Crustal nature and origin of the Russian Altai: implications for the continental evolution and growth of the Central Asian Orogenic Belt (CAOB)

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Abstract

The Central Asian Orogenic Belt is a gigantic tectonic collage of numerous accreted terranes. However, its geodynamic evolution has been hotly debated primarily due to incomplete knowledge on the nature of these enigmatic terranes. This work presents new detrital zircon U-Pb and Hf isotopic data to constrain the crustal nature and origin of the Russian Altai, a critical segment of Altai-Mongolian terrane. The youngest zircon $^{206}\text{Pb}/^{238}\text{U}$ ages of 470 Ma constrain that the Terekta Formation, previously envisaged as Precambrian basement, was actually deposited after the Middle Ordovician. As for the three more sedimentary sequences above the Terekta Formation, they have youngest zircon $^{206}\text{Pb}/^{238}\text{U}$ ages of 425 Ma, 440 Ma and 380 Ma, respectively, indicating their depositions likely in the Late Silurian to Devonian. From all analyses, it is noted that many zircon U-Pb ages cluster at ca. 520 Ma and ca. 800 Ma, and these zircons display oscillatory zoning and have subhedral to euhedral morphology, which, collectively, suggest that adjacent Neoproterozoic to Paleozoic igneous rocks were possibly dominant in the sedimentary provenance. Additionally, a few rounded Archean to Mesoproterozoic zircon grains are characterized by complex texture, which are interpreted as recycling materials probably derived from the Tuva–Mongolian microcontinent. Precambrian rocks have not been identified in the Russian Altai, Chinese Altai and Mongolian Altai so far, therefore, Precambrian basement may not exist in the Altai-Mongolian terrane, but this terrane

probably represents a large subduction-accretion complex built on the margin of the Tuva-Mongolian microcontinent in the Early Paleozoic. Multiple episodes of ridge-trench interaction may have caused inputs of mantle-derived magmas to trigger partial melting of the newly accreted crustal materials, which contributed to the accretionary complex. During accretionary orogenesis of the CAO, formation of such subduction-accretion complex is likely ubiquitous, indicating continental crust growth by both lateral accumulation and vertical basaltic injection.

Key words: Central Asian Orogenic Belt, Subduction-accretion complex, Altai-Mongolian terrane, Accretionary Orogenesis, Continent crustal growth

1. Introduction

Accretionary orogens occur in association with plate convergence, and provide valuable information on the origin and evolution of the continental crust (Cawood et al., 2009; Condie, 2007). The incorporation of numerous accretionary complexes is taken as one of the most prominent hallmarks of an accretionary orogen (e.g., Sengör et al., 1993; Sengör and Natal'in, 1996). Field investigations show that they are likely to have been subject to extensive and complex deformation, since long-lived accretionary process might have undergone multiple processes involving extension and collision, as well as long distance displacement (e.g., Sengör and Natal'in, 1996). Additionally, ridge–trench interaction, as one of the most important tectonic controls during the development of an accretionary orogen (Windley et al., 2007), may cause high-grade metamorphism (e.g., Sisson et al., 2003; Sengör and Natal'in, 1996). Consequently, these post-accretion processes might mask the original geological features of those accretionary complexes. Fortunately, these subsequent influences can hardly have impacts on the U–Pb and Lu–Hf isotopic systems of detrital zircons preserved in sedimentary rocks (Nelson, 2001; McLennan et al., 2001; Fedo et al., 2003). Therefore, a combined investigation of zircon U–Pb and Hf isotope geochemistry has been proved to be a valid approach to unravel crustal nature and origin of an altered accretionary complex.

The Central Asian Orogenic Belt (CAOB), being a typical of accretionary

orogen, provides a spectacular example of continental growth in an immense and fairly well-exposed region (Fig. 1). The growth of the CAOBS is attributed to laterally progressive assembly of exotic and/or juvenile geological units, including magmatic arcs, ophiolites, accretionary complexes, seamounts, as well as microcontinents (e.g., Sengör et al., 1993; Sengör and Natal'in, 1996; Badrach et al., 2002; Khain et al., 2002; Xiao et al., 2004, 2008, 2010, 2013; Windley et al., 2007; Kröner et al., 2011; Wilhem et al., 2012). Notwithstanding, mainly owing to the paucity of available geochemical and isotope geochronological data, the nature and origin of these diverse tectonic units have not been well understood. For instance, the Altai-Mongolian terrane consists mainly of metasedimentary strata and volcanic rocks of varied grade, intruded by extensive intrusions. Some rocks are highly deformed and underwent high-grade metamorphism, therefore the Altai-Mongolian terrane was previously considered as a Precambrian microcontinent (e.g. Dobretsov et al., 2003; Buslov et al., 2001; Dobretsov and Buslov, 2007; He et al., 1990; Hu et al., 2000; Li et al., 2006). However, this interpretation has been challenged by the detrital zircon U-Pb dating results for the Chinese Altai, because the high-grade rocks there actually have Early Paleozoic ages. The high temperature metamorphism was interpreted as a result of granitoid doming in the Late Paleozoic (e.g. Wei et al., 2007; Zhuang, 1994), or due to ridge subduction in the Devonian (e.g. Sun et al., 2008; Long et al., 2007; Jiang et al., 2010, 2011). In order to further explore the crustal nature and origin of the

Altai-Mongolian terrane, we performed a combined investigation of detrital zircon U-Pb and Hf isotopic analysis for the Russian Altai. Our new data place constraints on the depositional age and provenance of these crucial strata, providing useful information on the nature and origin of the crust of the region. This study also shed light on the tectonic evolution and continental growth of the CAOBS.

2. Geological background

The CAOBS is one of the largest and most complex tectonic collages and represents a typical region of Phanerozoic continental growth (Şengör et al., 1993; Jahn, 2004; Jahn et al., 2000a, 2000b; Windley et al., 2007). It extends from the Ural Mountain in the west to the Pacific coast in the east, and is bounded by the Siberian Craton to the north and the Tarim-North China Cratons to the south (Fig.1; Zonenshain et al., 1990; Mossakosky et al., 1993; Sengör et al., 1993; Sengör and Natal'in, 1996; Jahn et al., 2000a, 2000b; Windley et al., 2002, 2007). The CAOBS architecture includes a variety of tectonic units that are now delimited by large, mostly inactive strike-slip faults with very substantial and commonly poly-phase and poly-sense displacements (Sengör and Natal'in, 1996; Buslov et al., 2013; Windley et al., 2007; Wilhem et al., 2012). The Charysh–Terekta–Ulagan–Sayan suture-shear zone (CTUSSs) is one of these inactive strike-slip faults, and it separates the Altai-Mongolian terrane from the Gorny Altai to the north and from West-Sayan to the east,

139 respectively (Dobretsov and Buslov, 2007; Buslov, 2011).

140 The Altai-Mongolian terrane displays NW-SE striking and stretches for
141 more than 1000 km (Fig. 1). Rocks of this terrane consist mainly of extensive
142 marine sedimentary successions and massive granitoid intrusions. The
143 assumed oldest sedimentary sequence is made up of gneisses and schists,
144 which are metamorphic products of polymictic sandstone, silicolite, phyllite and
145 shale. In term of their strong deformed and metamorphosed features, these
146 high-grade rocks were presumed to be Precambrian. The so-called
147 Precambrian rocks are overlain by flysh-like sequences primarily consisting of
148 sandstone, siltstone and conglomerate. These rocks have spectacular
149 outcrops and are collectively named as Habahe Group in the Chinese Altai
150 (BGMRX, 1993; Long et al., 2007), which is overlain by Late Paleozoic
151 volcanoclastic rocks (Windley et al., 2002; Badarch et al., 2002; Glorie et al.,
152 2011). Igneous rocks also have extensive outcrops in the Altai-Mongolian
153 terrane, and show various lithology, chemical composition and mineralization
154 (Yuan et al., 2007; Wang et al., 2006, 2009; Kruk et al., 2011; Glorie et al.,
155 2011; Cai et al., 2011a, 2011b, 2012a, 2012b, 2014). A few NW-SE thrust
156 faults are subparallel to the Erqis fault that is considered as a boundary
157 between the Altai-Mongolian terrane and Junggar blocks (Fig. 1).

158 The Gorny Altai occurs as a triangle-shaped tectonic unit adjacent to the
159 Altai-Mongolian terrane. Its exposed rocks mainly include various
160 Neoproterozoic to Early Cambrian volcanic rocks, accretionary prisms

(including paleo-seamount), ophiolites and high-pressure metamorphic rocks (Volkova et al., 2005; Volkova and Sklyarov, 2007; Ota et al., 2007; Glorie et al., 2011). The Late Ediacaran - Early Ordovician Kuznetsk-Altai island arc have the oldest igneous rocks formed at ca. 555–525 Ma (Dobretsov and Buslov, 2007; Glorie et al., 2011). The distinctive rock associations constitute a typical intra-oceanic subduction-accretion system, representing a major site of the continental growth.

The West Sayan contains Vendian to Middle Cambrian ophiolites, back-arc and island arc rocks. The back-arc rocks consist mainly of Vendian to Middle Cambrian limestones, dolomites, clastic, volcanoclastic and volcanic rocks. There was a prolonged magmatic activity from the Early Cambrian (530–520 Ma) to Early Devonian (425–400 Ma), which may be related to the complex amalgamation of numerous terranes (Vladimirov et al., 2001; Rudnev et al., 2004, 2008; Glorie et al., 2011). A variety of granites were emplaced in the Middle-Late Devonian (390–360 Ma) and Carboniferous (340–310 Ma) (Vladimirov et al., 2001; Glorie et al., 2011, 2014).

3. Sample description

Our samples were collected from the Russian Altai (Table 1), a transitional region including a portion of the Altai-Mongolian terrane and the Gornyy Altai (Fig. 2), i.e., along the Charysh-Terekta-Ulagan-Sayan suture-shear zone (CTUSs). The following description is for samples collected bottom-up

183 according to the stratigraphic sequence (Fig. 3).

184 Sample 10HK50 is a schist from the Terekta Formation (Fig. 3). The
185 sample is composed mainly of quartz (40–60 modal%), plagioclase (20–30
186 modal%), biotite (10–25 modal%) and minor chlorite (<10 modal%). The rock
187 has strong foliation, shown by oriented biotite and chlorite.

188 Sample 10HK64 is a paragneiss from the Keam Formation (Fig. 3). It
189 consists of medium- to fine-grained quartz (45–55 modal%), plagioclase
190 (15–20 modal%), biotite (10–15 modal%), garnet (5-10 modal%) and
191 sillimanite (5-10 modal%), with minor opaque minerals. Garnet minerals are
192 subhedral and pink, sillimanite is acicular in shape and biotite occurs as dark
193 flakes that define the gneissosity.

194 Sample 10HK73 is a sandstone from the Gorny Altai Formation (Fig. 3).
195 The rock sample is gray and has medium-grained texture, consisting of
196 detritus (quartz and minor plagioclase) (50-60 modal%), lithic fragments (20-30
197 modal%) and cement (10-20 modal%). Quartz grains are polycrystalline and
198 align along certain orientation.

199 Three samples (10HK90, 10HK91 and 10HK92) were collected from the
200 Bar-burgaza Formation (Fig. 3). Sample 10HK90 is a schist from the lowest
201 portion of the formation, and it consists mainly of quartz (55–65 modal%),
202 plagioclase (15–20 modal%), biotite (10–20 modal%) and chlorite (10–20
203 modal%). Sample 10HK91 is a sandstone from the middle portion of the
204 formation, and it is gray and medium-grained, including 65–75 modal% quartz,

15–20 modal% feldspar and 5-10 modal% cement. Sample 10HK92 is a green conglomerate from the upmost of the formation, and comprises detrital minerals of quartz and plagioclase (65-75 modal%), lithic fragments (15-20 modal%) and cement (10-15 modal%).

4. Analytical methods

4.1 Zircon separation and CL imaging

After sample crushing, zircons were separated by standard heavy liquid and magnetic techniques. Crystal grains from the >25 μm non-magnetic fractions were hand-picked and mounted on adhesive tape, then enclosed in epoxy resin and polished to about half of their thickness. Cathodoluminescence (CL) images were taken using a JXA-8100 Electron Probe Micro-analyzer with a Mono CL3 Cathodoluminescence System for high resolution imaging and spectroscopy, at Guangzhou Institute of Geochemistry, Chinese Academy of Sciences, to investigate internal structures and to choose potential target sites for U-Pb and Hf analyses.

4.2 Zircon U-Pb dating

Zircon U-Pb isotopic analyses were performed at the Department of Earth Sciences, the University of Hong Kong. A 193 nm excimer laser ablation system (Resolution M-50) was used in connection with a Nu Plasma HR MC-ICP-MS. Helium gas, which carries the laser ablated sample aerosol from sample cell, was mixed with argon carrier gas and nitrogen as additional

di-atomic gas to enhance sensitivity and finally flows into the MC-ICP-MS torch for analysis. The detailed analytical procedure was described by Xia et al. (2011). The majority of the analyses were conducted with a beam diameter of 30 μm , 5 Hz repetition rate and energy of $\sim 5 \text{ J/cm}^2$ per pulse. 91500 and GJ-1 were used as the external standards for calibration. The mass fractionation correction and isotopic results were calculated by ICPMSDataCal (version 7.0, Liu et al., 2008). A common Pb correction was applied to all measured ratios using the interference and background-corrected ^{204}Pb signal intensity following the approach of Andersen (2002). The age calculation and concordia plots were processed using ISOPLOT 3.0 (Ludwig, 2003). Individual analyses are presented with 1σ error in Table 2 and in concordia diagrams. Uncertainties of mean age calculations are quoted at the 95% level. $^{206}\text{Pb}/^{238}\text{U}$ ages are used for grains $< 1.0 \text{ Ga}$, and $^{207}\text{Pb}/^{206}\text{Pb}$ ages are used for grains $> 1.0 \text{ Ga}$ according to Black et al. (2003).

4.2 Zircon Lu-Hf isotopic analyses

Zircon Hf isotope analyses was conducted using the laser ablation system Resolution M-50, attached to a Nu Plasma HR MC-ICPMS, at the Department of Earth Sciences, the University of Hong Kong. A spot size of 40 μm was used for most analyses and the ablation spot for Hf isotope analysis was sited at the same domain for the U-Pb dating. In the case of isotopic zoning or with intersecting cracks/inclusions, only the flattest and most stable portions of the

time-resolved signal were selected for integration. The standard zircon 91500 was used as a reference, with a weighted mean $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282298 ± 6 (2σ , $n = 50$), similar to the recommended ratio of 0.282306 ± 10 (Woodhead et al., 2004). Detailed instrumental settings and analytical procedures were described by Xia et al. (2011). All Hf isotope data were calculated using the decay constant of $1.865 \times 10^{-11} \text{ yr}^{-1}$ (Schärer et al., 2001). The chondritic values of $^{176}\text{Hf}/^{177}\text{Hf}=0.282772$ and $^{176}\text{Lu}/^{177}\text{Hf}=0.0332$ (Blichert-Toft et al., 1997) were used for the calculation of $\epsilon_{\text{Hf}}(t)$ values. Two types of Hf model ages, T_{DM} and T_{NC} , were calculated from the measured Lu-Hf isotope values and U-Pb ages of the dated zircon grains. The T_{DM} values were calculated using the measured $^{176}\text{Lu}/^{177}\text{Hf}$ ratios based on the assumption that the depleted mantle reservoir has a linear isotopic growth from $^{176}\text{Hf}/^{177}\text{Hf} = 0.279718$ at 4.55 Ga to 0.28325 at present, with $^{176}\text{Lu}/^{177}\text{Hf}=0.0384$ (Griffin et al., 2000). The T_{NC} values were calculated using the measured $^{176}\text{Lu}/^{177}\text{Hf}$ ratios based on the assumption that the new continental crust reservoir (island arcs) has a linear isotopic growth from $^{176}\text{Hf}/^{177}\text{Hf}=0.279703$ at 4.55 Ga to 0.283145 at present, with $^{176}\text{Lu}/^{177}\text{Hf}=0.0375$ (Dhuime et al., 2011).

5. Analytical results

Although all zircons in this study have comparable Th/U ratios (> 0.1) and obvious zoning structure, two subgroups of crystal morphology can be roughly

recognized on the basis of their distinguishable contour and internal structure (Fig. 4). The first type (Type-I) of zircons are dominant in quantity (~90%) and exhibit remarkable characters of euhedral prismatic shape and oscillatory zoning. In contrast, the other type (Type-II) displays rounded outline and most occur as inherited cores with sharp contact to the mantle. Both types have high Th/U ratios (> 0.1), and are considered to be igneous in origin (Hanchar and Rundnick, 1995; Corfu et al., 2003).

5.1. Sample of the Terekta Formation (10HK50)

Ninety-two analyses of zircons yield on or near the concordant curve (Fig. 5a). Fifty-one zircons give Early Paleozoic ages ranging from 480 to 540 Ma, with a prominent age peak at 520 Ma (Fig. 6). According to the CL images, all the Early Paleozoic zircons are typical of Type-I structure (Fig. 4). The remaining analyses give a quite wide age range from 542 to 2889 Ma (Fig. 5a), and they show two subordinate age peaks at 780 Ma and 1950 Ma (Fig. 6). Most Neoproterozoic zircons are Type-I, whereas, the old ones commonly have Type-II structure (Fig. 4). Forty analyses of the Early Paleozoic zircons yield $\epsilon_{\text{Hf}}(t)$ values of -13.5 to 15.8, and thirty-two Precambrian zircons give $\epsilon_{\text{Hf}}(t)$ values ranging from -14.0 to 18.2 (Table 3).

5.2. Sample of the Keam Formation (10HK64)

Forty-eight zircons were analyzed for this sample, and thirty-four analyses yield Early Paleozoic ages from 425 to 539 Ma (Fig. 5b), giving an age peak at 520 Ma (Fig. 6). The rest of analyses show scattered Precambrian ages (558

to 2237 Ma) and form a minor age peak at 800 Ma (Fig. 6). The Early Paleozoic zircons are characterized by an apparent dominance of Type-I structure and give $\epsilon_{\text{Hf}}(t)$ values ranging from -18.6 to 10.6, whereas, those Precambrian ones display relatively rounded shapes of Type-II and yield $\epsilon_{\text{Hf}}(t)$ values of -20.6 to 7.5 (Table 3).

5.3. Sample of the Gorny Altai Formation (10HK73)

Forty-six zircon spots were analyzed for U-Pb isotopic compositions (Table 2), and they all plot on or near the concordant curve (Fig. 5c). Twenty-eight zircons give a $^{206}\text{Pb}/^{238}\text{U}$ age cluster of 455 to 539 Ma, with a remarkable age peak at 520 Ma (Fig. 6). Other analyses yield Precambrian ages from 549 to 1940 Ma and most of them cluster between 700 and 900 Ma (Fig. 6). The Early Paleozoic zircons have Type-I internal structure and give $\epsilon_{\text{Hf}}(t)$ values ranging from -11.4 to 12.5 (Table 3). The Precambrian zircons yield comparable $\epsilon_{\text{Hf}}(t)$ values from -13.5 to 13.0 (Table 3).

5.4. Samples of the Bar-burgaza Formation (10HK90, 10HK91, 10HK92)

The three samples have comparable age spectra, and the predominant populations of the zircons with Type-I structure yield Early Paleozoic ages (460-540 Ma, 455-541 Ma, 380-541 Ma, for samples 10HK90, 10HK91 and 10HK92, respectively), sharing an age peak at 520 Ma. The remaining zircons have Type-I or Type-II structures and yield a wide age ranging from the Neoproterozoic to Archean, among which most have Neoproterozoic ages (680-950 Ma) whereas the rest eighteen analyses display pre-Neoproterozoic

ages (1.2-3.3 Ga) (Table 2). The Early Paleozoic zircons of the three samples give comparable $\epsilon_{\text{Hf}}(t)$ values from -16.3 to 16.3, and those Precambrian ones yield a similar range of $\epsilon_{\text{Hf}}(t)$ values (-14.5 to 18.0) (Table 3).

6. Discussion

6.1 Constraints on depositional timing

The depositional timings of the investigated sedimentary sequences in the Russian Altai have long been unclear. Mainly according to the association of fossils and lithology, these rocks are roughly assigned to Neoproterozoic to Early Paleozoic (Dergunov et al., 1980; Daukeev et al., 2008), and some high-grade rocks without fossil were assigned separately to Precambrian, such as the Terekta Formation in the Geological Map of the Gorny Altai (1:200,000).

The Terekta Formation gave the youngest detrital zircon age of 459 ± 5 Ma (Fig. 5a), indicating that the sequence was deposited after Middle Ordovician, rather than Precambrian as previously assumed. The youngest age of the Terekta Formation resembles that of the Early Ordovician sequence in the south lobe of the Altai Mongolian terrane (Chen et al., 2015a). The youngest detrital zircons from the Kaem Formation are 425 ± 10 Ma old, indicating its deposition after the Late Silurian. The youngest detrital zircon age of the Gorny Altai Formation is 455 ± 5 Ma, in conjunction with field relationship with the underlying Kaem Formation, suggesting the depositional age of the Gorny Altai

Formation is also younger than the Late Silurian. Accordingly, the depositional ages of the two formations are significantly younger than the assumed Ordovician (Daukeev et al., 2008) or Ordovician to Silurian. Rocks of the Bar-burgaza formation have wide distributions across the CTUSs (Fig. 2). The youngest detrital zircon U-Pb ages for the three samples from this formation are 460 ± 6 Ma, 455 ± 2 Ma and 385 ± 2 , respectively. Combined with the finding of diverse Devonian fossils (Daukeev et al., 2008), our data constrain that this formation was deposited in the Middle or Late Devonian.

Our zircon U-Pb data clearly constrain that the sedimentary sequences in the study area all deposited after the Middle Ordovician. Paleontological evidence of the Bar-burgaza formation (Daukeev et al., 2008) and stratigraphic relationship constrain that they all deposited prior to Middle or Late Devonian. This is consistent with the widespread intrusion of Late Paleozoic igneous rocks in the studied area (Glorie et al., 2011), and further supported by the absence of Late Paleozoic detrital zircons in our samples.

6.2 Sedimentary provenances

Although there are some minor differences in the age spectra of the investigated sedimentary sequences, all samples share two most prominent age peaks at ca. 520 Ma and ca. 800 Ma (Fig. 6). This remarkable signature suggests that the Early Paleozoic sedimentary setting could have been relatively stable, and common sources may have been persistent and served

as sustaining provenances for these sediments in a period between the Ordovician to Devonian. Additionally, these Neoproterozoic to Paleozoic zircons were likely derived from proximal sources on the basis of their subhedral to euhedral morphology (Fig. 5). A small quantity of zircons has Archean to Mesoproterozoic ages, and most of them are characterized by sub-rounded to rounded shapes (Fig. 5), probably implying a long distance transportation.

In the vicinity of the study area, there are voluminous rocks with the age of ~ 520 Ma as a result of intense Cambrian to Ordovician igneous activities, including those in the Tuva–Mongolian terrane, west Mongolia, west Sayan, and Gorny Altai. A few granitoids with ages of 536–465 Ma were emplaced in the margin of Tuva–Mongolian terrane (Sal’nikova et al., 2001). To the west of the Tuva–Mongolian terrane, minor ophiolitic fragments of the west Mongolia have zircon U-Pb ages of ca. 573–560 Ma, numerous I-type granitoids were emplaced during the period of 515–480 Ma (Jian et al., 2014; Yarmolyuk et al., 2011), and alkaline intrusive rocks have emplacement ages of 470–440 Ma (Yarmolyuk et al., 2011). In the west Sayan, rocks associated with island-arc magmatism were produced in the Early Cambrian (530–520 Ma) (Rudnev et al., 2009). Additionally, the Early Paleozoic (500–400 Ma) granitoid rocks are widespread in the entire region (Vladimirov et al., 2001, Rudnev et al., 2004, 2009; Glorie et al., 2011). In the Gorny Altai, the oldest island-arc igneous rocks formed at ca. 555–510 Ma (Dobretsov and Buslov, 2007; Glorie et al.,

2011; De Grave et al., 2011). Considering large strike-slip faulting and attendant unit repetition of the CAO, we suggest that the above igneous rocks may constitute the leading provenance to supply Early Paleozoic materials to the sedimentation of the investigated sedimentary sequences.

Igneous rocks with ages of ca. 800 Ma mainly crop out in the Tuva–Mongolian microcontinents (Badarch et al., 2002; Demoux et al., 2009; Levashova et al., 2010, 2011; Wang et al., 2001; Yarmolyuk et al., 2005) and in the collisional–accretional belts along the western and southern margins of the Siberian craton (Vernikovsky et al., 2003; Yarmolyuk et al., 2006). In addition, more coeval igneous rocks were identified in the Northeast China and Tarim North margin (Lu et al., 2008; Long et al., 2011; Zhang et al., 2009, 2011; Zhu et al., 2011; Wu et al., 2011). However, given that the Altai Orogen was likely separated from the Tarim block by a wide Paleasian ocean during the Early Paleozoic (Sengör et al., 1993; Sengör and Natal'in, 1996; Windley et al., 2007; Xiao et al., 2004; 2014; Levashova et al., 2010, 2011), we propose that the ca. 800 Ma zircons probably came from the Tuva–Mongolian microcontinent or its fragments.

A few of pre-Neoproterozoic (> 1.0 Ga) zircons are detected in our study and they have significant implications for crustal evolution of the CAO. Ancient rocks are widespread in the surrounding Precambrian basements of Tarim, North China and Siberian cratons (Lu et al., 2008; Rosen et al., 2006; Sal'nikova et al., 2001; Turkina et al., 2007; Zhai and Liu, 2003; Zhai et al.,

2005, Zhao et al., 2000, 2005, 2009; Long et al., 2011). It is worthy to note that ancient rocks also exist in the Tuva–Mongolian microcontinent and its fragments, for instance, a tonalite–trondhjemite gneiss was dated at 2833 ± 35 Ma (Kozakov et al., 2001), and other old basement rocks include the Paleoproterozoic granites with zircon U–Pb and Pb–Pb evaporation ages of ca. 2.36–1.82 Ga (Badarch et al., 2002; Kotov et al., 1995, Kozakov, 1993, 2001; Kröner et al., 2001; Demoux et al., 2009). The old zircons with complex internal structure in our study can be attributed to multi-phase recycling and long distance transportation. We tend to ascribe the recycling zircon grains to a derivation from the Tuva–Mongolian and associated microcontinent fragments in the vicinity.

6.3 Crustal nature and origin

Although the CAOBS is believed to be a gigantic tectonic collage by complicated amalgamations of magmatic arcs, ophiolites, accretionary complexes, seamounts and microcontinents (e.g., Sengör et al., 1993; Sengör and Natal'in, 1996; Badarch et al., 2002; Khain et al., 2002; Xiao et al., 2004, 2008, 2010, 2013, 2014; Windley et al., 2007; Safonova et al., 2010; Kröner et al., 2011, 2014; Wilhem et al., 2012), its geodynamic scenario has been much debated (for review see Wilhem et al., 2012). The controversial views are primarily due to incomplete knowledge of the numerous elusive terranes within the CAOBS. For instance, the Altai-Mongolian terrane has been considered as a

Gondwana-derived microcontinent with Precambrian basement (He et al., 1990; Zonenshain et al., 1990; Mossakovsky et al., 1993; Dobretsov et al., 2003; Buslov et al., 2004a, 2004b; Li et al., 2006; Dobretsov and Buslov, 2007; Glorie et al., 2011), or alternatively as a huge subduction-accretion complex like Japan island or Andean-type active margin (Şengör et al., 1993; Şengör and Natal'in, 1996; Xiao et al., 2004, Long et al., 2007, 2010; Sun et al., 2008; Cai et al., 2011a, 2011c). The hypothesis of microcontinent rifted from Gondwana in the Neoproterozoic is mainly based on geologic, lithologic, stratigraphic and paleomagnetic and few Sm-Nd isotopic data (Hu et al., 2000; Buslov et al., 2001; Li et al., 2006; Wang et al., 2009; Glorie et al., 2011). Virtually, rocks of the Altai-Mongolian terrane have undergone complex deformation and metamorphism (Buslov et al., 2001; Wei et al., 2007; Long et al., 2007; Sun et al., 2008; Glorie et al., 2011; Jiang et al., 2012), and the later geological events may easily obscure the original properties of the terrane (e.g. Sisson et al., 2003; Sengör and Natal'in, 1996). Our new dating results firmly document that the investigated sedimentary sequences in the Russian Altai were actually deposited during the Ordovician to Devonian, rather than in Precambrian as previously assumed (Dergunov et al., 1980; Daukeev et al., 2008).

Moreover, there are several lines of evidence to support the Altai-Mongolian terrane as a subduction-accretion complex formed in the Early Paleozoic. Firstly, our provenance analysis indicates that the majority of

447 detrital zircons with an age peak of 520 Ma were likely derived from the Early
448 Paleozoic igneous rocks possibly associated with subduction processes in the
449 surrounding regions, and the insignificant deviations between the
450 crystallization ages of most zircons and the depositional timing of the host
451 sedimentary rocks may imply a rapid and proximal accumulation in a
452 convergent tectonic setting (Cawood et al., 2012). Secondly, ~520 Ma detrital
453 zircon grains mostly possess positive $\epsilon_{\text{Hf}}(t)$ values (Fig. 6), indicating
454 derivations from juvenile crustal sources formed by subduction-related
455 magmatic events. Thirdly, geochemical studies suggest that the investigated
456 rocks of the Early Paleozoic sedimentary sequences are characterized by high
457 ICV values (1.18–2.53) and relatively low CIA values (37.9–76.3), suggesting
458 that the sediments have low maturity and underwent poorly chemical
459 weathering, similar to those of sedimentary rocks in the subduction-related
460 environment (Chen et al., 2014). Fourthly, in the Chinese Altai of the
461 Altai-Mongolian terrane, numerous NWW-SEE strike-extending thrusts have
462 moderate or sub-vertical dip angles and have commonly transported gneisses,
463 schists, and migmatites, volcanic rocks and other metasediments, which could
464 have been caused by large-scale southward-directed thrusting, orogen-parallel,
465 strike slip movements during the prolonged accretionary processes (Xiao et al.,
466 2009). Moreover, high grade rocks assumed as Precambrian basements in the
467 Chinese and Mongolian Altai have recently been dated at Late Paleozoic as
468 well (Long et al., 2007, 2010; Sun et al., 2008; Cai et al., 2011a; Jiang et al.,

2010). Therefore, increasing evidences converge at a conclusion that the Altai-Mongolian terrane lacks a crystallized Precambrian basement and was likely to be a subduction-accretion complex dominated by Early Paleozoic igneous materials (Şengör et al., 1993; Şengör and Natal'in, 1996; Xiao et al., 2004, Long et al., 2007, 2010; Sun et al., 2008; Cai et al., 2011a, 2011c).

6.4 Tectonic evolution and continental growth

The majority of Early Paleozoic sedimentary sequences on the periphery of Tuva-Mongolian microcontinent are commonly characterized by the major age populations at ~520 Ma and ~800 Ma, and minor age populations of ~1800 Ma and ~2500 Ma (Long et al., 2007, 2010; Sun et al., 2008; Safonova et al., 2010; Jiang et al., 2011; Rojas-Agramonte et al., 2011; Han et al., 2011; Buslov et al., 2013; Glorie et al., 2014). The above similarities enable us to conclude that the Tuva-Mongolian microcontinent may have served as the relatively stable and common provenances for sediment supply. On the other hand, new reliable paleomagnetic data suggest that, in the Ediacaran–Early Cambrian times, the Tuva–Mongolian microcontinent assembled near the equator, close to the position of the Siberian Craton (Levashova et al., 2010; Kravchinsky et al., 2010). Comparably, paleomagnetic data of the Early Devonian rocks from the Altai-Mongolian terrane suggest that their original locations were also near the equator, at 1–4°N (Buslov et al., 2004a). Recently, a Permo–Triassic 90° anticlockwise rotation in the southwest Mongolia has

491 been documented by palaeomagnetic and structural data (Bernard Edel et al.,
492 2014), accordingly, we suggest that the large scale strike-slip faulting system
493 may be responsible for the serious dismemberment and displacement of a
494 subduction-accretion system formed in the Early Paleozoic (Şengör et al.,
495 1993; Şengör and Natal'in, 1996). In combination with detrital zircon age
496 patterns of blocks in the peripheral region of the Siberian Craton, a huge and
497 continuous subduction-accretion system was suggested to have been built on
498 the margin of the Tuva-Mongolian microcontinent in the Early Paleozoic
499 (Fig.8).

500 It is worth to outline here that the Gorny Altai consists mainly of island arc
501 rocks and represents a typical intra-ocean subduction-accretion system prior
502 to the Devonian (Dobretsov and Buslov, 2007; Glorie et al., 2011; De Grave et
503 al., 2011). A common Devonian sedimentary sequence is exposed widely both
504 in the Gorny Altai and in the Altai-Mongolian terrane, suggesting a collisional
505 event to weld them prior to that time (Dobretsov and Buslov, 2007; Glorie et al.,
506 2011; Buslov et al., 2001, 2013). Glorie et al. (2011) related Devonian
507 magmatism to this collisional event. However, some diagnostic signatures
508 imply that ridge-trench interaction might have also played an important role in
509 tectonic evolution of the region, such as (1) the dramatic change in zircon Hf
510 isotope compositions (Sun et al., 2009; Cai et al., 2011a), (2) the generation of
511 adakitic rocks (Niu et al., 2006; Cai et al., 2010), (3) the opening of a backarc
512 basin (Xu et al., 2003; Wang et al., 2006), (4) the high T/P metamorphism with

a peak condition of ca. 720-800 °C and 0.5-0.7 kbar (Li et al., 2004; Wei et al., 2007; Jiang et al., 2010), and (5) the extensive mineral deposits (Han et al., 2004; Gao and Zhou, 2013). Ridge subduction can facilitate underplating and intraplate of mantle-derived magmas to consolidate the accretionary complex, and trigger partial melting of crustal materials to give rise to extensive emplacements of granitoids (Dickinson and Snyder, 1979; Sisson et al., 2003; Cai et al., 2010; Liu et al., 2012). This is compatible with the crustal architecture revealed by seismic refraction data, i.e., the crust of the Chinese part of the Altai-Mongolian terrane is up to 55 km thick and the middle-lower crust has mafic-intermediate compositions and is 46 km thick (Wang et al., 2003). Therefore, formation of subduction-accretion complexes has significant implications for continental crust growth by a composite style of lateral sediment accumulation and vertical basaltic injection.

7. Conclusions

(1) The Terekta Formation was deposited after the Middle Ordovician, rather than in the Precambrian era as previously assumed. The depositions of the Kaem Formtion, Gorny Altai Formation and Bar-burgaza formation are in the Late Silurian to Devonian.

(2) All the detrital zircon ages of the investigated sedimentary sequences share two most prominent age populations at ca. 520 Ma and ca. 800 Ma, and those zircon grains were likely derived from proximal igneous rocks.

Whereas, a few Archean to Mesoproterozoic ones were possibly from the Tuva–Mongolian microcontinent in the vicinity.

(3) The Altai-Mongolian terrane lacks a crystallized Precambrian basement and was likely a subduction-accretion complex formed in the margin of the Tuva-Mongolian microcontinent in the Early Paleozoic. Its formation bears significant implications for continental growth by both lateral sediment accumulation and vertical basaltic injection.

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Figure captions:

Fig.1 (A) Tectonic sketch in the northwestern CAO (after Dobretsov and Buslov, 2007; Glorie et al., 2011). Inset figure shows location of the CAO. SC- Siberian Craton, TC – Tarim Craton, NC – North China Craton. (B) Simplified geological map of the Russian Altai (from Glorie et al., 2011).

Fig.2: Simplified geological map of the studied area in the Russian Altai (from Dobretsov and Buslov, 2007; Glorie et al., 2011)

Fig.3: Major strata of the studied area in the Russian Altai, and the stratigraphic division is after the 1: 200,000 geological map.

Fig. 4: Cathodoluminescence images of the representative zircon grains separated from the investigated sedimentary sequences. The circles indicate U-Pb dating and Hf isotopic analyzed sites

Fig. 5: U-Pb concordia diagrams of zircon ages from the investigated sedimentary sequences in the Russian Altai.

Fig. 6: Diagrams of $\varepsilon_{\text{Hf}}(t)$ values vs. crystallizing ages for zircon grains from the investigated sedimentary sequences in the Russian Altai.

Fig. 7: Age histogram and relative probability plots of detrital zircon samples of the investigated sedimentary sequences in the Russian Altai.

Fig. 8: Pooled U–Pb age–frequency diagram for magmatic rocks in the Tuva-Mongolian microcontinent and western Mongolia (a) compared to the data compilations of the

Altai–Sayan region, Gorny Altai and the the Altai-Mongolian terrane. (b) The zircon age-spectrum for the sedimentary sequences in the Altai–Sayan region (after Glorie et al., 2014). (c) The zircon age-spectrum for the pre-Devonian sedimentary rocks in the Gorny Altai (after Chen et al., 2014b). (d) A compilation of detrital zircon U-Pb ages for sedimentary sequences in the Altai-Mongolian terrane (Data sources include this study, Long et al., 2007, 2010; Sun et al., 2008; Chen et al., 2015a and Jiang et al., 2012).

Fig. 9: Paleogeographic reconstruction of the huge Altai-Mongolian subduction-accretion complex, as well as the Gorny Altai intra-ocean subduction system, within the CAOB in the Early Paleozoic and Devonian time (modified from Meert and Lieberman, 2008; Levashova et al., 2010, 2011 and Xiao et al., 2010).

System	Formation	Lithostratigraphy	Thickness (m)	Lithology
Devonian	Bar-burgaza	Upper	800	Grey siltstone-conglomerate and limestone lens
		Lower	350-1000	Grey sandstone-siltstone-shale-conglomerate and minor limestone
Cambrian-Ordovician	Gorny Altai	Upper	1000-1300	Greenish sandstone-siltstone-phyllite and minor tuffic sandstone
		Lower	1400-1800	Grey siltstone-phyllite and locally conglomerate
Cambrian	Kaem	Upper	1200	Tuffic sandstone-siltstone-limestone
		Lower	800	Porphyry-tuff-sandstone-schist
Unknown	Terekta		> 2000	Gneiss and schist

Relative probability

Early Paleozoic

Devonian

■ Laurentia ■ Baltica ■ Gondwana ■ Siberia ■ Avalonia ■ Microcratons ■ East Africa ■ The Middle Atlantic oceanic crust