocks (after Herrington et al., 2002).

contribution of lead from older subducted oceanic crust in the melt. In contrast, those deposits in the Karamalytash and younger formations are more radiogenic with lead isotopes, reflecting a mantle wedge source, with model ages close to the age of the volcanics with little or no contribution to the melt from subducted sediment indicated.

4.2. Forearc basin sediments

In the central and southern part of the Magnitogorsk Zone, the Middle to Late Devonian Aktau Formation outcrops along the western margin of the Baimak–Buribai Formation. In the Aktau synform (Fig. 3), it consists of up to 150 m of uppermost Eifelian to lowermost Frasnian interbedded tuffaceous sandstone, chert and jasper, and minor conglomerate overlain by the Frasnian Mukas chert and Frasnian to Tournaisian volcaniclastics of the Zilair Formation (Fig. 8A) (Maslov and Artyushkova, 1991). The Aktau Formation is thought to be a remnant of the forearc basin to the Irendyk volcanic front that remained isolated from sediment input from the East Magnitogorsk volcanic front until the latest Frasnian (Brown et al., 2001).

The Magnitogorsk forearc basin consists of up to 5000 m of westward-thickening Givetian to Famennian chert and volcaniclastic sediments (Fig. 8B) that overlie the volcanic edifice. The lowest unit is the Givetian Jarlykapovo Formation, which consists of several tens of meters to approximately 200 m of jasper and chert that overlie the Irendyk Formation and is coeval with the Karamalytash Formation. The Bugulygyr jasper, which directly overlies the Karamalytash Formation, is correlated with the uppermost part of the Jarlykapovo Formation.

The Jarlykapovo Formation is stratigraphically overlain by the Givetian to lowermost Frasnian Ulutau Formation, which outcrops in the western part of the Magnitogorsk Zone (Fig. 3). The Ulutau Formation sediments were sourced from the volcanic sequences of the same name in the east. It ranges in thickness from several hundred to approximately 2000 m, and consists of volcanomictic sandstone, with thin interbeds of tuffaceous chert, and, locally, thin beds of siliceous shale. Outcrops of olistostromes on the order of several hundreds of square meters and containing Ulutau material, volcanics and, locally, blocks of limestone, are widespread (Kopteva, 1981; Brown et al., 2001).

Throughout the Magnitogorsk Zone, the tuffaceous Mukas chert forms a 10 to 100 meter thick unit that everywhere overlies the Ulutau Formation. The Frasnian-age Koltubanian Formation is composed of up to 1500 m of westward-thickening volcanomictic siliceous sandstone, with interbeds of tuffaceous chert, grading into siliceous siltstones in the upper part of the formation. The upper part of the Koltubanian Formation is marked by widespread syn-sedimentary deformation and olistostromes.

The latest Frasnian and Famennian Zilair Formation consists of up to 2000 m of westward-thickening clastics. It is composed predominantly of volcanomictic sandstone and siltstone. Internal sedimentary structures such as graded bedding, cross-bedding and ripple marks are found locally, but are not common. Very few current markers have been found. The Zilair Formation in the Magnitogorsk Zone is the eastern time equivalent of the Zilair Formation in the Southern Urals accretionary complex (see below).

The sediments of the forearc basin are overlain by Early Carboniferous conglomerate and limestone. Locally, the contact is a steeply east-dipping extensional fault (Brown et al., 2001). The forearc region is locally intruded by Middle to Late Carboniferous gabbroic dikes (Seravkin et al., 1992) (Fig. 3). Undeformed Jurassic clastic sediments unconformably overlie the forearc sediments, the Early Carboniferous sediments, and the Main Uralian Fault.

4.2.1. Syn-sedimentary deformation

Syn-sedimentary deformation is widespread in the Magnitogorsk forearc basin. The Jarlykapovo Formation (including the Bugulygyr jasper) often displays tight to isoclinal intraformational folds, disrupted bedding, veining, and brecciation that is not reflected in the beds or units immediately overlying or underlying it, nor is the fold style reflected in the large-scale structure of the basin (see below), indicating that these features are syn-sedimentary (Brown et al., 2001). Locally, decameter-scale and larger olistostromes are developed in the Ulutau Formation (Kopteva, 1981; Brown et al., 2001).

The Mukas chert often displays chaotic folding, brecciation, and slumping on a variety of scales. Locally, within the Koltubanian Formation, coherent slumps can be mapped for hundreds of meters in structural thickness, and up to a kilometer in length (Kopteva, 1981; Brown et al., 2001). Such slumps are characterized by folded and brecciated horizons underlain and overlain by undeformed beds that everywhere have the same regional dip. Olistostromes on the scale of 10 to 100 m wide and larger are common in the Koltubanian and Lower Zilair formations. A spectacular example is the Biyagodinskay olistostrome (Fig. 3), which is developed at the base of the Zilair Formation (Kopteva, 1981; Brown et al., 2001; Veymarn et al., 2004). The

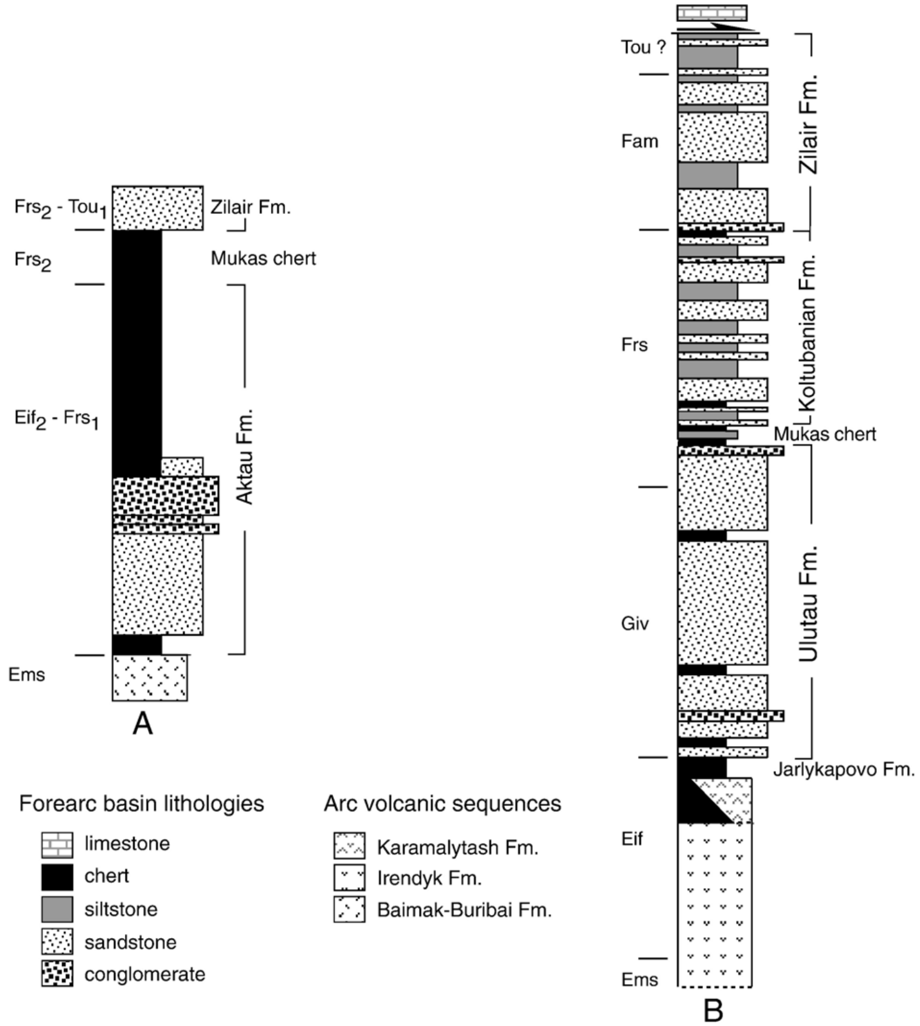

Fig. 8. A) Schematic stratigraphic column of the Aktau synform (after Maslov and Artyushkova, 1991). B) A simplified stratigraphic column of the suture forearc basin sediments (after Maslov et al., 1993; Brown et al., 2001).

Biyagodinskay olistostrome has an outcrop area of c. 250 to c. 300 km^2, and a thickness of several hundred meters. It consists of blocks of the Koltubanian, and possibly Ulutau formations, as well as several large blocks of volcanics, possibly the Karamalytash Formation, and limestone. Syn-sedimentary deformation is less common in the upper units of the Zilair Formation.

4.3. Structure of the Magnitogorsk Zone

The western part of the Magnitogorsk Zone is internally imbricated along the west-verging Baimak Thrust, which thrusts the Irendyk Formation in its hangingwall over the Baimak–Buribai, Aktau, and Zilair formations in its footwall (Fig. 3) (Brown et al., 2001). Northward, the Baimak Thrust merges with the Main Uralian Fault, and it can be traced in outcrop approximately 50 km south of the town of Baimak (Fig. 3). Farther south, the western part of the forearc basin is affected by open, west-verging, kilometer-scale folds that disappear northward, and the Baimak Thrust places Ulutau Formation on top of the Baimak–Buribai Formation (Fig. 3). Immediately west of Baimak, the eastern limb of the Aktau synform is overturned beneath a thrust that is interpreted to splay off the Baimak Thrust. To the east of the Baimak Thrust, the structure is that of a pair of elongate synforms, the Khudolas and Urtazym synclines (Fig. 3). These synclines are separated along the strike by a volcanic basement high that culminates in the Karamalytash Anticline. Both synclines are open and somewhat flat-bottomed, with gentle limb dips. The Karamalytash Anticline is a gently north–south plunging, slightly asymmetric fold, with nearly vertical limb dips along the eastern side of its volcanic core, and gentler dips on the western side (Fig. 3). On the basis of geometrical relationships between onlapping sediments and the volcanic rocks in the fold core, the Karamalytash Anticline has been interpreted to be a growth fold (Brown et al., 2001). Northward, the east-verging Murakaevo Thrust places Irendyk and Karamalytash formation rocks on top of the Zilair Formation. The Irendyk Formation volcaniclastics have been folded into a tight hangingwall anticline, and the Zilair Formation displays an open footwall syncline.

5. The Southern Urals accretionary complex

The Southern Urals accretionary complex(Brown et al., 1998; Brown and Spadea, 1999; Alvarez-Marron et al., 2000) is made up of the Timirovo thrust system, the Zilair Nappe, the Suvanyak Complex, the Uzyan Allochthon, the Sakmara Allochthon, and the Maksutovo Complex (Fig. 2). The internal structure of the individual units, their deformation style, metamorphic grade, and deformation mechanisms are distinct within each (Alvarez-Marron et al., 2000), reflecting deformation partitioning and the varied deformation conditions that each of the units underwent during its tectonic evolution. For this reason we describe them separately below.

The Timirovo thrust system (Figs. 2 and 3) outcrops continuously for nearly 100 km along the northwestern margin of the accretionary complex. It is composed of highly sheared Middle Devonian reef limestone with local intercalations of arkosic sandstone (Brown et al., 1997; Alvarez-Marron et al., 2000; Fernandez et al., 2004). Metamorphic grade in the Timirovo thrust system is low, and the regional conodont color alteration index ranges from 4.5 to 5 (Vladimir Baryshev, pers. comm.), suggesting a temperature of c. 250–300 °C. The Timirovo thrust system has the appearance of a duplex in which ductile shear zones display a well-developed, shallowly ESE dipping, foliation with an oblique, gently ENE plunging, stretching lineation (Ls) (Brown et al., 1997; Alvarez-Marron et al., 2000). Kinematic indicators, such as σ and δ clasts, S–C fabrics, and extensional crenulation cleavages, together with the stretching lineation, indicate a top-to-the-WSW sense of movement (Brown et al., 1997; Alvarez-Marron et al., 2000; Fernandez et al., 2004). Meso- and microscopic fabrics indicate a complex deformation path that involves pressure solution, brecciation and veining leading finally to low temperature ductile shearing (Fernandez et al., 2004). Measurements carried out on twinned calcite grains indicate that the differential stress affecting the Timirovo thrust system reached a peak of about 230 ± 40 MPa (Fernandez et al., 2004).

The Zilair Nappe (Fig. 2) is the largest and best-exposed unit within the Southern Urals accretionary complex. It consists of the latest Frasnian and Famennian polymictic and greywacke turbidites of the Zilair Formation. The Zilair Formation in the

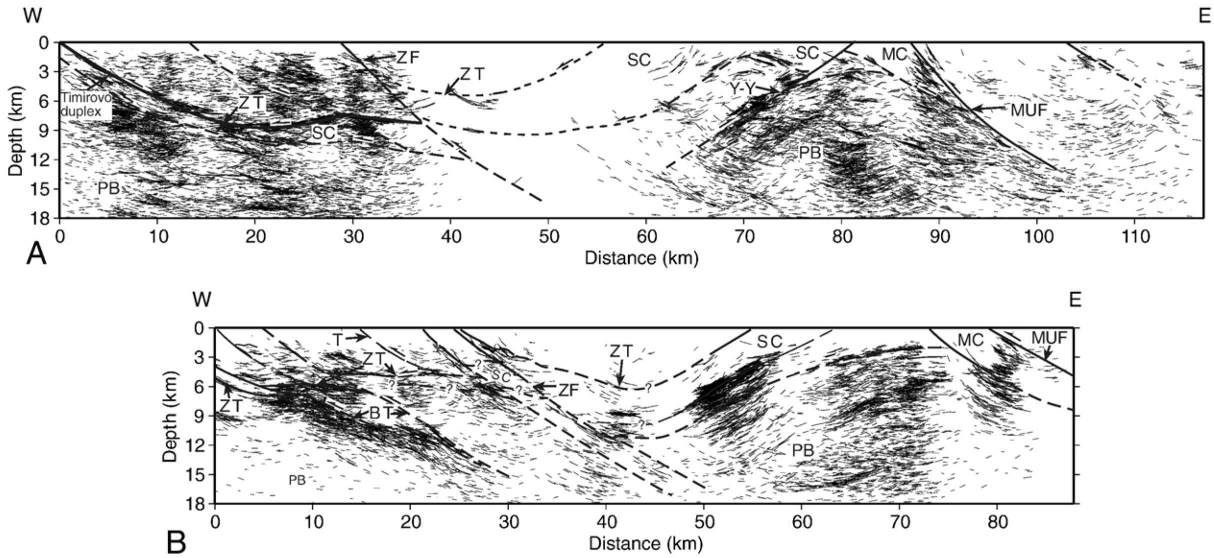

Fig. 9. A) Depth-migrated and interpreted automatic line drawings of reflection seismic profile R114. The missing areas represent data that could not be read from the original tapes. B) Depth-migrated and interpreted automatic line drawings of reflection seismic profile R115. ZT, Zilair Thrust; ZF, Zuratkul Fault; SC, Suvanyak Complex; PB, Precambrian basement; Y–Y, Yantyshevo–Yuluk Fault; T, unnamed thrusts; BT, breaching thrusts; MC, Maksutovo Complex; and MUF, Main Uralian Fault.

accretionary complex is thought to have been deposited in a foreland-basin setting to the arc–continent collision (Mizens, 1997; Brown et al., 1998; Brown and Spadea, 1999; Alvarez-Marron et al., 2000; Mizens, 2004). It has a structural thickness of 5 to 6 km (Keller, 1949; Pazukhin et al., 1996; Puchkov et al., 1998), and sedimentary structures such as graded bedding, parallel and cross-lamination, ripple marks and load casts are common, and provide good criteria for determining the way up. It is widely accepted that the Zilair Formation youngs westward, and that fine-grained, thin-bedded turbidites are more common in the west, whereas thicker beds of sandstones and conglomerates appear in the east. The metamorphic grade in the Zilair Nappe increases eastward from anchizone to greenschist facies (chlorite zone) (Bastida et al., 1997). The structure of the Zilair Nappe is that of a large-scale synform with a west-verging ramp anticline that, along its eastern margin, is affected by an east-verging thrust (Fig. 3). Asymmetric minor folds (F_1) are developed on a number of scales. In outcrop, meter-scale folds are generally tight (with a $\sim 60°$ interlimb angle), locally overturned to the west, and plunge gently NNE or SSW (Bastida et al., 1997; Alvarez-Marron et al., 2000). In many cases, the forelimbs of these folds are cut by minor thrust faults. In thin-bedded sandstone and shale, F_1 folds have a well-developed, moderately east-dipping, penetrative axial planar pressure solution cleavage (S_1) that, in the thick-bedded sandstones, is a spaced fracture cleavage. Along the eastern margin of the nappe, folds verge to the east and the cleavage fans until it dips westward (Fig. 3). Thrusts are typically wide zones of intense cleavage development that occupy the forelimb areas of kilometer-scale folds. A small number of minor thrusts have been identified within the nappe, and moderately east dipping, discontinuous reflections in seismic reflection profiles R114 and R115 may image thrusts (Fig. 9) (Alvarez-Marron et al., 2000; Brown et al., 2004). The western boundary of the Zilair Nappe is imaged on profile R114 (Fig. 9A) as a thin band of reflections that dips moderately eastward where it flattens into an openly concave downward band of reflections. In R115 (Fig. 9B), the base of the Zilair Nappe is imaged as bands of stepped, diffuse, subhorizontal reflections that we interpret to indicate that the Zilair Thrust is imbricated in several places.

The Suvanyak Complex outcrops along the east–central part of the accretionary complex (Figs. 2 and 3). It consists predominantly of moderately to strongly deformed chlorite — white mica metaquartzite, quartz— albite schist, and chlorite — white mica phyllite thought to represent the Paleozoic slope sediments of the Baltica continental margin (Zakharov and Puchkov, 1994; Puchkov, 1997). Poor exposure makes it impossible to determine the detailed structure and contact relationships within the Suvanyak Complex (Alvarez-Marron et al., 2000). Its regional-scale structure is that of an upright antiform (the Uraltau Antiform) that folds an earlier, locally penetrative foliation (Hetzel et al., 1998; Alvarez-Marron et al., 2000). The main fold phase found in the Suvanyak Complex is a kilometer-scale, moderately southwest-plunging, mildly non-cylindrical (and locally reclined) fold system (F_2) with an axial planar foliation (S_2) defined by chlorite and white mica. In fold hinges, a chlorite and or white mica-defined foliation (S_1) is preserved in lithons, indicating an earlier phase of deformation. Zones of intense shearing, with rare top-to-the-west shear sense indicators are found locally. Outcrop conditions hamper the correlation of these shear zones along the strike and determination of their relationship with the folding. The metamorphic grade appears to increase eastward from lower to middle greenschist facies. In R114 and R115 (Fig. 9), the Suvanyak Complex is imaged as diffuse reflectivity with thin bands of curved, east- and west-dipping reflections that extend beneath the Zilair Nappe. It is not clear how far westward the Suvanyak Complex extends, but on the basis of R114 and R115 we interpret it to extend to near the middle of the Zilair Nappe (Fig. 3).

In this paper we group the previously defined Uzyan Nappe (e.g., Brown et al., 1996; Puchkov, 1997) together with the Kraka ultramafic massif (in the same way that the Sakmara Allochthon contains ultramafic massifs and Paleozoic sediments) to form the Uzyan Allochthon (Fig. 2). The Uzyan Allochthon is made up of Ordovician to Middle Devonian quartz sandstone, siltstone, and shale, with layers of chert predominating in the top of the succession (Puchkov, 2002). These sediments are thought to have been deposited in a deep-water environment on the continental slope and rise. Basalt flows have been reported in the sedimentary sequence (Puchkov, 2002), but neither their type, nor their stratigraphic position is known for certain. Metamorphism in the Uzyan Allochthon sediments reaches lower greenschist facies. The Kraka massif structurally overlies the sediments of the Uzyan Allochthon. Kraka forms a 0.3 to 3 km thick thrust sheet that is composed predominantly of weakly metamorphosed lherzolite, but also contains dunite, harzburgite, pyroxenite, wehrlite, and gabbro, all folded into an open synform and with a well-developed foliation (Savelieva, 1987; Savelieva et al., 2002). The basal contact between Kraka and the underlying sediments is a several hundred meter-thick tectonic mélange. Basaltic pillow lava and chert occur as

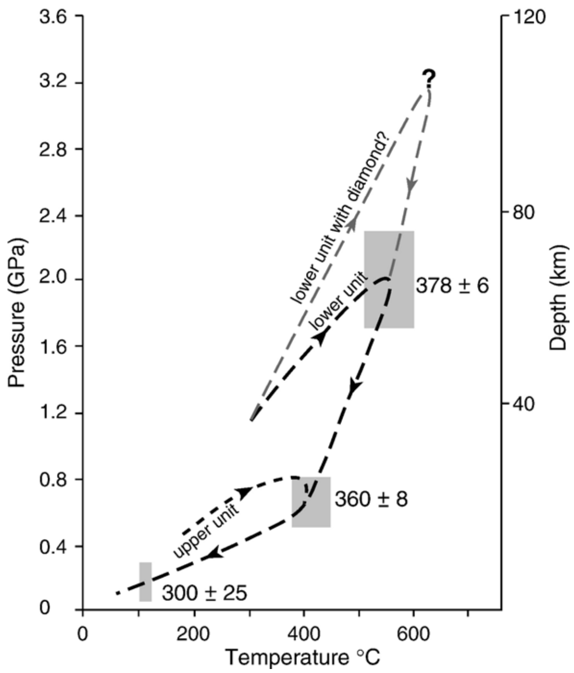

Fig. 10. Composite *P–T–t* diagram from the Maksutovo Complex. The data are from the references in the text. Ages are averages as determined by Hetzel and Romer (2000).

blocks in the mélange (Puchkov, 2000), but the crustal part of an ophiolite sequence is missing. Garnet amphibolites and garnet pyroxenite have also been found in blocks in the mélange (Savelieva et al., 1997) suggesting a dismembered metamorphic sole. On the basis of its normal mid-ocean ridge basalt REE signature, Savelieva et al. (1997) interpret Kraka as a fragment of oceanic mantle, although enrichment of large ion lithophile elements has led Fershtater and Bea (1996) to suggest that it may be a fragment of the continental mantle. Paleontological and radiometric age data are not available for Kraka. The structure of the Uzyan Allochthon is not well known because of poor exposure in the sedimentary rocks, but it appears to consist of several thin, west-verging thrust sheets, overlain by the Kraka massif.

The Sakmara Allochthon (Fig. 2) is made up of several imbricate thrust sheets consisting of Ordovician to Devonian shale, sandstone, chert and limestone interpreted to have been deposited on the continental slope and rise (Puchkov, 2002). Thrust sheets and a mélange consisting of pillow lava and chert, tuffaceous turbidites, polymictic olistostromes, belonging to oceanic and island arc crust, all in a serpentinite matrix, structurally overlie these. The Kempirsay and Khabarny ultramafic massifs form the uppermost thrust sheets. The best studied of these, Kempirsay, is composed of mantle tectonite of mostly harzburgite composition, accompanied by cumulates, gabbro, a sheeted dike complex, and basaltic pillow lavas capped with Middle Ordovician phtanites (Saveliev and Savelieva, 1991; Savelieva and Nesbitt, 1996; Savelieva et al., 1997; Melcher et al., 1999; Savelieva et al., 2002). Two Sm–Nd isochron ages (397±20 and 396±33 Ma) were determined from Kempirsay gabbros (Edwards and Wassenburg, 1985), but more recently Sm–Nd isochrons from two lherzolite samples yielded ages of 487±54 Ma, and a U–Pb zircon age of 420±10 Ma was obtained from dikes of pyroxene gabbro composition crosscutting ultramafic rocks (see Savelieva et al., 2002).

The Maksutovo Complex (Fig. 2), which crops out for ~180 km along the eastern margin of the accretionary complex, is composed of two tectono-metamorphic units that are structurally juxtaposed (Lennykh, 1977; Zakharov and Puchkov, 1994; Dobretsov et al., 1996). The structurally lowest unit consists predominantly of glaucophane-bearing metasediment and minor amounts of mafic eclogite and blueschist. Eclogite mineral phases record prograde growth zoning and a clockwise *P–T* path (Fig. 10), with pressure conditions having been estimated at about 2.0±0.3 GPa and 550±50 °C, followed by retrograde blueschist and greenschist facies conditions that indicate cooling during exhumation (Beane et al., 1995; Hetzel et al., 1998; Schulte and Blümel, 1999). Recently, microdiamond aggregates have been described from the Maksutovo Complex (Bostick et al., 2003), suggesting that even higher pressures (minimum of 3.2 GPa) and a temperature of 650 °C may have been reached (Fig. 10). However, Sr-isotope equilibrium among eclogite facies phases and among eclogite facies fluid veins and host rock, together with the prograde fabrics recorded in them, indicates that these rocks do not record any evidence of retrograde metamorphism from this apparent ultrahigh-pressure event (Hetzel et al., 1998; Glodny et al., 2002). The age of eclogite facies metamorphism in the lower unit has been constrained by various isotopic systems at a mean of 378±6 Ma (Matte et al., 1993; Lennykh et al., 1995; Beane and Connelly, 2000; Hetzel and Romer, 2000; Glodny et al., 2002), or Frasnian. Recent U–Pb SHRIMP dating of zircons has yielded slightly older ages, 388±4 (Leech and Willingshofer, 2004). The upper unit of the Maksutovo Complex is composed of Paleozoic metasedimentary and metavolcanic rocks. A serpentinite mélange at the base of the upper unit (and in contact with the lower unit) contains metarodingites that pseudomorph lawsonite. These rocks experienced peak metamorphic conditions of $P<0.8$ GPa and temperatures <450 °C (Dobretsov et al., 1996; Hetzel et al., 1998), although recent estimates by Schulte and Sindern (2002) suggest that P = 1.8–2.1 GPa and a T of 520–540 °C may have been reached (Fig. 10). The lower and upper units were structurally juxtaposed after eclogite facies metamorphism in the lower unit along a top-to-the-ENE retrograde shear zone (Hetzel, 1999) that yielded Rb–Sr and Ar–Ar ages of 360 ± 8 Ma (Beane and Connelly, 2000; Hetzel and Romer, 2000). Retrograde conditions in the upper unit have been determined to be >0.45 GPa at T < 440 °C at 339 Ma (Schulte and Sindern, 2002). Ar–Ar white mica ages from retrograde shear zones in the Maksutovo Complex range from 370 to 332 Ma (Beane and Connelly, 2000; Hetzel and Romer, 2000) and Rb–Sr ages cluster around 360 Ma (Hetzel and Romer, 2000). Apatite fission track data indicate that the Maksutovo Complex passed through the 110 °C geotherm at 300 ± 25 Ma (Leech and Stockli, 2000). The lower unit of the Maksutovo Complex is thought to have been derived from the Baltica continental margin, whereas the upper unit has been interpreted as containing remnants of the Uralian ocean crust (Hetzel, 1999) and fragments of island arc material (e.g., Dobretsov et al., 1996). The structure of the Maksutovo Complex is not well known, although Hetzel et al. (1998) and Leech and Ernst (2000) show it to be complexly deformed with, locally, two foliations developed during prograde metamorphism and two foliations developed during retrograde metamorphism (Hetzel et al., 1998). In seismic profiles R114 and R115 (Fig. 9) it is imaged as moderately east-dipping reflections against which the Suvanyak Complex and Precambrian basement reflectivity are truncated, suggesting that the Maksutovo Complex was thrust westward over these units. Final emplacement appears to have taken place between about 340 Ma (the young Ar–Ar ages) and the cooling through 110 °C at about 300 Ma given by apatite fission track modeling. The Maksutovo Complex is unconformably overlain by Lower Jurassic sediments.

6. Late overprinting

The Southern Urals accretionary complex was mildly overprinted by the Late Carboniferous to Triassic deformation event related to the continent–continent collision that took place between the Kazakhstan–Siberia and Baltica plates (Zonenshain et al., 1984, 1990). Baltica, by that time, included the accreted Magnitogorsk and Tagil arcs. Folds and thrusts in the accretionary complex, especially in the Zilair Nappe, are recognizable as Devonian by the pervasive and intense cleavage associated with them (Alvarez-Marron et al., 2000). This cleavage is not developed in the Carboniferous sediments that overlie the Zilair Nappe along its southwestern margin, indicating that deformation in the accretionary complex had stopped by the end of the Devonian (Brown et al., 1997, 2004). Thrusts that can be recognized from the seismic profiles as breaching the accretionary complex are generally associated with localized steepening (even overturning) and folding of the S_1 cleavage (Brown et al., 2004). The most significant of these, the Zuratkul Fault, is interpreted in the reflection seismic data (ZF in Fig. 9) to penetrate into the middle (and perhaps lower) crust and to offset the base of the accretionary complex. In profile R115 (Fig. 9B), two zones of moderately east-dipping reflectivity beneath the western part of the Zilair Nappe are interpreted to represent thrusts that breach the accretionary complex. The Yantyshevo–Yuluk Fault (Y–Y in Fig. 9), imaged in R114 (Fig. 9A) by folding and truncation of reflections, outcrops as an east-directed thrust that locally places the Suvanyak Complex on top of the Maksutovo Complex (Zakharov and Puchkov, 1994).

7. The Main Uralian Fault

The paleo-margin of Baltica, including the accretionary complex, is sutured to the Magnitogorsk Zone along the Main Uralian Fault (Figs. 1 and 2). In the Southern Urals, the Main Uralian Fault is an up to 10 km wide mélange consisting predominantly of serpentinite in which fragments of Devonian volcanics and sediments derived from the Magnitogorsk Arc are found. It also contains olistostromes and several lenticular or wedge-shaped slabs of mantle ultramafic rocks that measure several hundred m^2 to a km^2 in size (e.g., Nurali, Mindyak, Baiguskarovo and others). These bodies are mostly lherzolitic in composition, with an upper unit consisting of a crust–mantle transition zone made of mantle restites and ultramafic cumulates that are intruded by gabbro and diorite (Savelieva, 1987; Savelieva et al., 1997; Garuti et al., 1997; Savelieva et al., 2002). One of them, the Mindyak lherzolite, is of special interest since it was subducted and metamorphosed during the early stages of intra-oceanic subduction.

The Mindyak lherzolite massif has a geochemical signature that suggests that it formed in an intra-oceanic suprasubduction-zone setting (Scarrow et al., 1999). Garnet–tschermakite/pargasite–diopside-bearing blocks within a tectonic breccia in the upper mantle sequences of the Mindyak massif represent metamorphosed gabbros that originally crystallised at low pressures followed by low temperature rodingite alteration and

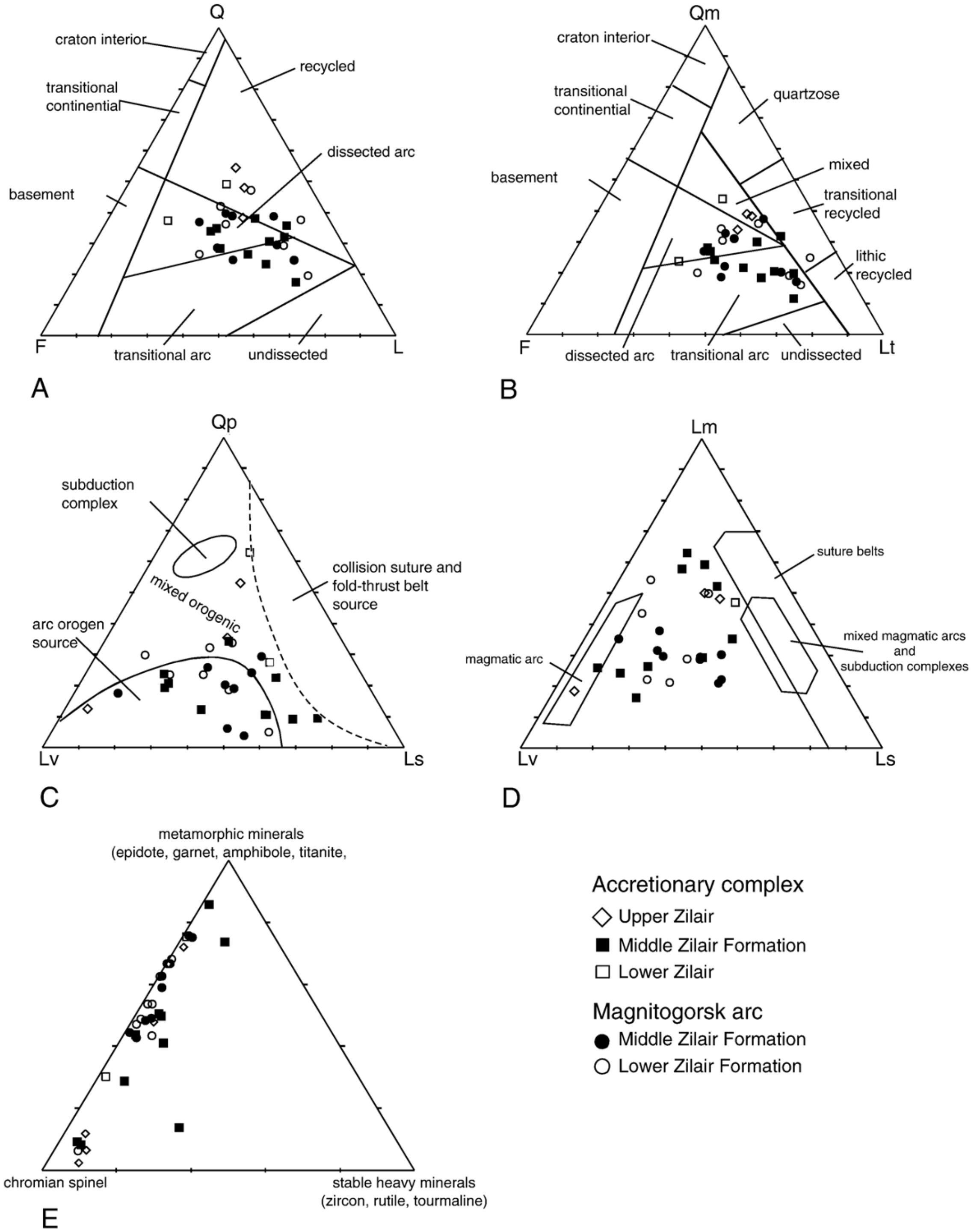

Fig. 11. The light mineral spectrum of the Zilair Formation in the accretionary complex and the Magnitogorsk Arc (redrafted from Willner et al., 2002). A to C) discrimination diagrams after Dickinson (1985). Qm = monocrystalline quartz, Qp = polycrystalline quartz, Lt = L+Qp, Ls = sedimentary lithoclasts, Lv = volcanic lithoclasts. D) Variation of sedimentary (L_S), metamorphic (L_m) and volcanic lithoclasts (L_V). Discrimination fields after Ingersoll and Suzcek (1979). E) Variation of metamorphic heavy minerals, stable heavy minerals, and chrome spinel.

then metamorphism under upper mantle conditions ($T\sim800$ °C and $P\sim1$ GPa) (Cortesogno et al., 1998) at approximately 415–410 Ma (Savelieva et al., 1999; Scarrow et al., 1999). This was followed by a retrograde re-equilibration to amphibolite/greenschist facies conditions at about 385 Ma (pers. comm., M. Cosca, 1999). Because of its mélange character, kinematic criteria that reliably indicate a consistent movement direction are lacking from the Main Uralian Fault. Reflection seismic profiling suggest that the Main Uralian Fault dips eastward and extends into the middle crust (Echtler et al., 1996; Knapp et al., 1996; Brown et al., 1998). In the northernmost part of the Southern Urals (c. 55° N), it has been intruded by rocks of the Syrostan batholith that has a deformed phase dated at 333±3 Ma, and an undeformed phase dated at 327±2 (Montero et al., 2000). The undeformed phase of the Syrostan batholith marks the end of tectonic activity along the Main Uralian Fault in the Southern Urals (Ayarza et al., 2000).

8. Provenance of syn-collisional sediments

The Zilair Formation, described above from both the Magnitogorsk Zone and the Southern Urals accretionary complex, represents sediments that were deposited during the late stage of arc–continent collision in a forearc and foreland-basin setting, respectively. The strong angularity of the majority of the clasts, together with the presence of low stability minerals such as K-feldspar, amphibole, pyroxene, biotite, garnet or carbonate, and the frequent local enrichment of specific heavy minerals, point to a very low degree of transport abrasion and therefore a nearby source area (Willner et al., 2002). Modification of the detritus during weathering and erosion from its source, its transport to the depositional basin, its deposition and its diagenetic interaction with pore waters and neighbouring minerals is relatively low. The composition of the Zilair Formation is therefore mainly related to sediment provenance.

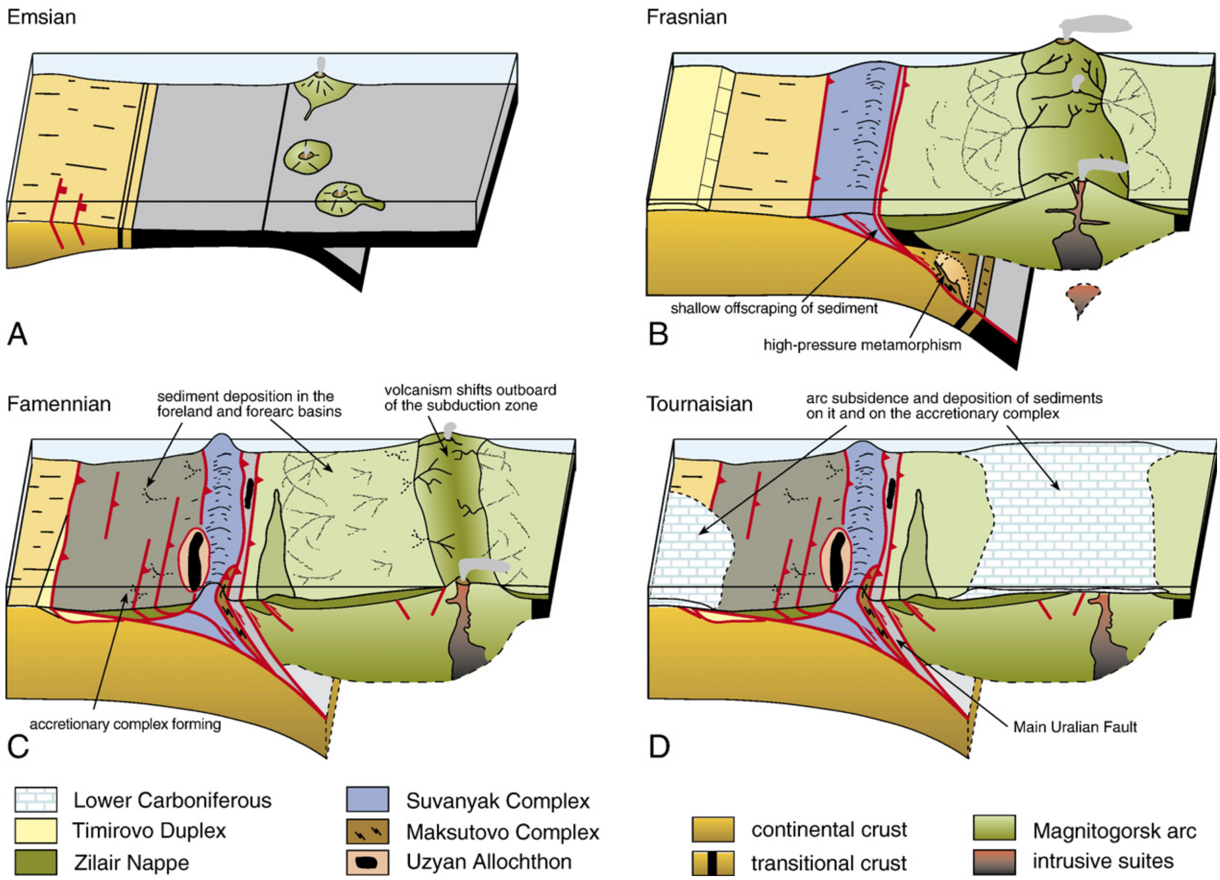

Fig. 12. Model outlining the tectonic processes through time that took place during arc–continent collision in the Southern Urals in the A) Eifelian, B) Frasnian, C) Famennian and D) Tournaisian.

Sedimentary clasts, in particular siliceous sedimentary lithoclasts and minor carbonate clasts, dominate in the Zilair Formation (Brown et al., 1996; Gorozhaninia and Puchkov, 2001; Willner et al., 2002). Metamorphic lithoclasts make up a large percentage of the clasts found in the Zilair Formation (Fig. 11A). Detrital light metamorphic minerals include white mica, chlorite, most quartz clasts, potassic feldspar and a minor part of the albite present. Among the transparent heavy minerals, metamorphic minerals dominate including epidote, garnet, tourmaline, amphibole, titanite and chloritoid, and possibly most or all of the rutile and apatite (Arzhavitina, 1978; Arzhavitina and Arzhavitin, 1991; Willner et al., 2002). Volcanic lithoclasts of basic to acid composition are abundant in the Zilair Formation. Most euhedral albite clasts were derived from a volcanic source, and approximately 10% of chromium spinel comes from a basaltic source or its cumulates (Willner et al., 2002). It is not clear if any lithoclasts were derived from plutonic rocks. Evidence of a magmatic source for the heavy minerals, with the exception of long, euhedral zircon, and possibly some apatite crystals, is weak. The Zilair Formation also contains an important percentage of clasts derived from a mafic to ultramafic source (Willner et al., 2002). Clasts derived from an ultramafic source, with the exception of chromium spinel, have a minor representation among light minerals. Chromium spinel is one of the dominant heavy mineral species found in the Zilair Formation, and its chemistry points to the entire spectrum of ultramafic and mafic rock types of island arc derivation (Willner et al., 2002).

When plotted on the discrimination diagrams of Dickinson (1985) the Zilair Formation falls across the

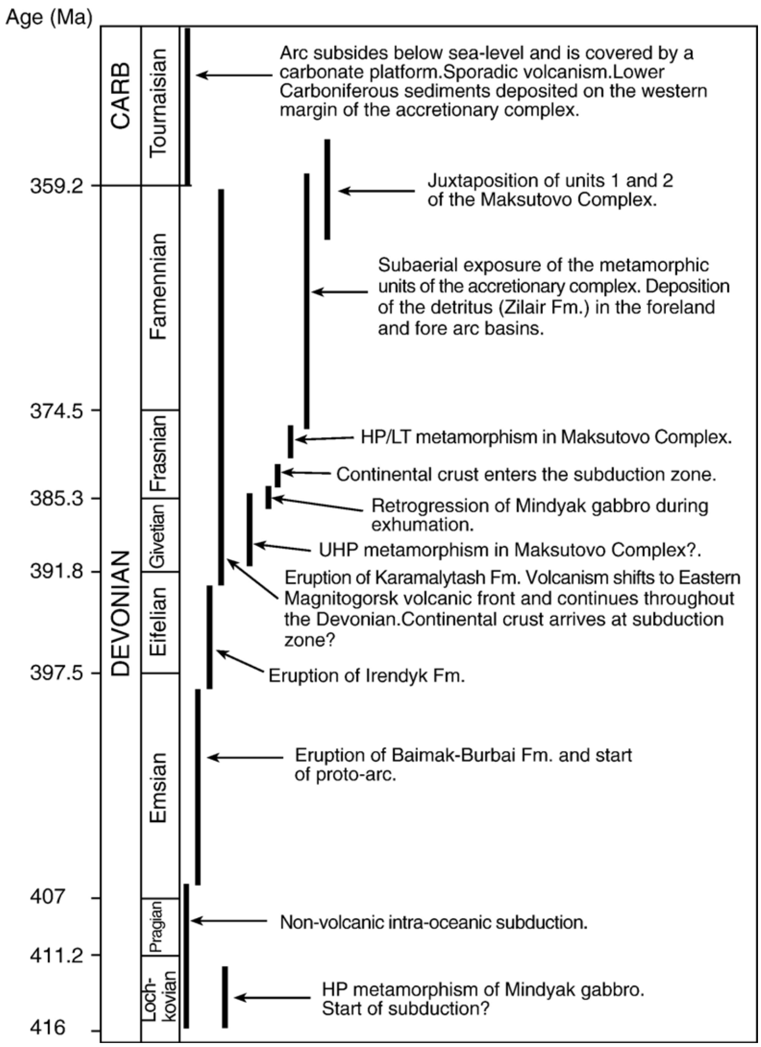

Fig. 13. Time/process evolutionary diagram for intra-oceanic subduction and arc–continent collision in the Southern Urals.

fields assigned to magmatic arcs, in accordance with the abundance of volcanic lithoclasts, and into the mixed orogen source field (Fig. 11B, C, and D). Similarly, a plot of the three major types of lithoclasts (sedimentary, metamorphic and volcanic) shows that the Zilair Formation does not conform with any one of the discrimination fields established by Ingersoll and Suzcek (1979) (Fig. 11E). In all three types of plot there is a broad scatter indicating a mixture of materials from different sources. The same clast signature is found in the Zilair Formation in both the accretionary complex and Magnitogorsk Zone (i.e. in the foreland and forearc basins, respectively). Also, there is no variation of the clast spectrum with the age of deposition. The source of volcanic lithoclasts in the Zilair Formation was the Magnitogorsk Arc and/or volcanic rocks that are found as blocks within the Main Uralian Fault zone. There was no other volcanic source area in the Southern Urals in the Late Devonian. An important source region for the Zilair Formation sediments in both the foreland and the forearc basins appears to have been dominated by metamorphic rocks in the greenschist to amphibolite facies. The presence of phengite with up to 3.4 to 3.45 Si per formula unit (suggesting minimum pressures of 0.5 to 1.2 GPa at 300 to 600 °C), glaucophane, and garnet with composition and zonation pattern matching that of the eclogite and garnet–micaschist of the Maksutovo Complex indicates that the source region also contained medium to high-pressure rocks (Willner et al., 2002). $^{40}Ar/^{39}Ar$ ages obtained from single grains of phengite are in the range 342–421 Ma (Willner et al., 2004). A logical choice for the source region for these clasts would be the metamorphic rocks in the accretionary complex. Greenschist and phyllite clasts were likely derived from the Suvanyak Complex, whereas garnet, chloritoid and rare glaucophane, and the dated phengite would have been sourced from an exhumed part of the Maksutovo Complex or its precursor (the presently exposed rocks of the Maksutovo Complex were exhumed to the surface after the end of the deposition of the Zilair Formation). The mafic to ultramafic source was most likely the ophiolite and peridotite units derived from the Main Uralian Fault. The metamorphic complexes and the ultramafic and volcanic rocks within the accretionary complex evidently formed part of an uplifting ridge that, during the Late Devonian, must have separated the forearc basin from the foreland basin.

9. Discussion

The convergent tectonic evolution of the Southern Urals started as the Magnitogorsk island arc, which had developed in the paleo-Uralian ocean, began to collide with the margin of Baltica (Zonenshain et al., 1984, 1990; Puchkov, 1997; Brown and Spadea, 1999; Puchkov, 2000; Alvarez-Marron, 2002) (Figs. 12 and 13). The geochemistry of the volcanic rocks and the fossil content of the intercalated cherts clearly indicate that the Magnitogorsk Arc formed in an intra-oceanic subduction setting during the Devonian (Maslov et al., 1993; Yazeva and Bochkarev, 1996; Spadea et al., 1998; Artyushkova and Maslov, 1999; Gusev et al., 2000; Herrington et al., 2002; Spadea et al., 2002). Prior to the onset of arc–continent collision in the Devonian, the margin of Baltica had evolved from the rift and drift phase of continental breakup in the Cambrian to Ordovician to a passive continental margin that comprised platform, slope and rise sediments deposited on a Precambrian basement. Below, we develop a detailed model for the arc–continent collision that took place in the Southern Urals (Figs. 12 and 13). Where possible we compare processes interpreted from the Urals data to those determined from currently active arc–continent collision zones and from modeling.

The Baimak–Buribai Formation boninites have been interpreted to have erupted in a forearc position, similar to other boninites worldwide (Crawford et al., 1989; Tatsumi and Maruyama, 1989; Arndt, 2003), and therefore to mark the early stage of subduction-related melting and the initiation of development of the Magnitogorsk Arc (Spadea et al., 1998; Brown and Spadea, 1999) (Fig. 12A). By the end of the Emsian, volcanism had changed from arc-tholeiite to the calc-alkaline suites of the uppermost Baimak–Buribai and Irendyk formations, and continued as a mature intra-oceanic arc well into the Eifelian. On the basis of geochemical and isotopic data, Spadea and D'Antonio (2006) suggest a depleted MORB-type mantle component in the Baimak–Buribai Formation and an enriched, subduction-related component (fluids and melts) with increased pelagic sediment input for the younger Irendyk and Karamalytash formations. While the geochemical and isotopic patterns in the Magnitogosk Arc volcanic rocks show similarities to those of the intra-oceanic Izu-Bonin subduction system (e.g., Pearce et al., 1999), they do not appear to reflect the arc–continent collision stage in the evolution of the Magnitogorsk Arc. For example, the increasing Nd-isotope ratios in the arc–continent collision-related Karamalytash Formation relative to those of the intra-oceanic Baimak–Burbai and Irendyk formations (Fig. 5) is contradictory to the pattern observed in the Australia–Banda Arc collision where a variable or decreasing Nd-isotope ratio in the arc–continent collision-related volcanic rocks relative to the intra-oceanic sequences is thought to be indicative of a continental crustal source in the former (e.g., Elburg et al., 2005).

The first indication of arc–continent collision in the geological record of the Southern Urals is the age of eclogite facies metamorphism in the Maksutovo Complex. If we accept the interpretation that the lower unit of the Maksutovo Complex was derived from the continental margin of Baltica (Hetzel, 1999), then part of the continental crust that had entered the subduction zone had reached at least 50 to 70 km depth (the recent find of microdiamond aggregates suggests that it may have gone deeper) by the middle part of the Frasnian (378 Ma). However, the earlier history is difficult to constrain since there are no time markers in the geological record that can be used. To make an estimate of when the leading edge of the Baltica margin could have entered the subduction zone, and hence arc–continent collision began, we look at published examples of the subduction of continental crust at various velocities and subduction angles (e.g., Ranalli et al., 2000; Li and Liao, 2002). These models show that, at subduction velocities ranging from 2.5 to 10 cm/yr and subduction angles of between 30° to 70°, approximately 0.5 and 3 My would have been needed for the leading edge of the Baltica continental margin to reach the 70 km depth recorded by the Maksutovo high-pressure rocks. Adding this to the age of the high-pressure metamorphism recorded in the Maksutovo Complex (378 Ma) we get a rough estimate of between 378.5 and 381 Ma (early Frasnian) for the entrance of the continental crust into the subduction zone.

Another possibly way to constrain the timing of the entrance of the continental margin into the subduction zone is to look at the shift of volcanism outboard, away from the subduction zone. In areas of active arc–continent collision such as Papua New Guinea, Timor, and Taiwan, volcanism stopped in the accreting volcanic front 1–3 My after the entry of the continental crust into the subduction zone but, in all three cases, volcanism still continues outboard of this (Cullen and Pigott, 1989; Teng, 1990; Snyder et al., 1996; Huang et al., 2006). In the Southern Urals, in the latest Eifelian and Givetian, the eruption of the Karamalytash Formation, in what appears to have been an intra arc rift (Herrington et al., 2002), marked the beginning of a significant change in the location and character of volcanism in the Magnitogorsk Arc. Island arc tholeiitic and basaltic volcanism in the Irendyk volcanic front stopped and volcanic activity shifted to predominantly andesitic to dacitic volcanism in the rift zone and the East Magnitogorsk volcanic front, where it continued until the end of the Devonian (Puchkov, 1997; Herrington et al., 2002). If we assume that this shift in volcanism reflects the entrance of the continental crust into the subduction zone, then arc–continent collision could have begun as early as the Eifelian, or approximately 10 My earlier than suggested by the time determined from the analyses of subduction rates. However, it must be stressed that, as indicated above, the Pb- and Nd-isotope ratios of the Karamalytash Formation do not appear to evidence a continental crustal source.

Following the entrance of the continental crust into the subduction zone, we suggest that the Paleozoic rise and slope sediments of the continental margin were shallowly underthrust beneath the crust and upper mantle of the forearc. The uppermost crust was subsequently detached within the sediments and thrust westward along with the overlying forearc mantle. These rocks are currently represented by the Uzyan and Sakmara allochthons. With continued subduction of the continental crust, parts of the slope and rise sediments were more deeply subducted, deformed and metamorphosed to lower and middle greenschist facies before being completely offscraped and exhumed to form the Suvanyak Complex (Fig. 12B). Similar processes to those suggested above have been interpreted in areas of active arc–continent collision such as Timor, Papua New Guinea, and Taiwan, where sedimentary offscraping and forearc crust and mantle obduction is currently taking place or has taken place in the recent past (Jahn et al., 1986; De Smet et al., 1990; Charlton, 1991; Abbott et al., 1997; Huang et al., 2000, 2006). These processes can also be inferred from analogue and thermo-mechanical modeling results, which suggest that shallow offscraping and exhumation of sediments is an integral part of arc–continent collision (van den Beukel, 1992; Chemenda et al., 2001; Boutelier et al., 2003, 2004). With continued subduction, the continental crust reached depths of at least 50 to 70 km (and perhaps as much as 120 km) by the late Frasnian, where it was metamorphosed to eclogite facies before starting its journey back up the subduction/exhumation channel (Fig. 12B). The deformation and metamorphic history of the Maksutovo Complex records conditions and processes that were active in the subduction channel, so these aspects are discussed in this context below.

The next indicator of arc–continent collision in the Southern Urals is the onset of deposition of the Zilair Formation sediments in the latest Frasnian (c. 376 Ma), which points to the presence of an evolving, and by then partially subaerial accretionary complex, and the exhumation to the surface of the first high-pressure rocks (Brown et al., 1998; Brown and Spadea, 1999; Willner et al., 2002) (Fig. 12C). The variety of clast types in the

Zilair Formation, the most abundant of which were derived from volcanic, ultramafic, low to medium metamorphic grade, and high-pressure rocks, indicate that by the late Frasnian the accretionary complex and the East Magnitogorsk volcanic front formed subaerial mountain ranges that provided sediments to the submarine foreland and suture forearc basins. The fact that both basins shared the same two source areas throughout their history suggests that they were linked and remained so until the end of arc–continent collision (Willner et al., 2002). The uplift and erosion of the Southern Urals accretionary complex is a similar process to that taking place in Papua New Guinea, Timor, and Taiwan where an uplift of the accretionary complex and forearc has been taking place at rates of 0.8–16 mm/yr^{-1} since the arrival of the continental crust at the respective subduction zones (Jahn et al., 1986; De Smet et al., 1990; Abbott et al., 1997; Huang et al., 2000). In Timor and Taiwan the offscraped continental margin sediments within the accretionary complex are the major sediment source for the foreland basin (Audley-Charles, 1986; Huang et al., 1997, 2000), whereas in Papua New Guinea the forearc is the major supplier of sediments. Uplift and subaerial exposure of the forearc in the Papua New Guinea and Timor has been interpreted to mark the arrival of the full thickness of the Australian crust at the respective subduction zones (De Smet et al., 1990; Abbott et al., 1994). Following this example, we interpret the uplift of the Southern Urals accretionary complex, the widespread syn-sedimentary deformation in the Koltubanian Formation, and the initiation of growth of the Karamalytash Anticline to indicate basin instability and deformation that was caused by the arrival of the full thickness of the Baltica continental crust at the subduction zone in the Frasnian (Brown et al., 1998; Brown and Spadea, 1999; Brown et al., 2001).

In the foreland basin, the overall younging of the Zilair Formation from east to west, together with the continuous transition from the deformed Late Devonian to undeformed Early Carboniferous stratigraphic sequence along much of the southwest margin of the basin (Brown et al., 2004), suggests that the Zilair Nappe grew by progressive accretion of younger sediments to its front throughout the Famennian (Alvarez-Marron et al., 2000). The wide spacing of thrusts in the Zilair Nappe (roughly 3 km) may be interpreted to indicate a high sediment input into the basin during convergence (c.f. Moore, 1979; Fuis and Clowes, 1993). In this scenario of frontal accretion, we interpret the Timirovo thrust system to represent the shallow underplating of a carbonate ramp to the frontal part of the accretionary complex late in the deformation history (Brown et al., 1997; Alvarez-Marron et al., 2000). Calcite fabrics developed in shear zones in the Timirovo thrust system indicate that deformation took place under low temperature ductile conditions (Fernandez et al., 2004). Measurements carried out on twinned calcite grains from these shear zones indicate that the differential stress at the base of the accretionary complex reached a peak of about 230± 40 MPa (Fernandez et al., 2004). This value is much higher than that measured for the Taiwan accretionary complex, and is more similar to those reported from the interior parts of orogens (Lacombe, 2001).

Meanwhile, as volcanism continued in the East Magnitogorsk volcanic front, thrusting, folding, and sediment deposition were taking place in the forearc region throughout the Frasnian and Famennian. This is indicated by the onlapping geometry of the forearc basin sediments over the Karamalytash Anticline, which is suggestive of a growth fold geometry and which indicates that deformation and sedimentation were active simultaneously (Brown et al., 2001). The irregular and non-linear juxtaposition of anticlines and synclines, together with the involvement of the volcanic substratum in the folding and thrusting, is suggestive of inversion of a pre-existing fault system (c.f. Cooper et al., 1989), possibly a system of trench-parallel faults such as those that commonly develop in the forearc region of volcanic arcs (Lallemand et al., 1999; Chemenda et al., 2000).

The Mindyak lherzolite and the Maksutovo Complex provide important information about the physical conditions and the processes that were active in the subduction zone during intra-oceanic subduction and arc–continent collision, respectively (Brown et al., 2000). First, the temperature conditions within the subduction zone were quite different during the early intra-oceanic stage, where the temperature reached about 800 °C at 30 km depth (data from the Mindyak lherzolite), compared to the arc–continent collision stage where temperatures of <440 °C were reached at around 15 km, 550 °C at 60 km and, possibly, about 650 °C at 110 km depth (Maksutovo Complex data). Numerical modeling of intra-oceanic subduction zones shows a pronounced cooling with time as geotherms are depressed (e.g., Peacock, 1996). Modeling shows that with the entrance of the continental crust into a subduction zone, relative heating may take place in the upper 50 km or so of the subduction zone (van den Beukel, 1992). Second, the variation of the ages and P–T conditions obtained from the Mindyak and Maksutovo Complex high-pressure rocks indicates that an important flux of material was taking place from deep within the subduction zone from the early stages of intra-oceanic subduction to the final stages of arc–continent collision.

This suggests that there was a mass flow of material in the subduction zone, along a subduction channel, in which concomitant subduction, high-pressure metamorphism, exhumation and subduction–erosion of material was continuously taking place. Third, there is widespread textural and isotopic evidence from both the prograde and retrograde Maksutovo Complex eclogite and blueschist that indicate that there was a free fluid phase in the subduction channel (Glodny et al., 2002). This fluid phase may have been the result of either influx from underplated and metamorphosed material or internal release, and it likely acted to trigger and catalyze reactions and growth of eclogite facies minerals (Glodny et al., 2002). And finally, the now exposed rocks of the Maksutovo Complex arrived at the surface a long time after arc–continent collision stopped. However, high-pressure rocks, similar to those of the Maksutovo Complex, were being eroded throughout the time of deposition of the Zilair Formation.

The Main Uralian Fault in the Southern Urals now forms the suture zone between the Uralide arc terranes and the paleo-continental margin of Baltica. It can be interpreted to reflect the position of the intra-oceanic subduction zone, and indeed many of the ultramafic units and volcanic rocks found within it have a geochemical signature that indicates that they occupied an intra-oceanic suprasubduction-zone position at some time in their history (Gaggero et al., 1997; Scarrow et al., 1999; Spadea et al., 2002, 2003). Currently active arc–continent collision zones, such as those in the southwest Pacific, are taking place along a suture zone in which the rigid volcanic arc forms either an arcward- or trenchward-dipping backstop (e.g., Abbott et al., 1994; Snyder et al., 1996; Huang et al., 1997). In the Southern Urals, reflection seismic profiling shows that the Main Uralian Fault dips arcward (Brown et al., 1998). We suggest that the Magnitogorsk forearc acted as an arcward-dipping, semi-rigid backstop to the accretionary complex. The rigidity of the forearc protected the forearc basin sediments from strong deformation. Nevertheless, as arc–continent collision progressed, at least some mechanical erosion of the forearc took place, and volcanic rocks and their overlying sediments were incorporated into the Main Uralian Fault and deformed.

Arc–continent collision stopped in the Southern Urals at the end of the Devonian, the arc massif subsided, and Early Carboniferous sediments were deposited on top of it and along the western margin of the accretionary complex (Fig. 12D). From beginning to end, arc–continent collision in the Southern Urals lasted about 20 My (or as much as 30 to 35 My if we assume that the continental crust arrived at the subduction zone in the Eifelian). In arc–continent collision zones such as Papua New Guinea, Timor, and Taiwan (Charlton, 1991; Abers and McCaffrey, 1994; Sibuet and Hsu, 2004; Tsai, 1986), active seismicity indicates that continental crust continues to underthrust the arc 3–9 My after its entry into the subduction zone. In somewhat older examples, such as along the northern Caribbean Plate margin, collision of the Greater Caribbean arc with the margin of North America began during the Late Paleocene (Mann et al., 1991) (i.e., in Cuba) and continues to be active today (i.e., Hispaniola), more than 55 to 60 My later. In other Paleozoic orogens, such as the Grampian Orogeny in the Caledonides of Britain and Ireland, or the Taconic Orogeny of the Appalachians, arc–continent collision appears to have lasted about 8 to 10 My (e.g., Cousineau and St-Julien, 1992; van Staal, 1994; Dewey, 2005). These examples, together with that of the Urals, suggest that arc–continent collision appears to be a short-lived event, generally lasting from 10 to 20 My. However, the duration of an arc–continent collision orogeny depends on factors such as subduction velocity and shape and thickness of the continental margin involved (e.g., Ranalli et al., 2000; Li and Liao, 2002). Furthermore, determining when the continental crust entered the subduction zone in a fossil orogen may be difficult. The Uralides case clearly shows that the sediments derived from the growing accretionary complex began to arrive in the foreland and suture forearc basins later than the peak high-pressure metamorphic event that affected the continental margin. The pressure and temperature conditions recorded in the high-pressure rocks, together with the age of the peak metamorphic event, indicates that the continental crust entered the subduction zone much earlier still. Finally, the geochemistry of the volcanic rocks do not record any clear evidence of the continental crust being in the subduction zone and supplying melt to the volcanism.

10. Conclusions

The geochemistry of the Magnitogorsk Arc volcanic rocks, the structure of the arc–continent collision accretionary complex and the forearc, the high-pressure rocks beneath and along the suture zone, the mafic and ultramafic ophiolitic material, and the syn-tectonic sediments show that the Paleozoic tectonic processes recorded in the Southern Urals can be favorably compared with those in currently active settings such as the west Pacific. For example, boninitic lavas found in the oldest arc volcanics provide a geodynamic marker that records the onset of intra-oceanic subduction and the development of a proto-arc. While the geochemical and isotopic signatures of the Magnitogorsk Arc volcanic rocks are indicative of intra-oceanic subduction, they do not record clear evidence of the subduction of the continental crust. High-pressure/low temperature rocks along the backstop of the accretionary complex were in part derived from continental margin material, and the Frasnian age of their high-pressure metamorphism provides a constraint for determining the timing of the entry of the continental crust into the subduction zone. The pressure, temperature, and thermochronology data of the Maksutovo Complex and high-pressure mafic rocks along the arc–continent suture, provide evidence for changing physical conditions over time in the subduction zone. The flux of material along it suggests the presence of a subduction channel along which material moved. The sediments overlying the volcanic arc record near surface processes such as growth folding. The widespread occurrence of debris flows within these sediments is thought to represent seismic events (seismites), and may be related to the arrival of the full thickness of the continental crust at the subduction zone. During the late stage of collision turbidites of the Zilair Formation were deposited in a forearc basin and a foreland basin separated by a subaerial part of the accretionary complex. These sediments represent eroded material from the magmatic arc and the growing accretionary complex. From beginning to end, the arc–continent collision in the Southern Urals lasted about 20 My and possibly as much as 30 My.

Acknowledgements

This paper is dedicated to Piera Spadea who sadly passed away in August of 2006.

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