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Geochronological, geochemical and isotopic study of detrital zircon suites from late Neoproterozoic clastic strata along the NE margin of the East European Craton: Implications for plate tectonic models

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Arctida

A B S T R A C T

U/Pb dating, REE geochemistry and Lu/Hf isotopic studies utilizing LA-ICPMS on detrital zircon suites from Neoproterozoic siliciclastic rocks from the northeast periphery of the East European Craton (sandstones of Djejim Formation, Djejim-Parma Hills, Southern Timan Ridge and sandstones of the Engane-Pe Formation, Northern Engane-Pe uplift, Polar Urals) are used to assess the provenance of sediments and test tectonic models for the late Precambrian assembly of continents. The data support the conclusion that Neoproterozoic complexes of the Timan–Pechora region (TPR) are composed mainly of sedimentary rocks (*SW Pre-Uralides–Timanides*) eroded from Baltica and deposited along the passive margin of Baltica. However, late Precambrian–Early Cambrian volcanic-sedimentary and volcanic rocks, granitoids, and rare ophiolites of the TPR (*NE Pre-Uralides–Timanides)* comprise more juvenile material developed some distance from Baltica. Important differences exist between the U/Pb ages and Lu/Hf isotopic systematics of zircons from rocks of the NE Pre-Uralides–Timanides and the Neoproterozoic complexes of the Peri-Gondwanan terranes and do not support a Peri-Gondwanan origin for the NE Pre-Uralides–Timanides. In our interpretation, the SW Pre-Uralides–Timanides were deposited in the Neoproterozoic along the passive Timanian–Uralian margin of Baltica, but the NE Pre-Uralides–Timanides were formed along the active (Bolshezemel) margin of a paleocontinent called Arctida and were caught in the collision zone between the two paleocontinents, Arctida and Baltica.

1. Introduction

The Timan–Pechora region (TPR) is a triangular-shaped area located along the northeastern periphery of the East European Craton. The TPR includes the Timan Ridge region, the Pechora Basin, the near-shore marine areas of the Barents, Pechora and Kara seas, and the western slopes of the North, Sub-Polar and Polar Urals (Figs. 1 and 2). The TPR lies between two large cratons of Northern Eurasia: the East European and the Siberian cratons. The Ural Mountains represent a younger superimposed Paleozoic fold-thrust belt, which is also located between those two cratons.

There is a general consensus that the northern part of Wegener's Pangea (the late Paleozoic framework of present-day northern Eurasia and its Arctic shelves) was formed as a result of the assembly of several large Precambrian and early Paleozoic paleocontinents — Baltica (the Precambrian shield of East European Craton), Laurentia (Precambrian shield of North America), Siberia (Precambrian shield of Siberian Craton), Kazakh–Kyrgyz (a composite middle Paleozoic paleocontinent) and smaller terranes of different origins (micro-continents, oceanic arcs, fragments of oceanic basins, etc.). Those smaller terranes are involved in Phanerozoic orogenic belts that lie between the more ancient cratonic shields and along their edges. At the beginning of the Mesozoic, Pangea began to fragment. Opening of the Atlantic Ocean first separated North and South America from the Pangea supercontinent, and then, in the late Mesozoic and Cenozoic, oceanic rifting in the Atlantic propagated into the Arctic where new ocean basins (the Eurasia and Amerasia) were formed.

Thus, the northern margin of the ancient (Precambrian) cratonic shield of northern Eurasia formed as a result of late Neoproterozoic to Phanerozoic accretion, collision and rifting. Part of the geodynamic history of this northern margin has been reconstructed, but many questions remain to be answered. For example: (1) How many oceanic arcs and back-arc basins were collapsed to form the Phanerozoic orogenic belts? (2) What was the nature (passive or active?) of the colliding margin of the continents involved? (3) Are there, and what is the origin of, exotic terrains in the collisional orogenic belts? Thus, the sequence of events and the details of those events involved in the formation of this margin (including the TPR)

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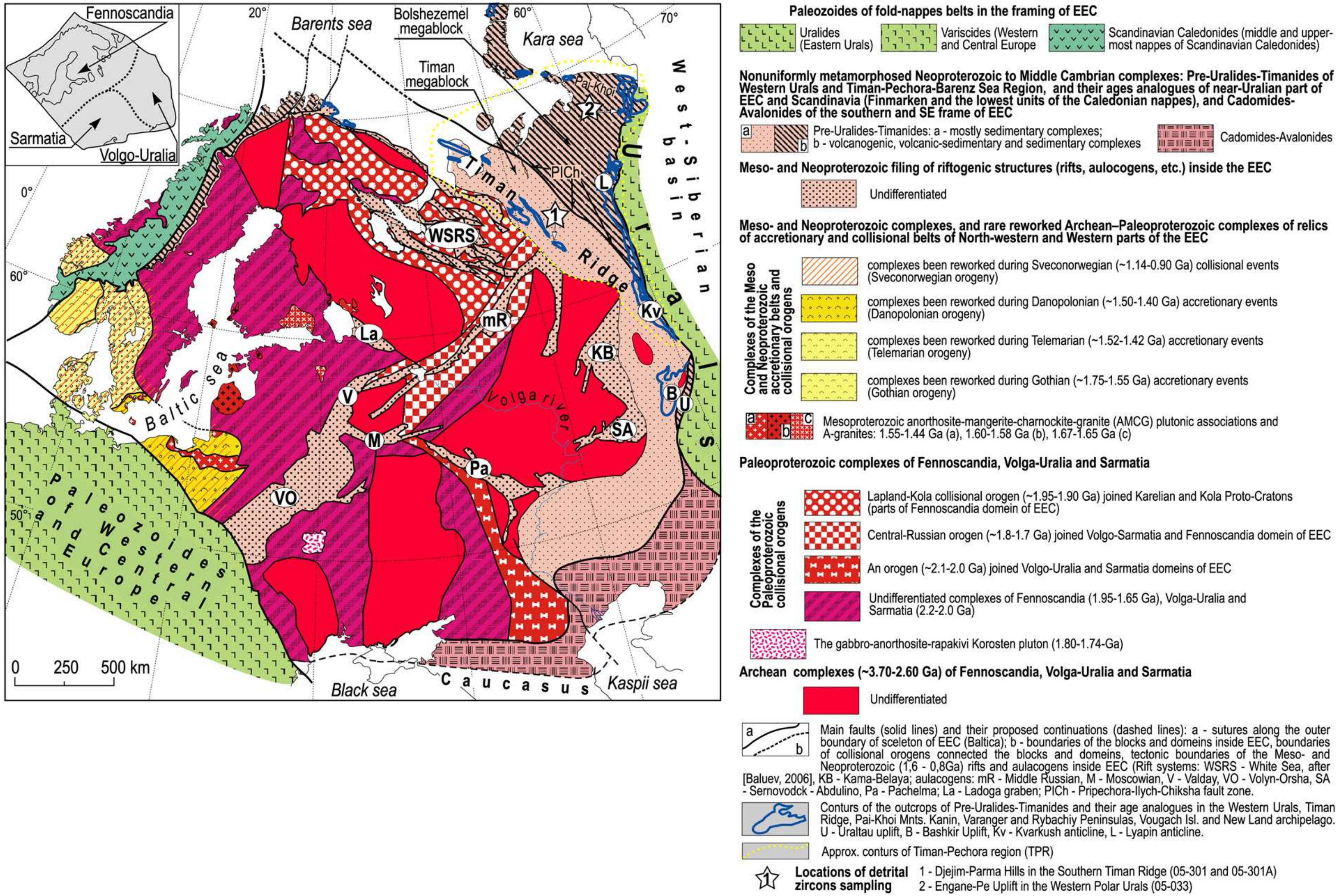

Fig. 1. Map of the main complexes and structures in the basement of the East European Craton and its periphery, including blocks of consolidated basement, rifted structures, and Neoproterozoic and Paleozoic fold-thrust belts. Late Paleproterozoic–early Neoproterozoic complexes of EEC from Bogdanova et al. (2008). Neoproterozoic–Middle Cambrian complexes at the eastern and northeastern periphery of the EEC after Kuznetsov et al. (2007a) and Kuznetsov (2009a,c). Configuration of the White Sea Rift System (WSRS) after Baluev (2006). Insert: Contours of EEC proto-Cratons Fennoscandia, Sarmatia and Volga-Uralia simplified from Bogdanova et al. (2008).

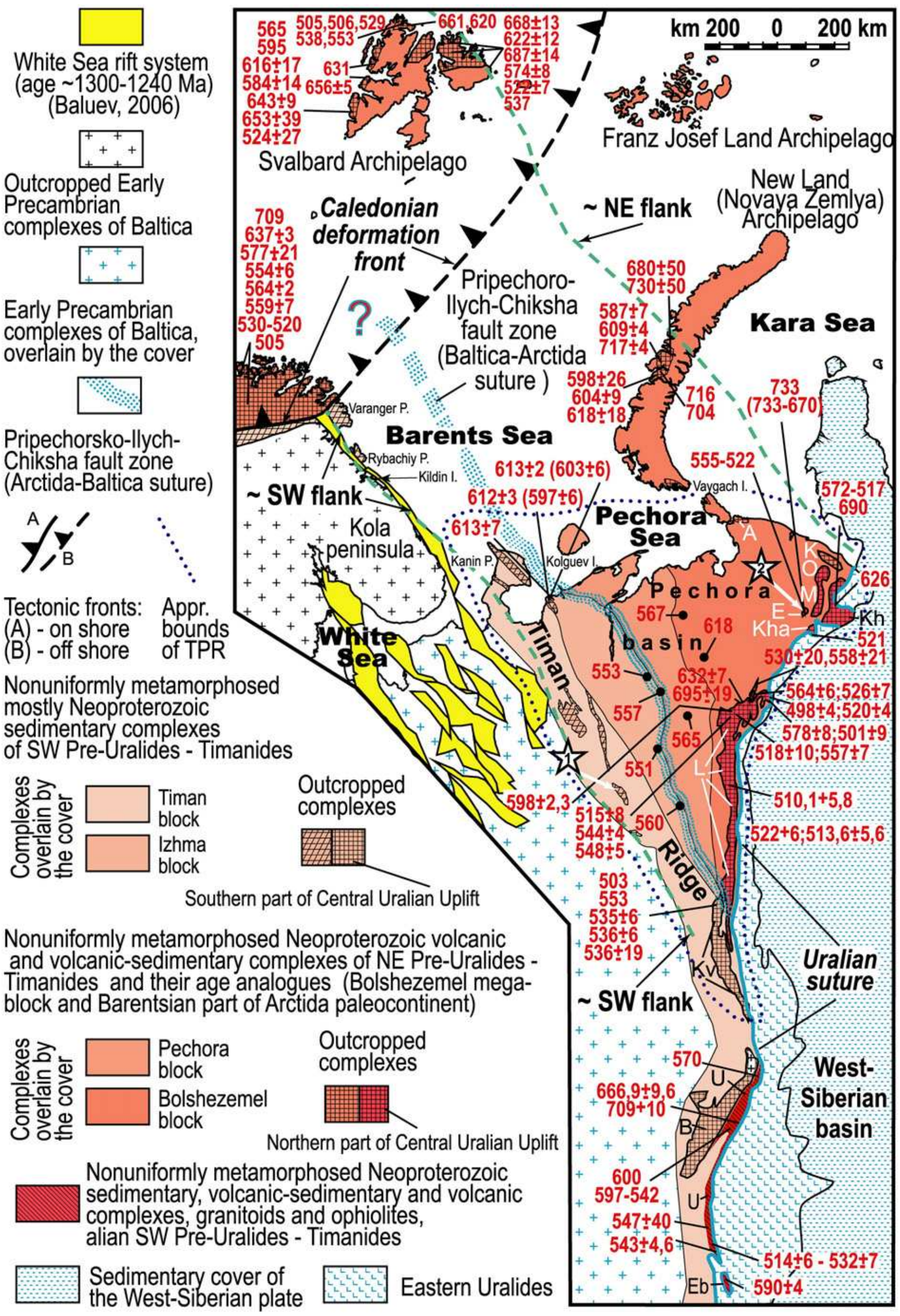

Fig. 2. Tectonic classification of the Pre-Uralide–Timanides and isotopic ages of its magmatic and metamorphic complexes from Kuznetsov (2009a,c), with minor additions. Map based on data from Bogdanov and Khain (1996) and Khain and Leonov (1999). Configuration of White Sea Rift System (WSRS) after Baluev (2006). A – Amderma, K – Kara, O – Ochenyrd, M – Manytanyrd, E – Engane-Pe, Kh – Kharbei, Kha – Kharamalataou, Ly – Lyapin, Kv – Kvarkush, U – Uraltau, B – Bashkir (including Taratash), E – Ebeta. *SW flank* and *NE flank* – approximate southwestern and northeastern boundaries of the Pre-Uralides–Timanides orogen. Stars – Sample locations (see Figs. 4 and 5). Numbers in red are isotopic ages of Pre-Uralides–Timanides magmatic and metamorphic rocks of TPR and their age corelatives in northwest Scandinavia, Svalbard and the Novaya Zemlya Archipelagos (after Gayer et al., 1966; Korago and Chuknonin, 1988; Nikiforov and Kaleganov, 1991; Rusin, 1996; Manechki et al., 1998; Andreichev, 1998; Andreichev and Yudovich, 1999; Andreichev and Larionov, 2000; Gee et al., 2000; Andreichev, 2003; Khain et al., 2003; Glodny et al., 2004; Johansson et al., 2004; Beckholmen and Glondy, 2004; Kuzenkov et al., 2004; Larionov et al., 2004; Leech and Willingshofer, 2004; Shishkin et al., 2004; Remizov and Pease, 2004; Soboleva, 2004; Soboleva et al., 2004a,b, 2005; Korago and Timofeeva, 2005; Udoratina et al., 2005, 2006; Andreichev and Litvinenko, 2007; Majka et al., 2007a,b; Samygin et al., 2007).

are still very much under discussion (see review of hypotheses in Kuznetsov et al., 2005). These topics are not only relevant for the exposed portions of the TPR but also for the vast portion of the orogen hidden by younger cover beneath the Arctic shelves (Fig. 2).

The study of the geochronology and isotopic characteristics of detrital zircon suites from sedimentary sequences can provide important information on the provenance of sediments, and is a technique that has been widely applied to solve stratigraphic, tectonic, paleogeographic, and other problems. We present the results of a study of detrital zircons from two localities in the TPR region (Figs. 1 and 2). Together with previously published data, the new data allows us to test tectonic hypotheses about the early stages of Pangea assembly and to propose an updated interpretation of the geodynamic (plate tectonic) evolution of the northeastern periphery of the East European Craton in the late Neoproterozoic to early Paleozoic.

2. Overview of the geologic and plate tectonic setting of the East European Craton, Timan–Pechora region and Urals

2.1. Geologic and tectonic setting of the East European Craton

The East European Craton (EEC) is broadly divided into basement and cover sequences. Basement rocks have been traditionally assigned to the early Precambrian (Archaean and Paleoproterozoic), and cover sequences to the late Precambrian (Meso- and Neoproterozoic) (Bogdanova, 1986 and references therein). However, it is not known if this two-part basement-cover division can be made in the western parts of the EEC where highly metamorphosed and deformed complexes may be younger in age (Neoproterozoic to 0.9 Ga; Bogdanova et al, 2008 and references therein).

The EEC was part of the paleocontinent Baltica (Proto-Baltica) in the late Neoproterozoic, and has been subdivided into three crustal blocks — Fennoscandia, Volgo-Uralia and Sarmatia (Bogdanova, 1991). Subsequent research revealed a complex inner architecture to these individual blocks (Glebovitskii, 2005; Bogdanova, 2005; Daly et al., 2006; Slabunov et al., 2006; Shchipansky et al., 2007; Mints et al., 2007; Bogdanova et al., 2008). The assemblage of the early Precambrian nucleus of the EEC began at ~2.1 Ga with the collision of Sarmatia and Volga-Uralia. As a result of this event, the Volga-Sarmatia proto-craton was formed (Shchipansky et al., 2007), which collided with Fenoscandia at ca. 1.8–1.7 Ga (Bogdanova, 2005; Bogdanova et al., 2008) to form an agglomerate of ancient continental blocks termed Proto-Baltica. Along the western margin of Proto-Baltica, accretionary and collisional processes took place intermittently until ca. 0.9 Ga. They are represented by the Gothian (1.73–1.55 Ga), Telemarian (1.52–1.48 Ga), Danopolonian (1.47–1.42 Ga) and Svekonorwegian (1.14–0.90 Ga) orogenic events.

In the late Neoproterozoic, these four orogens were added to Proto-Baltica and the erosional products derived from them accumulated in marginal (passive continental margin) and intra-cratonic sedimentary basins. The early intra-cratonic basins were rift basins spatially linked to older collisional belts of early Precambrian age (Bogdanova et al., 2008 and references therein). During the Mesoproterozoic, Proto-Baltica and other old continents are thought to have been amalgamated into the supercontinent, Rodinia (Meert and Powell, 2001; Li et al., 2008 and references therein; Santosh et al., 2009) or the supercontinent Paleo-Pangea (Piper, 2000; Baluev, 2006; Piper, 2007). During the late Neoproterozoic, this supercontinent broke up and the nucleus of the EEC was isolated as an independent continent, Baltica (Hartz and Torsvik, 2002; Bogdanova et al., 2008 and references therein). Baltica existed as an independent continent until the "Neoproterozoic–Cambrian" boundary (e.g., Puchkov, 1997, 2000; Kuznetsov et al., 2005, 2007a) or the beginning of the Middle Paleozoic (e.g., Cocks and Torsvik, 2005).

2.2. Geological setting of Timan Ridge and Pechora basin

The geology of the Pechora Basin region can be generally subdivided into platformal cover sequences, and deformed and folded basement rocks (Fig. 3). The folded basement – named the Timanides — is represented by deformed Neoproterozoic to Middle Cambrian stratigraphic units that are overlain unconformably by several kilometers of post-Cambrian sedimentary cover. Because the basement is deeply buried, it is poorly known, although several boreholes have reached basement rocks (Beliakova and Stepanenko, 1991; Dovzhikova et al., 2004). Syntheses of the geology the Timan–Pechora region have been published by Gee et al. (2000), Roberts and Siedlecka (2002), Dovzhikova et al. (2004) and Pease et al. (2004).

The upper age limit for Timanide deformation is Late Cambrian–? Early Ordovician, but its lower age limit is not well defined. The oldest ages obtained from Timanide units are Neoproterozoic, although some authors have argued for a late Mesoproterozoic age for some Timanide units (Olovyanishnikov, 1998; Roberts and Olovyanishnikov, 2004).

The basement rocks of the Pechora Basin crop out in the western Urals (see below) and to the southwest in the Timan Ridge region. The Timan Ridge can be traced southeast to the Middle Urals (Kvarkush anticline) and northwest to the Kanin Peninsula. Beneath the Phanerozoic sedimentary cover, outcrops along the Timan Ridge are represented by Neoproterozoic complexes (Fig. 2).

Complexes that are age equivalent to the Timanides are known in some neighboring regions: on the Yugorsky peninsula (Pai-Khoi ridge), on Vaygach Island, in the south part of the Novaya Zemlya archipelago, along the north border of the Baltic shield (Kildin Island and on the Sredny, Rybachiy and Varanger peninsulas), and on the islands of Svalbard (Fig. 1.2).

2.3. Geological setting of the Urals

The Urals take the form of a N–S trending late Paleozoic fold-thrust belt, which exposes Late Cambrian to late Paleozoic rocks that collectively form the *Uralides*. In contrast, older rocks complexes in and west of the Urals are collectively named the *Pre-Uralides* (cf. Puchkov, 2003; Stern, 2008) (Fig. 3). The Pre-Uralides are clearly separated from the Uralides by a tectonic contact. The subdivision between the two was first made by Kheraskov (1948).

The Uralian fold-thrust belt is characterized by prominent N–S trending tectonic zones. Along strike, the Urals are subdivided into the Eastern-Uralian and Western-Uralian (~western slope of the Urals) megazones by the Main Uralian Fault (MUF) or Uralian Suture. The eastern Uralides are mostly covered by younger sediments, but are generally thought to be allochthonous with respect to EEC (Baltica). The Western-Uralian megazone contains both the Uralides and the Pre-Uralides.

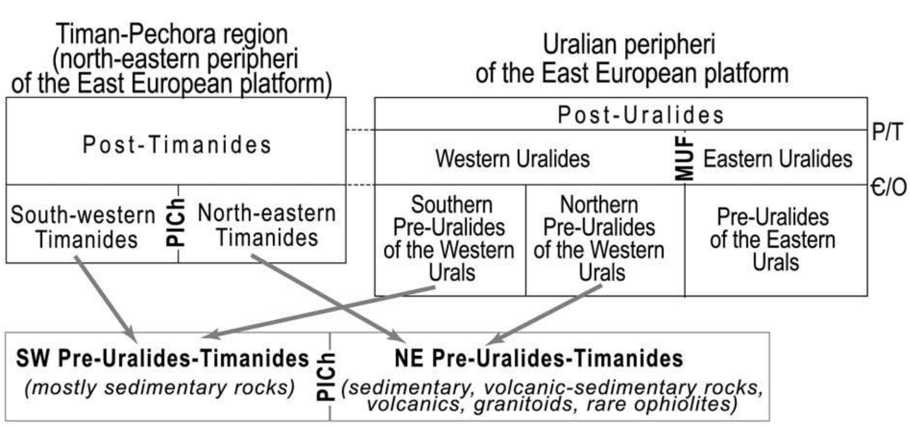

Fig. 3. General historical-geological characterization of complexes from the eastern and northeastern periphery of the East European Platform. MUF – Main Uralian Fault (Uralian suture), PICh – Pripechora-Ilych-Chiksha fault zone (Baltica – Arctida suture). See Figs. 1 and 2 for location.

The Uralides of the Western-Uralian megazone consists of Upper Cambrian to Upper Paleozoic sedimentary (carbonate, siliceous and terrigenous) rocks. Their Pre-Uralide basement consists of variably metamorphosed sedimentary, volcanic-sedimentary and volcanogenic rocks, intrusions (granites and gabbros), and rare ophiolites of Neoproterozoic to Middle Cambrian age. The Pre-Uralides of the Western-Uralian megazone are mostly stratigraphically overlain by Paleozoic strata. The exposed complexes of the Western-Uralian Pre-Uralides form a chain of uplifts along the entire length of the Urals, called the Central Uralian Uplift (CCU). CCU-related structures include (from south to north): the Ebeta antiform, the Uraltau and Bashkir uplifts, the Kvarkush and Lyapin anticlinoriums, the Sob' (including the Kharanatalow, Engane-Pe, Manitanyrd-Paipudyna, and Kharbei uplifts and anticlines), and the Ochenyrd (Malaya Kara) uplifts (Fig. 2). The Pre-Uralides of the Western Urals in the North, Sub-Polar and Polar Urals are Timanides involved in late Paleozoic (Uralian) orogenesis.

The Neoproterozoic complexes of the Ebeta antiform and Uraltau uplift consist mainly of non-uniformly faulted and metamorphosed (up to eclogitic faces) volcanic, volcanic-sedimentary rocks, granites and ultramafic rocks (Lennykh et al., 1995; Hetzel, 1999; Ivanov, 1998; Leech and Willingshofer, 2004; Samygin et al., 2007). There is some indication that the complexes of the Ebeta antiform and Uraltau uplift are exotic to Baltica (the neighboring parts of the EEC) (Kuznetsov et al., 2005; Kuznetsov, 2009c; Puchkov, 2000; Willner et al., 2003, etc.).

In contrast, the Pre-Uralides of Bashkir uplifts and Kvarkush anticlinorium consist predominantly of sedimentary rocks (Maslov, 2004 and references therein). Only rare horizons of alkali and sub-alkali basaltic lavas are exposed (Karpukhina et al., 1999; Nosova, 2007). There is strong stratigraphic, sedimentalogical and paleogeographic evidence that the Pre-Uralides of the Bashkir uplift and Kvarkush anticlinorium have affinities with the Neoproterozoic complexes of the Kama–Belaya aulacogen system of the Eastern part of the EEC (Maslov, 2004). We refer to the Neoproterozoic complexes of the Bashkir uplifts and Kvarkush anticlinorium as the *Southern Pre-Uralides* (Fig. 3).

The Pre-Uralides complexes in the Lyapin anticline and the northern structural elements of the CUU are predominantly volcanic and volcanic-sedimentary rocks with granites and rare ophiolites (Mizin, 1988; Dushin, 1997; Scarrow et al., 2001; Khain et al., 2003; Remizov and Pease, 2004; Soboleva, 2004; Kuznetsov et al., 2007a etc.). We name these Neoproterozoic complexes the *Northern Pre-Uralides* (Fig. 3).

2.4. Classification and tectonic origin of the Pre-Uralian Timanide complexes

Collectively, the Pre-Uralides of the Western-Uralian megazone and the Timanides and their corelatives beneath the Barents and Kara shelves and exposed on Svalbard are referred to as Pre-Uralide–Timanides. Following Shatsky (1946), Kheraskov and Perfil'ev (1963), Zhuravlev and Gafarov (1959), these Pre-Uralide–Timanide complexes belong to two groups that differ both compositionally and in terms of their spatial position (Figs. 2 and 3):

(1) The Neoproterozoic complexes of Timan Ridge and adjacent part of the Pechora basin, together with Neoproterozoic complexes of the southern parts of the western slope of the Urals (Kvarkush anticlinorium and Bashkir Uplift of the CUU), are composed mainly of sedimentary rocks (Offman, 1961; Olovyanishnikov, 1998; Maslov et al., 2002, Maslov, 2004; Roberts and Olovyanishnikov, 2004 and references therein). They are collectively termed the *SW Pre-Uralides–Timanides* (Figs. 2 and 3).

(2) The late Precambrian complexes of the northeastern part of the basement of the Pechora basin and the northern Pre-Uralides (pre-Late Cambrian–Early Ordovician rocks of the Lyapin anticlinorium and all the more northern structures of the CUU) are made up of volcanic-sedimentary and volcanic rocks, granitoids and rare ophiolites (Mizin, 1988; Dushin, 1997; Scarrow et al., 2001; Khain et al., 2003; Soboleva, 2004; Dovzhikova et al., 2004, Dovzhikova, 2007; Kuznetsov et al., 2005, 2007a,b; and references therein). They are collectively called the *NE Pre-Uralides–Timanides*.

The SW and NE Pre-Uralides–Timanides are separated by the Pripechoro-Ilych-Chiksha (PICh) fault zone (Fig. 2) which can be traced by its gravity and magnetic signatures (Kuznetsov et al., 2007a). Many aspects of the geology of the SW Pre-Uralides–Timarides (e.g., relations of individual sequences of different scale, their biostratigraphic and isotopic-geochronological characteristics) are presently well understood. In contrast, the NE pre-Uralides–Timanides are poorly known. This is because of the poor exposure of the NE Pre-Uralides–Timanides complexes, in particular, their burial beneath thick (up to 6–12 km) Phanerozoic cover in the Pechora Basin and the relatively difficulty in accessing their outcrops in the Polar Urals. As a result, many problems of the geology and stratigraphy of the NE Pre-Uralides–Timanides remain unresolved.

3. Study of detrital zircons from Timanides–Pre-Uralides

3.1. Sampling of rocks and preparation of zircons

Rocks were sampled in two locations (Fig. 2). One location was at the western boundary of TPR close to the proposed boundary of the Precambrian craton of the EEC – Baltica. Here, we sampled rocks of the *SW Pre-Uralides–Timanides*. The second location was on the opposite or eastern side of the TPR and far from Baltica. Here we sampled rocks of *NE Pre-Uralides–Timanides*.

3.1.1. Location 1

The Djejim-Parma Hills are located in the southern part of the Timan Ridge (Fig. 4). Here, Neoproterozoic rocks are represented by red sandstones and siltstones of the Djejim Formation and by limestones (sometimes dolomitized) and silty-argillites of the Pavyuga Formation. The inner structure of the Neoproterozoic complexes of the Djejim-Parma Hills is poorly known because there are few outcrops in streams and in the Asyvvozh quarry. Nevertheless, the two formations are believed to be tectonically juxtaposed. Generally, Neoproterozoic complexes occur in the cores of large anticlines, the limbs of which consist of Late Devonian and younger Paleozoic sedimentary units. We collected two samples in the eastern part of the Asyvvozh quarry (61°47′11.5″N, 54°06′35.2″E.) from different lithologies but in adjacent beds of the Djejim Formation. Sample 05-301 was taken from red silty sandstones with well-developed ripple marks and sample 05-301A was taken from red cross-bedded sandstones.

3.1.2. Location 2

The Engane-Pe Uplift is located at the junction of the Polar Urals with the northeastern part of the Pechora Basin. The Pre-Uralides–Timanides form the core of a large (~60 × 20 km) NE-trending anticline (Fig. 5), the limbs of which are composed of unconformably overlying ?Upper Cambrian–Tremadocian to Upper Paleozoic strata. The basal horizons above the unconformity (?Late Cambrian–Tremadocian Manitanyrd Formation) are composed mainly of red colored cross-bedded arkosic sandstones. The underlying Pre-Uralide–Timanide complexes include faulted and metamorphosed (up to greenschist facies) sedimentary, volcanic-sedimentary, and volcanic and sub-volcanic rocks (Mizin, 1988; Dushin, 1997; Scarrow et al., 2001; Khain et al., 2003, Kuznetsov, 2009c) intruded by rare gabbro-dolerite and rhyolite bodies. SHRIMP isotopic dating of single zircon crystals from rhyolites of several isolated sub-volcanic bodies on the southern edge of the uplift yielded an age of 555–522 Ma (Shishkin et al., 2004), which provides an upper age limit for the Pre-Uralides–Timanides in the Engane-Pe Uplift.

Stratified rock units of the Engane-Pe Uplift are represented by two Neoproterozoic sequences: (i) basalts, andesites, dacites, and rhyolites with interbedded tuffs and tuffaceous sediments that are

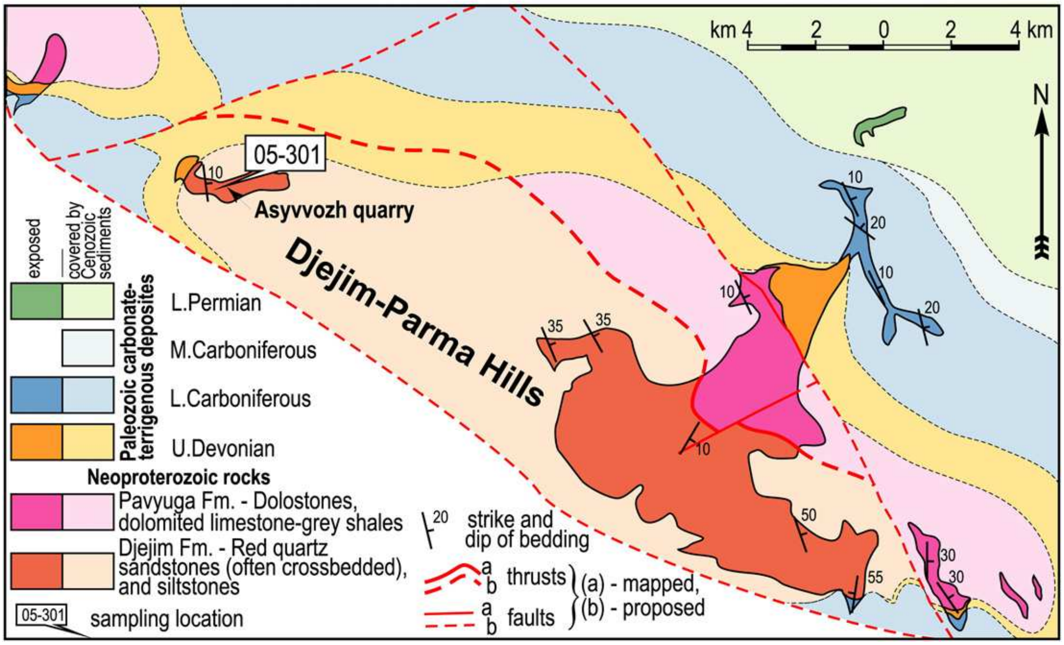

Fig. 4. Geological map of the Djejim-Parma Hills, southern Timan Ridge (compiled using data from Slutskiy et al., 1984).

sometimes altered to greenschist grade and are known in the regional geological literature as the Bedamel Formation; (ii) siliceous and pelitic rocks, siliceous mudstones and siltstones, quartz and sub-arkosic sandstones with rare horizons of conglomerates (diamictites) collectively named the Engane-Pe Formation (Stratigraphical Committee of Russia, 1993). The contact between these two sequences in the Manyukuyakha and Yaneskewlektalba river watersheds (north-west of the Engane-Pe Uplift) is marked by a serpentine mélange with ultramafic, gabbroic, and plagiogranite blocks. These rocks represent the mélange of an ophiolite assemblage (Scarrow et al., 2001) located within several hundred meters of the sample location. Plagiogranite from this complex was dated at 670 Ma (Khain et al., 2003), which is inferred to be the crystallization age of the ocean crust protolith.

The core of the Engane-Pe anticline encloses the Pre-Uralides–Timanides within the fault-bound NNE-striking Izyavozh antcline, the core of which (upper reaches of the Levyi and Pravyi Izyavozh Rivers, and upper reaches of the Enganeyakha River) includes tuffaceous sediments (in the lower part) and volcanogenic and volcanic-sedimentary rocks (in the upper part) of the Bedamel Formation. This formation is overlain structurally by the Engane-Pe Formation consisting of alternations of siltstones with rare thin layers of sandstone, both containing a common tuffaceous component. At the northern margin of the Izyavozh anticline (on the left wall of the Kamashor River valley) and in its eastern limbs (at the source of the Shervozh River and on the left valley wall of the left tributary of the Enganeyakha River) the sedimentary rocks of the Engane-Pe Formation are overlain structurally by the Bedamel volcanic rocks. Hence, rocks of the Bedamel Formation occur at different structural levels within the Engane-Pe Uplift, being structurally underlain and overlain by sedimentary rocks of the Engane-Pe Formation. Hence, the entire structure can be interpreted as an antiformal packet of large-amplitude recumbent isoclinals folds. We think (Kuznetsov et al., 2007b): (i) that these heterogeneous complexes were first tectonically juxtaposed into a single nappe complex, which was later folded into a packet of flat-lying isoclinals folds, and (ii) that the Bedamel and Engane-Pe Formations are approximately the same age. Ultimately, the whole packet was folded along with the Uralides into a simple antiform during the Uralian tectonic stage.

The sample 05-033 was collected at 67°21′30.4″N, 64°47′54.4″E from greenish-grey quartzose sandstones (with subordinate greywacke and tuffaceous sediments) in the upper part of the Engane-Pe Formation exposed on the right bank of the lower reaches of Tumannyi Brook, a right upper tributary of the Manyukuyakha River in the north part of the Engane-Pe uplift.

3.2. Sample preparation

Initial processing of all rock samples and the separation of zircons concentrate was carried out by N.B. Kuznetsov at GIN RAS. Samples were crushed manually in an iron crucible to prevent contamination common in mechanized rock crushers. The crushed material was sifted with standard sieve sizes. Crushed 0.3 mm and finer materials were washed to remove clay-size material and dried. A hand magnet was used to separate magnetic minerals. Material less dense than 2.7 gm/cm^3 was removed with heavy liquids. The final $>$2.7 gm/cm^3 heavy mineral concentrates were mainly composed of grains of zircon, garnet and apatite. The heavy fractions were sent to the GEMOC Centre, where the concentrates were further processed by E. Belousova and L. Natapov using slope magnetic separation and higher density heavy liquids. Zircon grains were mounted in epoxy discs and polished. All grains were imaged in the GEMOC CAMEBAX SX50 electron microprobe (EMP) using back-scattered electron/cathode-luminescence (BSE/CL) techniques; images were captured digitally using a LINK analysis system.

3.3. Analytical methods

Isotopic and geochemical studies of detrital zircons were conducted at the GEMOC Center, and included (1) U/Pb dating of zircons, (2) trace-element characterization of the zircons allowing determination of the composition of the parent igneous rock, and (3) analysis of zircon Lu/Hf isotopic systematics in order to estimate the model (T_{DM}) age of the parental protolith of the magma that originally contained the zircons. Our analysis of the data follows the TerraneChron^{TM} approach (O'Reilly et al., 2004). All measured values were processed using the Isoplot program (Ludwig, 2003).

3.3.1. U/Pb dating

The samples were acid-washed before being analysed to remove possible surface Pb contamination. The grains were analysed using a

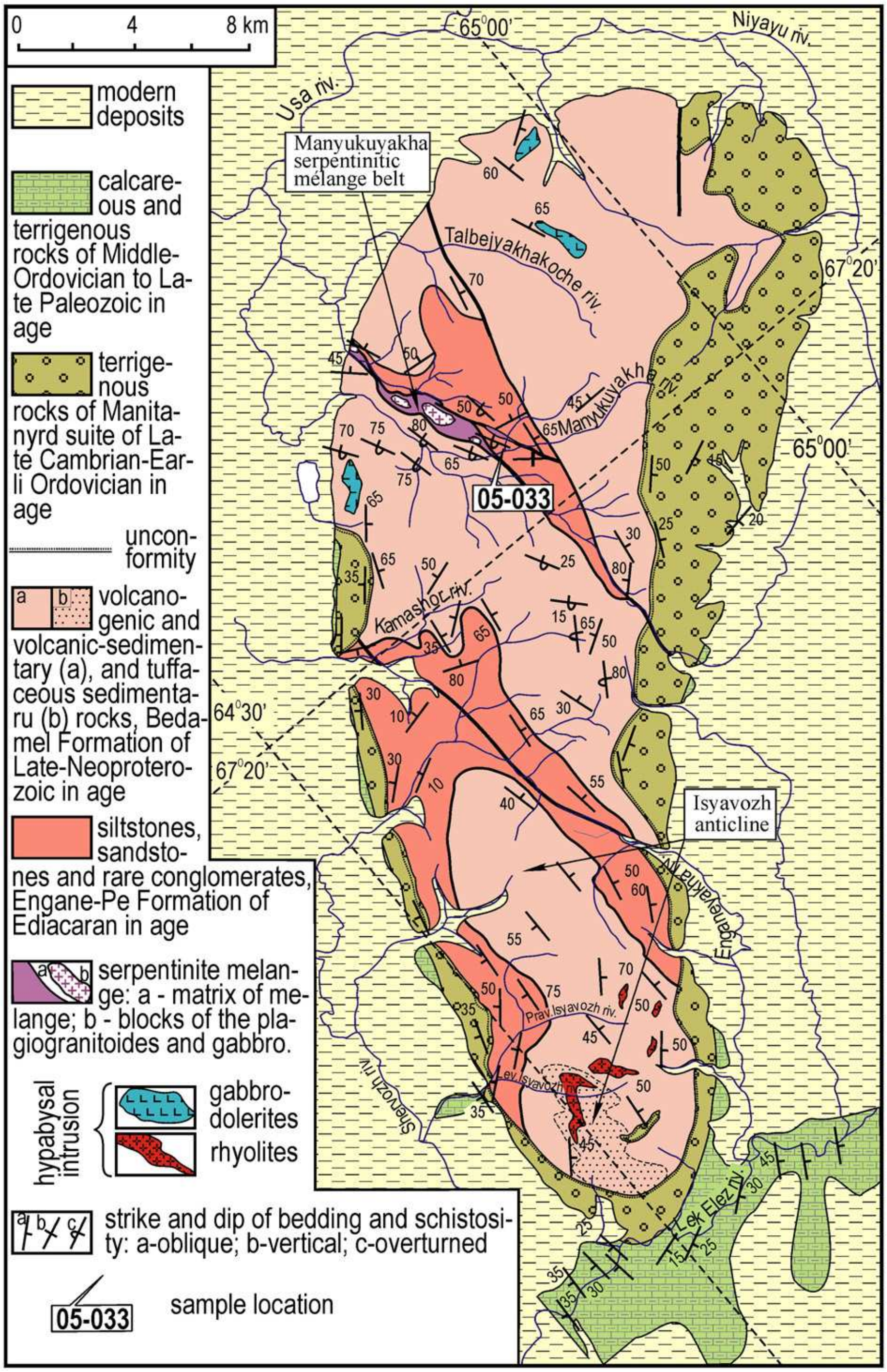

Fig. 5. Geological map of the Engane-Pe Uplift (compiled using data from K.G. Voinovski-Kriger, V.N. Gesse, A.A. Savelev, B.Ya. Dembovskii, M.A. Shishkin, O.N. Malykh, I.M. Malykh, P.E. Popova et al. and the author's field observations).

commercial LUV213 laser ablation system ($\lambda = 213$ nm) (New Wave/Merchantek), attached to an Agilent 7500s ICPMS (Jackson et al., 2004). All ablations were carried out in He. Ablation pits were about 50 µm in diameter. The time-resolved signals were processed using the GLITTER interactive software to select the portions of the grains that had suffered least lead loss, or gain of common Pb, and were thus closest to being concordant.

The standard used in this work is the GEMOC-GJ-1 gem zircon with a TIMS age of 608.5 Ma. This standard is run 2 times before and after each ten unknowns. Cross-analysis of other international standards also gives good results. Two analyses of Mud Tank zircon (734 ± 32 Ma; Black and Gulson, 1978) and two of zircon 91500 (1064 Ma; Wiedenbeck et al., 1995) were run during this work, and their mean values are within 2 s.d. of the recommended values (Jackson et al., 2004).

3.3.2. Parental magmas for detrital zircons

An extensive study of the trace-element patterns in zircons (Belousova et al., 2002) has shown good correlations with the composition of the magmatic host rocks. The zircon database for U, Th, Y, Yb, Lu and Hf acquired during the EMP, U/Pb and Hf-isotope analyses were evaluated using CART statistical software (Breiman et al., 1984). This analysis creates a classification tree based on simple binary switches, which allows classification of any individual zircon grain in terms of its parental rock type. Zircons derived from a

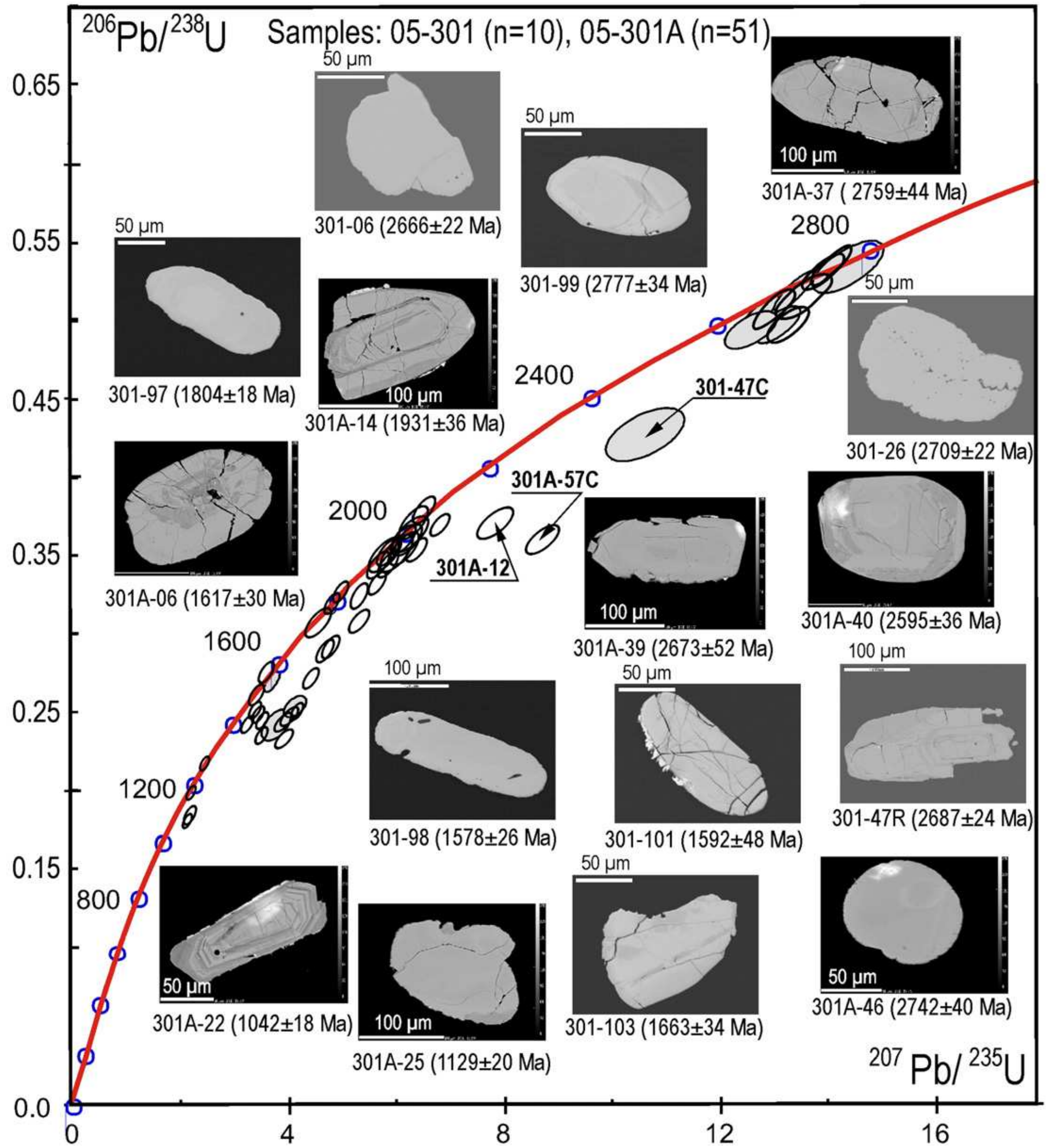

Fig. 6. Concordia diagram for zircons from samples 05-301 and 05-301A of the Djejim Formation, Djejim-Parma Hills, southern Timan Ridge. Back-scattered electron images of several zircons, their numbers, and U/Pb isotopic ages (in Ma) are also shown.

parental rock classified as a high-Si granitoid (>70–75% SiO_2) or its volcanic equivalent are termed "granitic" or derived from "granites", whereas those derived from a parental rock classified as a low-Si granitoid (<65% SiO_2) or its volcanic equivalent are termed "dioritic". Similarly, we refer to zircons derived from mafic rocks as "mafic" or derived from "gabbro", and those with trace-element contents similar to zircons from alkaline rocks (e.g., syenite and carbonatite) as being derived from "syenite".

3.3.3. Hf-isotope analyses

Hf-isotope analyses were carried out in situ with a New Wave UP 213 nm laser ablation microprobe attached to a Nu Plasma multi-collector ICPMS. The analyses were carried out with a beam diameter of ca 50 µm, a 5 Hz repetition rate, and energies of about 0.1 mJ/pulse and 0.6 J/cm^2. Typical ablation times were 80–120 s, resulting in pits 40–60 µm deep. The methodology and analyses of standard solutions and standard zircons are described by Griffin et al. (2000).

The measured ^{176}Lu/^{177}Hf ratios are used to calculate initial ^{176}Hf/^{177}Hf ratios. The typical 2SE uncertainty on a single analysis of ^{176}Lu/^{177}Hf is $\pm$1–2%, reflecting both analytical uncertainties and the spatial variation of Lu/Hf across many zircons; at the Lu/Hf ratios considered here, this contributes an uncertainty of <0.1 ε_{Hf} unit. For the calculation of ε_{Hf} values, we have adopted the chondritic values of Blichert-Toft and Albarede (1997). To calculate model ages (T_{DM}) based on a depleted mantle source, we have adopted a model with (^{176}Hf/^{177}Hf)i = 0.279718 and ^{176}Lu/^{177}Hf = 0.0384; this produces a value of ^{176}Hf/^{177}Hf (0.28325) similar to that of average MORB over 4.56 Ga. There are currently three proposed values for the decay constant for ^{176}Lu: 1.93 × 10–11 yr^{-1} proposed by Blichert-Toft and Albarede (1997); 1.865 × 10–11 yr^{-1} by Scherer et al. (2001); and 1.983 × 10–11 yr^{-1} by Bizzarro et al. (2003); calculations using all three are provided in the data tables. The values of the ε_{Hf} and model ages used in the figures were calculated using the value proposed by Scherer et al. (2001).

T_{DM} ages, which are calculated using the measured ^{176}Lu/^{177}Hf of the zircon, can only give a minimum age for the source material of the magma from which the zircon crystallised. Therefore we also have calculated for each zircon a "crustal" model age (T^C_{DM}), which assumes that the parental magma was produced from average continental crust ($^{176}Lu/^{177}Hf = 0.015$) originally derived from depleted mantle.

3.4. Results

3.4.1. Description of zircons

Samples 05-301 and 05-301A: Sample 05-301 (silty sandstones) contained a few (10) zircons in the size range 70–160 µm. Sample 05-301A (sandstone) contained many zircon grains about 200 µm in size. All 10 zircons from sample 05-301 were picked for analysis, and 51 zircons were selected from sample 05-301A. All zircons were semi-clear rounded grains (Fig. 6).

Sample 05-033 contained about 100 small (50–100 µm), cloudy to semi-clear, prismatic and rare sub-rounded zircon grains, forty eight of which were dated. Oscillation zoning was observed to a variable extent in all back-scattered images, suggesting a magmatic origin for these zircons (Fig. 7). The images revealed that only some grains contained cores overgrown by thin (about 10–20 µm) rims, whereas the majority show no evidence of zoning. Cores and rims were analysed in 12 grains. However, the rim analyses were rejected because they were too narrow (i.e. thinner than the diameter of the ablation crater) to produce reliable data.

3.4.2. U/Pb dating

3.4.2.1. Samples 05-301 and 05-301A. The U/Pb isotopic ages of 61 dated zircons from these two samples vary from 2.972 to 1.175 Ga with no younger ages represented (Figs. 6 and 8). Three results (05-301A-12, 05-301A-57C and 05-301-47C) differ significantly from concordia and were excluded, the last one (2.651 Ga) coming from the core of a grain, the rim of which (05-301-47R) yielded an older concordant age (2.748 Ga). A fourth grain (05-301-16) was excluded because of significant differences in its $^{207}Pb/^{206}Pb$, $^{207}Pb/^{235}U$ and $^{206}Pb/^{238}U$ ages. Of the remaining zircons, 4 grains are Mesoproterozoic and 7 grains are Archean (Neoarchean). Zircons from both samples have similar age ranges. Analysis of grain 05-033-60 was also rejected because it produced a very discordant age.

3.4.2.2. Sample 05-033. Analyses of 47 zircons from sample 05-033 show a range in U/Pb age from 1143 to 590 Ma (Figs. 7 and 9). One zircon yielded a Mesoproterozoic age of 1143 ± 20 Ma. The remaining zircons fall into two populations — “Population A” zircons (~65%), which fall in the range 760–675 Ma with a distinct peak at ~704 Ma, and “Population B” zircons (~35%), which fall in the range of 670–590 Ma with minor peaks at ~656 and 628 Ma.

3.4.3. Trace-element analysis

3.4.3.1. Samples 05-301 and 05-301A. Satisfactory trace-element concentrations were obtained for 47 of the 61 dated grains (Figs. 8 and 10). Of these, twenty-two (about half) have trace-element concentrations compatible with an origin from “granitic” rocks, 12 grains are from “dioritic” rocks, 5 grains from rocks of “syenitic” composition and 8 from “mafic” rocks. Thus, the majority of zircons (>75%) were sourced from “granites” and “diorites” or their volcanic equivalents, with “granitic” zircons predominating.

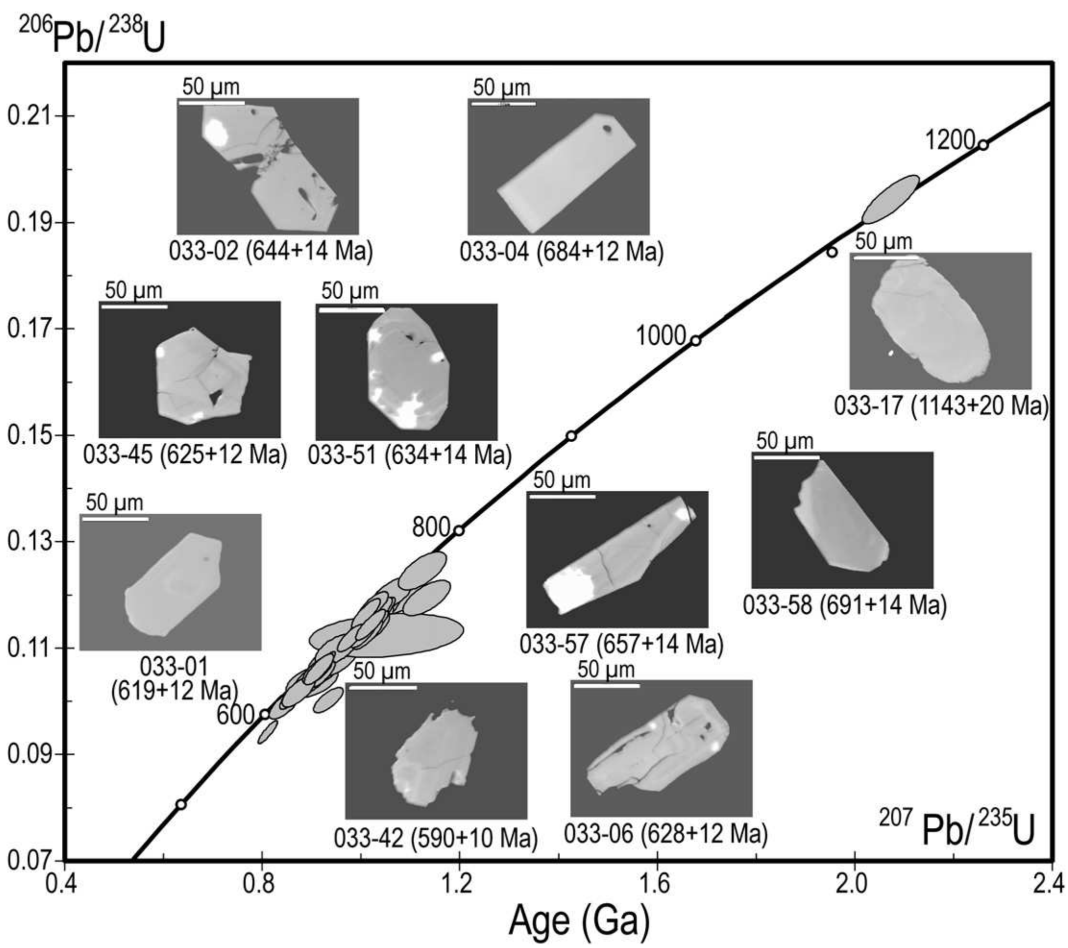

Fig. 7. Concordia diagram for zircons from sample 05-033 of the Engane-Pe Formation, northern part of the Engane-Pe Uplift. Back-scattered images of several zircons, their numbers, and U/Pb isotopic ages (in Ma) are also shown.

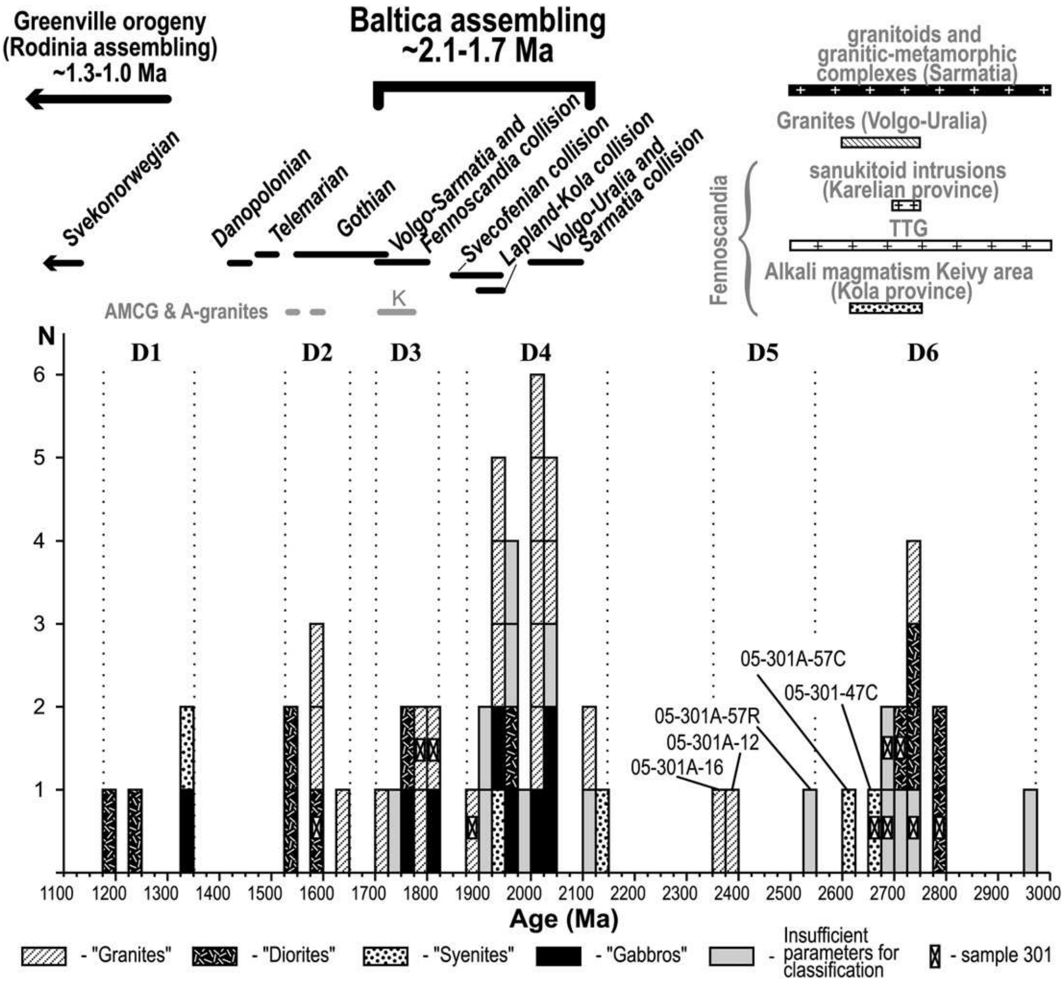

Fig. 8. Histogram of U/Pb isotopic ages of detrital zircons (61 analyses) from sandstones of the Djejim Formation, Djejim-Parma hills, southern Timan Ridge (samples 05-301 and 05-301A). Classification of parental magma types based on method described by Belousova et al. (2002). D1–D6 are zircon groups described in the text. Time intervals of the main tectonic events in the basement of the EEC shown as black bars in the upper part of the figure. Ages of AMCG and A-granites (including Rapakivi granites) are shown by grey bars. K = age of Korosten pluton.

3.4.3.2. Sample 05-033. Trace-element concentrations show that the majority (~55%) of the zircons from this sample (26 grains) were derived from "diorites", with 11 grains (~25%) from "granites", only 5 grains (~8%) from "mafic" rocks, and a single grain from "syenite" (Figs. 9 and 11). Three grains were not classified. There is no obvious correlation between source rock type and zircon age in either of the two (A and B) age populations. However, it is worth noting that no "granitic" zircons are present in grains older than 700 Ma. A single Mesoproterozoic grain (1143 Ma) shows a distinct composition suggesting derivation from an alkalic igneous source. Thus, the majority of zircons (>80%) from this sample were derived from "granites" and "diorites" or their volcanic equivalents with "diorites" prevailing.

Hence, the zircons from the two sample areas not only display different age ranges, but were also derived from compositionally different sources.

3.4.4. Lu/Hf-isotope analyses

3.4.4.1. Samples 05-301 and 05-301A. Satisfactory Lu/Hf data were obtained from 47 of the 61 grains dated (Fig. 10). The zircons show a wide range of ε_{Hf} from positive values (+8) characteristic of magmas derived from a depleted mantle source, to negative values (−15) that suggest crustal involvement or a crustal component in the magmas yielding these zircons. Model ages of the source areas of the parental magmas yielding the zircons vary from 1.40 to 3.24 Ga. For many of the zircons, the U–Pb and T^{C}_{DM} ages are very close (points lie close to DM line in Fig. 10). This means that the crust involved in the magma genesis of the parental rocks was juvenile. Some zircons show high values of radiogenic materials of the Lu/Hf isotopic system, indicating input of ancient remobilized crust in the magmatic source areas of the parental rocks.

3.4.4.2. Sample 05-033. The older (763 to 675 Ma) age population (A) of sample 05-033 plots mostly between the CHUR (line of isotopic evolution of a chondrite homogeneous reservoir) and depleted mantle (DM) curves (Fig. 11), with some grains plotting slightly below the CHUR line. A few zircons (4 grains) plot close to the DM line and have ε_{Hf} values ranging from about +11 to +13. The data indicate that the parental magmas for zircons of this age were juvenile with a minor contribution from recycled older crust. The T^{C}_{DM} crustal model age for zircons of this age population is ~0.84 Ga.

Some zircons plot below the CHUR line and have ε_{Hf} values ranging from −2.5 to +2.6. The T^{C}_{DM} crustal model age for the least radiogenic zircons from this cluster is ca. 1.6–1.8 Ga. All remaining zircons have ε_{Hf} values between +2.6 and +6.4.

The younger (590–670 Ma) age population (B) plots mostly between the CHUR and depleted mantle (DM) curves with none falling below CHUR. The ε_{Hf} values of this population vary from +3 to +6. The population is characterised by more homogeneous Hf isotopic signatures and a less significant contribution of recycled older crust compared

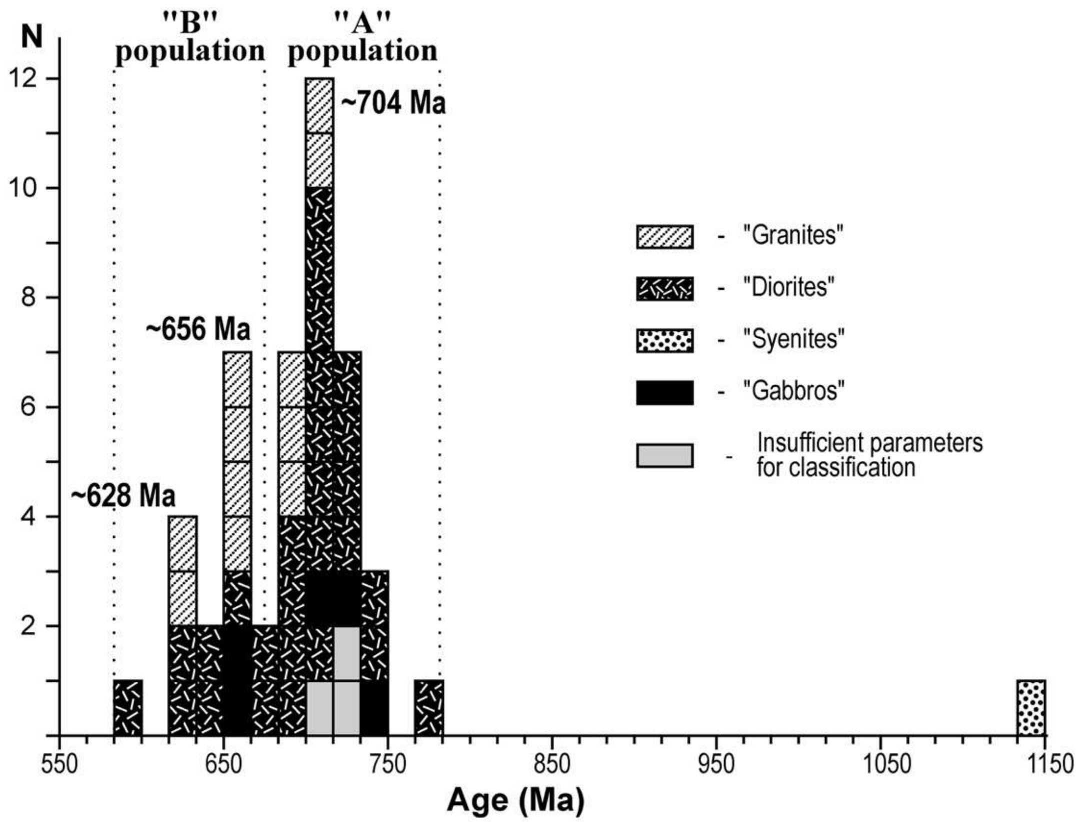

Fig. 9. Histogram of U/Pb isotopic ages of detrital zircons (47 analyses) from sandstones of the Engane-Pe Formation (sample 05-033), northern part of the Engane-Pe Uplift. Classification of parental magma types based on method described by Belousova et al. (2002).

to the older population A. The mean T_{DM}^{C} crustal model age for this population is ca. 1.28 Ga.

Well-rounded, Mesoproterozoic (1143 ± 20 Ma) grains have $\varepsilon_{Hf} = +2.3$ and plot between the CHUR and DM lines. The model age of the source regions for the magmas hosting these zircons (T_{DM}^{C}) is ~1.76 Ga, which is equal to lower limit of model ages for zircons of age population A.

4. Classification of studied zircon grains

4.1. Location 1. Southern Timan (Djejim Formation, samples 05-301 and 05-301A)

Zircons from the Djejim Formation fall into six groups (Figs. 8 and 10). Grains of the youngest (Mesoproterozoic — 1.18 to 1.35 Ga) group (D1 in Figs. 8 and 10) are derived from "gabbro" and "diorite" (3 grains) and "syenite" (1 grain) from magmas with juvenile sources (Fig. 10). The second group (1.53–1.65 Ga) (D2 in Figs. 8 and 10) was derived from "dioritic" to "granitic" source rocks (6 zircons) with ε_{Hf} values ranging from −3 to +5 and model $T_{DM}{}^{C}$ ages of 1.82–2.12 Ga.

The third group (1.70–1.83 Ga) (D3 in Figs. 8 and 10) includes zircons derived from "gabbroic" (2 zircons), "dioritic" (1 zircon) and "granitic" (4 zircons) sources. Although the composition of the parental rocks of the zircons in this group varies, their Lu/Hf isotopic systematics are similar: values of ε_{Hf} range from 0 to +5 and the model ages for magmatic source regions is ca. 2.12 Ga.

The largest group (1.88–2.15 Ga) (D4 in Figs. 8 and 10) includes 27 zircons derived from magmatic rocks with a wide range of compositions and isotopic values. For example, two zircons from "dioritic" rocks have ε_{Hf} values ranging from −12 to −15, indicating that the magmas they crystallized from were derived from isotopically old (model age 3.08–3.25 Ga) crustal material. In contrast, most of the zircons derived from "granitic" rocks contain only small amounts of radiogenic material and their source regions are essentially juvenile (Fig. 10). About half of the zircons from this group have isotopic parameters close to CHUR and model ages of ca. 2.6–2.8 Ga.

Group 5 includes 2 zircons with crystallization ages of ~2.35 and ~2.40 (strongly discondant) Ga derived from "granitic"sources, and 1 unclassified zircon with a crystallization age of ~2.5 Ga. Both "granitic" zircons define model ages almost 1 Ga older than the age of zircons themselves. Radiogenic sources for magmas are supported by their low values of ε_{Hf} (−8).

Group 6 comprises Archean zircons with ages of 2.53 to 2.80 Ga and 1 zircon with an age of 2.972 ± 0.064 Ga. Four zircons in this group are derived from "dioritic" sources, 2 from "syenitic" sources, and one from a "granitic" source. Zircons derived from "dioritic" sources have lower amounts of radiogenic material, and the "granitic" zircon can be classified as having a juvenile source (Figs. 8 and 10).

4.2. Location 2. Engane-Pe Uplift (Engane-Pe Formation, sample 05-033)

Except for a single zircon with a Mesoproterozoic age of 1143 ± 20 Ma derived from "syenitic" sources, the younger zircons in this sample belong to two populations, A and B, as previously discussed (Fig. 9 and 11).

Zircons of population A (760–675 Ma) are represented by weakly rounded crystals or fragments of rounded grains. Most (>65%) were derived from "dioritic" and lesser "granitic" compositions and "mafic" rocks. Population A zircons show varying ε_{Hf} values**,** and $T^{C}{}_{DM}$ ages that range from ~1.76 to 0.84 Ga. Four zircons (1 "granitic" and 3 "dioritic") form a sub-cluster containing lower quantities of radiogenic material and their model ages (~0.84 Ga) are just a little older than the age of zircons themselves (~0.70 Ga).

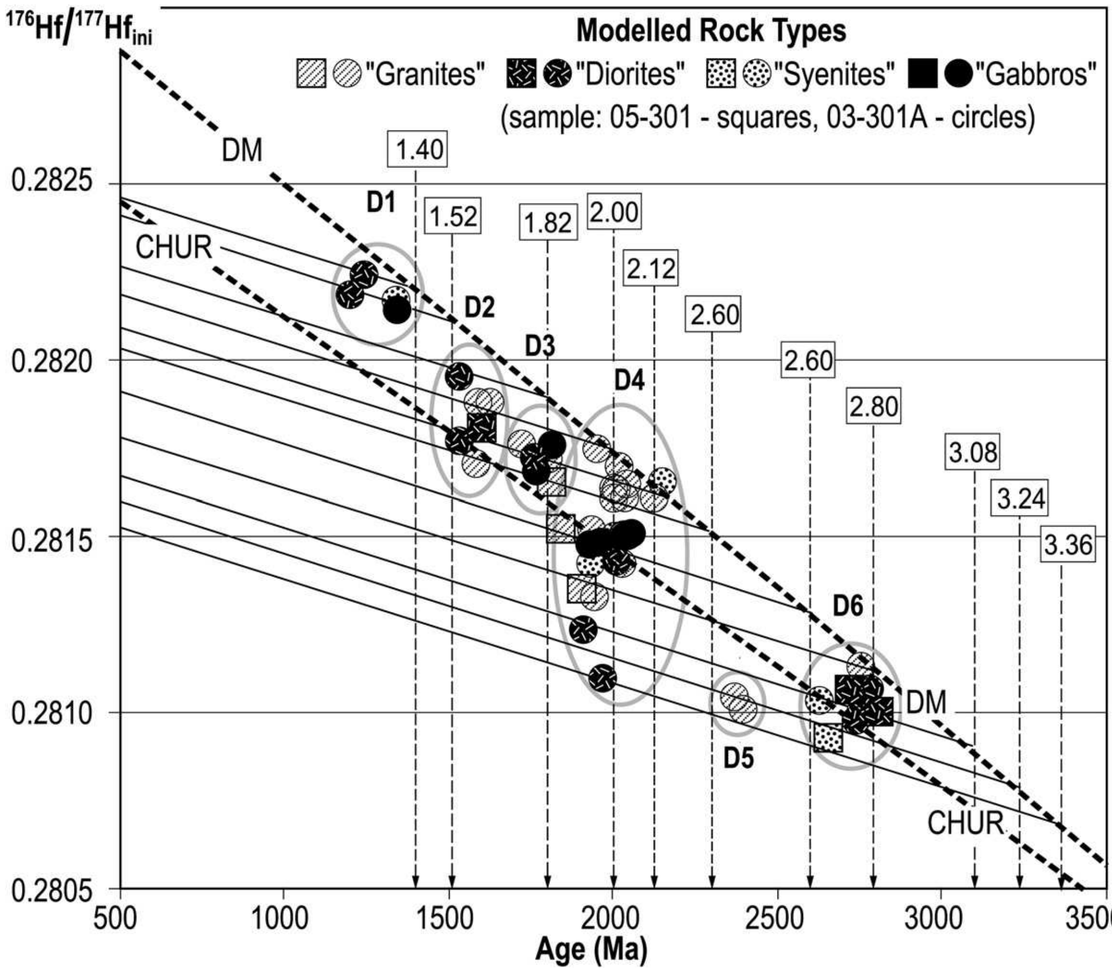

Fig. 10. Model ages of parental magmas for zircons from sandstones of the Djejim Formation, Djejim-Parma Hills, southern Timan Ridge (samples 05-301 and 05-301A). Grey ellipses mark zircon groups D1–D6. See text for further explanation.

Zircons in population B (670–590 Ma) are euhedral prismatic crystals or fragments thereof. About half of the zircons of this population were derived from "dioritic" source rocks and most of the rest come from "granitic" sources. Two zircons were derived from grabbroic source rocks. Population B zircons show similar isotopic characteristics that suggest a moderate input of recycled crust into source regions of the parental magmas, and give T^{C}_{DM} ages of ~1.28 Ga**.**

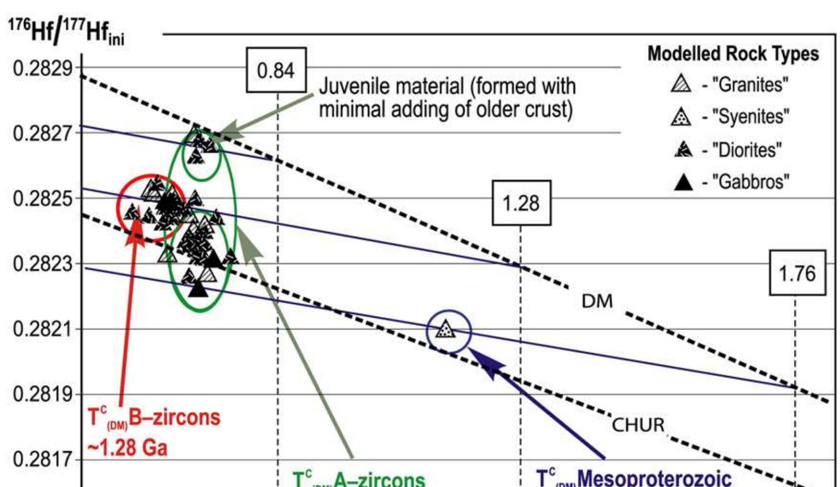

Fig. 11. Model ages of parental magmas for zircons from sandstones of the Engane-Pe Formation (sample 05-033), northern part of the Engane-Pe Uplift.

5. Discussion of source regions for detrital zircons and testing tectonic models

5.1. Characteristics of detrital zircons from Djejim Fm

Many authors have proposed that the Neoproterozoic rocks of the Timan Ridge region represent deposits of the Timan passive margin (Maslov, 2004 and references therein). The U/Pb ages and Lu/Hf isotopic systems of detrital zircons from the Djejim Formation allow us to test this hypothesis. Most of the detrital zircons fall into the age interval 1.70–2.05 Ga, during which Baltica was assembled (Fig. 8). During this period, Sarmatia and Volgo-Uralia collided to form Volgo-Sarmatia, which then collided with Fennoscandia. As a result of the collisions, large volumes of "granitic" rocks were generated, metamorphic complexes were formed, and a volcanic arc was built along the active margin of Sarmatia. Detrital zircons in the Djejim Formation have clear matches with source regions that are now part of the Baltic Shield.

The ages of the youngest zircons (4 grains) do not correspond to any well-known Baltic Shield events but match the onset of the Grenville orogeny (Fig. 8). Zircons from "gabbroic" to "syenitic" sources and juvenile parental magmas in this age group could have originated from mantle-derived magmatism during the earliest stages of this orogenic event. Those derived from "granitic" to "dioritic" sources in the D2 group, with model ages 1.82–2.12 Ga, could reflect crustal melting during the Gothian accretionary event along the western edge of the Baltic Shield (Figs. 1 and 8). Zircons from "granitic" sources could also have been eroded from the Rapakivi granites, which are widely distributed in the northwestern part of Baltica (Bogdanova et al., 2008; Bingen et al., 2008 and references therein) (Fig. 8). Zircons derived from "gabbros" to "granites" in the D3 group may reflect sources related to a prominent magmatic event of this age in Baltica (including intrusion of the Korosten pluton in Sarmatia), composed of gabbroic rocks, anorthosites and Rapakivi-like granites (Bogdanova et al., 2004 and references therein). However, if both the "granitic" and "gabbric" zircons were derived from the rocks of this massif, it is difficult to explain their identical model age (~2.12 Ga).

Zircons from the 2.5–2.8 Ga D6 group were most likely derived from rocks of the eastern and northern parts of Fennoscandia (Karelian, Belomorian, Kola and Murmansk provinces). Archean rocks in these provinces are dominated by TTG associations that cover about 80% of the area and are widely represented by ages of 2.6–3.1 Ga (Hölttä et al., 2008; Slabunov, 2008 and references therein). However, ages of 2.6–2.75 Ga have been obtained for granitoids and granitic-metamorphic complexes of the Volgo-Uralia (Bogdanova et al., 2005) and are well-known in Sarmatia (Bogdanova et al., 2008 and references therein). Zircons derived from "syenitic" sources might have come from the Keivy area (Kola province of the Fennoscandia), where a distinctive suite of alkaline granitoids and gabbro-anorthosites, together with spectacular coarse-grained kyanite-, staurolite- and garnet schists, have yielded ages of 2.63–2.75 Ga (see review Hölttä et al., 2008; Slabunov, 2008).

Thus, all the studied zircons from the Djejim Formation probably have magmatic or/and metamorphic sources within Baltica, and were mostly derived from Fennascandia and, to a lesser degree, from Sarmatia and Volgo-Uralia. However, the oldest ages in Sarmatia (up to 3.7 Ga, Bogdanova, 2005 and references therein) and Fennoscandia (up to 3.5 Ga in the Siurua area, Hölttä et al., 2008 and references therein) do not appear in the detrital zircons suites of the Djejim Formation, nor are any model ages older than 3.36 Ga. This may indicate that the Djejim Formation was formed predominantly by the erosion of the northern and central parts of Baltica — mostly eastern and northern Fennoscandia and adjacent areas of Sarmatia and Volgo-Uralia.

We therefore interpret the red sandstones and silty sandstones of the Djejim Formation (SW Pre-Uralides–Timanides) as a part of a clastic sequence deposited along the passive Timanian margin of Baltica. There is no reason, based on our data, to suggest any additional sources of clastic material.

Detrital zircon ages have been recently obtained from Paleozoic metasedimentary strata (Rogis quartzite) in the Mid-Germany Crystalline Zone (Zeh, 2008; Zeh and Gerdes, 2010–this issue). Approximately 90% of these ages lie between 0.9 and 1.8 Ga with only few zircons giving older ages. Thus, the quartzite is interpreted to have been derived from SW Baltica. Meso- and Neoproterozoic complexes, which are relicts of accretionary–collisional orogens (Fig. 1), are widely distributed over this part of Baltica, and are the likely source for most of the detrital zircons. The "provenance-signal" of SW Baltica differs significantly from that of NE Baltica, in which ages of 1.9–2.15 and 2.6–2.8 Ga are common as demonstrated by the age spectrum of the Djejim Formation. Typical SW Baltica ages of 0.9–1.8 Ga are poorly represented the Djejim Formation. The limited presence of Grenville age detritus in the Djejim Formation also suggests that the Timanian sequence represented the distal passive margin of Baltica when it was a part of Rodinia.

5.2. Characteristics of detrital zircon suites from the Engane-Pe Formation

U/Pb (0.59–1.143 Ga) ages and Lu/Hf isotopic data from detrital zircons from sandstones of the Engane-Pe Formation (NE Pre-Uralides–Timanides) differ dramatically from those (1.197–2.972 Ga) of the Djejim Formation (SW Pre-Uralides–Timanides). Firstly, no Paleoproterozoic or older zircons are found in the Engane-Pe Formation and, more importantly, no Lu/Hf model ages are older than 1.8 Ga. This suggests that the rocks of the Engane-Pe uplift were not derived from Baltica, the basement of which consists of magmatic and metamorphic complexes with widespread ages of 1.7–2.1 Ga and older. In addition, zircons ages in the interval 0.60–0.75 Ga are unknown in Baltica. Thus, our results support the idea that the SW and NE Pre-Uralides–Timanides were not derived from the same source, and that the NE Pre-Uralides–Timanides likely developed far from Baltica.

Prior to this study, sedimentary rocks of the Engane-Pe Fm were interpreted to be either Cryogenian–Ediacaran or Ediacaran–Early Cambrian in age (see review and discussion in Kuznetsov et al., 2009a; Kuznetsov, 2009c). Our results of provide a maximum depositional age of 590 Ma. The age of sub-volcanic rhyolites that intrude the assemblage of large-amplitude recumbent isoclinals folds in the Bedamel and Engane-Pe formations in the southern Engane-Pe Uplift varies between 555 and 522 Ma (Shishkin et al, 2004), giving a minimum age limit for these formations of 555 Ma. Thus, the Engane-Pe Formation was deposited between 590 and 555 Ma, which corresponds to the Ediacaran.

All zircons from the Engane-Pe Formation have ages older than the late-tectonic granites of the Pre-Uralides–Timanides orogen[1] (~560–500 Ma; Gee et al., 2000; Puchkov, 2003; Pease et al., 2004; Kuznetsov et al., 2005, 2007a; Kuznetsov, 2009c). Hence, the Engane-Pe Formation cannot have formed from the erosion of this orogen.

The well-faceted detrital zircons of population B in the Engane-Pe Formation have not been subjected to significant mechanical abrasion, as would be expected if they were involved in long-distance sedimentary transport. Hence, they were presumably supplied to the basin from a proximal source. The fact that these sediments are tuffaceous indicates a volcanic origin for some of the zircons. We suggest (Kuznetsov et al., 2009a) that they were included in the sandstones as a result of contemporaneous volcanism, which produced the basalt-basaltic andesite–andesite–dacite–rhyolite volcanic complex of the Bedamel Formation that is closely associated with the Engane-Pe Formation. Based on the detrital zircon ages, the igneous source rocks of the detrital zircons in the Engane-Pe Formation were intruded or extruded mainly between 760 and 590 Ma.

[1] Some authors refer this orogenesis as Timanian (see book "In: Gee, D.G., Pease, V. (Eds.), The Neoproterozoic Timanide Orogen of Eastern Baltica. Geological Society, London, 2004, Memoirs 30).

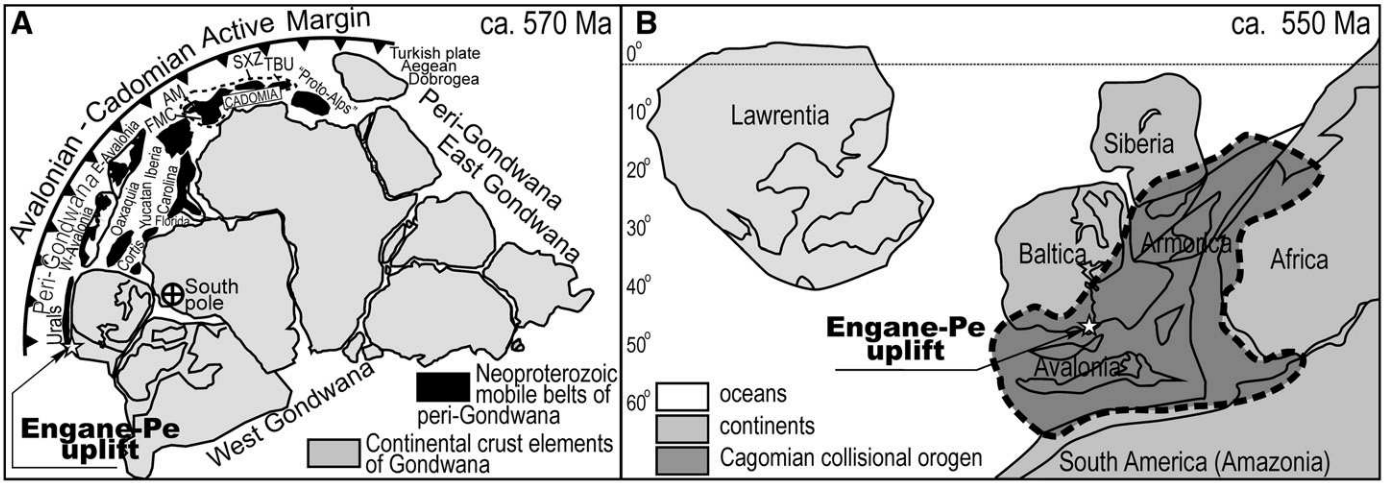

Fig. 12. Tectonic position of Pre-Uralides–Timanides as (A) a western continuation of the late Neoproterozoic Cadomian–Avalonian (Peri-Gondwanan) subduction belt (Linnemann et al., 2007), and (B) as a relict of the late Neoproterozoic Cadomian collisional orogen belt (Puchkov, 2000). AM – Armorican Massif; FMC – French Massif Central; SXZ – Saxo-Thuringian zone (part of the Bohemian Massif); TBU – Teplá-Barrandian unit (part of the Bohemian Massif).

The NE Pre-Uralides–Timanides are believed to be a relict of the western continuation of the late Neoproterozoic Cadomian–Avalonian (Peri-Gondwanan) subduction zone (Fig. 12A), which formed along a margin of Gondwana (Linnemann et al., 2007), or a relict of the late Neoproterozoic Cadomian orogen (Fig. 12B), which formed as a result of the collision of Peri-Gondwana with the Timanian–Uralian margin of Baltica (Mossakovskii et al., 1996; Puchkov 1997, 2000). It is widely accepted that parts of the structural basement of the Appalachians and the Paleozoic orogens of Western and Central Europe were located along the northern edge of Gondwana in the Neoproterozoic–earliest Cambrian, as part of the Peri-Gondwanan (Avalonian–Cadomian) belt (Murphy et al., 2006; Linnemann et al., 2007, etc.). Peri-Gondwanan terrains contain complexes formed as a result of Neoproterozoic and Early Cambrian tectonic–magmatic events (arc magmatism, accretionary deformation, etc.). The time interval of this tectonic–magmatic activity is estimated to span ~760–530 Ma (Murphy et al., 2006; Linnemann et al., 2007, etc.), which is comparable to the time interval of tectonic–magmatic activity in NE Pre-Uralides–Timanides (Fig. 2). Moreover, in both the Peri-Gondwanan terranes and parts of the NE Pre-Uralides–Timanides, a synchronous early Paleozoic unconformity is developed (Fig. 3) (Puchkov, 2000; Bogolepova and Gee, 2004; Linnemann et al., 2007; Kuznetsov, 2009c). These data form the main arguments used to correlate the Pre-Uralide–Timanides with the Peri-Gondwanan complexes (Puchkov, 1997, 2000, 2003, 2005).

In the last decade, hundreds of detrital zircons from Peri-Gondwanan Neoproterozoic clastic units have been dated in Western and Central Europe and in eastern North America. Estimates of model ages based on Sm/Nd isotopic data from Peri-Gondwanan Neoproterozoic–Early Cambrian felsic magmatic complexes are also available. This data has allowed the Peri-Gondwanan terranes to be subdivided into those of Cadomian (Cadomides) and Avalonian (Avalonides) affinity. The Cadomides include North Armorica, Ossa-Morena, Saxo-Thuringia and Moldanubia, whereas the Avalonides include Western and Eastern Avalonia, Carolinia, Moravia–Silesia, NE Iberia and probably that part of Armorica south of the North Armorica fault zone (Murphy et al., 2004, 2006; Linnemann et al., 2007, etc.). In Cadomian detrital zircon suites, ages peaks have been established at ~0.57 Ga, ~0.59 Ga, ~0.65–0.60 Ga, 0.79–0.70 Ga, 1.05–0.9 Ga, 2.2–1.8 Ga, 2.4 Ga and 2.6 Ga (Fernandez-Suarez et al., 2002; Samson et al., 2003; Gutierrez-Alonso et al., 2005; Linnemann et al., 2007). However, no zircons with ages in the interval ~1.75–1.05 Ga are present in these rocks (Fig. 13). Avalonian zircons

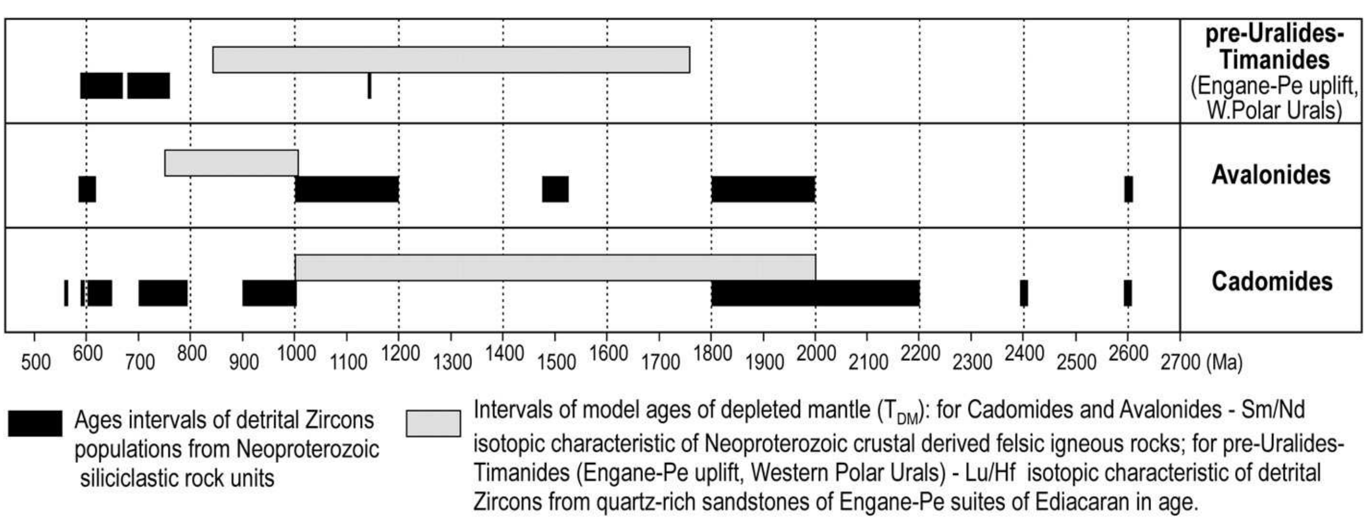

Fig. 13. Comparison of the ages of detrital zircons populations from late Neoproterozoic siliciclastic rocks from NE Pre-Uralides–Timanides (pU–T), Avalonides (A) and Cadomides (C), and intervals of depleted mantle model ages (T_{DM}) for felsic magmatic rocks from the Cadomides and Avalonides, and parental felsic magmas for detrital zircons from sandstones of the Engane-Pe Formation (Kuznetsov, 2008a). Data for Avalonides and Cadomides from Keppie et al. (1998), Fernandez-Suarez et al. (2002), Samson et al. (2003), Linnemann et al. (2004, 2007), Gutierrez-Alonso et al. (2005) and Murphy et al. (2006).

show peaks at: ~0.60 Ga, 1.2–1.0 Ga, ~1.5 Ga, 2.0–1.8 Ga and 2.6 Ga (Keppie et al., 1998; Linnemann et al., 2004; Murphy et al., 2006). Sm/Nd isotopic data from felsic rocks shows that, for the Cadomides, ε_{Nd} is low (from −9.9 to +1.6) and depleted mantle model ages (T^{Nd}_{DM}) (for $t = 610$ Ma) fall in the interval 1.0–2.0 Ga (Fig. 13) (Samson and D'Lemos, 1998; Linnemann and Romer, 2002), whereas the Avalonides are characterized by higher ε_{Nd} (from −1.0 to +5.0) and younger model ages (T^{Nd}_{DM}) of 1.1 to 0.75 Ga (Murphy et al., 2000, 2006).

T^{Nd}_{DM} model ages and U/Pb detrital zircon ages from Neoproterozoic complexes of the Peri-Gondwanan terranes (Linnemann et al., 2007; Murphy et al., 2006) differ from the $T_{DM}{}^{C}$ model ages and U/Pb detrital zircon ages in the NE Pre-Uralides–Timanides (Fig. 13). Thus our data does not support a Peri-Gondwana origin for the NE Pre-Uralides–Timanides.

Instead, we prefer a tectonic interpretation of the Pre-Uralides–Timanides as relicts of an orogen that did not have any connection to Gondwana. In our interpretation, the SW Pre-Uralides–Timanides were deposited in the Neoproterozoic along the passive Timanian–Uralian margin of Baltica, but the NE Pre-Uralides–Timanides were formed at the active Bolshezemel margin of the paleocontinent Arctida (cf. Chandra et al., 2007; Kuznetsov et al., 2007a) and were involved in the collision of Baltica and Arctida in latest Ediacaran to earliest Cambrian time (Fig. 14A,B). The suture zone between Baltica and Arctida is represented by the Pripechoro-Ilych-Chiksha fault zone in the basement of the Pechora basin (Fig. 2), which can be clearly traced with gravity and magnetic data (Kuznetsov et al., 2005, 2007a etc.). The Pre-Uralide–Timanides divergent fold-thrust collisional orogen was formed between the colliding continents of Baltica and Arctida (Kuznetsov et al., 2007a; Kuznetsov, 2009a,c). The SW Pre-Uralides–Timanides are relicts of the « Baltic » flank of this orogen, whereas the NE Pre-Uralides–Timanides are relicts of the « Arctida » flank. Space-time variations of magmatic/metamorphic activity throughout the Pre-Uralides–Timanides (Fig. 2), together with the structural features of the Pre-Uralides–Timanides (Kuznetsov, 2009a), support a collisional interpretation of the orogen (Kuznetsov et al., 2007a; Kuznetsov, 2009a,c).

There are very few data to help us understand the structural setting of the active Bolshezemel margin of Arctida, the relicts of which are now the Bolshezemel and Pechora blocks of the Pechora basin (Fig. 2), before its collision with Baltica. However, a Japan or South-Kuril subduction system model might be a present-day analog.

Based on the isotopic ages of magmatic rocks in the NE Pre-Uralides–Timanides (Fig. 2), and the detrital zircon data from the Engane-Pe Formation (Figs. 9 and 11), we propose that subducted-related magmatist along the Bolshezemel margin of Arctida lasted from ~730 Ma until the time of Baltica-Arctida collision at ~560 Ma.

Continental crust formed part of the basement of the Bolshezemel subduction complex, because granites of the Bolshezemel block commonly contain zircons with old inherited cores. The ages of zircon cores in the Pre-Uralides–Timanides granites range from ~0.9 to ~2.7 Ga (Korago and Chuknonin, 1988; Gee et al., 2000; Pease et al, 2004; Larionov and Teben'kov, 2004; Korago et al., 2004; Pease et al, 2004; Johansson et al., 2004; Udoratina et al., 2005), whereas the $T_{DM}{}^{C}$ ages for detrital zircons from the Engane-Pe Formation span the interval ~0.84–1.76 Ga.

Following Scarrow et al. (2001) and Khain et al. (2003), we further believe that the Manyukuyakha serpentinitic mélange belt (Fig. 5) is a relict of a back-arc basin positioned behind the Bolshezemel subduction complex (Kuznetsov, 2009a,b). We think that the Engane-Pe Formation was deposited on the slope of this Manyukuyakha back-arc basin close to the Bolshezemel subduction zone. This is supported by the presence of zircons interpreted to be first cycle grains derived from tephra of early Neoproterozoic age that have Mesoproterozoic model basement ages (Kuznetsov et al., 2009b).

5.3. The Arctida paleocontinent and the earliest stage of the assembly of Pangea

The vast Arctic shelves and some northern parts of North America and Eurasia are thought to be underlain by Precambrian crust. Shatsky (1935) was the first to propose that they represented fragments of an ancient Hyperborean continent in the Arctic region. Zonenshain and Natapov (1987) and Zonenshain et al. (1990) later proposed the first plate tectonic reconstruction of this ancient continent (named Arctida), which was inferred to have existed as a separate continent unrelated to the East European Craton until the Middle Paleozoic. The paleogeography of Arctida has been restored by adding Barentsia, including the NE part of the Pechora basin, the Western slope of the Sub-Polar, Polar and Northern Urals, and Svalbard (Borisova et al., 2002; Kuznetsov et al.,

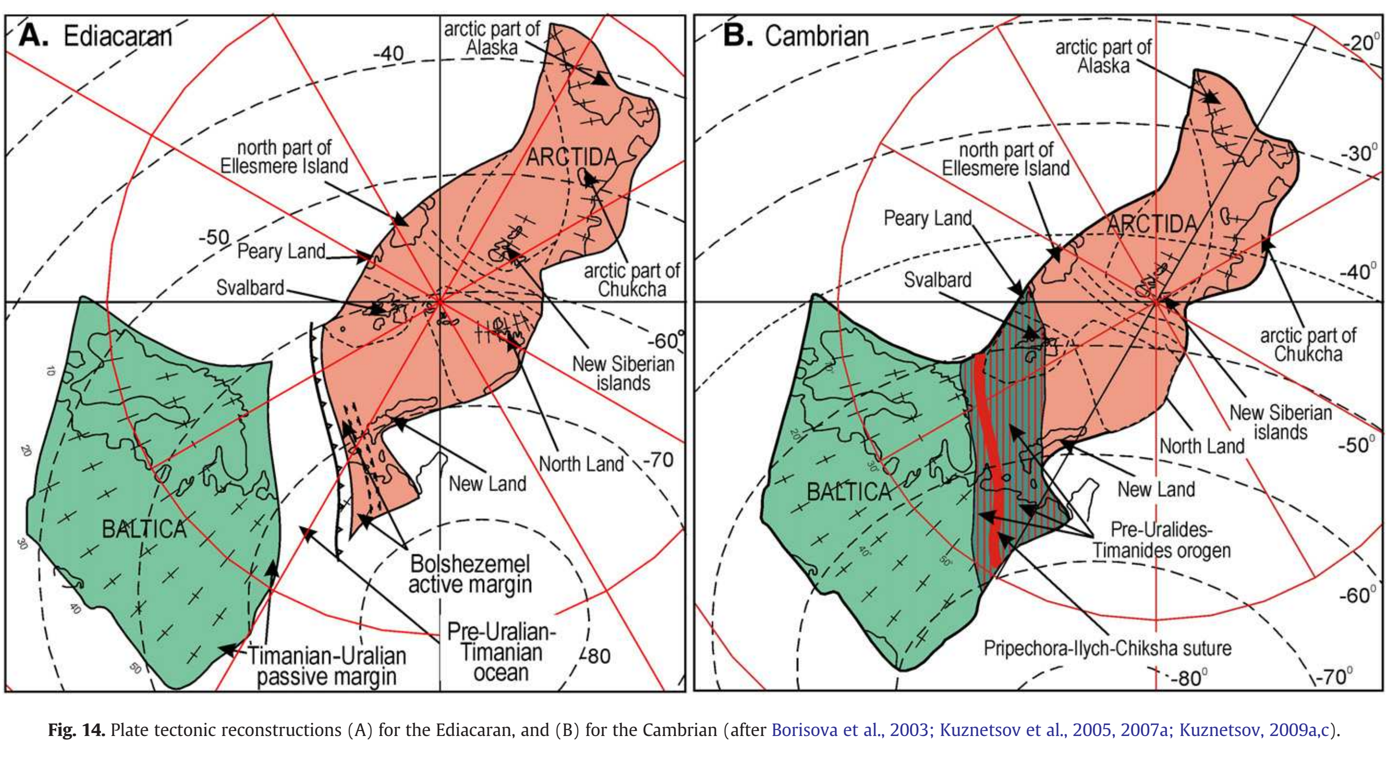

Fig. 14. Plate tectonic reconstructions (A) for the Ediacaran, and (B) for the Cambrian (after Borisova et al., 2003; Kuznetsov et al., 2005, 2007a; Kuznetsov, 2009a,c).

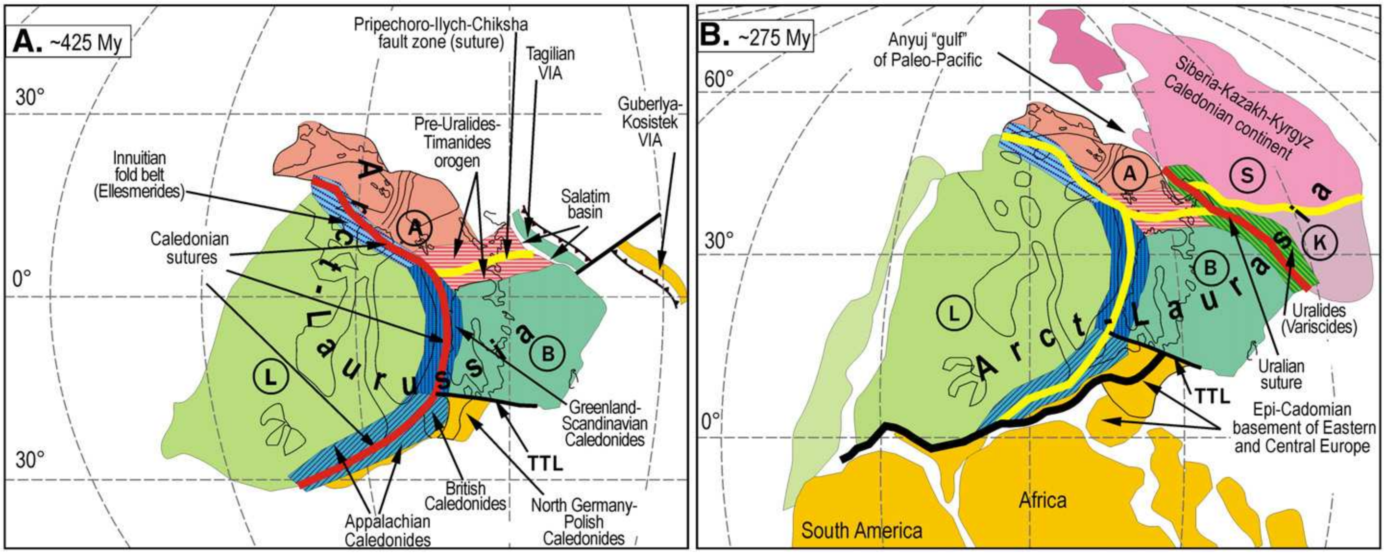

Fig. 15. Plate tectonic reconstructions of Paleozoic composite continents. (A) Epi-Caledonian continent Arct-Laurussia, and (B) Epi-Hercinian continent Arct-Laurasia (Kuznetsov, 2008b, 2009a,c). L – Laurentia, A – Arctida, B – Baltica, S – Siberia, K – Kazakh–Kyrgyz, TTL – Teiseir-Tornqwist Line. Meridians are at 30 degree intervals. Configuration and position of Laurentia, Baltica, Siberia, Kazakhstan, South America and Africa were specified using PC program « TRACKER » by K Ckotize.

2005; Kuznetsov et al., 2007a; Kuznetsov, 2009a,c). Late Neoproterozoic detrital zircon populations in Paleozoic and early Mesozoic rocks of Eastern Arctica have recently added powerful evidence for links between the Chukotka–Alaska crustal block (Northern Chucotka, Arctic Alaska and adjacent shelves) and Barentsia (Baltica) from the beginning of Paleozoic (Amato et al., 2009; Miller et al., in press). This supports our proposal for the existence of the composite continent Arct-Europe, at least from the early Paleozoic (Fig. 14). The collision of Baltica and Arctida near the Neoproterozoic–Cambrian boundary formed the Pre-Uralides–Timanide orogen, which was a source area for late Neoproterozoic detrital zircons for the entire Arctic region, including Eastern Arctica (Kuznetsov, 2006). The new detrital zircon data do not support tectonic models that consider crustal blocks of the Eastern Arctic to have been assembled into an isolated continental landmass, separate from Barencia and Baltica, in the Paleozoic. However, the existence of a vast paleocontinent Arctida does not contradict our understanding of the later stages of northern Pangea's assembly (Fig. 15).

At the Precambrian–Cambrian boundary, Arctida collided with Baltica to form the composite paleocontinent Arct-Europe (Baltica + Arctida). Between the Silurian and Devonian, this composite paleocontinent collided with Laurentia to create the Caledonian belt of Westerm Greenland and Scandinavia, and the Innuitian belt of Canada, forming the composite paleocontinent Arct-Laurussia (Arct-Europe + Laurentia). This composite continent subsequrently collided with the composite paleocontinent Siberia–Kazakh–Kyrgyz during the Uralian orogeny, at which time the composite paleocontinent Arct-Laurasia (Arct-Laurussia + Siberia–Kazakh–Kyrgyz composite) was created to assemble the northern part of Wegener's Pangea.

During these collisional events, the relicts of the northwestern edge of the Pre-Uralides–Timanides orogen were involved in the Caledonian orogeny and the relicts of the southeastern edge of the orogen were involved in the Uralian orogeny. Thus, the collision of Baltica and Arctida was the earliest continent–continent collisional event in the assembly of northern Pangea (Fig. 16). Much later, in the late Mesozoic–early Cenozoic, the northeastern parts of Arctida (as part of Pangea) were fragmented as a result of opening of the North Atlantic and Arctic oceans. As a result, fragments of Arctida underlie the northern periphery of North America and North Eurasia and their Arctic shelves, and likely constitute part of the continental Lomonosov Ridge and the possibly continental Mendeleev and

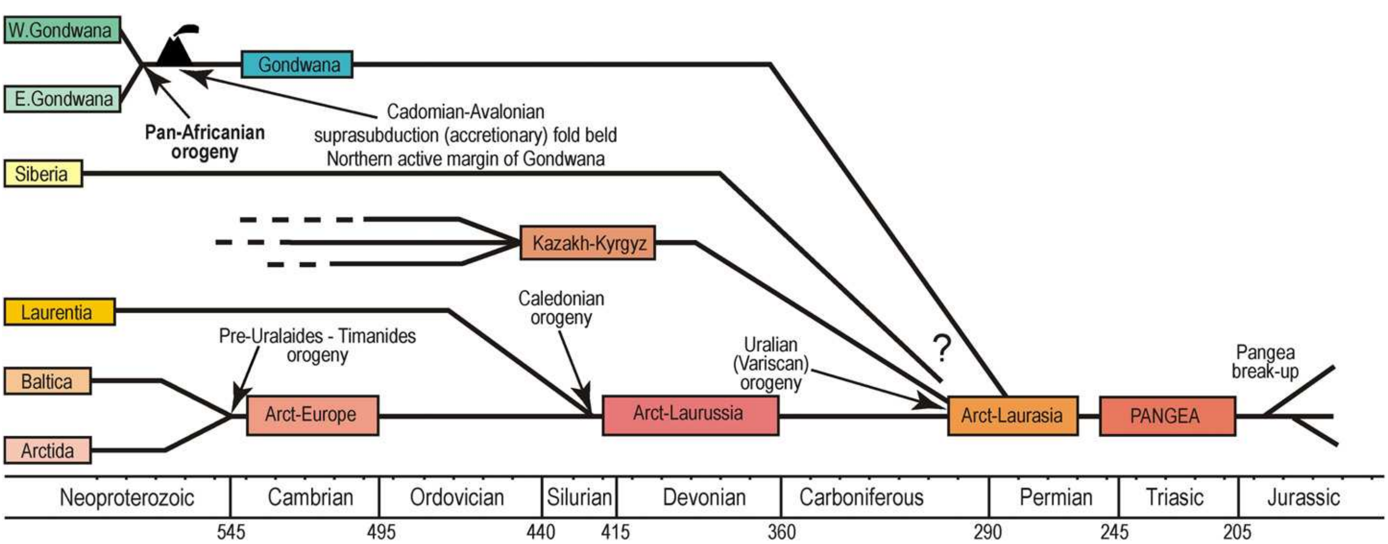

Fig. 16. Dendrogram for late Precambrian–Paleozoic assembly of northern Pangea (Kuznetsov, 2009a, c).

Alpha Ridges, which lie between the deeper oceanic basins of the Arctic Ocean.

6. Conclusions

Neoproterozoic complexes of the TPR are composed mainly of sedimentary rocks and are collectively classified as the SW Pre-Uralides–Timanides. In contrast, Neoproterozoic to Early Cambrian volcanic-sedimentary and volcanic rocks, granitoids, and rare ophiolites of the TPR are collectively classified as the NE Pre-Uralides–Timanides. Our main conclusions are listed below:

1. Comparison of the characteristics of detrital zircons from the Djejim Formation of the SW Pre-Uralides–Timanide with the well-known ages of magmatic/metamorphic events in basement rocks of Baltica shows that the clastic units the of the SW Pre-Uralides–Timanides were derived from Baltica, and were likely deposited along what was once its passive margin.
2. Comparison of the characteristics of the detrital zircons from the Djejim Formation with those the Engane-Pe Formation in the NE Pre-Uralides–Timanides reveals important differences in both the age and geochemistry of the source region. Zircons in the Engane-Pe Formation are mostly Neoproterozoic and were derived from magmas emplaced in a non-cratonic setting. These data suggest that the rocks of the NE Pre-Uralides–Timanides likely evolved far from the Baltic Shield.
3. Comparison of the U/Pb ages and the Lu/Hf isotopic systematics of detrital zircons from rocks of NE Pre-Uralides–Timanides show them to differ from those of Neoproterozoic complexes in the Peri-Gondwanan terranes. This suggests a non-Peri-Gondwanan origin for the NE Pre-Uralides–Timanides.
4. We interpret the SW Pre-Uralides–Timanides to have formed in the Neoproterozoic on the passive Timanian–Uralian margin of Baltica, while the NE Pre-Uralides–Timanides were formed at the active Bolshezemel margin of the paleocontinent Arctida and were caugth in the collision zone between Arctida and Baltica (the Pre-Uralide–Timanide orogen).

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.gr.2009.08.005.

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