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The formation of gneisses in the Southern East Uralian Zone—a result of Late Palaeozoic granite ascent and emplacement

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Abstract

Geochemical and radiogenic age data are incompatible with the idea that the gneiss-complexes of the East Uralian Zone are microcontinental blocks that collided with the Magnitogorsk island arc during the Uralian orogeny. Field and microstructural data from the Dzhabyk and Suunduk complexes presented in this paper enable us to suggest an alternative model that interprets the gneisses as deformed margins of the batholiths of the Uralian Granite Axis.

The gneisses have been formed from a granitic protolith by a continuous fabric transformation on a retrograde metamorphic path. All structures in the gneisses can be assigned to the ascent and emplacement of batholiths in an active strike–slip belt.

Keywords: East Uralian Zone; Granite–gneiss complexes; Magma ascent; Metamorphism

1. Introduction

The Uralian orogeny has been interpreted as a continent–microcontinent collision (Peive et al., 1977; Zonenshain et al., 1984). This geotectonic model is in conflict with the Altaid model of Sengör et al. (1993), which interprets that the Mongol–Okhotsk, Altai–Sayan, Kazakhstan, Tien Shan and Ural mountains have been dominantly formed by subduction–accretion processes.

What is the reason for this different interpretation? Comparing the architecture of the Urals with an idealized collision orogen after van der Pluijm and Marshak (1997) we can recognize an important similarity. Several gneiss complexes can be found in the hinterland of the volcanic arc, in a structural position that is usually occupied by the colliding continent.

The Uralian gneiss complexes have been interpreted as microcontinental blocks that broke off the East European Craton in the Ordovician, collided with the Uralian island arc in the Givetian/Visean and have been intruded by granitic melts in the Late Carboniferous and Permian (Peive et al., 1977; Zonenshain et al., 1984).

This interpretation is still used (Kimbell et al., 2002; Yazeva and Bochkarev, 1996), but it is not compatible with the geochemical data: Nd-isotope data show (Gerdes et al., 2002; Popov et al., 2001) that the crustal residence time of the granites is much too short for melts derived from the East European Craton. Isotopic age data yield a Late Carboniferous crystallization age and a Permian deformation age for the orthogneisses of Kartali east of the Dzhabyk granite (Görz et al., 2004).

Field data are also inconsistent with the microcontinental model: typical compressional structures that might be expected between colliding microcontinental blocks are lacking in the East Uralian Zone.

Furthermore, the low percentage of gneisses in the granite–gneiss complexes (approximately 10%) is also inconsistent with the model of Zonenshain et al. (1984), which interpreted the gneisses as the country rock intruded by the granites.

The microcontinental model does not give a satisfactory explanation of the geological and geochemical data from the

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East Uralian Zone. Thus, the question arises how the gneisses have developed?

In order to address this question, we present field and microstructural data from two granite-gneiss complexes from the Southern Urals: the Dzhabyk and Suunduk complexes.

2. Regional geology

The Uralian orogen is the result of a long-lasting tectonic evolution at the eastern margin of the East European Craton (Fig. 1). The Archean to Paleoproterozoic cratonic basement consolidated before 1.6 Ga (Giese et al., 1999; Puchkov, 1997).

The West Uralian Zone consists of sedimentary sequences that have been deposited at the continental margin of the East European Craton from the Riphean to the Carboniferous. These sedimentary sequences have been folded during the Uralian orogeny and form the foreland thrust and fold belt (Peive et al., 1977).

In the Devonian the intraoceanic subduction at an east dipping subduction zone led to the development of the Magnitogorsk island arc (Ivanov et al., 1986). With the arrival of the East European Craton margin at the subduction zone during the middle Devonian an arc-continent collision started (Ivanov, 1998a) and the subduction ended. An accretionary complex developed between the Magnitogorsk island arc and the East European Craton. Sediments were scraped off and thrust westward onto the East European Craton forming the nappes and duplex structures of the Central Uralian Zone (Alvarez-Marron et al., 2000).

The Main Uralian Fault, the suture zone of the orogen, is an eastward dipping tectonic melange zone that was active as a subduction zone, thrust and strike–slip fault from the Devonian to the Triassic (Ivanov, 1998b; Kisters et al., 1999). Plutonic intrusions such as the Syrostan pluton (Montero et al., 2000) and fault related gold deposits (Kisters et al., 1999) can be found in the Main Uralian Fault.

In the Magnitogorsk Zone that lies east of the Main Uralian Fault, spectacular well preserved Middle Devonian island arc complexes are exposed. These units collided with the East European continental margin in the Frasnian/Early Tournasian (Ivanov, 1998a). The orogenic crust was thickened to 57 km and preserved as crustal root

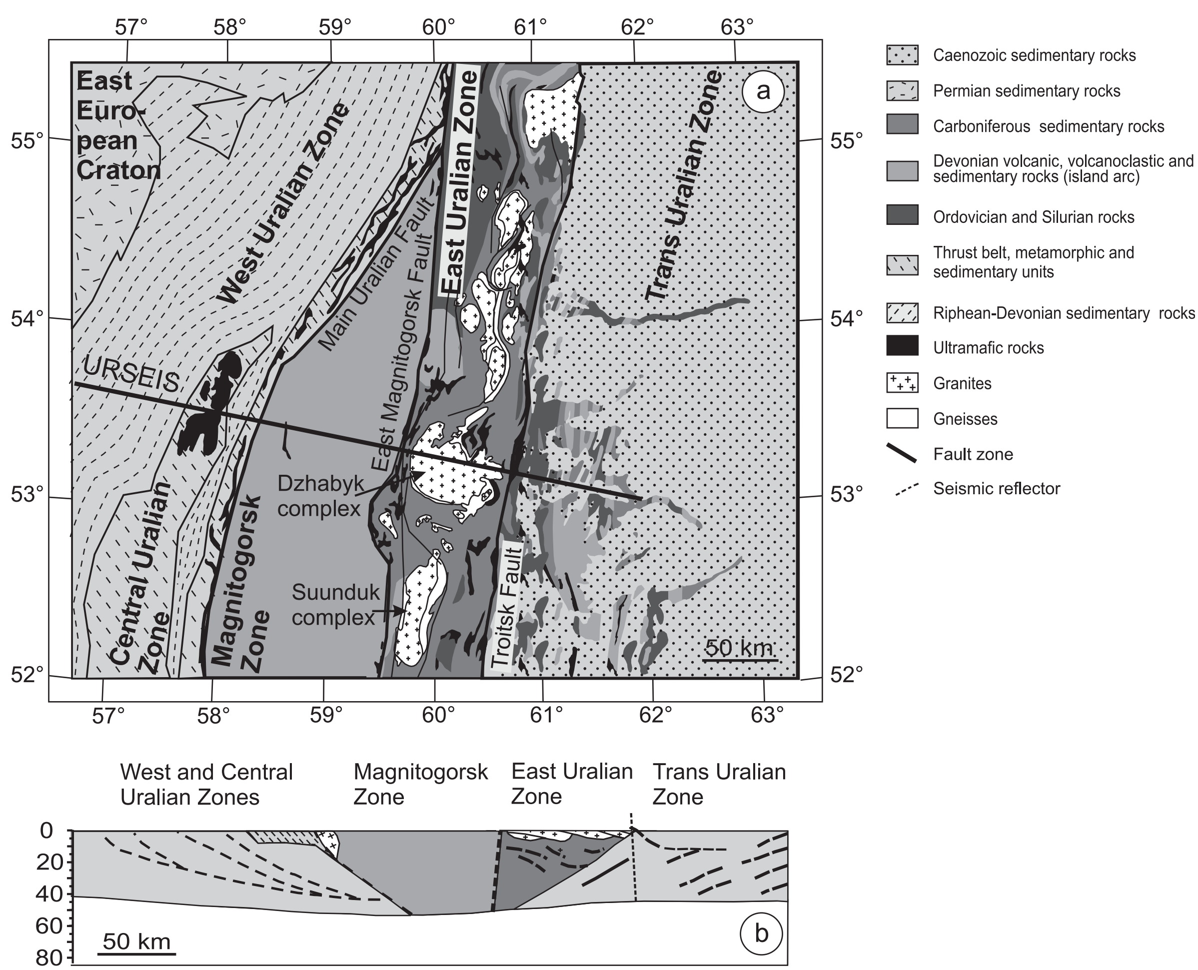

Fig. 1. (a) Sketch map of the southern Urals after Petrov and Shatov (2000), (b) Cross-section along the URSEIS-line after Tryggvason et al. (2001).

(Knapp et al., 1996). The Magnitogorsk Zone is juxtaposed to the East Uralian Zone at the East Magnitogorsk Fault, a steeply westward dipping fault.

The East Uralian Zone is made up of tectonic lenses consisting of rocks of different metamorphic grade from non-metamorphic (Tevelev and Kosheleva, 2002) to greenschist and amphibolite facies (Kisters et al., 1999; Bankwitz et al., 1997). Tectonic lenses of island arc rocks and ophiolite complexes can be found (Seravkin et al., 1992).

The country rocks of the Dzhabyk and Suunduk granite–gneiss complexes consist of associations of serpentinites and amphibolites and of a 5 km thick monotonous sequence of Early Carboniferous sedimentary rocks. These sediments are fine-grained greywackes, slates and fine-grained volcaniclastic rocks. There is no evidence for island arc activity, which is verified by the composition of the siliciclastic succession, as well as in the lack of island arc related mineral deposits (Petrov and Shatov, 2000) (Fig. 2). Therefore, we can conclude that these rocks represent a relic of an oceanic basin.

Several granite–gneiss complexes are embedded in these oceanic units (Fig. 1). The granites form the Uralian Granite Axis, a train of intrusive bodies, which strikes over an N–S distance of 800 km. The principal architecture of all the granite–gneiss complexes is similar. They consist of gneisses that surround granite plutons and are associated with N–S striking shear zones. All gneisses from the East Uralian Zone that have been dated so far, have formed during the Uralian orogeny: the Salda Metamorphic Complex at around 350 Ma (Frigberg et al., 2000), the Sysert complex at 355±5 Ma (Echtler et al., 1997), the gneiss-plate of Kartali at 290±3 Ma (Görz et al., 2004).

Two types of granites can be distinguished: Early small volume granites and granodiorites with an age range of 360–320 Ma and large granite batholiths with an age range of 290–260 Ma (Fershtater et al., 1997).

The granitic melts were generated approximately 90 Ma after the subduction had finished and can be classified as post-collisional (Gerdes et al., 2002). Because the granitoids have a geochemical island arc signature, Ronkin et al. (1988), Ivanov et al. (1995) and Gerdes et al. (2002) suggest melted island arc complexes as the magma source.

Although compressional tectonics lasted until the Triassic (Kisters et al., 1999), the East Uralian Zone behind the Magnitogorsk back arc and in front of the Valerianov fore-arc, did not collide with the Magnitogorsk Zone. No typical compressional structures such as duplexes or thrust sheets and no indications for orogenic thickening of the upper crust can be found.

Instead, the tectonic evolution of the East Uralian Zone was characterized by sinistral N-directed strike–slip movements in N–S trending shear-belts. The strike–slip movements were contemporary with the formation of the granite–gneiss complexes in the Late Carboniferous and Permian (Bankwitz et al., 1997). A Permian temperature emphasized metamorphism has affected the sedimentary and the volcanic rocks and overprints the dykes in the granite–gneiss complexes (Kisters et al., 2000).

The East Uralian Zone has been juxtaposed tectonically with the Magnitogorsk Zone, which is indicated by strongly

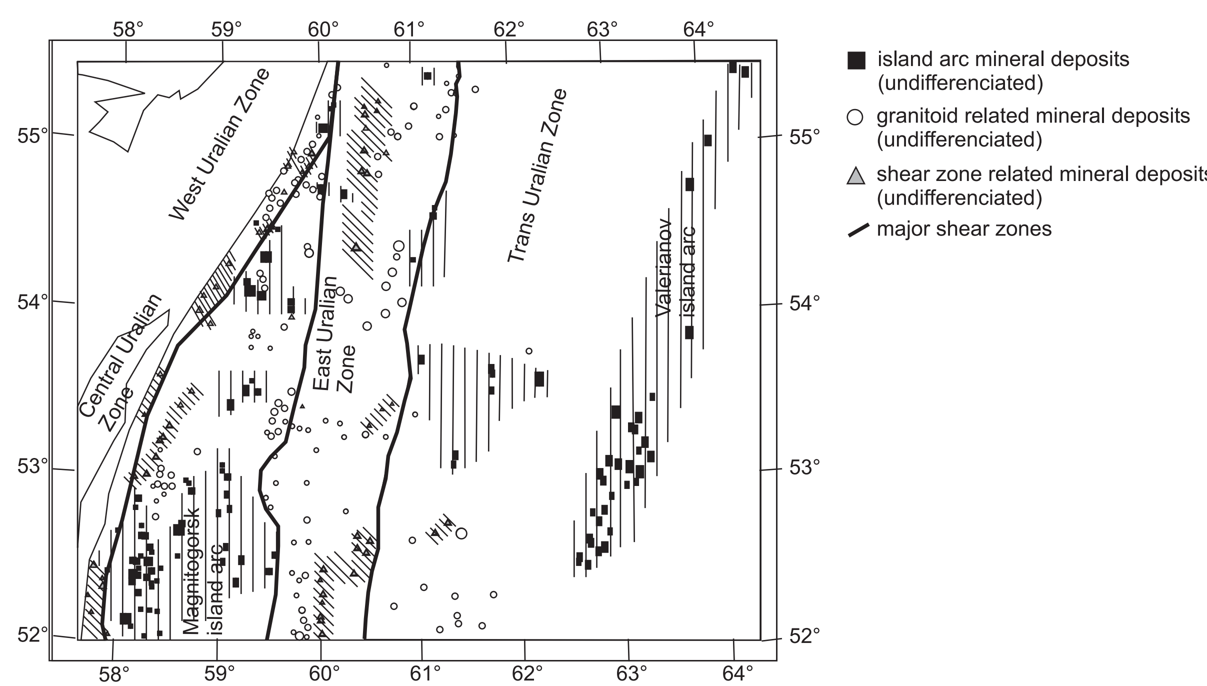

Fig. 2. Distribution of ore-deposits in the southern Urals after Petrov and Shatov (2000). In the Magnitogorsk and Trans Uralian Zone island arc deposits can be found. In the Main Uralian Fault and East Uralian Zone granitoid and shear-zone related mineral deposits occur.

differing sedimentary successions east and west of the East Magnitogorsk Fault (Shalaginow, 1984).

The East Uralian Zone and the Trans Uralian Zone are separated by the steeply dipping Troitsk Fault.

The Trans Uralian Zone consists of westward imbricated rocks such as oceanic crust, Devonian limestones, sandstones and island arc rocks (Kimbell et al., 2002).

These rocks are exposed in narrow N–S striking zones with tectonic contacts (Fig. 1). In the east the so-called, 'Valerianov Island Arc' was active from the Late Visean (Zonenshain et al., 1984). Related island-arc ore deposits have been described (Petrov and Shatov, 2000). Since HP-rocks can be found in the west of the Valerianov arc (Puchkov, 1997), an east-directed subduction has been assumed.

3. Results

3.1. The Dzhabyk complex

The Dzhabyk is the largest granite–gneiss complex of the southern East Uralian Zone. It has an E–W extension of 50 km and an N–S extension of 35 km. It is situated east of the city of Magnitogorsk at 60° longitude and 53° latitude.

The Dzhabyk pluton is situated in the centre of the granite–gneiss complex. Traditionally, the Dzhabyk pluton has been divided into five sub-complexes (Fershtater et al., 1994). These complexes are coeval with an intrusion age of 291 ± 4 Ma (Montero et al., 2000) Gerdes et al.(2002) show that two geochemical trends exist, one peraluminous that can be found in approximately 80% of the pluton and one metaluminous occurring in some marginal parts (Fig. 3). The Nd-isotope ratios are very homogenous in the whole pluton (Gerdes et al., 2002). The initial $^{87}Sr/^{86}Sr$ ratios range from 0.7037 to 0.7062 (Gerdes et al., 2002, Görz et al., 2004).

The granites are continuously transforming into the gneissic margin. Sharp contacts and contact metamorphic mineral associations cannot be found. The strain intensity increases toward the margin. The foliation strikes concordantly with the rim of the Dzhabyk granite–gneiss complex and dips towards its margin at angles of $<30°$ (Fig. 3). The stretching lineations plunge eastward at angles of $<20°$ indicating a tectonic transport with top to the east (Fig. 3). These structures are discordant with the regional framework of the East Uralian Zone, which strikes N–S.

Two major strike–slip zones with greenschist facies mineral associations can be found in the east and west of the Dzhabyk post-dating the gneisses.

In the east the gneiss margin adjoins the ocean floor rocks of the East Uralian Zone. They have been affected by doming and appear in a condensed profile, which exposes the succession from serpentinites, amphibolites and sediments within 2 km. At the contact with the gneisses syntectonic dykes with a granitic filling intrude the country rock serpentinites and amphibolites (Fig. 4(1)). The growth of sillimanite in the amphibolites, andalusite and biotite in the Early Carboniferous sediments near the contact confirms the hot emplacement of the gneisses in the sediments and the following exhumation. Thus, one of the most important results of the field campaign is that the country rocks of the granites are not the gneisses, but the ocean floor rocks of the East Uralian Zone.

If the gneisses have not been the country rock of the granites, which relation do both rocks have to each other?

Microstructures document a continuous fabric transformation of the granites (Fig. 4(2)) to the gneisses on a simple retrograde metamorphic path. The deformation pressure and temperature can be estimated using synkinematic mineral associations. Undeformed samples contain primary crystallized muscovite and quartz with chessboard subgrain boundary patterns indicating deformation under β-quartz conditions (Kruhl, 1996). Minimum crystallization conditions of 650 °C and 4 kbar can be deduced from this mineral assemblage. The absence of chlorite in recrystallized biotite-rich domains, the replacement of syn-kinematic sillimanite by syn-kinematic andalusite and the exclusive occurrence of prism-plane parallel subgrain boundaries in recrystallized quartzes indicate deformation temperatures of 550–600 °C and a deformation pressure of 2–3 kbar (Spear, 1993; Kruhl, 1996). The deformation was restricted to the solid state and ended in the lower amphibolite facies.

The deformation was accompanied by a metamorphic mineral reaction: the complete myrmekitization and albitization of K-feldspar. In the Dzhabyk complex myrmekitization is the dominant mechanism of K-feldspar replacement. Myrmekites grew along the grain boundaries parallel to the S-surfaces (Fig. 4(3)). The myrmekite grains recrystallized to a polygonal mosaic of quartz and plagioclase. Quartz ribbons have formed sub-parallel to the S-faces (Fig. 4(3)). They show evidence for grain boundary area reduction: The grain size of the recrystallized quartz corresponds with the width of the quartz layer.

With progressing recrystallization the primary granitic fabric has completely been destroyed. Ultramylonites with a parallel foliation, coarse quartz ribbons and a fine-grained plagioclase–quartz matrix have formed (Fig. 4(5)). However, even in the ultramylonitic samples relics of microcline can be found (Fig. 4(6)). They prove that the gneisses have a granitic protolith.

3.2. The Suunduk granite–gneiss complex

The Suunduk granite–gneiss complex is situated NE of the city of Uralskoje (59° 50′longitude and 52° 20′latitude). It is an elongated plutonic intrusion with an N–S extension of approximately 70 km and an E–W extension of 15–20 km (Fig. 5). The Suunduk complex consists of highly differentiated peraluminous rocks with initial $^{87}Sr/^{86}Sr$

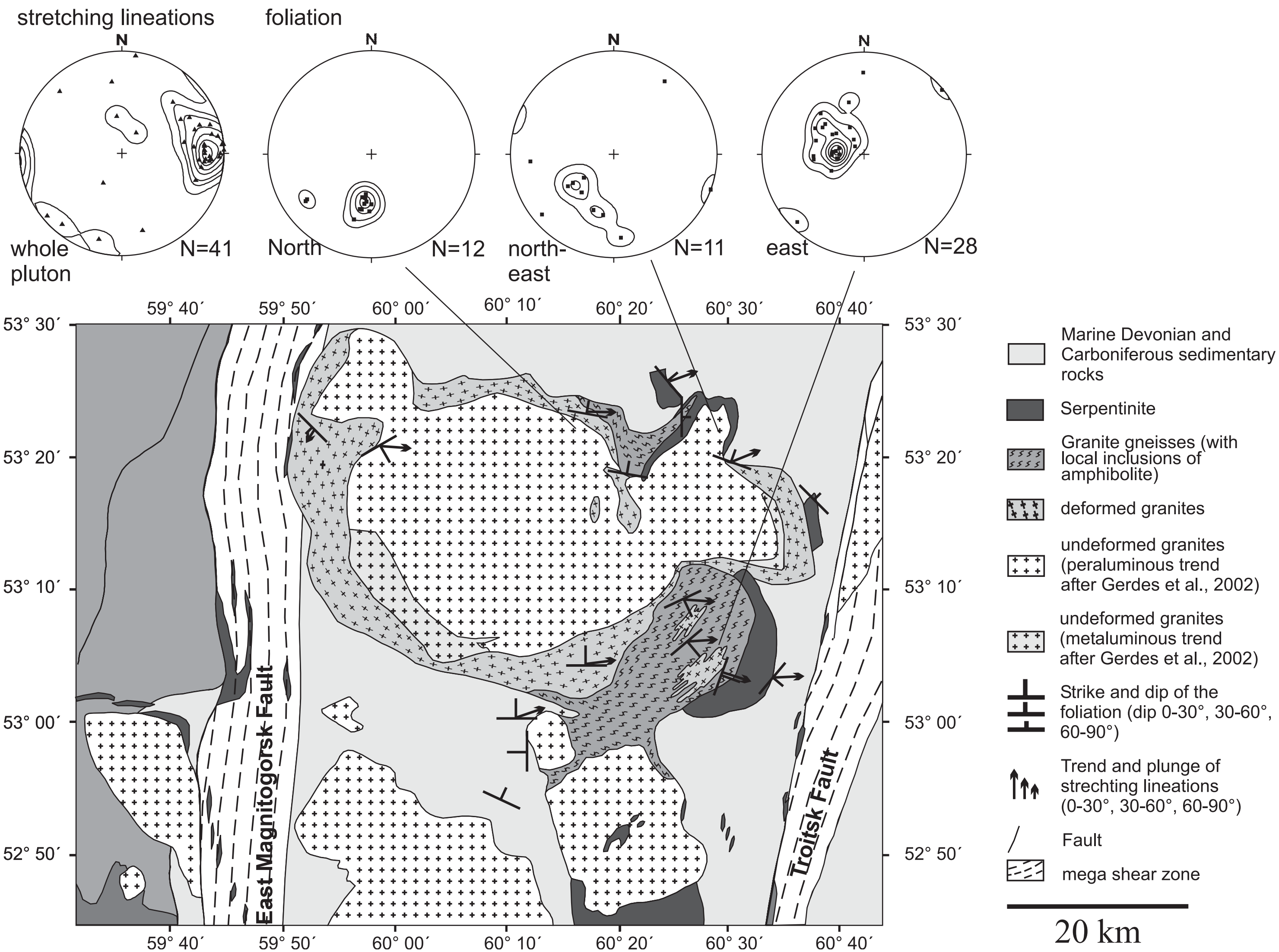

Fig. 3. Geological map of the Dzhabyk granite–gneiss complex and stereographic projections of the foliation and stretching lineations. While the foliation strikes circular, the stretching lineation in the whole pluton plunges eastward.

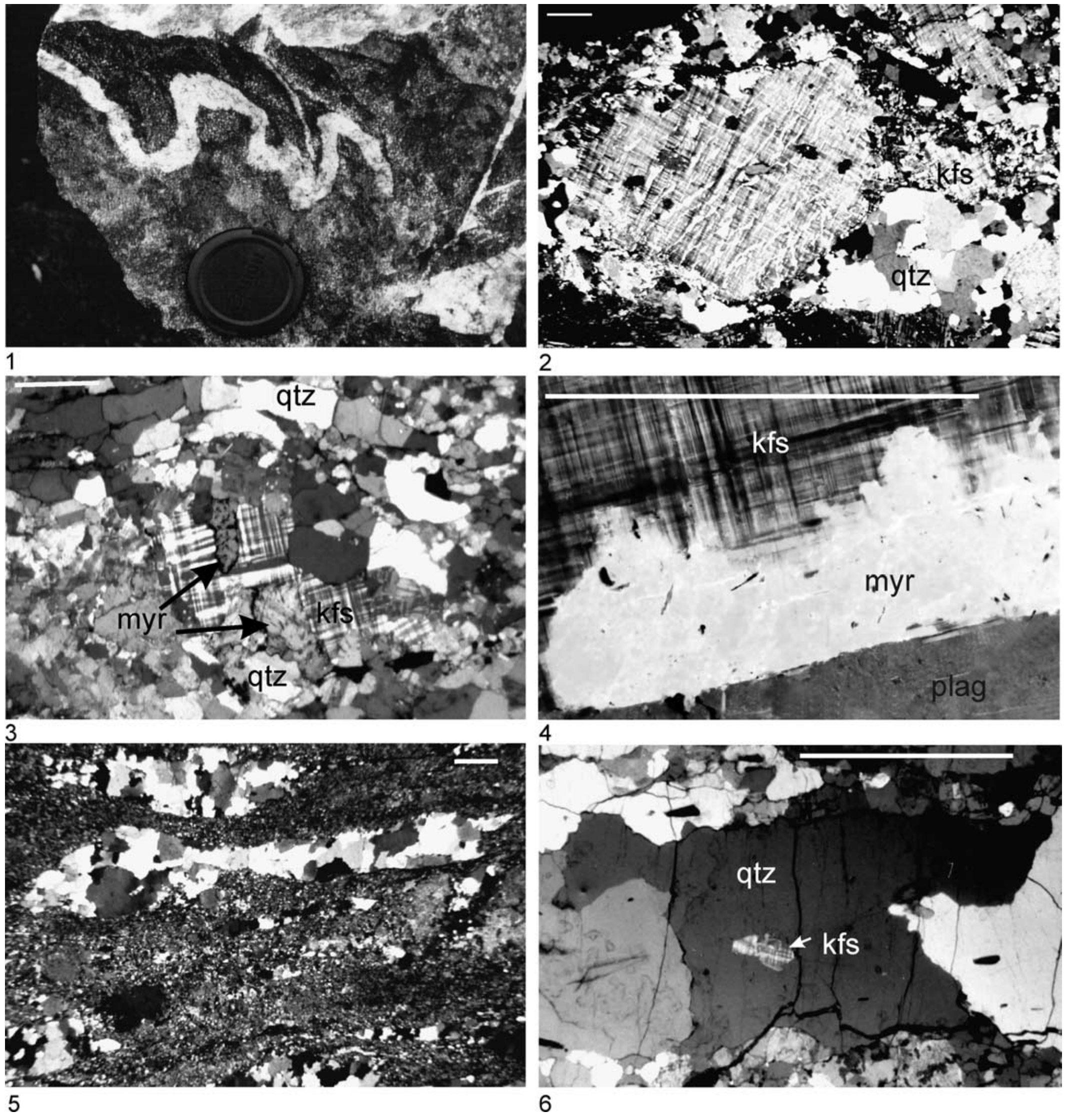

Fig. 4. (1) Syn-kinematic dykes with a granitic filling can be found in the country rock amphibolites of the Dzhabyk gneisses and verify the intrusive relationship of both rocks. (2–6) microphotographs with crossed nicols, white bar: 1 mm. (2) A granite with deformation beginning. The dynamic recrystallization of the microcline by subgrain rotation causes a significant grain size reduction, while the quartzes form ribbon structures without grain-size reduction that give evidence for a dynamic recrystallization by grain boundary migration. (3) Syn-kinematic myrmekitization of a microcline porphyroclast and formation of quartz ribbons around it. (4) Detail of a myrmekitization front between microcline and plagioclase, (5) an ultramylonitic gneiss with coarse quartz ribbons and a fine grained quartz–plagioclase matrix. (6) Relics of microcline can be found in the ultramylonitic samples and testify that the protolith of the gneisses was a granite.

isotope ratios ranging from 0.702 to 0.711 and initial εNd values between 0 and 4 (Görz et al., 2004).

The gneissic show margin shows different deformation fabrics at the eastern and the western margin of the Suunduk pluton.

The eastern gneisses are dominated by lit-par-lit intrusions, which are characterized by a close interaction of brittle and plastic deformation. Brittle fractures are common in magmatic rocks that crystallize in an active shear zone, since pore-fluids reduce the brittle strength of rocks dramatically (Davidson et al., 1992). In the field, fractures can usually be seen as contact faces between magmatic layers (Fig. 6(1)). The foliation of the gneisses is parallel to the magmatic veins and follows two main directions. A NNE striking and perpendicular dipping swarm has stretching lineations that trend sub-horizontally to SSW, an N striking swarm with a dip of about 20° shows stretching lineations that trend sub-horizontally to the S (Fig. 5). These structures indicate that the pluton intruded an active strike–slip system. Sinistral kinematic indicators such as boudins, shear lenses and porphyroclasts dominate in the whole pluton. The tectonic transport occurred top to the south. Brittle fractures dip at angles $>40°$ and have striations that plunge with $>40°$.

No evidence of doming, such as radial structures and a condensed metamorphic profile are found in the country

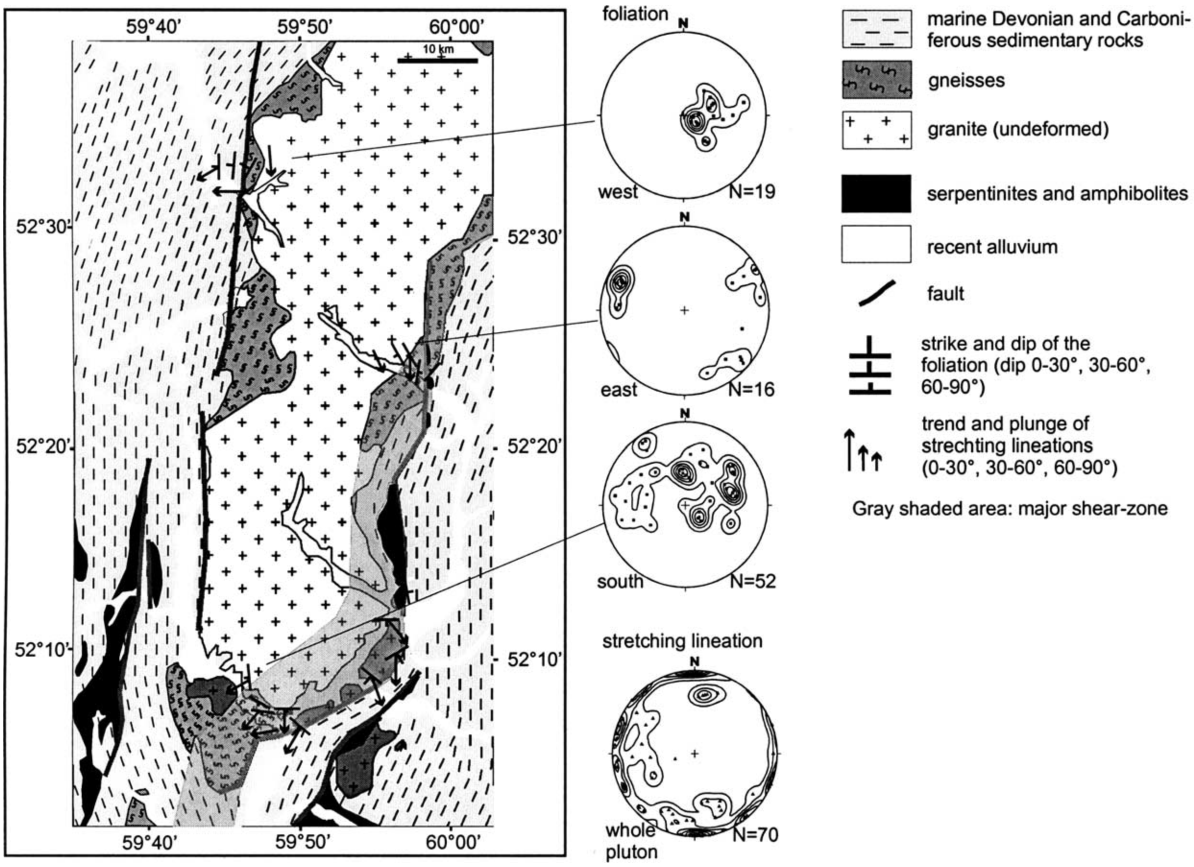

Fig. 5. Geological map of the Suunduk granite–gneiss complex and stereographic projections of the foliation and stretching lineations.

rocks of the Suunduk complex. Serpentinites are absent and amphibolites are exposed only along faults. The country rock sediments contain contact metamorphic minerals, especially cordierite (Fig. 6(2)).

In thin-sections the continuous transition of granites to gneisses is well documented. Muscovites show resorbed crystal boundaries and overgrowth by K-feldspar, suggesting that the granitic melt ascended during the crystallization and passed the muscovite stability boundary (Fig. 6(3)).

In the incompletely crystallized rock melt reduces the brittle fracture strength (Davidson et al., 1992) and the syn-magmatic fabrics are dominated by brittle material behaviour. Heterogeneous plagioclase glomerocrysts with fractures and overgrowth (Fig. 6(4)), equally grained cataclasites of K-feldspar (Fig. 6(5)), asymmetric plagioclase crystals with wings in the pressure shadows, as well as plagioclase and quartz that crystallized in fractures (Fig. 6(5)) can be found. These structures provide evidence that the eastern Suunduk granite crystallized in an active shear zone.

After the complete crystallization ductile intracrystalline deformation structures have been formed. Lobate grain boundaries between quartz and feldspar (Fig. 6(6)) that develop by diffusion creep at deformation temperatures of more than 630 °C (Rutter, 1976) are typical. These structures provide evidence that the deformation continued in the solid-state. At temperatures lower than 600 °C perthites, microcline lamellae and recrystallizates have been formed. The metamorphic reaction of K-feldspar to plagioclase occurred predominantly by albitization along the microcline twinning lamellae. Myrmekitization played only a subordinate role in the eastern Suunduk complex. Chlorite grew in syn-kinematic shear bands, documenting that the deformation lasted to the greenschist facies.

The gneissification was associated with the destruction of K-feldspar, the formation of quartz ribbon structures and the complete recrystallization of the granitic fabric. The ultramylonitic gneiss is characterized by a fine polygonal quartz–plagioclase–chlorite matrix and coarse quartz ribbons. Relics of microcline with uneven grain boundaries can be found (Fig. 6(7) and (8)).

Pressure, temperature sensitive mineral associations and intracrystalline structures show that the plastic deformation occurred in a wide temperature range, but under almost constant pressure conditions: Syn-kinematic sillimanite is absent and has not been described within any Russian literature. The coupled appearance of chessboard quartz subgrains and syn-kinematic andalusite gives a pressure estimate of <2 kbar for deformation temperatures of >600 °C (Kruhl, 1996; Spear, 1993). The combined presence of recrystallized plagioclase with syn-kinematic chlorite allows the estimation of the deformation pressure at

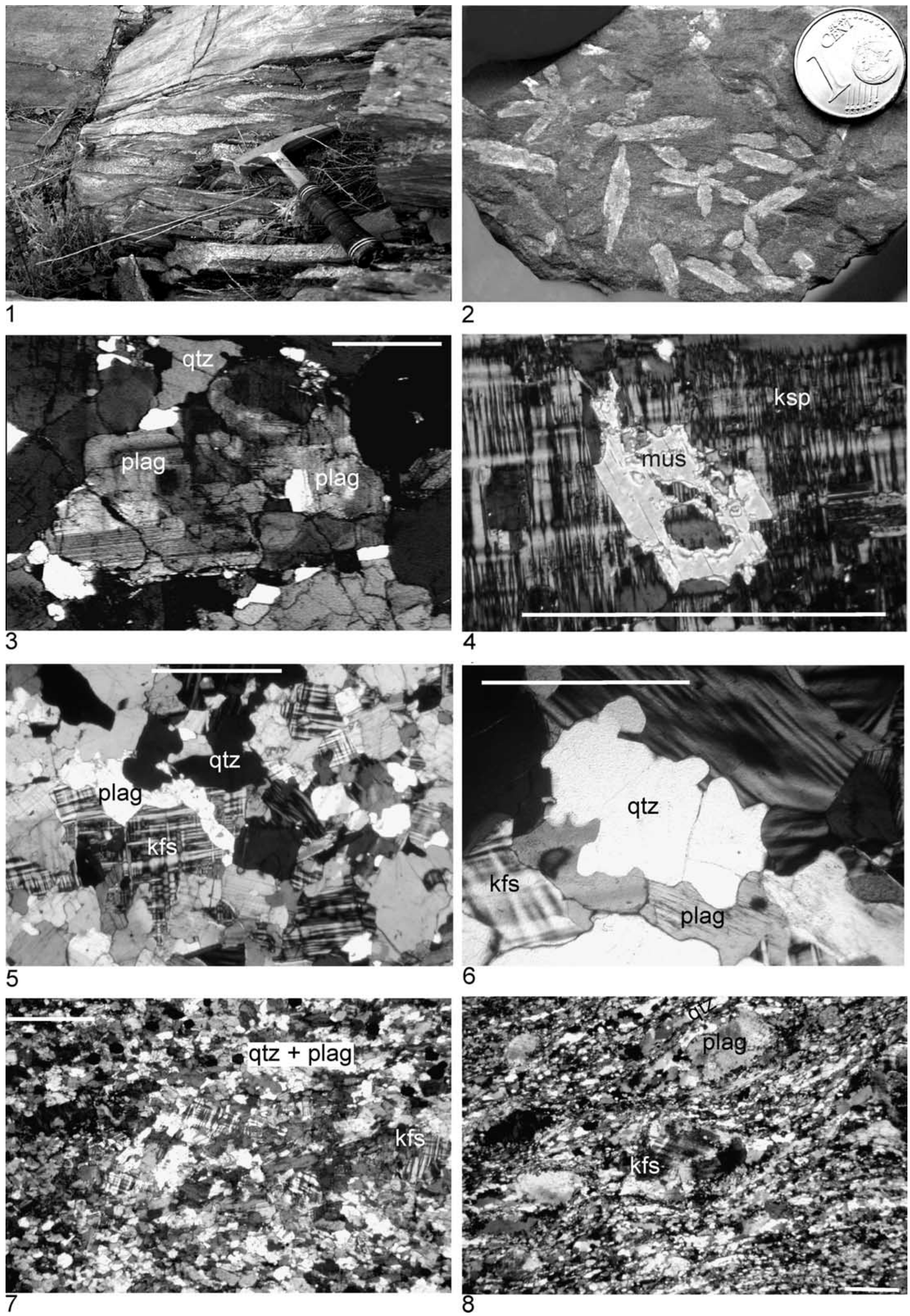

Fig. 6. (1) Lit-par-lit intrusions in the eastern gneiss-mantle of the Suunduk complex. Granitic veins are surrounded by gneisses and have been folded. (2) Cordierite-bearing shale of the country rock, (3–8) microphotographs with crossed nicols, white bar: 1 mm, (3) plagioclase glomerocryst as a result of syn-magmatic deformation and overgrowth, (4) resorbed muscovite with an atoll-like shape providing evidence for the ascent of the granite during crystallization, (5) syn-magmatic deformation producing equal-sized cataclasites of feldspar. In the fractured microcline in the middle of the picture plagioclase has crystallized, (6) Lobate grain boundaries between quartz and feldspars show evidence that diffusion creep acted as crystal deformation process at high temperatures, (7 and 8) the solid state deformation during the cooling of the granites produces mylonitic gneisses with relics of microcline.

1–2 kbar at temperatures of ~500 °C (Spear, 1993). These pressure estimates indicate that the deformation of the eastern Suunduk gneisses was not associated with the exhumation of the complex, but with the strike–slip tectonics in the shear zone.

The structures in the western gneissic margin are similar to the Dzhabyk gneisses. Lit-par-lit intrusions are lacking. The transition from granites to gneisses is continuous and took place under solid-state conditions. K-feldspar has been replaced predominantly by myrmekitization. The foliation of the gneisses has the same orientation as in the eastern gneiss-mantle, the tectonic transport occurred top to the west.

4. Discussion

4.1. The gneisses: the deformed margins of the plutons

The gneiss-mantles of the Dzhabyk and Suunduk complexes show no indication of the existence of a microcontinent. Age data (Görz et al., 2004) suggest that the gneisses have formed during the granite intrusion.

The gneisses are not the host-rock of the granites, but ortho-rocks that intruded the ocean floor rocks of the East Uralian Zone. They have formed from granites of the Uralian Granite Axis in the Late Palaeozoic by a continuous fabric transformation on a retrograde metamorphic path.

For these reasons, the gneisses should be regarded as deformed marginal parts of the granite bodies. All structures in the gneiss-mantles should be related to the ascent and emplacement of the plutons.

4.2. Magma ascent models

Two scenarios describe the ascent of granites. Either the space for the intrusion can be created by the pluton itself removing its host rock laterally or downward (Clarke, 1992). In this case, the pluton ascends by buoyancy as a diapir (Weinberg, 1995). Diapir models predict intense roof-parallel strains, a concentric granitic foliation and parallelism of host rock structures in the aureole with respect to the pluton (Clarke, 1992). Because diapirs expand during their emplacement, a solid-state deformation of the frozen margin is typical (Ramsay, 1981).

If the space for an intrusion is created by regional tectonic movements displacing the host rock along faults, the pluton can ascend through dykes (Yoshinobu et al., 1998). Then fault-related plutons are formed. They can be found in extensional zones (Hutton et al., 1990) and in strike–slip belts (Hutton and Reavy, 1992). In active shear zones the magma ascent occurs in a network of structurally controlled channels (Creaser, 1995). Lit-par-lit intrusions and sheeted margins are typical (Yoshinobu et al., 1998). Fault-related plutons are mostly elongated and parallel to the main fault direction (Schmidt et al., 1995).

The magma ascent in dykes is much faster than in diapirs (mm/s and m/y according to Petford, 1995). Therefore, diapirs ascend only to middle crustal levels, while fault related plutons can rise to upper crustal levels as Edwards and Harrison (1997) describe for the example of the Kuhla Kangri and Laghoi Kangri granites in the Himalayas.

The East Uralian Zone offered good conditions for both: diapirism because it is composed of mafic rocks with a high density and dyking, because it was an active strike–slip belt with a high amount of lateral displacement.

4.3. Reconstruction of the kinematic history of the Dzhabyk and Suunduk complexes

The Dzhabyk pluton shows typical characteristics of a diapir. A large magma chamber had to develop until the buoyancy of the melt was high enough for ascent. Different magma types had the possibility to mix and homogenize in the magma chamber, which is consistent with the uniform isotopic composition of the Dzhabyk magmatic rocks. The country rock resisted the ascending magma. Thus, the magma chamber ascended only to a middle crustal level of 2–3 kbar pressure and a pluton concordant gneiss-mantle with a circular striking foliation has formed. The strain intensity increased continuously towards the margin. The deformation is exclusively connected with the diapiric ascent and is restricted to a limited temperature range. Mineral associations of the amphibolite facies are typical.

K-feldspar was replaced predominantly by myrmekitization suggesting subordinate fluid mobility in the Dzhabyk pluton. The country rocks were affected by doming and ring folding resulting in an anticlinal structure, a continuation of pluton concordant structures and a condensed crustal profile in the host rocks. All these structures developed independently on the stress field caused by the plate motions and are discordant with the regional framework. Two major shear zones that were localized at the margins of the Dzhabyk pluton (Fig. 3) overprinted the radial structures and produced a local greenschist metamorphism.

The Suunduk pluton intruded into an active strike–slip fault. The outline of the pluton is concordant with the regional structures. It is strongly elongated in an N–S direction and tapers off towards its northern and southern ends. The Suunduk pluton ascended in numerous dykes that strike in an N and NNE direction and can be identified in the field as lit-par-lit intrusions. Using pre-existing fractures, the Suunduk body could have risen to a shallow crustal level with 1–2 kbar pressure. Numerous pulses of magma intruded the shear system resulting in highly variable Sr isotope ratios.

During the magma ascent various syn-magmatic deformation structures formed. The deformation of the rocks continued after their emplacement. Ductile mineral associations formed under amphibolite and greenschist facies conditions. A foliation developed parallel to the strike–slip faults striking NNE and N. K-feldspar was replaced by albitization indicating high fluid mobility in the Suunduk complex.

Metamorphic layers that are characterized by a constant strain intensity and separated by intrusive contacts are typical for the lit-par-lit intrusions of the Suunduk complex.

Magmatic body forces acted only secondarily in the Suunduk complex. Evidence for doming such as radial structures and a condensed crustal profile in the country rock have not been found. The shear system parallel to the respective pluton margin was preferentially used for the intrusion indicating that the body forces were not totally negligible.

The western gneissic margin is similar to the Dzhabyk complex margins and shows only evidence of solid-state deformation, an absence of lit-par-lit intrusions and the destruction of K-feldspar by myrmekitization. This data provides evidence for a diapiric ascent of shear-zone parallel magma sheets. It indicates that the magmas of the Suunduk complex migrated from the west to the shear zone and filled it.

In summary: each pluton has formed by an interaction of two ascent mechanisms. The Dzhabyk granite was a diapir. At its margins, shear zones were localized. The Suunduk was mainly a fault-related pluton that intruded an active strike–slip zone, but in the west shows characteristics for diapiric ascent.

All structures in the gneissic margins can be explained as a result of these magma ascent and emplacement processes, except the east-directed stretching lineations in the Dzhabyk complex. These we will allude to in the following section.

4.4. Feedback and iteration processes

Since the gneisses are not the magma source of the granites, another question has to be addressed: Where have the melts been generated?

The southern East Uralian Zone at the end of the Early Carboniferous was a marine sedimentary basin that had accumulated 5 km thick sedimentary rocks. We assume that this crust had been in isostatic equilibrium at the end of the Early Carboniferous before the strike–slip tectonics and the magmatism started. The water depth of the East Uralian Zone can be deduced from ammonite shells found in the sediments (Shalaginow, 1984) because the Calcite Compensation Depth in the Early Carboniferous was 1–2 km (Boss and Wilkinson, 1991). In order to estimate the thickness of the East Uralian Zone we assume a maximum water depth of 1 km and a minimum water depth of 0 km.

The crustal thickness of the isostatically equilibrated East Uralian Zone was calculated after Stuewe (2002) comparing the East Uralian Zone with a ‘mantle sea’ (Fig. 7). The maximum possible crustal thickness was 13 km. This crust is too thin for the generation of melts.

Furthermore, the granites have an island arc signature (Gerdes et al., 2002); although in the southern East Uralian Zone no indications of island arc activity can be found. The magma source had to be situated outside of the East Uralian Zone in a thickened island arc.

Field data indicate that the Suunduk magmas migrated from the west to the strike–slip zone used for ascent. In the Dzhabyk pluton stretching lineations show a tectonic transport from the west.

In the west, the East Uralian Zone is juxtaposed with the Magnitogorsk island arc. The density model of Döring and Götze (1999) shows that the Magnitogorsk Zone consists of mafic and intermediate rocks at all crustal levels. Its present thickness is 57 km, but was probably higher in the Carboniferous, since the isostatic equilibration of the Urals started in the Permian (Seward et al., 1997). On both flanks of the Magnitogorsk Zone, the Main Uralian Fault and East Uralian Zone, post-collisional granites and

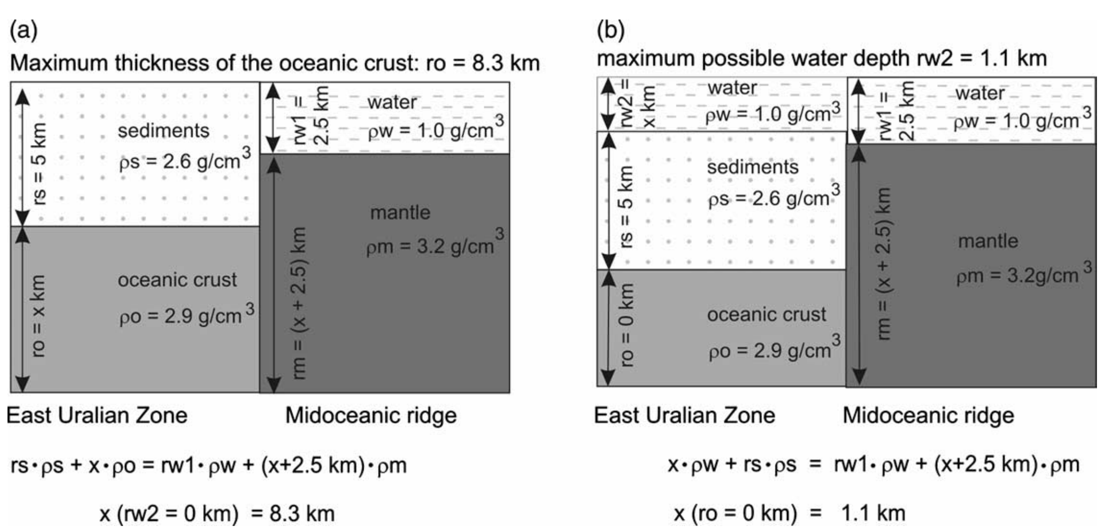

Fig. 7. Isostatic equilibrium of the oceanic units in the East Uralian Zone at the end of the early Carboniferous calculated after Stuewe (2002), the thickness of the sedimentary succession after Shalaginow (1984), rock densities after Lillie (1999). (a) The maximum possible crustal thickness can be reached at 0 km water depth, (b) At 1 Km water depth the oceanic crust reaches a thickness of 0 km and the sediments + water are in isostatic equilibrium with the “mantle sea”.

granitoid-related gold mineral deposits can be found (Kisters et al., 1999). We interpret these observations as evidence that melts were generated in the Magnitogorsk Zone after the Uralian collision. Field data, as well as the regional geology indicate that the Magnitogorsk Zone has been the magma source of the East Uralian Zone batholiths.

Why did these melts migrate into the East Uralian Zone?

In the Permian strike–slip tectonics dominated the deformation style in the eastern Urals. The lens like shape of the Magnitogorsk Zone suggest that this zone was affected by northward shearing of its eastern part. As a consequence, the thin unconsolidated lithosphere of the East Uralian Zone is suggested to have been juxtaposed against the extremely thick crust of the Magnitogorsk Zone creating a steep eastward directed gravitational and thermal gradient that triggered the migration of the melts.

The East Uralian Zone was an active strike–slip belt and offered good conditions for the magma ascent. The architecture of the shear belt essentially influenced the ascent mechanism of each pluton. Diapirs were associated with the low strain domains, fault-related plutons have formed in active deformation zones.

Since fluids cannot support shear stress and reduced the shear stiffness of the crust dramatically, new shear zones formed in the domains weakened by magmas. These shear-zones offered good pathways for the ascending magmas. Thus, the formation of the Uralian Granite Axis is a good example of the positive feedback between magmatism and tectonics.

4.5. The final exhumation

The East Uralian batholiths were emplaced in a depth of more than 6 km and cooled there, as documented by the ductile deformation fabrics. Today, they are exposed at the surface and surrounded by unmetamorphosed rocks of the upper crust. In the Suunduk granite the ductile stretching lineations plunge sub-horizontally, but the striations on brittle faults plunge steeply. This is interpreted as indication that the exhumation of the batholiths was a late process that took place after the plutons had cooled. What mechanism could have caused the exhumation?

A question also remains about how the East Uralian Zone reached its present crustal thickness of 53 km (Tryggvason et al., 2001), although at outcrops it shows features of thin crust.

Since the density of the lower levels of the East Uralian Zone is high (Döring and Götze, 1999), the crustal thickness cannot be explained by the intrusion of the granites.

Compressional structures are lacking in the upper plate of the East Uralian Zone. An eastern foreland basin that would have been depressed by the load of thickened upper crust is absent. This data indicates that the thickening of the East Uralian Zone was decoupled from the upper plate. This leads us to suggest that lower crustal flow may have been the mechanism that caused the crustal thickening in the East Uralian Zone. Lower crustal flow has been simulated by Roy and Royden (2000), Hopper and Buck (1996) and Henk (2000). These models give indications concerning pre-requisites for lower crustal flow. Lower crust flows from regions with a thick crust to regions with a thin crust (Hopper and Buck, 1996). It decouples completely from the upper crust and the mantle. Furthermore, it can only flow, if the lower crust and the foreland are weak. The result is a gentle gradient in the topography (Clark and Royden, 2000).

In the East Uralian Zone and the Magnitogorsk Zone suitable conditions for lower crustal flow existed during the Permian. The Magnitogorsk Zone was thickened and placed next to the thin East Uralian Zone, so that a lateral gravitational gradient existed. The melts generated in the Magnitogorsk Zone migrated into the East Uralian Zone, so both zones were weakened.

Several geophysical models support the assumption that lower crustal flow has occurred from the Magnitogorsk Zone to the East Uralian Zone. The model of Knapp et al. (1996) shows a flat MOHO topography at the eastern flank of the Magnitogorsk Zone and below the East Uralian Zone. This contrasts with a steep MOHO topography at the western flank of the Magnitogorsk Zone. In the gravity model of Döring and Götze (1999) the lower crust of the East Uralian Zone has the equal density as rocks of the same crustal level in the Magnitogorsk Zone (2.83 kg/m^3). Furthermore, patches of high reflectivity are visible in the seismic model of Tryggvason et al. (2001) below the Dzhabyk complex. These patches have not been disturbed by the emplacement of the granites (Tryggvason et al., 2001) suggesting that these structures formed after the intrusion of the granites.

Lower crustal flow, like magma migration, is a mechanism for the post-collisional equilibration in an orogen. It seems to be the only mechanism that can have produced the crustal structures of the East Uralian Zone. It can also have caused the final exhumation of the granite–gneiss complexes, since it produced an upward directed force. The advective heat transfer connected with this process provides a good explanation for the Permian temperature emphasized metamorphism that is only evident in the East Uralian Zone.

5. Conclusions

The field and microstructural data presented in this paper show that there is no evidence for a microcontinent in the gneiss-complexes of the East Uralian Zone of southern Russia. This data is consistent with geochemical data from Popov et al. (2001) that indicates that the existence of a microcontinent can be excluded at least for the area between Cheljabinsk and the Kazakh border.

We find that the gneiss complexes represent the deformed margin of the plutons of the Uralian granite axis. The emplacement of the granite–gneiss complexes in the East Uralian Zone is the result of processes acting during the post-collisional stage of the Uralian orogeny.

The Magnitogorsk Zone was affected by folding and thrusting during the Uralian collision. The crustal thickness of this island arc complex increased to at least 57 km. Because of the development of a new subduction zone in the Trans Uralian Zone, the East Uralian Zone was not affected by collisional tectonics. Its crustal thickness was a maximum of 13 km.

During the post-collisional stage of the orogeny strike–slip movements occurred. The shearing of the back-arc regions of the Magnitogorsk island arc juxtaposed the thin crust of the East Uralian Zone against the thick crust of the Magnitogorsk Zone. Magmas that have formed in these Magnitogorsk Zone migrated into the East Uralian Zone. Since melts and fluids cannot support shear stress, they facilitated the strike–slip tectonics in the East Uralian Zone. The strike–slip zones offered pathways for magma migration. Thus, the Uralian Granite Axis provides an example of an association of strike–slip belts and granites caused by links between magma accumulation and tectonics.

Two mechanisms of magma ascent were most favourable in the East Uralian Zone: diapirism, because the host rock was composed of basic rocks and dyking because deep-reaching, steep-dipping shear zones were available.

It can be assumed that the strain distribution in the Permian East Uralian Zone was very heterogeneous. The local strain rate dictated which ascent mechanism was used by each single magma body. Diapirs were restricted to low strain lenses. Since they require large magma chambers,

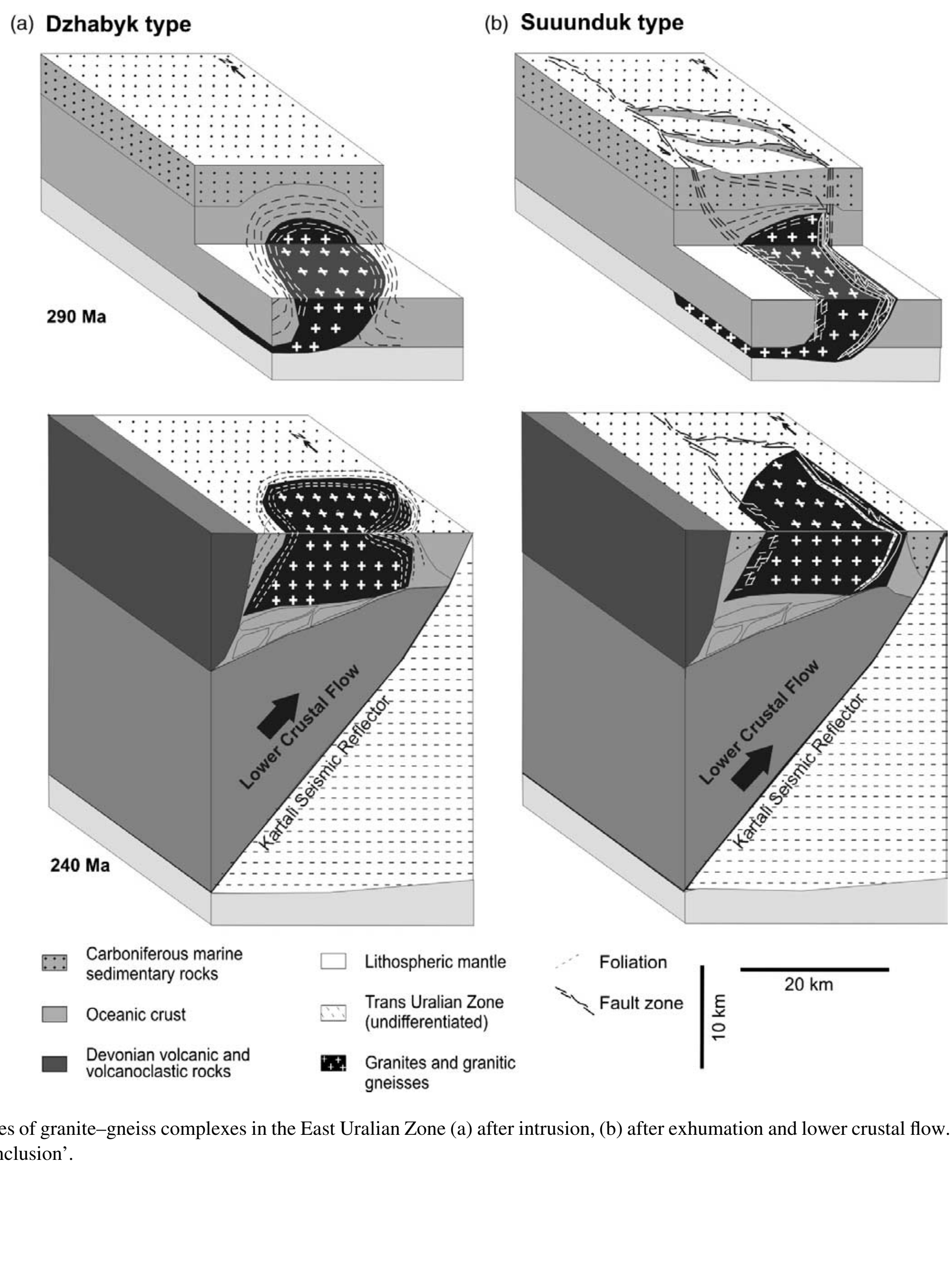

Fig. 8. The two types of granite–gneiss complexes in the East Uralian Zone (a) after intrusion, (b) after exhumation and lower crustal flow. Further explanation in the chapter 'Conclusion'.

they should be younger than fault-related plutons. In active shear belts magma can ascend in pre-existing fault zones by dyking. Fault-related plutons have been formed these environments.

Both ascent mechanisms were related with deformational processes in the batholith and the country rock and have produced characteristic structures that can be identified in the field or in microstructures (Fig. 8a).

Two types of plutons can be distinguished. The Dzhabyk type shows typical properties of diapirism: a concentric pluton geometry, a homogeneous chemistry, the development of pluton concordant and radial structures, the restriction of metamorphic mineral assemblages to a narrow temperature range and the doming, detachment and thinning-out of the roof and the country rock.

The Suunduk type shows typical characteristics of a fault-related pluton: an elongated pluton geometry, a heterogeneous chemistry, evidence for high fluid mobility, lit-par-lit intrusions, the formation of metamorphic mineral associations in a wide temperature range, the development of structures concordantly with the shear zone.

East-directed lower crustal flow from the Magnitogorsk zone probably caused the final exhumation of the granite–gneiss complexes and their exposition between unmetamorphic sedimentary rocks (Fig. 8b).

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