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GR focus review

Juvenile versus recycled crust in the Central Asian Orogenic Belt: Implications from ocean plate stratigraphy, blueschist belts and intra-oceanic arcs

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ABSTRACT

New or "juvenile" crust forms and grows mainly through mafic to andesitic magmatism at Pacific-type or accretionary type convergent margins as well as via tectonic accretion of oceanic and island-arc terranes and translation of continental terranes. During the last decades the juvenile or recycled nature of crust has been commonly evaluated using whole-rock isotope and Hf-in-zircon isotope methods. However, evidence for the accretionary or Pacific-type nature of an orogenic belt comes from geological data, for example, from the presence of accretionary complexes (AC), intra-oceanic arcs (IOA), oceanic plate stratigraphy units (OPS), and MORB-OIB derived blueschist belts (BSB). The Central Asian Orogenic Belt (CAOB) represents the world's largest province of Phanerozoic juvenile crustal growth during ca. 800 m.y. between the East European, Siberian, North China and Tarim cratons. From geological point of view, the CAOB is a typical Pacific-type belt as it hosts numerous occurrences of accretionary complexes, intra-oceanic arcs, OPS units, and MORB-OIB derived blueschist belts. In spite of its accretionary nature, supported by positive whole rock Nd isotope characteristics in CAOB granitoids, the Hf-in-zircon isotope data reveal a big portion of recycled crust. Such a controversy can be explained by presence of accreted microcontinents, isotopically mixed igneous reservoirs and by the tectonic erosion of juvenile crust. The most probable localities of tectonic erosion in the CAOB are the middle and southern Tienshan and southern Transbaikalia because these regions comprise a predominantly recycled crust (based on isotope data), but the geological data show the presence of intra-oceanic arcs, blueschist belts and accreted OPS with oceanic island basalts (OIB) and tectonically juxtaposed coeval arc granitoids and accretionary units. This warrants combination of detailed geological studies with isotopic results, as on their own they may not reflect such processes as tectonic erosion of juvenile crust and/or arc subduction.

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1. Introduction: formation and transformation of continental crust in Asia

New or "juvenile" continental crust forms and grows through tonalite–trondhjemite–granodiorite (TTG-type) and mafic magmatism at convergent margins mostly in intra-oceanic arcs and through intra-plate mantle plume magmatism. In addition, the volume of continental crust can be increased via accretion of oceanic crust, island-arcs and/or microcontinents (Fig. 1). The juvenile crust can be then re-worked and/or recycled at convergent margins by andesitic to felsic magmatism in continental arcs and by collision-related granitoid magmatism and metamorphism. Evaluation of proportions of juvenile continental crust versus the crust which was reworked in orogenic belts during the suturing of oceans and continental collisions is extremely important in academic and applied (mineral exploration and prospecting) sciences to understand major stages of Earth's geological history and secular changes in dominant types and settings of mineral deposits.

Asia is the world's most complex area of continental crust formation and its interaction with the oceanic lithosphere of the Pacific plate at the western Pacific as well as in the India-Sumatra subduction zones, both producing large volumes of TTG-crust (Fig. 2). Asia can be viewed as a probable core of the next supercontinent (Maruyama et al., 2007; Safonova and Maruyama, 2014). The Asian part of the Eurasian continent includes five major cratonic blocks (Siberia, Tarim, North China, South China and India), two large younger continental domains (Kazakhstan, Indochina), numerous microcontinents (MCs) and orogenic belts. It has been formed by multiple events of oceanic subduction and suturing in the Paleo-Asian, Paleo-Pacific, Tethyan and Pacific oceanic realms, accompanied by multiple continental collisions (e.g., Maruyama et al., 1989; Zonenshain et al., 1990; Sengör and Natal'in, 1996; Jahn et al., 2000; Yakubchuk, 2004; Windley et al., 2007; Safonova and Santosh, 2014; Dong et al., 2016). Oceanic closures and related continental collisions form two types of orogenic belts: accretionary or Pacific-type (also known as Turkic-type) and Himalayan-type (also known as collision-type) (Fig. 3). The Pacific-type belts form above the subduction zones, where oceanic lithosphere is submerged under intra-oceanic arcs or active continental margins, e.g., the Circum-Pacific orogens (Fig. 3A) and their fossil equivalents of the Central Asian Orogenic Belt (CAOB; Fig. 2). The Pacific-type belts are major sites of continental growth through TTG-type magmatism and accretion (e.g., Maruyama et al., 1989; Stern, 2010; I. Safonova et al., 2011; Safonova and Maruyama, 2014; Xiao and Santosh, 2014). The Himalayan-type belts result from collisions of continental blocks accompanied by strong re-working of already existing continental crust through granitoid magmatism and extended medium-pressure (MP) to high-pressure–high-temperature (HP-HT) metamorphism (Fig. 3B).

The CAOB is located between the East European, Siberian, North China and Tarim cratons, encompassing an immense area from the Urals in the west, through Altai-Sayan and Transbaikalia in Russia, Kazakhstan, Kyrgyzstan, Uzbekistan, north-western China, Mongolia, and north-eastern China to the Sea of Okhotsk in the Russian Far East (Figs. 2, 4). It is the world's largest accretionary orogen that evolved during some 800 m.y. The CAOB includes predominantly Pacific-type orogens, but Himalayan-type orogens can be also recognized. Evaluation of proportions between juvenile and recycled crust, i.e., proportions of Pacific-type versus Himalayan-type orogens, is a key element in reconstructing major stages and dynamics of continental growth. As syn-orogenic and post-orogenic tectonics may have disturbed or destroyed the original relationships of rocks units, direct "in-field" evaluation of those proportions can be problematic.

During recent years, the focus in the studies of orogenic belts has shifted from the field geology to the application of methods of radiogenic isotope research, the most reliable and popular of which are whole-rock Nd and Hf-in-zircon systematics (Jahn, 2004; Kröner et al., 2014). The Nd systematics can provide a rough estimate of the proportions of juvenile versus recycled older crustal material for samples with determined crystallization ages. The Hf systematics, if combined with geochronology, has a potential not only to fix "juvenile versus recycled" crust proportions but to reconstruct a history of magmatism through age vs isotope correlations. The numerous Nd and Hf isotope data, obtained during the last 10 years, show from 20 to 80% of recycled crust in the CAOB (Jahn, 2004; Kovalenko et al., 2004; Condie and Kröner, 2013; Turkina et al., 2012; Kröner et al., 2014) in contrast to its geologically recorded predominantly Pacific-type nature (e.g., Windley et al., 2007; Kruk et al., 2011; Yarmolyuk et al., 2012; Safonova and Maruyama, 2014) (Fig. 5). This paper is an attempt to understand the reasons for the discrepancy and possible limitations of the isotopic methods in differentiating Himalayan-type and Pacific-type orogens as well as in evaluating the proportions of juvenile and recycled crust in the CAOB. The four key components of Pacific-type orogens – accretionary complexes (AC), intra-oceanic arcs (IOA), oceanic plate stratigraphy (OPS), and MORB-OIB derived blueschist belts (BSB) – will be systematically assessed on detailed geological studies accompanied by well-justified application of up-to-date high-precision analytical methods for reconstructing the settings of formation of continental crust within the structurally complex orogenic belts. A special focus

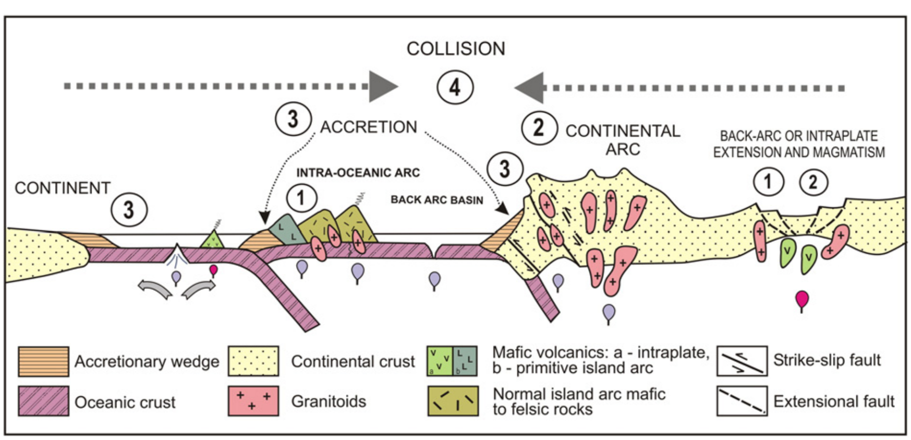

Fig. 1. A cartoon, illustrating a principal model of continental construction at Pacific-type or accretionary orogens, including formation of juvenile and recycled continental crust at major tectonic domains and associated types of magmatism (modified after Groves et al., 1998; I. Safonova et al., 2011). Numbers in circles show the main magmatic (1, 2) and tectonic (3, 4) stages of continental construction: 1 — formation of juvenile crust in intra-oceanic arcs (TTG-type and mafic magmatism); 2 — formation of juvenile and recycled crust at continental arcs (mafic to felsic magmatism) and in intra-plate settings (mafic and bimodal magmatism related to rifting/mantle plumes); 3 — accretion of oceanic, island-arc terranes and microcontinental terranes; and 4 — collision of terranes.

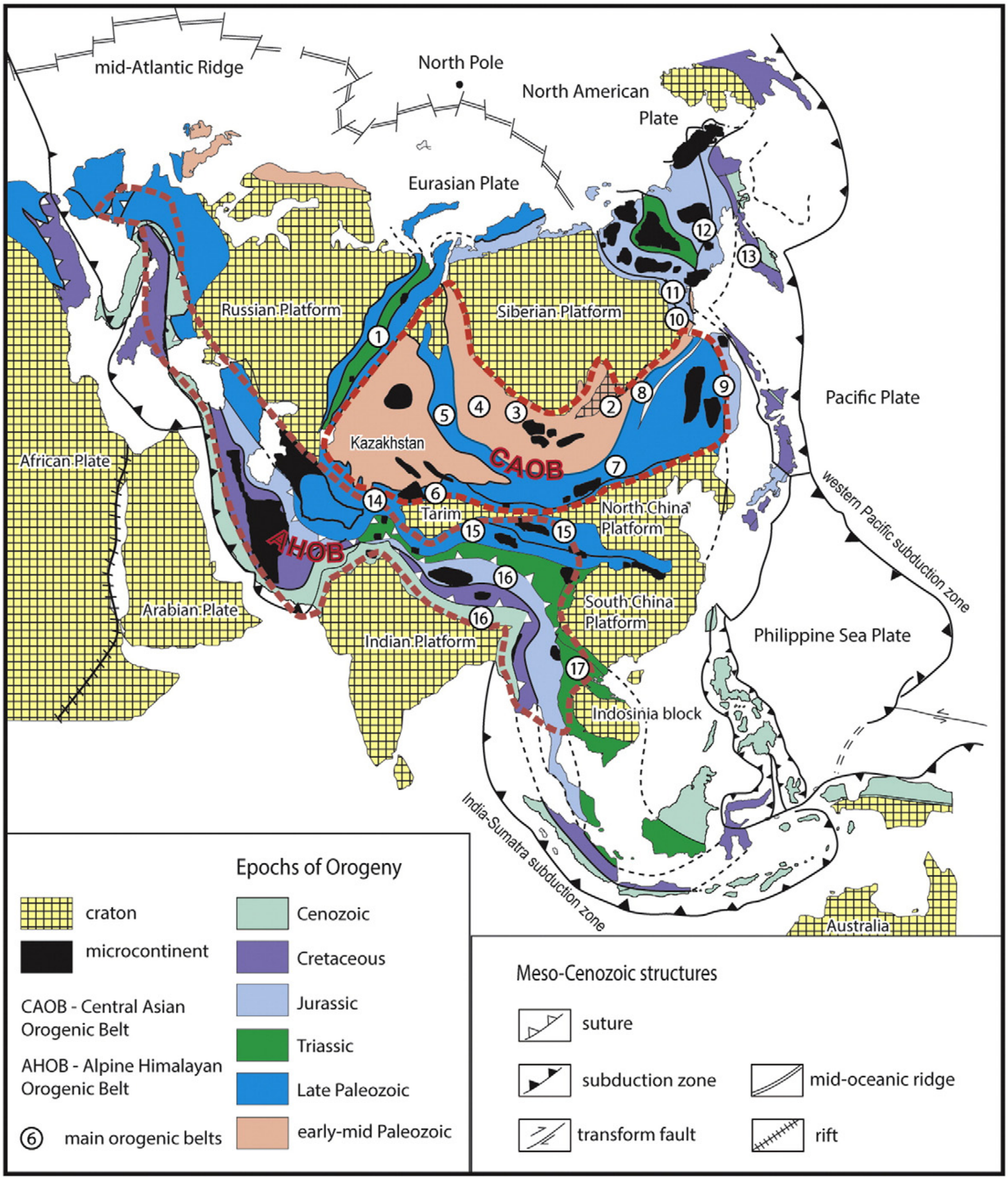

Fig. 2. A general tectonic outline of Asia: major cratonic blocks, microcontinents, Pacific and collision-type orogenic belts (modified from Maruyama and Sakai, 1986; Maruyama et al., 1989; Safonova and Maruyama, 2014). Numbers in circles for orogenics belts: 1 = Uralian, 2–6 = Central Asian (2 = Baikal–Muya, 3 = Yenisey–Transbaikalia–North Mongolia, 4 = Altay–Sayan–NW Mongolia, 5 = Irtysh–Zaisan, 6 = Tienshan), 7 = South Inner Mongolia, 8 = Mongol–Okhotsk, 9 = Sikhote–Alin, 10 = Verkhoyansk, 11 = Okhotsk–Chukotka, 12 = South Anyui, 13 = West Kamchatka, 14 = Pamir–Hindukush, 15 = Kunlun–Qinling, 16 = Tibet–Himalaya, and 17 = South China. CAOB, Central Asian Orogenic Belt; and AHOB, Alpine-Himalayan Orogenic Belt. For details about CAOB units see Figs. 4 and 10. The outlines of main microcontinents are very relative and shown out of scale.

will be given to the links between geological, geochemical and isotope data.

Formation and transformation of continental crust in Central and East Asia is a focus of Project#592 "Continental construction in the Central Asian Orogenic Belt compared to actualistic examples in the western Pacific" of the International Geoscience Program (2012–2016) supported by UNESCO and International Union of Geosciences (https://sites.google.com/site/igcp592/), which the author of this paper is a leader of together with Profs. Reimar Seltmann, Min Sun and Wenjiao Xiao. This paper reviews results of the 4 years of author's project activity.

2. Pacific-type orogens as Earth's principal sites of continental growth: definition and recognition

The Pacific-type orogenic belts form above the subduction zones, where oceanic lithosphere is submerged under active continental margins (Fig. 3A). The Pacific-type belts are the most important sites of juvenile crust formation through subduction-related TTG-type granitoid magmatism (i.e., calc-alkaline andesitic volcanism) and M- and I-type granitoid magmatism and through accretion of various fragments of oceanic lithosphere (i.e., oceanic islands, plateaux and ridges) and continental crust, such as island arcs and microcontinents

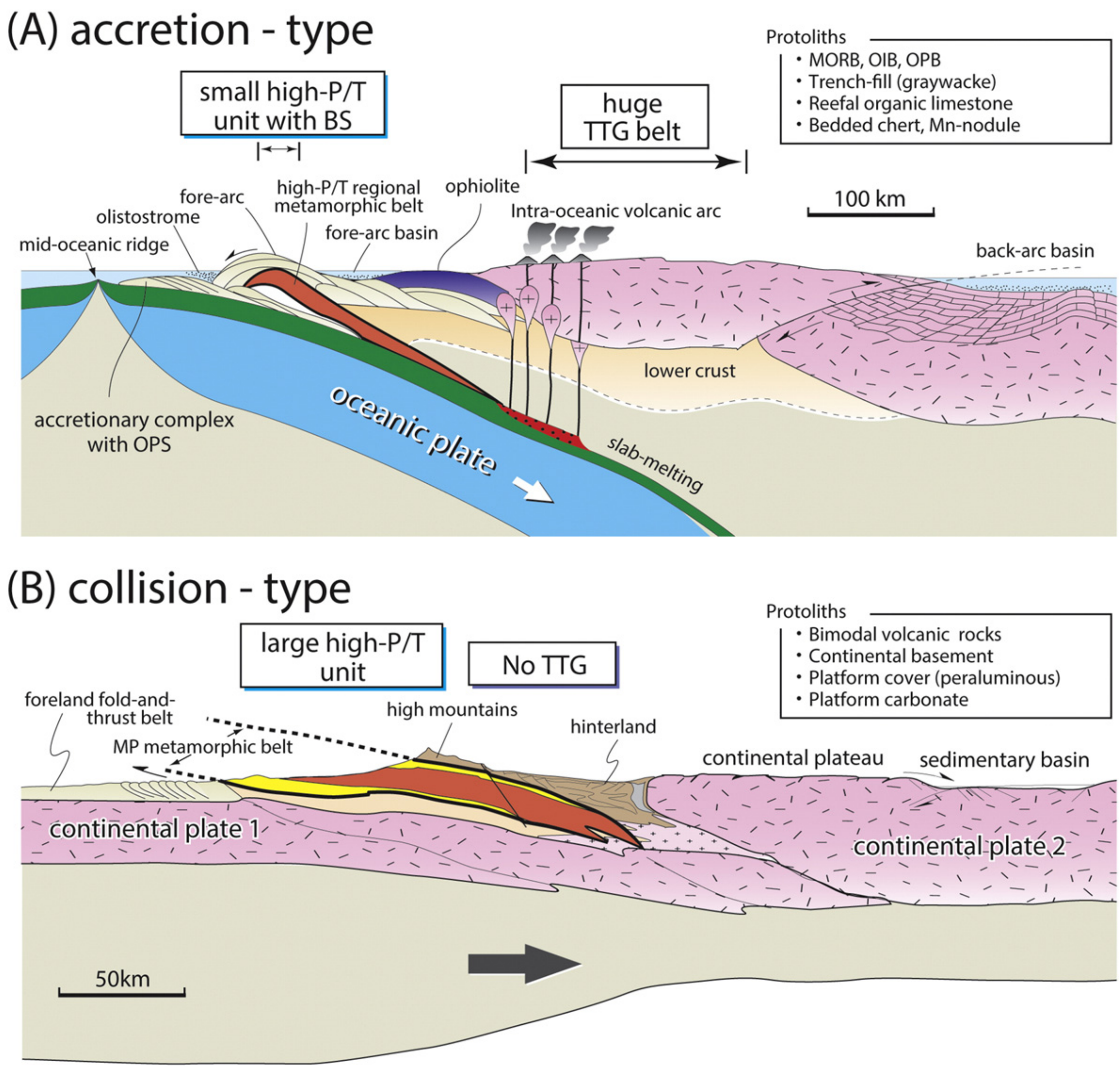

Fig. 3. Major features of accretionary or Pacific-type (P-type) and Alpine-type or collisional (C-type) orogenic belts (modified from Maruyama et al., 1996; Santosh et al., 2010). Abbreviations: MORB, mid-oceanic ridge basalt; MP, medium pressure; high-P/T, high temperature–high pressure; OIB, oceanic island basalt; OPB, oceanic plateau basalt; OPS, ocean plate stratigraphy; and TTG, tonalite-trondhjemite-granodiorite.

(Maruyama et al., 1996; Santosh et al., 2010; Maruyama et al., 2011). As a result, the subduction forms accretionary complexes, which, together with accreted terranes, finally enlarge the continents. Consequently, the most important "diagnostic" constituents of Pacific-type orogens are intra-oceanic arcs, ocean plate stratigraphy (OPS), blueschist belts and accretionary complexes, which may include fragments of the former three (Table 1; Fig. 6).

The intra-oceanic arcs form at convergent margins, in the upper "stable" plate, when the subducting plate submerges to the depths of melting, i.e., to ca. 50–100 km (Fig. 6A). Many scientists believe that intra-oceanic arcs are most important sites of juvenile continental crust formation and that the subduction-related TTG-type granitoid magmatism is a main contributor to continental growth (e.g., Clift et al., 2003; Stern, 2010; Maruyama et al., 2011; Safonova and Maruyama, 2014). The intra-oceanic arcs may consist of volcanic flows of dominantly basalt-andesite-(boninite) composition and tonalite-diorite-granodiorite plutons. The composition of both types of magmatic rocks depends on the amount of water released from the subducting hydrated oceanic slab, which, in turn, is determined by the distance from the trench. Consequently, at early stages of IOA evolution, the subduction produces fore-arc depleted tholeiitic basalts, andesibasalts and boninites. In time and with distance away from the subduction zone, i.e., continentward, the supra-subduction magmatism proceeds at lower degrees of melting and produces calc-alkaline, alkaline and shoshonitic andesitic to felsic series, more enriched in incompatible elements.

The ocean plate stratigraphy represents a regular succession of sedimentary and magmatic rocks that were deposited on the sea floor as the underlying oceanic basement made its inevitable journey from its formation at a mid-ocean ridge to its demise at a deep sea trench (e.g., Isozaki et al., 1990; Maruyama et al., 2010; Kusky et al., 2013). The "standard" model for OPS includes typical Penrose type ophiolite, grading down from pelagic sediments to basalts, gabbros, and ultramafics formed on the oceanic floor (Fig. 6B). As the oceanic plate moves towards the trench, the pelagic sedimentation continues until the plate enters the trench. At this point, the pelagic sediments and their underlying lavas become covered by hemipelagic shale/mudstone, greywacke and turbidite (Kusky et al., 2013). If a topographic rise or oceanic rise (seamount, island, plateau) is present on the oceanic floor, the OPS will be added by oceanic island carbonate cap and carbonate-siliceous-volcanic epiclastic slope facies and breccia. At convergent margins, OPS units may be accreted to island arcs/continental margins or tectonically eroded and subducted, carrying water and carbonate-rich materials deep into the mantle. The presence of accreted seamount OPS units in foldbelts is a key signature of a Pacific-type orogen.

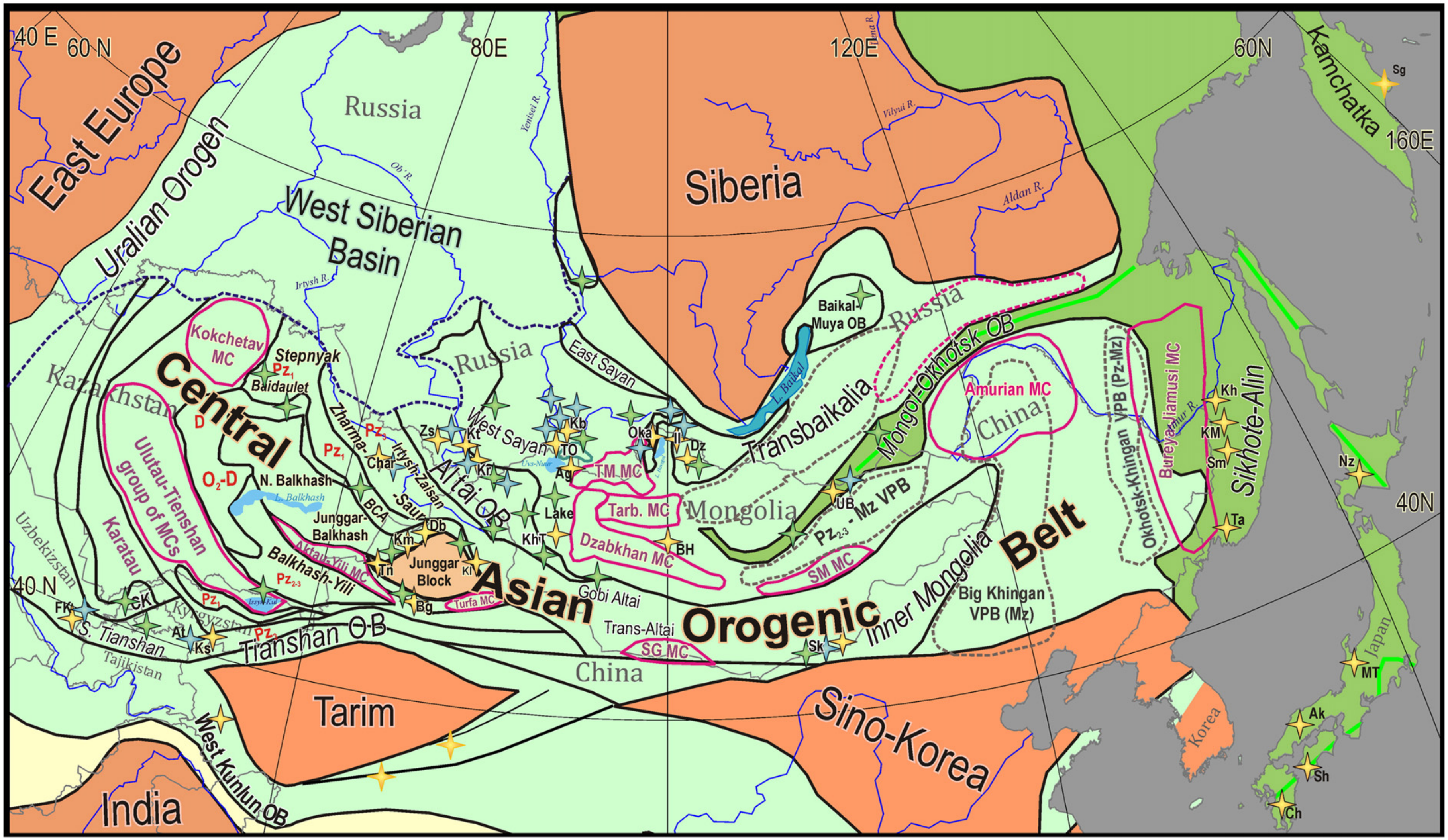

Fig. 4. A general scheme of the Central Asian Orogenic Belt showing major orogenic belts (italic font) and microcontinents or their groups (pink lines) plus surrounding tectonic structures (compiled using Ren et al., 1999; Windley et al., 2007; Yarmolyuk et al., 2011; Kröner et al., 2014; Safonova and Santosh, 2014; Guy et al., 2015; Buriánek et al., in this issue; Degtyarev et al., in this issue; Dolgopolova et al., 2017-in this issue). Stars: accreted OIB (golden), blueschist (blue), intra-oceanic arc (green). Accretionary complexes: Ag, Agardag; Ak, Akiyoshi; Bg, Bayingou; BH, Bayanhongor (Burg Gol); Ch, Chichibu; Dr., Darbut; Dz, Dzhida; FK, Fan-Karategin; Il, Ilchir; Kb, Kurtushibin; Kl, Kalamaili; KhT, Khan-Taishir; Km, Karamay; Kh, Khabarovsk; Kr, Kurai; Ks, Kokshaal; Kt, Katun; MT, Mino-Tamba; KM, Kiselevka-Manoma; Nz, Naizawa; O, Oka (Hug, Shishged, Shishkhid); Sg, Smagin; Sh, Shimanto; Sk, Solonker; Sm, Samarka; Ta, Taukha; Tn, Tangbale; TO, Tannu-Ola; UB, Ulaanbaatar (Adaatsag); Zs, Zasur'ya. Other abbreviations: BCA, Bozshakol-Chingiz arc; CK, Chatkal-Kurama continental arc; MC, microcontinents (TM, Tuva-Mongolian; SG, South Gobi; SM, South Mongolia; Tarb., Tarbagatai); N, north; OB, orogenic belts; S, south; and VPB, volcano-plutonic belts. Dashed lines: dark blue — boundary of the West Siberian Basin; pink — a terrane of debatable origin (MC or another); and gray — superimposed VPB. The shapes and sizes of the fields are relative in most cases.

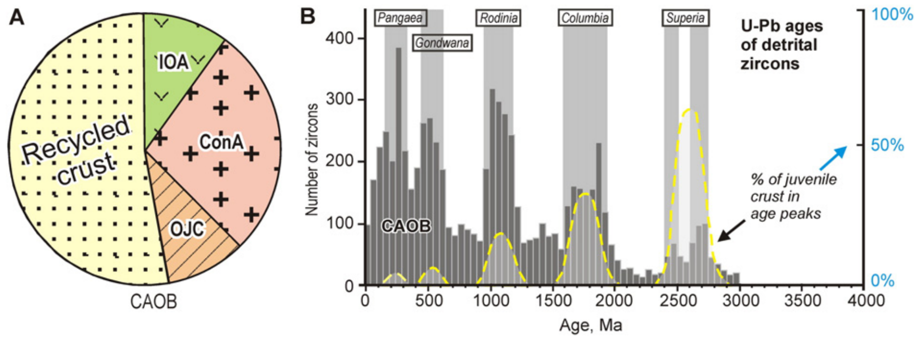

Fig. 5. Isotope-based proportions of juvenile and recycled crust in the CAOB. A, Nd and Hf isotopic data based estimates of the volume distribution of juvenile continental crust (modified from Condie and Kröner, 2013); B, the peaks of U–Pb crystallization ages for over 7000 detrital zircons obtained worldwide (Campbell and Allen, 2008) and the respective portions of juvenile crust (modified from Hawkesworth et al., 2010). Settings of juvenile crust formation (for A): IOA, intra-oceanic arcs; ConA, continental arc; OJC, other juvenile crust terranes (e.g., intra-plate igneous fields). The names of supercontinents (for B) are after Santosh et al., 2014.

Blueschist is a metabasite, containing Na-amphibole stable at relatively high-P and low-T conditions. The term "blueschist belt" (BSB) refers to the blueschist-assemblage-bearing part of an orogen, which may also include metagraywacke and marble, as well as eclogite-facies rocks. BSBs are markers of fossil subduction zones, because they form during the stop/cessation of subduction often accompanied by exhumation of buried metamorphic rocks, and therefore are of high tectonic significance (Miyashiro, 1973; Xiao et al., 1994; Maruyama et al., 1996). BSBs may be part of paired metamorphic belts, which are spatially related, i.e., located side-by-side, metamorphic belts of low dP/dT (greenschist, andalusite-sillimanite) and high dT/dP (blueschist, eclogite) (Miyashiro, 1961; Oxburgh and Turcotte, 1971; Brown, 2010; Fig. 6C). The paired metamorphic belts were first recognized in Japan (Miyashiro, 1961; Maruyama, 1997); similar belts in different Phanerozoic orogenic belts have been subsequently identified and are thought to indicate destructive convergent margins (e.g., Zhang et al., 1993; Maruyama et al., 1996; Brown, 2010). BSBs are classified into two types according to the protoliths. In Pacific-type orogens, the protoliths for blueschists are rocks forming an accretionary complex, i.e., OPS units, including MORB, OIB, seamount (pure) limestone and turbidite. The protoliths of blueschists in C-type orogens are continental basement granite-gneiss complexes and their overlying peraluminous sediments, impure limestones and bimodal volcanic series (Table 1).

The accretionary complexes consist of an accretionary wedge or prism and subduction-related metamorphic units. The accretionary wedge forms by off-scrapping and underplating accretion to the non-subducting tectonic plate or the landward slope of trench. Lithologically, accretionary wedge is dominated by OPS units plus terrigeneous trench sediments, oceanward olistostrome, tectonic and serpentinite mélanges. After cessation of accretion and subduction, accretionary wedge together with blueschists and other exhumed metamorphic rocks form accretionary complex (Fig. 6D). During subduction, slices and sheets of scrapped-off rocks pile next to each other along reverse or thrust faults resulting in formation of a very complicated structure. Structurally, an accretionary complex is dominated by duplexes and thrust imbricates characterized by inward younging of beds within each sheet and overall outward younging. A key element of accretionary complex is a duplex structure formed in the hanging wall of the trench; its geometry and the direction of bed younging can be used to determine the direction of subduction or "subduction polarity" (e.g., Isozaki et al., 1990; Maruyama et al., 2010; Wakita, 2012; Safonova and Santosh, 2014; Fig. 6D).

The field and laboratory recognition of OPS, IOAs, BSBs and ACs is necessary for differentiation of Himalayan-type or Pacific-type belts. Pacific-type orogens can be recognized by the presence of: (i) blueschists formed after mid-oceanic ridge basalt (MORB) and oceanic island or plateau basalt (OIB/OPB); (ii) dominantly mafic boninite-bearing island arcs; (iii) accreted oceanic rises, especially those capped by carbonates; (iv) paired metamorphic belts; and (v) huge supra-subduction granitoid batholiths (see Table 1 and Safonova and Maruyama, 2014). The accretionary complexes, OPS units and blueschist belts of the CAOB have been reviewed in Volkova and Sklyarov (2007) and Safonova and Santosh (2014). The list of OIB-bearing OPS and intra-oceanic arcs with

Table 1
Main features of accretion-type and collision-type orogenic belts (modified from Maruyama et al., 1996; Safonova and Maruyama, 2014).

Rock types/other aspects	P-type: subduction-related orogens with accretionary complexes	H-type: collision-related orogens built over continental basement
Key lithologies		
Shallow-marine sediments	Reefal limestone, volcanogenic-carbonate breccia, often Z-folded, siliceous shale and mudstone (OPS)	Thick platform carbonate and clastics
Deep-sea sediments	Bedded chert often with manganese-nodules (OPS)	Absent
Turbidite	thick and immature trench-fill graywackes (mafic to andesitic)	Platform cover (peraluminous and carbonate-rich deposits)
Terrigenous fan	deep-marine flysch/forearc covering OPS	continental molasse and/or foreland basin
Volcanic units	Mafic to andesitic and felsic lavas	Bimodal (basalt and dacite) lavas
Intrusive rocks	Huge subduction-related granitoid batholiths	Syn-collisional (I and S-type) and post-collisional (A-type) granitoids
Related metamorphic belts		
Key metamorphic assemblages	Blueschist, greenschist, HP-UHP	Blueschist, MP (kyanite-sillimanite), UHP
Blueschist protoliths	MORB, OIB	Peraluminous, pelitic, bi-modal
Metamorphosed peridotites	Strongly serpentinized spinel- and plagioclase-bearing peridotites	Strongly metamorphosed garnet- and spinel-peridotite
Paired metamorphic belts	May be present	Absent
Other aspects		
Continental basement	Absent	Abundant granite-gneiss complexes
Ore	black smokers, Fe-Mn and volcanogenic-sedimentary deposits	Petroleum (foredeep troughs), gold from quartz veins; gems
Intra-oceanic arc boninites	May present	Always absent
Structural features	Duplex OPS structures; dominantly asymmetric folds and small thrusts	Variable deformation styles to form buckling, huge thrusts, etc.

Abbreviations: HP-UHP — high pressure-ultra high pressure; MORB — mid-oceanic ridge basalt; MP — medium pressure; OIB — oceanic island basalt; and OPS — Ocean Plate Stratigraphy.

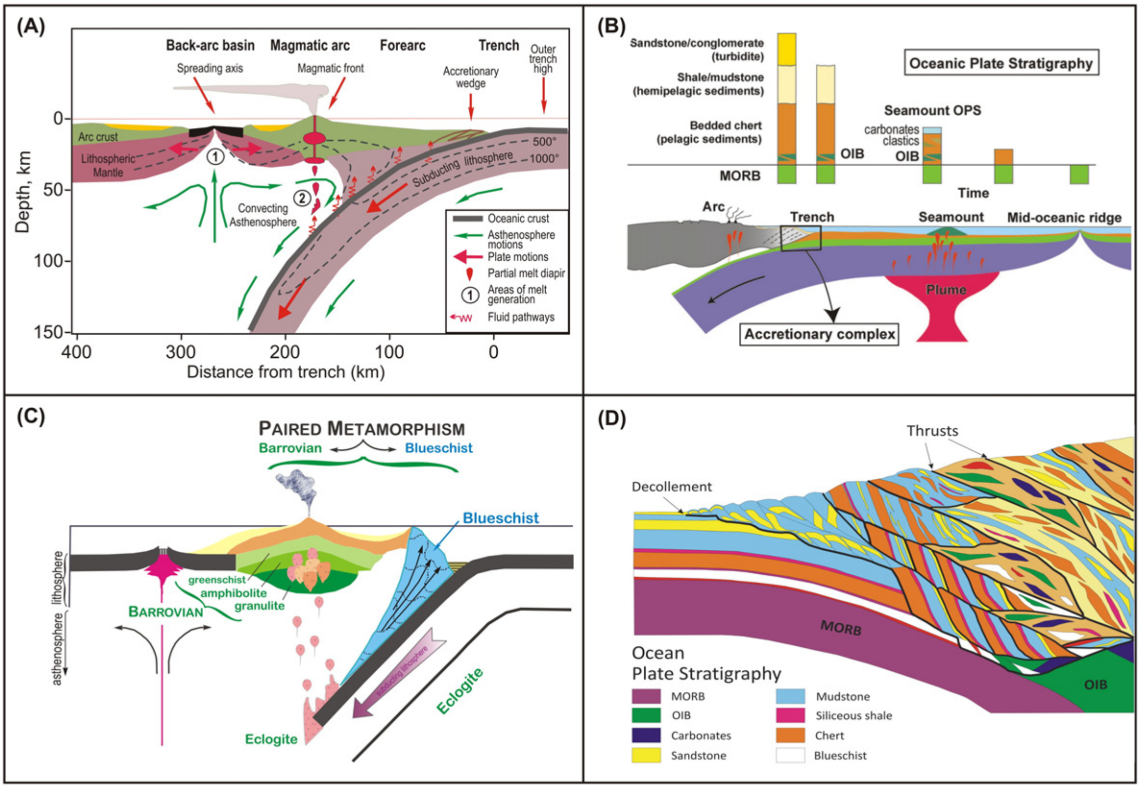

Fig. 6. Major diagnostic features of Pacific-type orogenic belts: Intra-oceanic arcs (A), Ocean Plate Stratigraphy including accreted seamounts (B), paired metamorphic belts (C), and accretionary complexes (D). A, a typical intra-oceanic arc system with adjacent back-arc basin showing major crustal and upper mantle components and major sites of melt generation: 1 — at mid-oceanic ridges, 2 — above the subducting slabs (modified from Stern, 2010); B, OPS with seamounts (modified from Maruyama et al., 2010; Safonova et al., 2016b); C, principal scheme of formation of paired metamorphic belts, whose key component is the MORB-OIB derived blueschist (modified from the website of the Society for Sedimentary Geology, http://www.sepmstrata.org/); D, structurally complex accretionary wedges (modified from Wakita, 2012).

or without blueschists in the CAOB, including their location and age, are shown in Tables 2, 3.

3. The Central Asian Orogenic Belt: world's largest Pacific-type orogen

3.1. Definition, brief history and key terranes

The Central Asian Orogenic Belt (CAOB) is the world's largest accretionary orogen, formed during the Neoproterozoic to Paleozoic evolution of the Paleo-Asian Ocean and multi-stage collisions of the Siberian, Tarim and North China cratonic blocks, Kazakhstan and Mongolian Precambrian terranes (Fig. 4). In late Paleozoic time, the amalgamating CAOB was also affected by the collision with the East European Craton and formation of the Uralian Orogenic Belt (Puchkov, 1997). Formation of CAOB started from accretion of numerous Neoproterozoic intra-oceanic arc terranes to the southern margin of the Siberian Craton, reaching its peak during the late Neoproterozoic to the early Paleozoic (e.g., Buslov et al., 2001; Turkina, 2002; Gordienko et al., 2010) (Table 2). During the rest of the Phanerozoic, accretion of oceanic to intra-oceanic arc terranes and probably Gondwana-derived microcontinents continued at the southern active margin of the Siberian Craton with contribution to the intra-plate mafic and granitoid magmatism (e.g., Kovalenko et al., 2004; Windley et al., 2007; Safonova, 2009; Xiao et al., 2009, 2010; Safonova et al., 2011; Yarmolyuk et al., 2012; Donskaya et al., 2013; H. Yang et al., 2015).

The subduction ceased diachronously in the late Paleozoic-Mesozoic after the closure of the Paleo-Asian Ocean and its Turkestan, Junggar and Mongol-Okhotsk branches with relevant rocks occurring in East Kazakhstan (Safonova et al., 2012; Safonova and Santosh, 2014), South Tienshan (Biske and Seltmann, 2010; Wang et al., 2011; Safonova et al., 2016a), Junggar (Wang et al., 2003; G. Yang et al., 2015) and Far East (Dobretsov et al., 2003; Donskaya et al., 2013; H. Yang et al., 2015), and collision of the Siberian Craton with the Tarim and North China cratons and the Kazakhstan continent to become part of Laurasia (e.g., Safonova and Maruyama, 2014). Formation of the central part of CAOB is related to the Paleozoic collisions of the Tuva-Mongol, Dzabkhan (Baydaric, Baydrag), Tarbagatai and South Gobi microcontinents with the Siberian Craton (e.g., Didenko et al., 1994; Salnikova et al., 2001; Kozakov et al., 2007; Demoux et al., 2009). Western part of CAOB was formed during the Middle Paleozoic collision of the Kazakhstan continent with the Siberian active continental margin. This produced the Altai orogen, extending from Russia across eastern Kazakhstan and China to Mongolia (e.g., Buslov et al., 2001; Xiao et al., 2010; Glorie et al., 2011; Safonova, 2014). Collisions between the Kazakhstan continent and Tarim Craton and smaller microcontinents formed the Tienshan orogenic belt (e.g., Gao et al., 1998; Chen et al., 1999; Xiao et al., 2004; Charvet et al., 2007; Biske and Seltmann, 2010; Wang et al., 2010). The eastern part of CAOB was formed during the Late Permian to Triassic approach of the North China and Siberian Cratons and collision of the former with the island-arc terranes and microcontinents at the southern margin of the latter (e.g., Zhao et al., 1990; Xiao et al., 2004; Zhao et al., 2013; Zhou and Wilde, 2013; H. Yang et al., 2015; Li et al., 2016).

During the last 15 years, there has been much discussion about the mechanisms of the CAOB formation, e.g., extended long-living arc versus numerous arcs and microcontinents (e.g., Sengör and Natal'in, 1996; Didenko et al., 1994; Buslov et al., 2001; Windley et al., 2007; Xiao et al., 2010; Safonova et al., 2011 and references therein), the nature of its continental crust, i.e., juvenile versus recycled, the

Table 2
Major intra-oceanic arcs of the Central Asian Orogenic Belt.

No.	Terrane name	Location	Age	Bon	BSB	References
1	Dunzhugur	NW Mongolia–eastern Tuva, Russia	Neoproterozoic	Yes	Yes	Kuzmichev et al. (2001)
2	Aksu	Western Chinese Tienshan	Neoproterozoic		Yes	Zhu et al. (2011)
3	Shishkhid (Oka-Hugein)	NW Mongolia–eastern Tuva, Russia	Late Neoproterozoic	Yes	Yes	Kuzmichev et al. (2005)
4	Kurtushibin	East Sayan, Russia	Late Neoproterozoic	Yes	Yes	Dobretsov et al. (1992), Volkova et al. (2009)
5	Agardag	NW Mongolia-SE Tuva, Russia	Late Neoproterozoic			Pfänder et al. (2002)
6	Dariv	Western Mongolia	Late Neoproterozoic			Dijkstra et al. (2006)
7	Khan-Taishirin	Western Mongolia	Late Neoproterozoic	Yes		Izokh et al. (1998)
8	Tannu-Ola	Western Tuva, Russia	Late Neoproterozoic			Mongush et al. (2011)
9	Lake	Western Mongolia	ca. 570 Ma			Kozakov et al. (2007); Kovach et al. (2011)
10	Dzhida	N. Mongolia-Transbaikalia (Russia)	l. Neoproterozoic–Early Cambrian	Yes		Gordienko et al. (2007)
11	Kurai	Southern Russian Altai	598 ± 25 Ma	Yes	Yes	Simonov et al. (1994); Buslov et al. (2002)
12	Katun	Northern Russian Altai	early Cambrian	Yes		Buslov et al. (1998), Safonova et al. (2011a)
13	Bozshakol-Chingiz	East Kazakhstan	late Cambrian	Yes		Shen et al. (2015), Degtyarev (2011)
14	Dzhalair-Naiman	Northern Kyrgyz Tienshan	Cambrian			Ryazantsev et al. (2009)
15	Selety/Severny Segment	Northern Kazakhstan	Late Cambrian-early Ordovician	Yes		Windley et al. (2007), Degtyarev (2011)
16	Fan-Karategin	Southern Tienshan, Tajikistan	Ordovician		Yes	Volkova and Budanov (1999)
17	Tangbale	Western Junggar, NW China	Ordovician		Yes	X. Xiao et al. (1994)
18	Kentash arc	Western Kyrgyz Range, northern Tienshan	Middle Ordovician			Degtyarev et al. (2012)
19	Gobi Altai	SW Mongolia	Late Ordovician—early Silurian	Yes		Dergunov et al. (2001), Guy et al. (2015)
20	Kyzylkum (Chatkal) arc	Turkestan-Alai, Uzbekistan	Silurian			Biske and Seltmann (2010), Dolgopolova et al. (2017-in this issue)
21	Dananhu-Tousuquan arc	West Junggar	Late Silurian			Xiao et al. (2015)
22	Dulate arc, Shaerbulake Fm.	East Junggar	Early Devonian	Yes		Zhang et al. (2005)
23	Kuerti	NE Junggar, NW China	Middle Devonian	Yes		Wang et al. (2003)
24	Kangbutiebao	Southern Chinese Altai	Devonian	Yes		Niu et al. (2006)
25	Char arc	East Kazakhstan	Late Devonian or Early Carboniferous	Yes	Yes	Volkova et al. (2008), Kurganskaya et al. (2014)
26	Bogda arc	N. Chinese Tienshan	Late Carboniferous			Xie et al. (2016)
27	Solonker arc	Inner Mongolia	Early Permian	Yes	Yes	Jian et al. (2010)

Bon, boninite; and BS, blueschist.

proportions of juvenile and recycled crust (e.g., Jahn et al., 2000; Safonova et al., 2011; Yarmolyuk et al., 2012; Kröner et al., 2014) and/or the size, age and constituents of the continents and the number of microcontinents incorporated into the CAOB (Dobretsov and Buslov, 2007; Levashova et al., 2010). A contribution to understanding the nature of the CAOB can be made by recognition of IOAs, accreted terranes of seamount OPS and BSBs hosted by and/or associated with the numerous late Neoproterozoic to Mesozoic accretionary complexes present in Central and East Asia (Safonova and Santosh, 2014).

3.2. Intra-oceanic arcs

Identification of intra-oceanic arcs is of primary importance for recognizing the Pacific-type orogenic belts because they have the highest rate of TTG generation, i.e., of juvenile crust. The key indicative features of IOAs are: (i) the presence of boninites; (ii) predominantly (about 70%) mafic (basalt-andesibasalt) composition of rocks in modern IOAs (e.g., Gil, 1981; Kelemen et al., 2003); and (iii) immature turbidite (greywacke) and deep-sea flysch (Fig. 6A). The CAOB hosts widespread intra-oceanic arcs, many of them with boninites (Tables 1, 2; Fig. 4). The intra-oceanic arcs of CAOB are related to the evolution of the Paleo-Asian Ocean including its probable two major branches, Turkestan Ocean (or South Tienshan) and Mongol-Okhotsk Ocean, and active margins of the Siberian, Kazakhstan, and North China continents. Numerous late Neoproterozoic to Carboniferous intra-oceanic arcs of the Paleo-Asian Ocean (most of them include boninites) outcrop in Transbaikalia and northern Mongolia, Tuva and northwestern Mongolia, western Mongolia, Russian-Kazakh-Chinese-Mongolian Altai, western and eastern Junggar, southern Mongolia and Inner Mongolia (e.g., Simonov et al., 1994; Izokh et al., 1998; Buslov et al., 1998, 2001; Pfänder et al., 2002; Dobretsov et al., 2003; Kovalenko et al., 2004; Dijkstra et al., 2006; Zhang et al., 2005; Niu et al., 2006; Gordienko et al., 2007; Ota et al., 2007; Gordienko et al., 2010; Jian et al., 2010; Yang et al., 2014) (Table 2). Early Paleozoic intra-oceanic arcs of the Turkestan Ocean have been identified in the South Fergana zone of the western Tienshan, e.g., the Ordovician Taldyk Formation (Burtman, 2008). Subduction of the Mongol-Okhotsk Ocean, an eastern branch of the Paleo-Asian Ocean (Didenko et al., 1994; Dobretsov et al., 1995; Donskaya et al., 2013; Ruppen et al., 2014), produced Carboniferous to Permian IOAs in Inner Mongolia, northeast China and Russian Far East; the better known are the Early Carboniferous Baolidao arc (309 Ma) and Late Carboniferous-Permian Hegenshan arc in Inner Mongolia (e.g., Chen et al., 2000; Dergunov et al., 2001; Miao et al., 2008). The amount of the late Neoproterozoic intra-oceanic arcs is much greater than that of middle and late Paleozoic arcs (Fig. 4; Table 2). Western part of the Paleo-Asian Ocean, including the Turkestan branch, was progressively sutured during the Carboniferous and Permian, whereas its eastern part, i.e., the Mongol-Okhotsk Ocean, remained opened until the Jurassic (e.g., Zonenshain et al., 1990; Gordienko, 1994; Dobretsov et al., 1995; Buslov et al., 2001; Biske and Seltmann, 2010).

3.3. Accreted oceanic islands/seamounts/plateaus

The second important feature of the Pacific-type orogens is accreted OPS of oceanic rises or seamount OPS, which is characteristic of Pacific convergent margins only (e.g., Isozaki et al., 1990; Kanmera and Sano, 1991; Safonova, 2009; Kusky et al., 2013; Safonova and Santosh, 2014). Seamount OPS includes OIB, OPB, MORB, foothill siliceous sediments (chert, shale, mudstone), carbonate-siliceous-volcanic epiclastic slope facies and breccia, and oceanic island carbonate cap (Fig. 6B). The oceanic rises typically form during the passage of subducting oceanic plate over the mantle plume head to be finally accreted and/or partially and fully eroded at convergent margins and subducted deep into the mantle (e.g., Hofmann, 1997; Maruyama et al., 2007; Clift et al., 2009; Scholl and von Huene, 2007; Safonova, 2009; Safonova and Maruyama, 2014). The CAOB hosts numerous accreted oceanic rises tracing almost continuous archives of seamount OPS from

Table 3
Accretionary complexes of the Central Asian Orogenic Belt with accreted OIB-OPS (modified and upgraded from Safonova and Santosh, 2014).

Terrane[a]	Geography	Age of seamount (or associated OPS)	Age of accretion	References[b]
Oka	NW Mongolia–E. Tuva, Siberia (Russia)	Late Neoproterozoic	Latest Neoproterozoic	Kuzmichev et al. (2005)
Ilchir (Il)	N. Mongolia-W. Sayan, Siberia	Late Neoproterozoic	Cambrian–Early Ordovician	Kuzmichev et al. (2001), Dobretsov et al. (1992)
Kurtushibin (Kb)	E. Sayan, Siberia (Russia)	Late Neoproterozoic	Cambrian–Early Ordovician	Dobretsov et al. (1992), Volkova et al. (2009)
Agardag (Ag)	NW Mongolia-SE Tuva, Siberia (Russia)	569 ± 2 Ma	Early Cambrian	Pfänder et al. (2002)
Tannu-Ola (TO)	western Tuva, southern Siberia	Late Neoproterozoic	Cambrian–Early Ordovician	Mongush et al. (2011)
Lake	western Mongolia	ca. 570 Ma	Middle–Late Cambrian	Yarmolyuk et al. (2011), Kovach et al. (2011)
Dzhida (Dz)	N. Mongolia-Transbaikalia (Russia)	L. Neoproterozoic-E. Cambrian	Early Ordovician–Early Devonian	Belichenko (1969), Gordienko et al. (2007)
Bayanhongor (BH)	Central Mongolia	569 ± 21 Ma	Late Cambrian	Kepezhinskas et al. (1991), Buchan et al. (2001)
Kurai (Kr)	Southern Russian Altai	598 ± 25 Ma	Middle Cambrian	Safonova (2009)
Katun (Kt)	Northern Russian Altai	Early Cambrian	Middle Cambrian	Safonova et al. (2011a)
Zasur'ya (Zs)	Northwestern Russian Altai	Late Cambrian	Early Devonian	Safonova et al. (2011b)
Tangbale (Tn)	West Junggar, NW China	E.-M. Ordovician,	L. Devonian	Xiao et al. (1994), G. Yang et al. (2015)
Maile	West Junggar, NW China	L. Silurian (Wenlock-Pridoli)	L. Devonian	G. Yang et al. (2015)
Karamay (Km)	West Junggar, NW China	E.-M. Devonian	Late Carboniferous	G. Yang et al. (2015)
Darbut (Db)	West Junggar, NW China	middle-late Devonian	Late Carboniferous	G. Yang et al. (2015)
Fan-Karategin (FK)	S. Tienshan, Tajikistan	Middle-Late Devonian	Carboniferous	Volkova and Budanov (1999)
Kokshaal (Ks)	central Tienshan, Kyrgyzstan	Middle-Late Devonian	Late Carboniferous	Safonova et al. (2016a)
Ulaanbaatar (UB)	Hentey Range, N. Mongolia	L. Silurian-E. Devonian	Early Carboniferous	Safonova and Santosh (2014)
Kalamaili (Kl)	East Junggar, E. China	L. Devonian - E. Carboniferous	Late Carboniferous	Zhang et al. (2009)
Char	East Kazakhstan	L. Devonian - E. Carboniferous	Late Carboniferous	Safonova et al. (2012)
Bayingu (Bg)	Southern Junggar	L. Devonian - E. Carboniferous	Late Carboniferous	Gao et al. (1998), Xu et al. (2006)
Solonker (Sk)	Inner Mongolia	295 ± 15, 298 ± 9 Ma	Late Permian	Miao et al. (2008), Jian et al. (2010)
Akiyoshi	SW Japan	M. Carboniferous	Early Permian	Isozaki et al. (1990), Kanmera and Sano (1991)
Khabarosvk (Kh)	Western Sikhote-Alin	Carboniferous-Permian	Early Permian	Safonova (2009)
Mino-Tamba (MT)	Central Japan	Permian	Early–Middle Jurassic	Isozaki et al. (1990), Safonova et al. (2016b)
Samarka (Sm)	Central Sikhote-Alin	Carboniferous-Permian	Middle–Late Jurassic	Golozubov (2006)
Chichibu (Ch)	SE Japan	Triassic	Jurassic–Early Cretaceous	Wakita (2012), Ishizuka et al. (2003)
Taukha (Ta)	Southern Sikhote-Alin	M.-L. Triassic	Jurassic–Early Cretaceous	Golozubov (2006)
Kiselevka-Manoma (KM)	Northern Sikhote-Alin	M. Jurassic	Early Cretaceous	Filippov et al. (2010)
Naizawa (Nz)	Hokkaido, Japan	L. Jurassic	Early Cretaceous	Ueda et al. (2000)
Shimanto (Sm)	SE Japan	E. Cretaceous	Late Cretaceous	Taira et al. (1988)
Smagin (Sg)	East Kamchatka	L. Cretaceous	Late Cretaceous	Saveliev (2003)

a — abbreviations in parentheses match those in Fig. 4; and b — key selected references for geology and geochemistry are shown in this column. E., early; M., middle; and L., late.

the Neoproterozoic to the Permian and tracking oceanic plume magmatism (Safonova and Santosh, 2014; Table 3). Most localities of accreted seamount OPS in the CAOB can be divided into two major groups, Late Neoproterozoic-Cambrian and Devonian-Permian, plus a less abundant group of Ordovician-Silurian units. The Late Neoproterozoic-Cambrian seamounts, formed in the Paleo-Asian Ocean, were accreted to the IOAs and active margins of the Siberian and Tarim cratons and Tuva-Mongolian and Dzabkhan microcontinents. The Ordovician-Silurian oceanic intra-plate magmatism, probably less intensive and/or extensive, produced oceanic rises in the Turkestan Ocean. They were accreted to the active margins of the Kazakhstan continent and Tarim craton. The Devonian-Permian seamount OPS recorded the final stages of the Paleo-Asian Ocean and the intra-plate environments in the Mongol-Okhotsk Ocean; those oceanic rises were accreted to the active margins of the Siberia and Tarim cratons and Kazakhstan continent (Safonova, 2008, 2009; Safonova et al., 2011a, 2011b, 2012; Safonova and Santosh, 2014). The seamounts formed in the Paleo-Pacific Ocean in late Paleozoic and Meso-Cenozoic time are currently hosted by the accretionary complexes of the Russian Far East and Japanese Islands. A part of those accretionary complexes is often considered as previously single units later separated by the Sea of Japan, which opened at ca. 15 Ma (Kojima et al., 2000; Safonova, 2009; Filippov et al., 2010; Safonova and Santosh, 2014).

3.4. Blueschist belts

The third important characteristics for recognition of the Pacific-type belts is blueschists formed after oceanic basalts, as they mark fossil subduction zones and are therefore key indicators of Pacific-type convergent margins. In the CAOB, localities of blueschists, formed after MORB and OIB, have been discovered in many accretionary complexes of Transbaikalia-northern Mongolia, Altai-Sayan, eastern Kazakhstan, Russian and Chinese Altai, Tajik, Kyrgyz and Chinese Tienshan and in Inner Mongolia (Fig. 4; Table 2; e.g., Gao et al., 1995; Buslov et al., 1998; Volkova and Budanov, 1999; Volkova and Sklyarov, 2007; Volkova et al., 2009; Gao et al., 2011; Safonova and Maruyama, 2014; Safonova and Santosh, 2014). In general, three groups of BSBs can be recognized according to their age and location: (1) Late Neoproterozoic-Early Cambrian BSBs of Transbaikalia-northern

Mongolia (North Muya, Oka, Kurtushiba) and Altai-Sayan regions (Borus, Kurai, Uimon); (2) Ordovician BSBs of eastern Kazakhstan (Char or Chara or Charsk), West Junggar (Tangbale) and Inner Mongolia (Ondor Sum); and (3) Devonian-Carboniferous BSBs of southern Tienshan (Fan-Karategin, Atbashi, Tonghuashal, Kekesu-Akeyazi). They occur as slices and/or lenses, e.g., the Oka and Kurtushiba BSBs of western Transbaikalia and the Uimon BSB of the Russian Altai, or as blocks in mélange, e.g., the Chara BSB of eastern Kazakhstan and Borus BSB of North Sayan. The BSBs are spatially associated with coeval or older OPS and/or accretionary complexes. The protoliths of blueschists are mostly N-MORB or enriched OIB basalts (e.g., Volkova and Budanov, 1999; Gao and Klemd, 2003; Ai et al., 2006; Volkova and Sklyarov, 2007; Volkova et al., 2009). The blueschist mineral assemblages include Na–Ca amphiboles (glaucophane, lawsonite), epidote, clinozoisite, chlorite, albite, pumpellyite, calcite, quartz, stilpnomelane and phengite/muscovite plus accessory rutile, titanite and magnetite. Blueschists typically occur at the base of accreted OPS or in lower units of accretionary complex (Wakita, 2012) (Fig. 6D). Their presence in an accretionary complex suggests that a part of subducted OPS (basaltic) was separated from the descending slab and then quickly uplifted. Possible mechanisms of exhumation could be either the reverse currents in accretionary wedge caused by seamount – island arc or island arc – active margin collisions, or the release of large amounts of water due to strong dehydration of subducting serpentinite-rich slab (Volkova and Sklyarov, 2007).

The widespread Neoproterozoic to late Paleozoic tholeiite-boninite intra-oceanic arcs, Neoproterozoic to Mesozoic seamount OPS units and Paleozoic MORB-OIB-derived blueschist belts in the CAOB clearly indicate that the latter was initiated and evolved as a Pacific-type orogen (Buslov et al., 1998; Pfänder et al., 2002; Gordienko et al., 2007; Volkova and Sklyarov, 2007; Yarmolyuk et al., 2012; Safonova and Maruyama, 2014; Safonova and Santosh, 2014). Consequently, the CAOB consists of numerous smaller P-type orogenic belts with subordinate collision-type belts formed by local continent-microcontinent collisions, e.g., Kazakhstan and North Tianshan, Kazakhstan and Tarim, Siberia and Tuva-Mongolia, Siberia and Dzabkhan (Mossakovsky et al., 1993; Didenko et al., 1994; Fedorovskii et al., 1995; Kheraskova et al., 2003; Kozakov et al., 2005; Charvet et al., 2007; Konopelko et al., 2008; Alexeiev et al., 2009; Biske and Seltmann, 2010; Levashova et al., 2010; Alexeiev et al., 2011; Han et al., 2011; Zheng et al., 2013). Thus, the Central Asian Orogenic Belt was a major supplier of juvenile crust in Asia during the Phanerozoic.

4. Juvenile versus recycled crust in the CAOB

4.1. Methodological background

The juvenile vs recycled (reworked) origin of continental crust has been commonly evaluated using the methods of isotope geochemistry: whole-rock Nd (e.g., Jahn et al., 2000; Turkina et al., 2007; Jahn, 2010; Kruk et al., 2011; Long et al., 2011; Urmantseva et al., 2012) or Hf-in-zircon (e.g., Long et al., 2007; Belousova et al., 2010; Cai et al., 2011; Kröner et al., 2014; Chen et al., 2015). The advent, fast distribution and wide application of the Hf-in-zircon isotope methodology during the last 10 years allowed more reliable evaluation of the proportions of juvenile to recycled crust in older orogenic belts. Both methods are typically applied for magmatic and sedimentary rocks or for their hosted zircons, in particular, mafic rocks, granitoids and metasediments. Mafic rocks of supra-subduction complexes are studied by the whole-rock Nd method (if their ages are known) for differentiating intra-oceanic and active continental margin arcs, because the former are unambiguously juvenile as dominated by boninites, tholeiitic basalts and andesitic magmas with positive epsilon Nd values, whereas the latter may be formed by both juvenile igneous complexes consisting of tholeiitic to calc-alkaline basalts and andesites with positive epsilon Nd values and recycled magmas, i.e. shoshonitic, andesitic to felsic with negative positive epsilon Nd values. In addition, microcontinents,

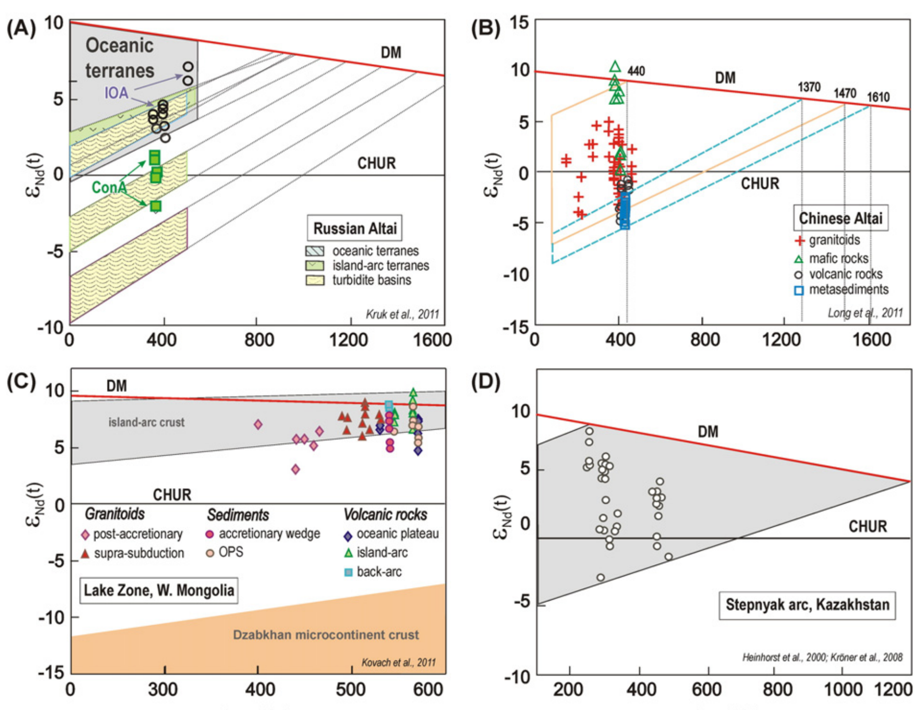

Fig. 7. Plots of whole-rock Nd isotope characteristics of different segments of the CAOB. A, late Neoproterozoic-Early Paleozoic mafic and granitic rocks of the Russian Altai (Kruk et al., 2011; Safonova et al., 2011a, 2011b; IOA, intra-oceanic arc; ConA, continental arc); B, Early–Middle Paleozoic granitoids, metasediments and volcanic rocks of the Chinese Altai (Long et al., 2011); C, Early Paleozoic granitoids, metasediments and volcanic rocks of the Lake Zone, western Mongolia (Kovach et al., 2011); D, Early Paleozoic of the Stepnyak island-arc terrane in northern Kazakhstan (Heinhorst et al., 2000; Kröner et al., 2008).

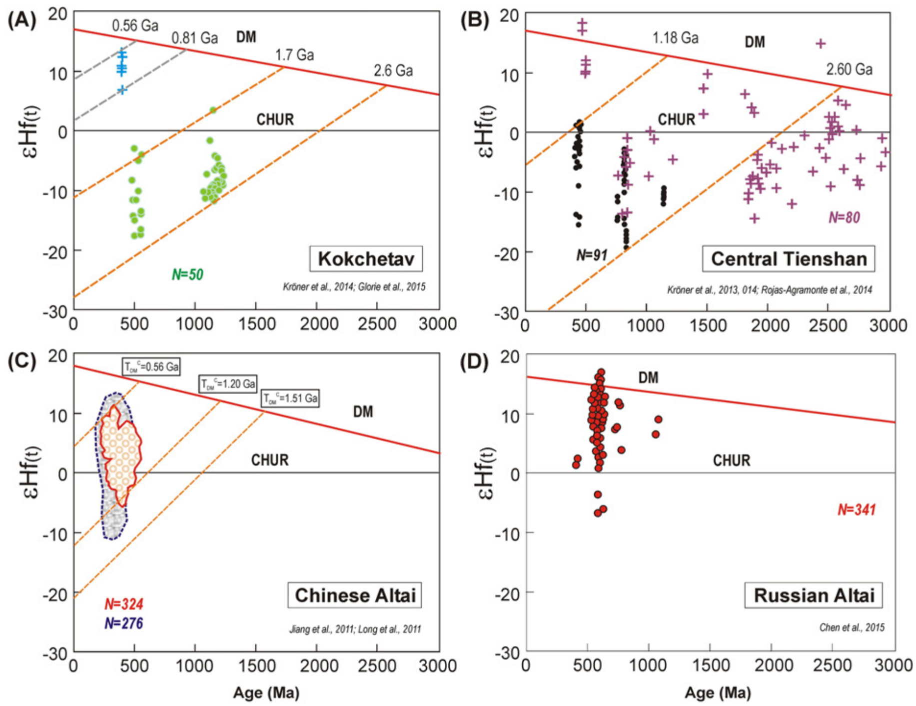

Fig. 8. Plots of Hf-in-zircon isotope characteristics of several CAOB segments. A, recycled crust of the Kokchetav microcontinent (granitoids in circles, Glorie et al., 2015) and juvenile crust of the Baidaulet-Akbastau intra-oceanic arc (granitoids in crosses, Kröner et al., 2014); B, dominantly recycled crust of the Central Tienshan microcontinent (granitoids, Kröner et al., 2013, 2014; metasediments, Rojas-Agramonte et al., 2014); C, juvenile, mixed and recycled segments of the Chinese Altai (metasediments, Jiang et al., 2011; Long et al., 2011); D, dominantly juvenile crust of the Russian Altai inferred from metasediments (Chen et al., 2015).

when approaching and colliding an active margin, may serve as sources of crustal crust, typically recycled, which can be involved into supra-subduction magma generation. Granitoids of active margin terranes can be analyzed by both methods (whole-rock Nd and Hf-in-zircon) also to differentiate intra-oceanic and continental margin magmatic arcs or to define a dominant isotopic type of magma source in the latter:

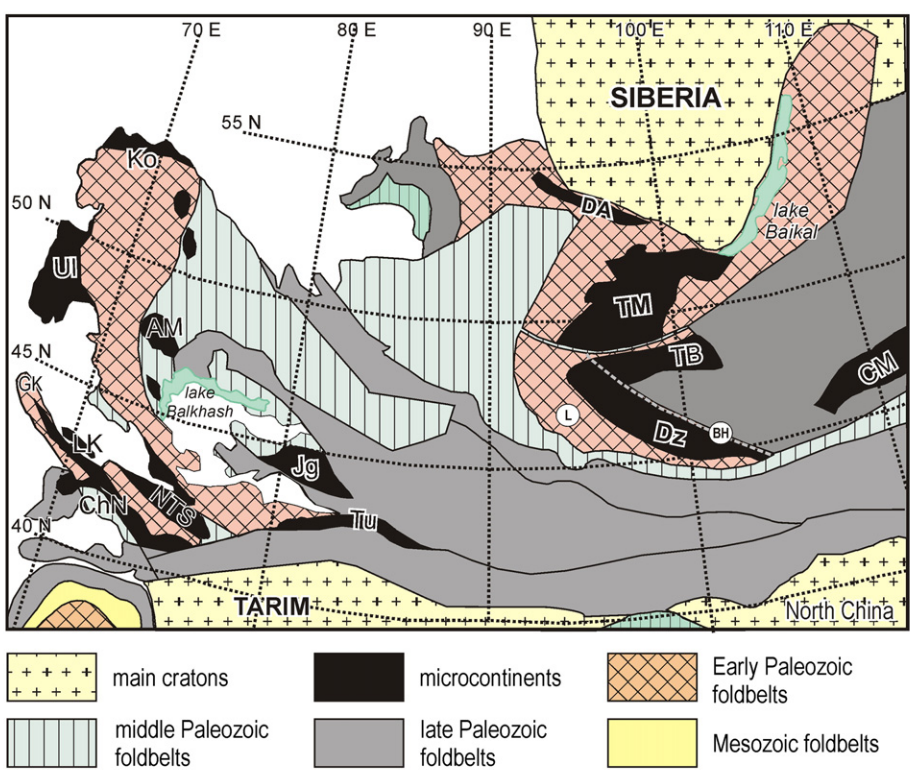

Fig. 9. Microcontinents in the western and central CAOB (modified from Levashova et al., 2010). AM, Aktau–Mointy; Dz, Dzabkhan (including Baydrag); CM, Central Mongolian; ChN, Chatkal–Naryn; DA, Derba–Arzybey; GK, Greater Karatau; Jg, Junggar; Ko, Kokchetav; LK, Lesser Karatau; NTS, North Tien Shan (also referred to as Central Tien Shan); TB, Tarbagatai; TM, Tuva–Mongolian; Tu, Turpan (Turphan, Turfan); UK, Ulutau. In circles: BH, Bayanhongor ophiolite belt; L, Lake island-arc terrane. The boundaries and subdivisions of foldbelts are shown schematically — for more details see Fig. 4 and related references.

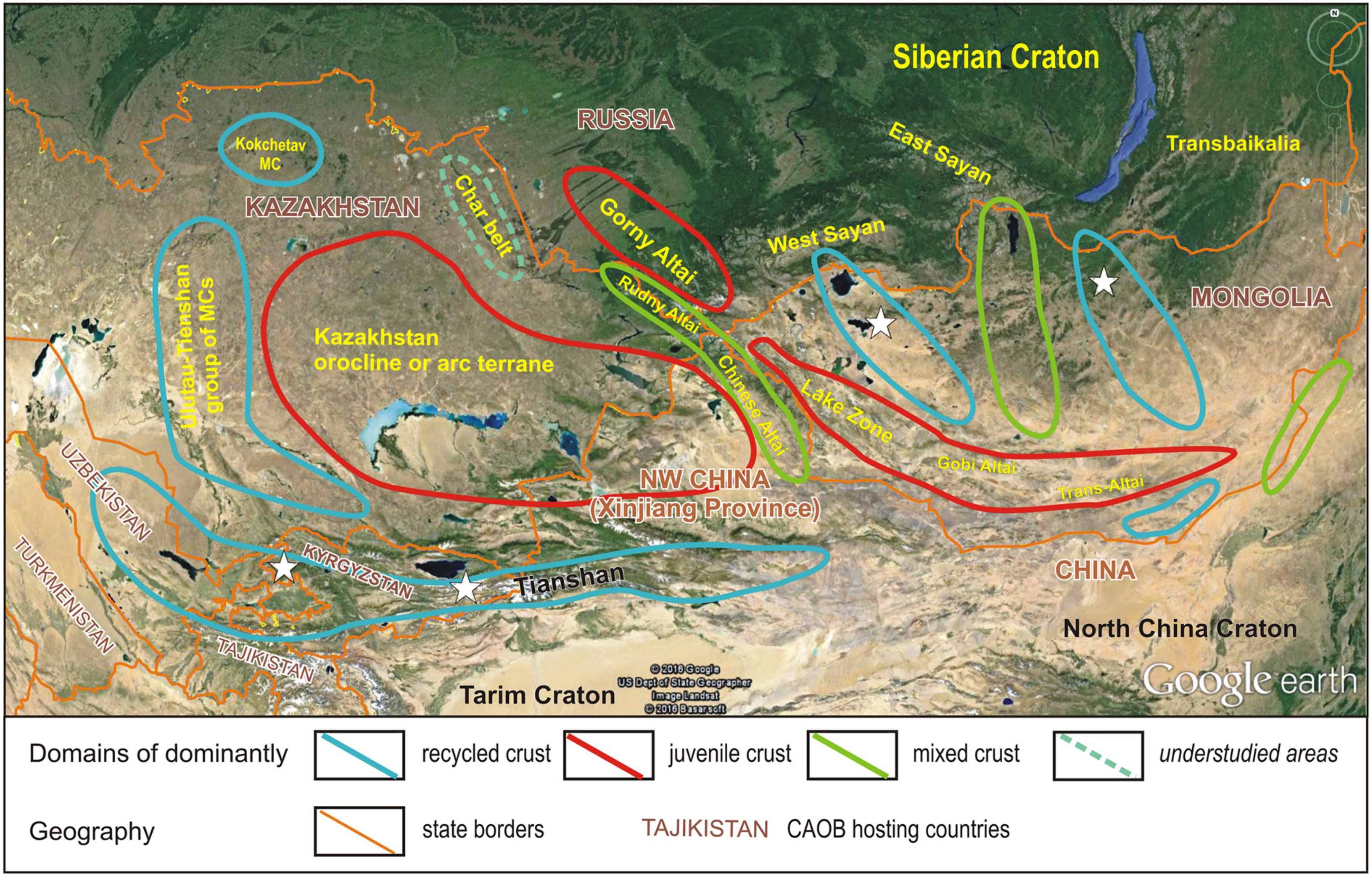

Fig. 10. The CAOB segments possessing isotope features of juvenile, mixed and recycled crust (based on Kröner et al., 2014, with modifications from Shen et al., 2015; Guy et al., 2015; R. Seltmann, pers. comm.). White stars mark an approximate location of the accretionary complexes with OPS-hosting OIB and OIB-MORB-derived blueschists, suggesting Pacific-type (accretionary) orogeny and, therefore, dominantly juvenile crust, but within the "recycled segments". The original fields are for the geology and Nd-isotope based juvenile crust areas of the Russian Altai (Gorny and Rudny Altai) and eastern Kazakhstan (dashed line). The eastern Kazakhstan area bears signs of tectonic erosion (see Section 5 and Fig. 14).

juvenile, mixed or recycled. Metasedimentary basins also can form in various tectonic settings: intra-oceanic arcs (greywacke turbidite basins), active continental margin or continental arcs (felsic to mafic turbidites) and microcontinents. Island-arc turbidite basins can be reliably identified by dominantly mafic composition and Nd isotopes (e.g., Kruk et al., 2011; Long et al., 2011) (Fig. 7). Continental arc terranes and related metasediments can be both juvenile or mixed, whereas old microcontinental terranes typically possess recycled isotope characteristics. Therefore, the differentiation of microcontinents from continental arcs needs the analysis of the Hf isotope ratios in detrital zircons of metasedimentary basins with an emphasis on the proportions of juvenile to recycled crust (e.g., Turkina et al., 2007; Sun et al., 2008; Jiang et al., 2011; Kruk et al., 2011; Salnikova et al., 2013) (Fig. 8). A special issue is the idea of mixed reservoirs of magma generation, which has been used to explain the isotopic heterogeneity of some magmatic complexes (Jahn, 2010; Cai et al., 2011; Kruk et al., 2011; Konopelko et al., 2013; Wang et al., 2013, 2014). However, some researchers think that the whole-rock Nd isotope systematics does not reflect source heterogeneity and that such a heterogeneity of Hf-in-zircon isotope characteristics may originate from initially mixed magma sources (Belousova et al., 2011; Kröner et al., 2014). At the same time, it is hard to prove whether those isotopically mixed magma sources and their derived melts can exist in principle. Of course, best results could be obtained by combining the Hf-in-zircon with whole-rock Nd isotopic systematics, but this is not always possible, as all mafic and metamorphic rocks may contain few, if not zero, zircons and such a "combined" approach would be very expensive and not always necessary or justified.

4.2. Whole-rock Nd vs Hf-in-zircon data

The analysis of Nd and Hf isotope ratios in rocks and zircons of the CAOB is very important because the belt has a very complex history since formed by multiple episodes of oceanic subduction and accretion, collision and amalgamation of arcs, microcontinents and continents (see Section 3). They produced huge amounts of mafic to felsic magmas and sedimentary basins possessing both juvenile and mixed/recycled isotope characteristics. The Nd isotope data obtained from CAOB granitoids since the 2000s allowed researchers to introduce and confirm an idea that the CAOB was the most important site of juvenile crustal formation since the Neoproterozoic and to develop a general scenario of massive juvenile crust production in the CAOB with limited contribution of old microcontinents to the genesis of Phanerozoic granitoids (e.g., Chen et al., 2000; Jahn et al., 2000; Jahn, 2004; Jahn et al., 2004; Kovalenko et al., 2004; Yuan et al., 2007; Kröner et al., 2008; Yarmolyuk et al., 2012). Mafic rocks have been studied to a much lesser degree (e.g., Kovach et al., 2011; Safonova et al., 2011a, 2011b, 2012, 2016a). However, the previous geology- and geochemistry-based recognition of Precambrian terranes in the CAOB (e.g., Buslov et al., 2001; Dergunov et al., 2001; Kuzmichev et al., 2001; Salnikova et al., 2001; Badarch et al., 2002; Kozakov et al., 2005; Li et al., 2006), which has been supported by the recent Hf-in-zircon isotope data from the CAOB (e.g., Sun et al., 2008; Belousova et al., 2010; Kröner et al., 2012a, 2013) coupled with re-assessment of Nd isotope data (e.g., Rytsk et al., 2011) showed that the CAOB continental crust includes from 50 to 80% of recycled or re-worked crust (Condie and Kröner, 2013; Kröner et al., 2014 and references therein). This is typically explained by the presence of microcontinents (Figs. 9A, B; 10) or by the mixed sources of primary magmas (Fig. 7A, B; 8C).

The presence of numerous microcontinents in the CAOB is reflected in the whole-rock-Nd and Hf-in-zircon data obtained, for example, from the Barguzin terrane of the Baikal-Muya belt (Rytsk et al., 2011) and the Angara-Kan terrane of the Yenisey Range (Turkina et al., 2007) at the southern and south-western margins of the Siberian Craton, respectively, from the Kazakhstan terranes (Kröner et al., 2008, 2012a, 2012b, 2013) and from the Tuva-Mongolian, Dzabkhan and Tarbagatai terranes in Mongolia (e.g., Kozakov et al., 2005, 2007, 2011) (Fig. 9). Evidence for mixed magma sources comes from many recent Hf-in-zircon data from CAOB granitoids (e.g., Sun et al., 2008; Jahn, 2010; Cai et al., 2011; Konopelko et al., 2013; Kröner et al., 2013; Wang et al., 2013; Kröner et al., 2014; Wang et al., 2014).

The choice of suitable objectives for the Nd bulk rock and Hf-in-zircon isotope studies is important as illustrated by the Nd and Hf data from the Neoproterozoic to early Paleozoic magmatic arcs in the Russian and Chinese Altai, western Mongolia and northern Kazakhstan in the western CAOB (Figs. 7, 8). The juvenile Nd bulk rock characteristics were obtained from granitoids, mafic volcanics of intra-oceanic arcs and oceanic islands as well as from their related turbidite basins in the

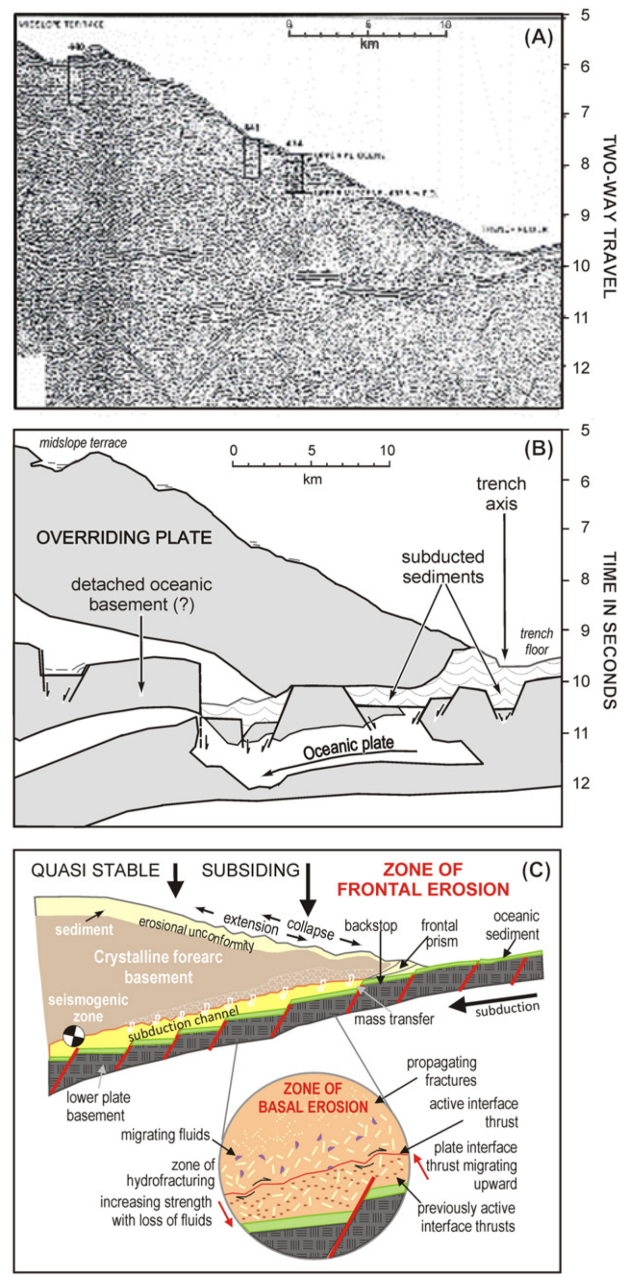

Fig. 11. Tectonic erosion at Pacific-type convergent margins. A and B, multi-channel seismic reflection profile plate (A) and its interpretation (B) across the axis and lower slope of the Japan trench along the 35°45 N′ latitude, showing the subducted oceanic slab (modified from Hilde, 1983). The data for the profile were collected by the Japan National Oil Corporation for the IPOD Japan Trench transect (Shipboard Scientific Party, 1980). C, subduction-erosion model indicating areas of most severe fracturing and dragging of dislodged fragments into subduction channel (from von Huene et al., 2004).

Lake zone in Mongolia (Kovach et al., 2011; Fig. 8A) and Chinese and Russian Altai (Kruk et al., 2011; Long et al., 2011; Safonova et al., 2011a, 2011b) (Fig. 7B, C). Active continental margin granitoids of the Russian and Chinese Altai have positive to negative εNd and εHf values (Figs. 7B, C, 8C, D) (e.g., Sun et al., 2008; Cai et al., 2011; Jiang et al., 2011; Kruk et al., 2011; Long et al., 2011; Chen et al., 2015). The thick metasedimentary units of the southern Russian and Chinese Altai have been previously interpreted as Altai-Mongolian microcontinent (e.g., Hu et al., 2000; Buslov et al., 2001; Li et al., 2006; Glorie et al., 2011; Safonova, 2014), but their recent Hf-in-zircon study clearly showed they have mixed isotope characteristics implying rather an active continental margin (Sun et al., 2008; Cai et al., 2011; Jiang et al., 2011; Chen et al., 2015). Thus, whole rock Nd isotopes are sufficient for studying intra-oceanic arcs with boninites, blueschist and OPS, because they match the geological and lithological data suggesting formation of juvenile crust (see Section 2). The Hf isotopes can be used for the study of metasedimentary units and granitoids of unclear origin, for example, microcontinent vs active margin, intra-oceanic vs continental arc, active margin vs post-collisional granitoids. The Chinese Altai, representing a granitoid-rich terrane, appeared to be an Early Paleozoic continental magmatic arc based on both whole-rock Nd (Wang et al., 2009; Long et al., 2011; Fig. 7B) and Hf-in-zircon data (Sun et al., 2008; Cai et al., 2011; Jiang et al., 2011) (Fig. 8C). On the contrary, the Russian Altai includes boninite-bearing IOA and accretionary wedge with seamount OPS terranes, blueschists and turbidite basins with positive εNd values (Buslov et al., 1998; Safonova, 2008, 2009; Kruk et al., 2011) (Fig. 7C). Consequently, there was no need to analyze Hf-in-zircon isotopes (Chen et al., 2015) for understanding the nature of crust in that area.

Another example of disagreement between Hf-in-zircon and Nd isotope data comes from the Gissar terrane of the South Tien Shan and the Chatkal-Kurama terrane of the Middle Tien Shan in Uzbekistan. Carboniferous granitoids from these terranes show mostly negative εNd values, but dominantly positive εHf values (Dolgopolova et al., 2017-in this issue). In addition, the problem may be due to the gap between the DM and CHUR lines, which becomes greater as the rocks get younger.

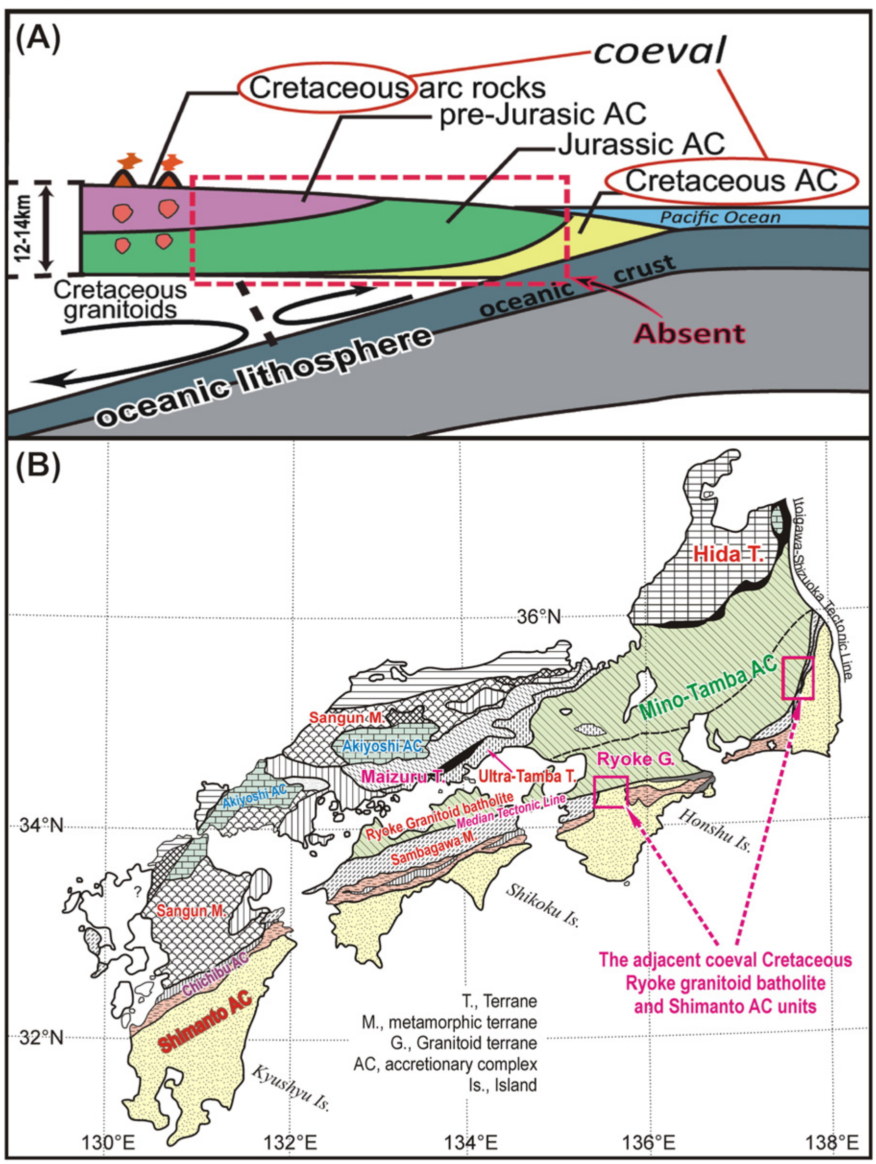

Fig. 12. Examples of tectonic erosion and arc subduction from Japan. A, a typical succession of accretionary complexes of different ages reconstructed across the Shikoku island in southwestern Japan, showing that coeval arc granitoids and accretionary units must be separated by older accretionary wedges (from Safonova et al., 2015a, based on Nakajima, 1994); B, areas of probable tectonic erosion in Japan (outlined by red rectangles): there, the Cretaceous Ryoke granitoids are spatially adjacent to the coeval units of the Shimanto accretionary complex (from Isozaki et al., 2010; Safonova et al., 2015b).

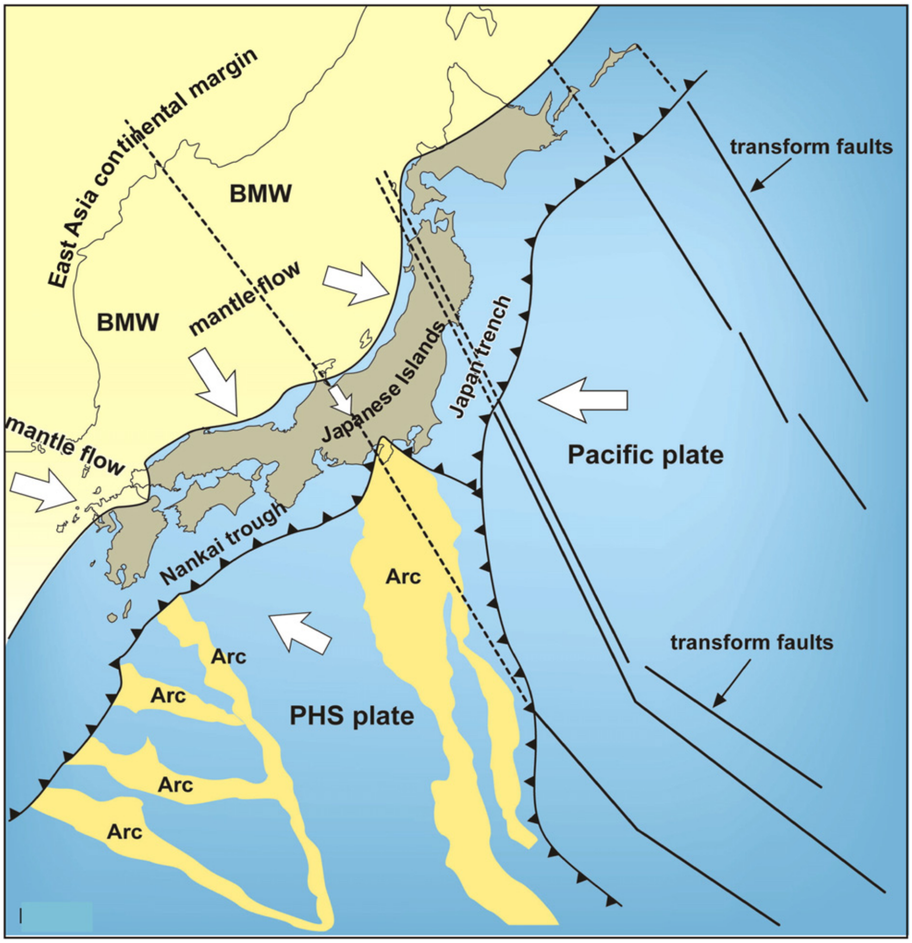

Fig. 13. Intra-oceanic arcs directly subducting into the Nankai trough with minor or nil accretion over the subduction zone (modified from Yamamoto et al., 2009).

Therefore, we cannot expect the same "juvenile vs recycled proportions" if inferred from epsilon Hf and epsilon Nd data. Anyway, the reasons for those abnormal isotope behaviors remain controversial and require still focused study.

The "mixing" source model for the origin of the CAOB crust was mentioned in connection with the Kokchetav-Tienshan (or Kokchetav-Ishim) "superterrane" and Bochshakol-Chingiz arc in the western CAOB (Belousova et al., 2011; Degtyarev, 2011; Kröner et al., 2012b). However, this "superterrane" actually consists of geologically different terranes: the Stepnyak arc and the Kokchetav microcontinent in the north, Baidaulet-Akbastau arc in the center and the North Tienshan and Aktau-Junggar (or Aqtau-Mointy) microcontinents and the Middle Ordovician Kentash arc in the south (Fig. 9) (Bakirov and Maksumova, 2001; Obut et al., 2006; Windley et al., 2007; Biske and Seltmann, 2010; Degtyarev et al., 2012). The granitoids of the "superterrane" yielded variable, i.e., positive to negative, Hf isotope ratios, which were interpreted as mixing between juvenile magmas and older crustal protoliths (Belousova et al., 2011; Kröner et al., 2014), but the geological position, age and rock associations of the Kentash arc suggest its intra-oceanic origin (Degtyarev et al., 2012). The Stepnyak arc formed on a sialic basement, but lithologically its intrusions are dominated by TTG-type granitoids (tonalite and granodiorite) and showing mostly positive Nd isotopes (Kröner et al., 2008; Fig. 7D). The Baidaulet-Akbastau arc also displays juvenile Hf isotope characteristics (Kröner et al., 2014; Fig. 8A). The Kokchetav and North Tienshan microcontinents are dominated by granitoids with recycled isotope characteristics (Kröner et al., 2013; Glorie et al., 2015) (Fig. 8A, B). However, the metasediments from spatially distinct terranes of the northern Tianshan yielded both juvenile and recycled Hf isotope characteristics (Rojas-Agramonte et al., 2014; Fig. 8B) also suggesting a "multi-terrane" provenance or mixed results as well, which may require reconsideration against more careful analysis of tectonic units. Therefore, at least two explanations are possible for the wide range of Hf isotope ratios: (i) mixed juvenile and recycled magma protoliths; and (ii) sampling of both juvenile and recycled terranes from a tectonically composite superterrane (Figs. 2, 4; 8A). This particular area is considered in details in Degtyarev et al. (in this issue).

4.3. Isotopes vs geology

Why do the Hf isotope data show a big portion of recycled crust in the CAOB in spite of the presence of numerous intra-oceanic arcs, MORB/OIB protolith blueschist and accreted seamount OPS (Table 1; Section 3; Figs. 4, 5, 7, 8)? Fig. 10 shows the CAOB segments of predominantly juvenile, mixed and recycled (reworked) crust, which were recognized by Kröner and co-authors based on Hf-in-zircon and whole-rock Nd isotope data (Kröner et al., 2014 and references therein). In addition, it shows Russian Altai, whose juvenile nature was proved many years ago by geological data (Simonov et al., 1994; Buslov et al., 1998, 2001, 2002) and by more recent Nd isotope data (Kruk et al., 2011; Safonova et al., 2011a, 2011b). However, several segments of re-cycled crust in the Tienshan orogen, northwestern and northern Mongolia (Fig. 10) include accretionary complexes with blueschist belts derived from oceanic basalts (OIB and MORB) and accreted oceanic plate stratigraphy units with OIBs (e.g., Volkova and Budanov, 1999; Ai et al., 2006; Volkova and Sklyarov, 2007; Volkova et al., 2009; Ruppen et al., 2014; Safonova and Santosh, 2014; Safonova et al.,

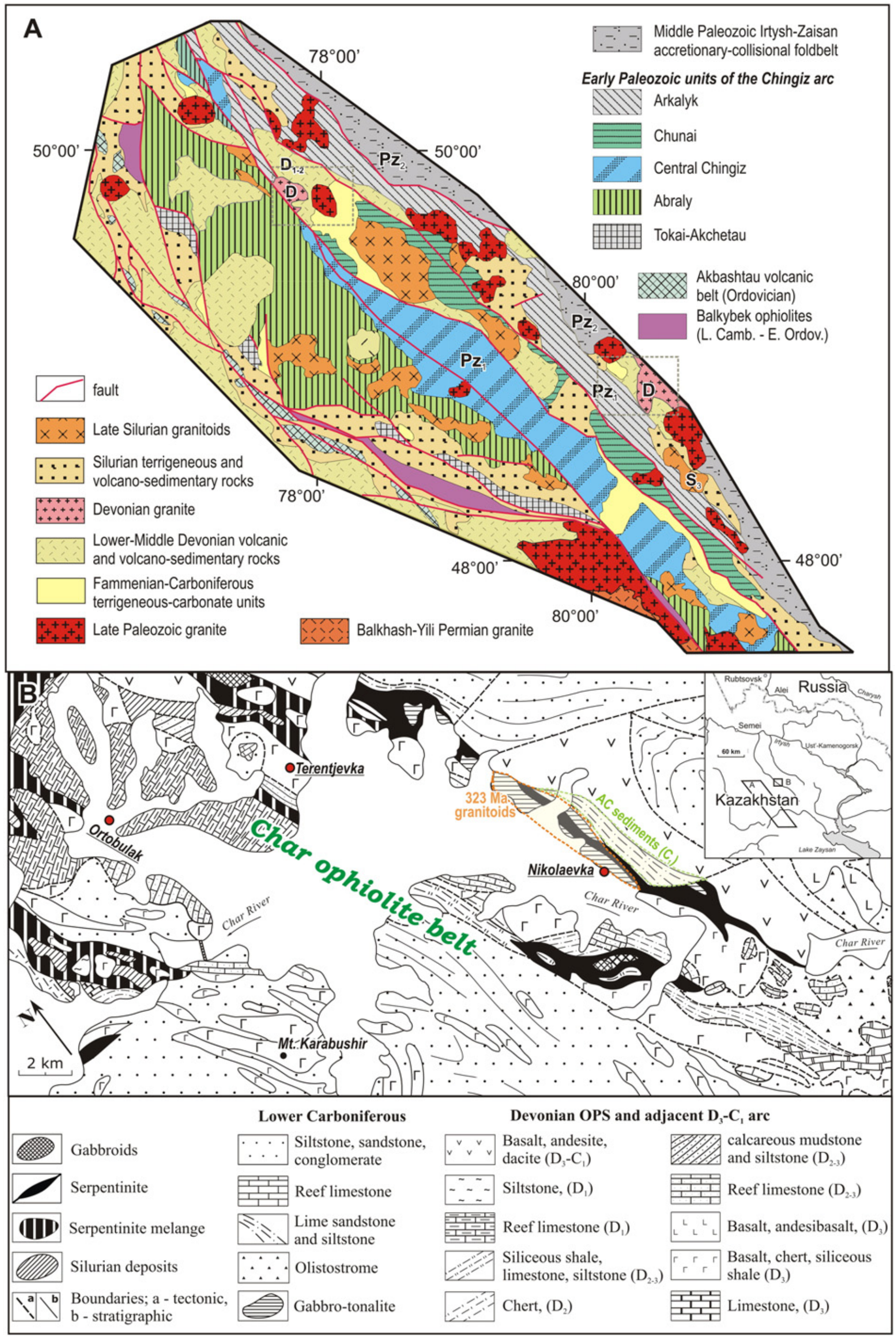

Fig. 14. Areas of probable tectonic erosion in the western CAOB showing close location of coeval accretionary (terrigeneous-volcanogenic) units and arc granitoids (dashed outlines). A, the Chingiz arc in central-eastern Kazakhstan (modified from Degtyarev, 2011); B, the Char belt of eastern Kazakhstan (modified from Polyanskii et al., 1979; Safonova et al., 2012).

2012, 2016a). More specifically, the juvenile nature of the Altai intra-oceanic arc (Kuznetsk-Altai arc in Russian literature, Buslov et al., 2001) is confirmed by the presence of boninites, OIB/MORB-protolith of blueschists and accreted seamount OPS (Buslov et al., 1998, 2002; Volkova et al., 2005; Safonova, 2008, 2011a, 2011b). Therefore, there is an objective disagreement between geological and isotopic data.

From the above examples one can infer that while studying by isotope methods the nature of the CAOB crust, juvenile versus recycled, some researchers might have worked with biased concepts or with biased datasets, for example, those dominated by granitoids (e.g., Jahn et al., 2000; Jahn, 2004), although the CAOB crust includes many metasedimentary domains or those dominated by mafic rocks, both accreted, post-accretionary and/or post-collisional (intra-plate). Such a disproportional sampling or ignorance of gabbro, basalt and metabasic rocks, which are typical parts of OPS, accretionary, intra-oceanic arc and intra-plate igneous and metamorphic complexes, often occupying huge territories, resulted in pseudo-correlations due to the artificial sampling gap or bias. Data obtained from different terranes, for example, from island arcs and microcontinents, should not be merged; otherwise we come to conclusions of "mixed source of magma", whereas it can be just a product of an artificial mix-up of juvenile and recycled characteristics. All these "biases" logically would lead to wrong or incomplete conclusions, actually representing a half of the truth. Consequently, while sampling the rock units and planning their investigation all researchers working in those structurally and historically complex orogenic belts must be absolutely sure what type of terrane they sample and choose the methods that better suit the tasks and problems they are going correspondingly to fulfill and to solve. “Geology must be brought back into equation” (White et al., 2013).

5. Tectonic erosion of juvenile crust and arc subduction: evidence from the CAOB

In previous sections I emphasized several disagreements between Hf-in-zircon and whole-rock Nd isotope data and between isotope and geological data. First, the Hf-in-zircon isotope data show the dominantly recycled crust (see Section 4) although the whole-rock Nd isotope data suggest its dominantly juvenile character (Jahn et al., 2000; Jahn, 2004). Second, the isotope data from several segments of the CAOB crust suggest their recycled crust (Fig. 10) and their dominant Pacific-type orogenic character implying juvenile crust as inferred from geological data (Figs. 2-4; Table 1; Section 3). In addition to the presence of microcontinents in the CAOB crust and the mixed characters of magma reservoirs of most granitoids, these disagreements can be explained by the tectonic erosion of juvenile arc crust and accreted oceanic terranes (OPS, OIB, MORB, OPB). Although, the Pacific-type belts are sites of continental growth, they are also places of strong plate interactions accompanied by tectonic erosion and arc subduction (Scholl and von Huene, 2007; Clift et al., 2009; Yamamoto et al., 2009; Stern and Scholl, 2010; Safonova et al., 2015a, 2015b). Tectonic erosion of juvenile crust and arc subduction at the convergent margins of the Paleo-Asian Ocean was quite probable. The question is whether the big portion of recycled crust in some segments of the CAOB can be due to tectonic erosion?

The first evidence for tectonic erosion at P-type active margins was obtained from seismic reflection profiles across the Tonga (Hilde and Fisher, 1979) and Japan trenches (von Huene and Uyeda, 1981; Hilde, 1983), showing prominent sediment-filled grabens and horsts with thinned sediments (Fig. 11A, B). The mechanism of tectonic erosion includes destruction of oceanic slab, island arcs, accretionary prism and fore-arc by thrusting, the prominent oceanic floor relief as horsts, grabens and oceanic rises scratching the hanging wall, and by (hydro)fracturing (von Huene et al., 2004; Yamamoto et al., 2009) (Fig. 11C). At subduction-related convergent margins, the TTG-type crust and material of accretionary wedges can be eroded and subducted with oceanic slab into the deep upper mantle. Evidence for this comes from the Cretaceous Shimanto accretionary complex of Shikoku Island in Japan, where accretionary units are spatially adjacent to the coeval granitoids of the Ryoke belt, suggesting that older accretionary complexes have been eroded (Fig. 12; Safonova et al., 2015a, 2015b, 2016b), and from the present-day tectonic erosion and ongoing subduction of several oceanic arcs of the Philippine Plate under the Japanese Arc (Fig. 13).

The modern Pacific Ocean is surrounded by 75% of eroding (narrowing or non-accreting) convergent margins and 25% of growing or accreting margins. The global long-term rate of subduction-related erosion is much greater than that of crustal additions (Senshu et al., 2009; Stern, 2011). As the present Western Pacific is a most probable analog of the CAOB (Safonova et al., 2011), one can suggest that processes of tectonic erosion also could have been active at the consuming margins of the Paleo-Asian Ocean, which was responsible for formation of the CAOB.

There are several areas in the CAOB, which host intra-oceanic arcs and/or accreted seamount OPS, and/or blueschists formed after MORB/OIB, on the one hand, and the isotopically-proven recycled crust (Kröner et al., 2014), on the other hand: Tienshan, northwestern and northern Mongolia (Fig. 10). In addition, eastern Kazakhstan remains poorly studied in terms of Hf-in-zircon isotopes in granitoids as so far whole rock Nd isotope data have been obtained for Late Devonian oceanic basalts only (Safonova et al., 2012).

In terms of geology, a good indication of probable cases of tectonic erosion at old Pacific-type convergent margins is the close spatial location or structural juxtaposition of coeval accretionary units and TTG-type granitoids (andesites, tonalites, granodiorites), suggesting that older accretionary and/or convergent margin units were eroded (Safonova et al., 2015a, 2015b). Several probable cases of tectonic erosion can be proposed in the western CAOB, more specifically, in the middle to late Paleozoic convergent margin units of eastern and southeastern Kazakhstan and southern Kyrgyz Tienshan. For example, the Devonian terrigeneous-carbonate and volcano-sedimentary continental margin units, superimposed onto the eastern part of the Chingiz arc terrane are, in places, adjacent to Devonian granites of probably arc origin (Degtyarev, 2011) (Fig. 14A). Second, the Chatkal-Atbashi Ordovician arc in the Kyrgyz Middle Tienshan includes well dated coeval Early Devonian arc granitoids, ophiolites and accretionary units (Alexeiev et al., 2016) and therefore represents a relatively well-proven locality of tectonic erosion. Third, the Char accretionary complex in eastern Kazakhstan hosting late Devonian-early Carboniferous OPS units (Safonova et al., 2012) includes Early Carboniferous limy sandstones and siltstones, spatially adjacent to Early Carboniferous tonalites (Polyanskii et al., 1979; Kuibida et al., 2016) (Fig. 14B). Consequently, in addition to the Chatkal-Atbashi arc in the Kyrgyz Tienshan (Alexeiev et al., 2016), these three areas (Chingiz, Kyrgyz South Tienshan, Char) are best “candidates” for the cases of probable tectonic erosion at active margins of the Paleo-Asian Ocean, but both detailed geological survey and precise dating of key lithologies are necessary to prove it.

There are also several indirect evidences for the middle-late Paleozoic tectonic erosion in the western CAOB, in general, and Kyrgyz Tienshan, in particular. First, the number of middle Paleozoic accreted OPS with OIB in the Kyrgyz Tienshan is smaller than in the other regions of the CAOB (Fig. 4). There are several localities of accreted OPS-OIB units in the Kyrgyz Tienshan (Safonova et al., 2016a), but they are notably less abundant than those of the same age in the adjacent western Junggar (G. Yang et al., 2015). Second, the amount and size of accreted middle Paleozoic OIB-bearing OPS units in the whole CAOB are, in general, much less than those of late Neoproterozoic, early and late Paleozoic ages (see Tables 2, 3, Section 3.3 and Safonova and Santosh, 2014). Third, the amount of middle Paleozoic intra-oceanic arcs is also, in general, smaller than that of late Neoproterozoic and early Paleozoic (Fig. 4, Table 2), which suggest direct arc subduction like in the modern north-western Pacific (Fig. 13). All these facts may indicate either lesser plume activity (Safonova and Santosh, 2014), or tectonic and/or subduction erosion of OPS-OIB and IOAs or direct arc subduction in middle Paleozoic time.

In spite of the dominantly Pacific-type style of the CAOB, its continental crust includes a big portion of recycled crust (Kröner et al., 2014 and references therein). This can be due to the presence of numerous microcontinents in the CAOB (Fig. 2) and due to destruction of juvenile crust formed at convergent margins. According to various evaluations, from 80 to 100% of the juvenile crust of initially Pacific-type orogenic belts, including those in the CAOB, could have been removed by later tectonic erosion and arc subduction (Clift et al., 2009; Yamamoto et al., 2009; Stern and Scholl, 2010; Ichikawa et al., 2013). The identification of cases of tectonic erosion is critical for the CAOB as Pacific-type convergent margins are places of probably most active and fast destruction of continental crust and they are the only ways to deliver the crustal material to the deep mantle and to metasomatize it (Litasov, 2011; Safonova et al., 2015a, 2015b). The eroded/subducted TTG-type juvenile continental crust could be either returned to the surface as supra-subduction granitoids with recycled isotope signatures or accumulated at the Mantle Transition Zone and served as a source of heat, which induced mantle upwelling, trigger plumes and surface rifting (Maruyama et al., 2011; Safonova et al., 2015a, 2015b). Thus, not only the presence of certain rocks or formations may prove this or that tectonic process or setting – their absence also can be a key for understanding how an accretionary orogen formed and evolved.

6. Conclusions

1. Accretionary or Pacific-type orogenic belts are major sites of continental growth through TTG-type juvenile granitoids magmatism, which takes place at intra-oceanic and continental margin arcs, and accretion of oceanic and island-arc terranes. Pacific-type belts can be recognized by the presence of: (i) dominantly mafic, in places boninite-bearing, intra-oceanic island arcs; (ii) accretionary wedges, hosting the Ocean Plate Stratigraphy units, in particular, those representing oceanic islands/seamounts/plateaus (OIB, OPB); (iii) blueschists formed after MORB, OIB and/or OPB; (iv) huge supra-subduction granitoid batholiths; and (v) paired metamorphic belts.
2. The Central Asian Orogenic Belt is the world's largest accretionary or P-type orogen and the major site of Phanerozoic juvenile crustal growth. Evidence for its juvenile character comes from numerous late Neoproterozoic to late Paleozoic accretionary complexes, hosting the OPS units with OIBs and blueschist belts, formed after MORB and OIB throughout the whole CAOB, and from the late Neoproterozoic to middle Paleozoic boninite-bearing intra-oceanic arcs in the Altai-Sayan area, western Mongolia, Transbaikalia and northern Mongolia.
3. However, in spite of the geology-based accretionary nature and positive whole rock Nd isotope characteristics in most of granitoid localities of the CAOB, the Hf-in-zircon isotope data show a big portion of recycled crust. Such a discrepancy can be explained by the presence of accreted microcontinents, the isotopically mixed nature of igneous reservoirs, the tectonic erosion of juvenile crust and/or by the biased rocks sampling or interpretation of their setting (e.g., focused on granitoids only). A very important problem, while using Hf-in-zircon methods, is the disregard or underestimation of primary geological information.
4. The Hf-in-zircon method works well for understanding a type of magma source (juvenile, mixed or recycled) for a certain granitoid complex. The Hf-in-zircon data may be helpful for evaluating the proportions of juvenile and recycled crust at localities lacking clear geological evidence for this or that type of orogen, e.g., in metasediments, or for large fields of intra-plate or/and post-orogenic granitoids. The whole-rock Nd isotope systematics works well for intra-oceanic arcs (mafic rocks and TTG-type granitoids) and/or their related turbidite basins and post-orogenic formations.
5. There are discrepancies between the Hf-in-zircon data, indicating dominantly recycled crust in the middle and southern Tienshan and southern Transbaikalia, and the geological data, showing the presence of intra-oceanic arcs, OIB/MORB derived blueschists and accreted oceanic islands in the same regions. In addition, there are tectonically juxtaposed coeval arc granitoids and accretionary units. This suggests that the Middle Paleozoic juvenile continental crust, which formed at convergent margins in Kazakhstan, Tarim and Siberia, underwent tectonic erosion.
6. Identification of Pacific-type orogens, as places of major crustal growth, requires a detailed geological study, because the isotopic results may not reflect many posterior processes, e.g., tectonic erosion of juvenile crust and/or arc subduction. Geology must be “put back into equation” as field and laboratory recognition of OPS units, blueschist belts and intra-oceanic arcs.

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