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GR Focus Review

Intra-oceanic arcs of the Paleo-Asian Ocean

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ABSTRACT

The paper reviews and integrates geological, geochronological, geochemical and isotope data from 21 intra-oceanic arcs (IOA) of the Paleo-Asian Ocean (PAO), which have been identified in the Central Asian Orogenic belt, the world largest accretionary orogeny. The data We discuss structural position of intra-oceanic arc volcanic rocks in association with back-arc terranes and accretionary complexes, major periods of intra-oceanic arc magmatism and related juvenile crustal growth, lithologies of island-arc terranes, geochemical features and typical ranges of Nd isotope values of volcanic rocks. Four groups of IOAs have been recognized: Neoproterozoic – early Cambrian, early Paleozoic, Middle Paleozoic and late Paleozoic. The Neoproterozoic – early Cambrian or Siberian Group includes eleven intra-oceanic arcs of eastern and western Tuva-Sayan (southern Siberia, Russia), northern and southwestern Mongolia and Russian Altai. The Early Paleozoic or Kazakhstan Group includes Selety-Urumbai, Bozshakol-Chingiz and Baydaulet-Aqastau arc terranes of the Kazakh Orocline. The Middle Paleozoic or Southern Group includes six arc terranes in the Tienshan orogen, Chinese Altai, East-Kazakhstan-West Junggar and southern Mongoia. Only one Late Paleozoic intra-oceanic arc has been reliably identified in the CAOB: Bogda in the Chinese Tienshan, probably due to PAO shrinking and termination. The lithologies of the modern and fossil arcs are similar, although the fossil arcs contain more calc-alkaline varieties suggesting either their more evolved character or different conditions of magma generation. Of special importance is identification of back-arc basins in old accretionary orogens, because boninites may be absent in both modern and fossil IOAs. The three typical scenarios of back-arc formation - active margin rifting, intra-oceanic arc rifting and fore-arc rifting were reconstructed in fossil intra-oceanic arcs. Some arcs might be tectonically eroded and/or directly subducted into the deep mantle. Therefore, the structural and compositional records of fossil intra-oceanic arcs in intracontinental orogens allow us to make only minimal estimations of their geometric length, life span, and crust thickness.

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1. Introduction: Pacific-type convergent margins and intra-oceanic arcs

Intra-oceanic arcs (IOAs) form at Pacific-type convergent margins, in the upper "stable" plate, when the subducting plate submerges to the depths of melting, i.e., to ca. 50–100 km. A typical IOA system, such as Mariana-Bonin and the Philippines Sea, consists of subduction zone, fore-arc region with accretionary prism, frontal or active arc, marginal basin with spreading center, and, in some cases, one or more remnant arcs and inactive marginal basin (Fig. 1). The IOAs are very important elements of Pacific-type convergent margins as they represent major sites of juvenile continental crust formation (e.g., Clift et al., 2003; Stern, 2010; Maruyama et al., 2011). Most of the modern IOAs are located around the Circum-Pacific (Fig. 2). A consensus exists that new or "juvenile" crust today is mostly generated above subduction zones with subordinary imputs from intra-continental rifts and/or plume-related volcanism, and from rifted or tranform margins. During the last hundreds of million years, the annual amount of juvenile continental crust generated at Pacific-type margins was estimated to be ca. 2.7 km^3 (Scholl and van Huene, 2007).

Juvenile IOA crust is formed by magmatism triggered by water-induced melting of mantle above subduction zones (e.g., Coats, 1962; Dimalanta et al., 2002). The melting of the mantle produces mostly mafic (basaltic or boninitic) magmas followed by intermediate or andesitic rocks compositionally close to the average bulk continental crust (Rudnick and Gao, 2003; Tatsumi et al., 2008). Accordingly, IOAs are typically dominated by basalt-andesite-(boninite) lavas and tonalite-diorite-granodiorite plutons (e.g., Gill, 1981; Kelemen et al., 2003; Osanai et al., 2006; Hacker et al., 2008). 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 high-degree melted boninites, tholeiitic basalts, and andesibasalts. In time and with distance away from the subduction zone, the magmatism proceeds at lower degrees of melting and produces calc-alkaline, alkaline and shoshonitic andesitic to felsic series, more enriched in incompatible elements (Fig. 3). The specific geochemical characteristics of subduction-related magmas are increased contents of silica and large-ion lithophyle elements (LILE; K, U, Sr and Pb) and low contents of high field strength elements (HFSE; Nb, Ta and Ti). The variations of trace elements in arc rocks are similar to those of bulk continental crust thus confirming the idea that the subduction-related magmatism contributes largely to continental crustal growth (Kelemen 1995; Rudnick and Gao, 2003). Thus, the formation of IOA crust at Pacific-type convergent margins includes two major processes: mantle wedge melting to generate mafic magmas and their further evolution to produce granitic crust and refractory residue, which typically sinks back to the mantle (Hawkesworth and Kemp 2006). Pacific-type margins are not only sites of new crust formation, but are also the most important sites of crust removal by sediment subduction and tectonic/subduction erosion (Stern and Scholl, 2010).

Therefore, identification of intra-oceanic arcs within fossil, currently intra-continental orogenic belts is of primary importance. Criteria for identificaiton of IOAs may differ though because during the last 2–3 decades researchers have been using different terms: ensimatic arc, primitive arc, and intra-oceanic arc. Term "IOA" is more modern that ensimatic arc, i.e. built over Mg—Fe rich oceanic crust, or "primitive arc", i.e. dominated by primitive Mg-rich magmas. In this paper we follow the definition of R. Stern (2010, p. 7):"...IOAs are constructed on thin, mostly mafic crust, and consequently these magmas are not as contaminated by easily fusible felsic crust as are magmas from Andean-type margins, which are built on much thicker and more felsic continental crust". In addition, a very important feature of IOA is the presence of "true" graywackes consisting of clasts of olivine, pyroxene and Ca-plagioclase, which may be absent though. Geologically, IOA must be surrounded by a back-arc basin and oceanic ophiolites, which also may no be present. Thus, term "ensimatic arc" is equal to IOA as it suggests arc formation over an oceanic crust. Term "immature arc" is still used by ophiolite researchers and typically means an early or boninitic stage of IOA development. There is also a special type of IOA, which rock units formed at an Andean-type active margin, i.e. over a mature crust, but later were separated from the continent by back-arc rifting, like the Japanese arc, which formerly was part of the East Asia active margin, but split off it in Miocene time (Khanchuk et al., 1989; Maruyama et al., 1997; Martynov et al., 2017). Those arcs typically consist of mature, geochemically recycled crustal basement and younger juvenile rocks (Moreno et al., 2016).

The major site of fossil IOAs in Asia, except for the modern western Pacific, is the Central Asian Orogenic Belt (CAOB), the world's largest accretionary orogenic belt. In this paper we consider the CAOB as equivalent to the Altaids of Sengör et al. (1993), i.e. excluding the Urals as there is no concensus on the relation between the Urals and the orogenic belts of Central Asia (Puchkov, 2003; Windley et al., 2007; Vladimirov et al., 2015; Yakubchuk, 2017). The CAOB formed by the suturing of the Paleo-Asian Ocean (PAO) and multi-stage collisions of the East European, Siberian, North China and Tarim cratons (Figs. 4, 5). Many scientists believe that the PAO was evolving similarly to the present-day Pacific (e.g., Zonenshain et al., 1990; Dobretsov et al., 1995; Buslov et al., 2002; Safonova et al., 2016a; Safonova, 2017). In fossil orogenic belts IOA units must be differentiated from the units of Ocean Plate

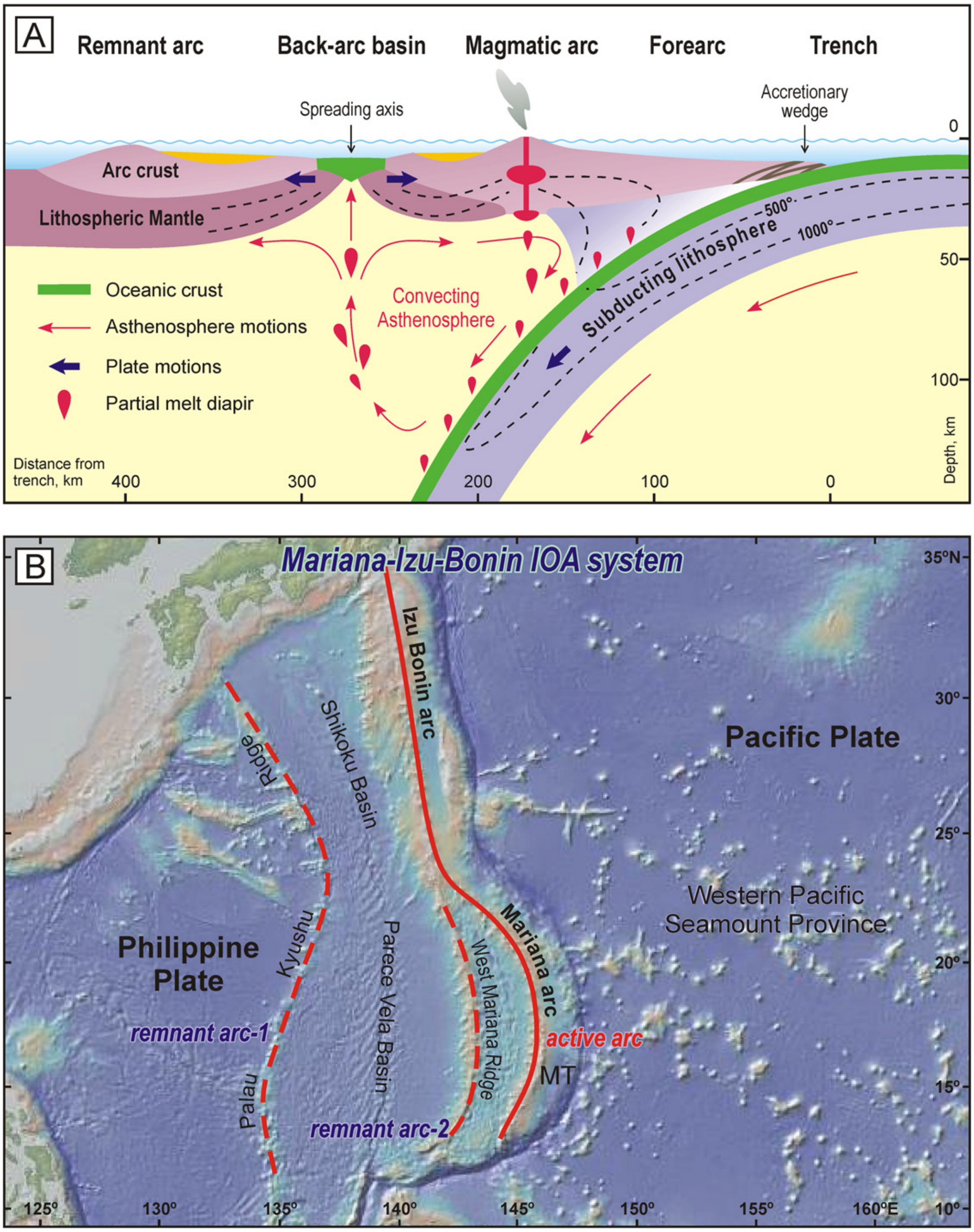

Fig. 1. A - schematic model of intra-oceanic arc system including a back-arc basin (modified from Stern, 2010). B - major topographical and tectonic features of the Izu-Bonin and Mariana arc systems (modified from Straub et al., 2015). The bold continuous red line indicates the active arc. Bold dashed lines denote remnant arcs: the Eocene protoarc (Palau–Kyushu Ridge) and the Miocene early arc (West Mariana Ridge). MT, Mariana Trough.

Stratigraphy (OPS), which consist of typical ophiolites (peridotite, serpentinite, gabbro), mid-oceanic ridge basalt (MORB) and overlapping sediments of oceanic floor (chert), shelf (siliceous shale/mudstone/siltstone), and trench (turbidites) plus OPS of seamounts (OIB, carbonate cap, volcanogenic-carbonate slope breccia and foothill limy and siliceous sediments) (Kusky et al., 2013; Safonova and Santosh, 2014; Safonova et al., 2016a, b, c). The present paper reviews and integrates main features of modern IOA systems (dataset of 76 analyses) and geological, geochronological and geochemical data from major island-arc terranes of CAOB (dataset of 150 analyses). Those arcs formed in the Paleo-Asian Ocean during a period from the late Neoproterozoic to the middle Paleozoic and their fragments have been preserved in the central, western and southwestern segments of the CAOB (Fig. 4).

2. Magmatism of modern IOAs

The magmas erupted at IOAs show a gradual evolution in composition with time, for example, the Mariana arc system, which is a world type locality of IOAs, was initiated in Eocene time (Figs. 1, 3A, B). The whole arc system evolved entirely within the oceanic environment, i.e. no continental crust or sub-continental lithosphere has been involved. In general, the Mariana arc is special for at least three distinct

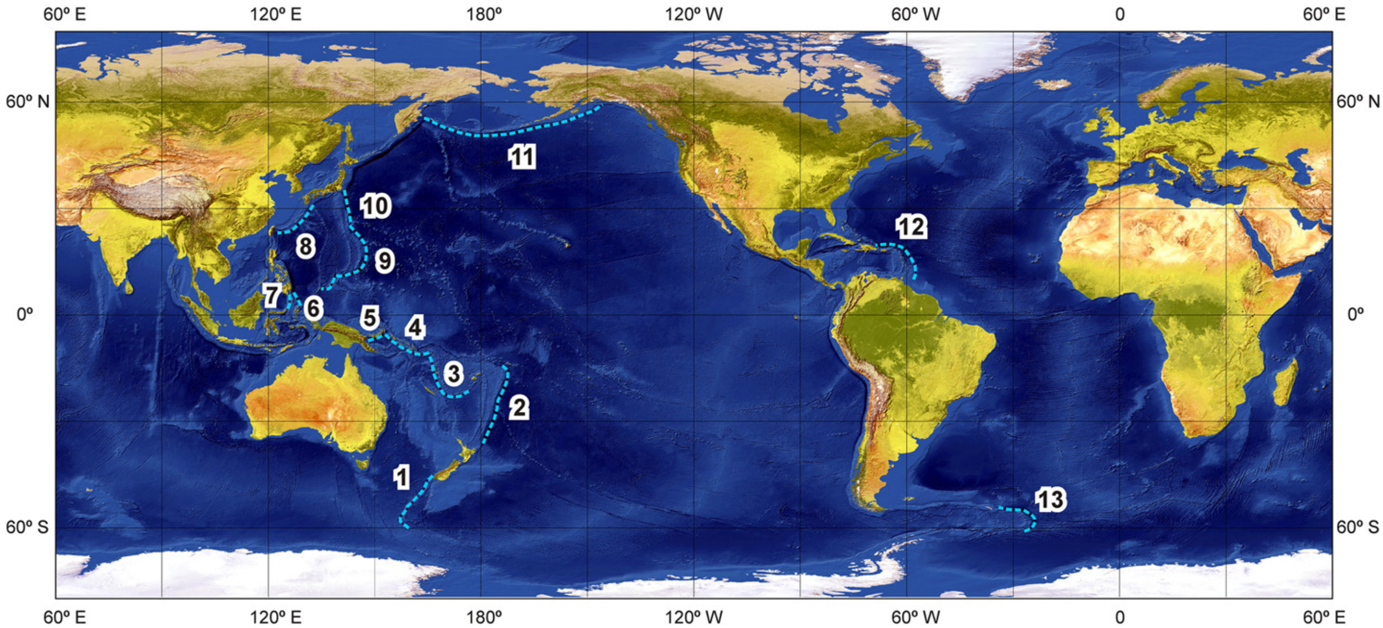

Fig. 2. Modern intra-oceanic arc systems of the world; blue dashed lines show the subduction zones producing these arcs (Leat and Larter, 2003): 1 – MacQuarie; 2 – Tonga-Kermadec; 3 – Vanuatu; 4 – Solomon; 5 – New Britain; 6 – Halmahara; 7 – Sangihe; 8 – Ryuku; 9 – Mariana; 10 – Izu-Bonin; 11 – Aleutian; 12 – Lesser Antilles; 13 – South Sandwich.

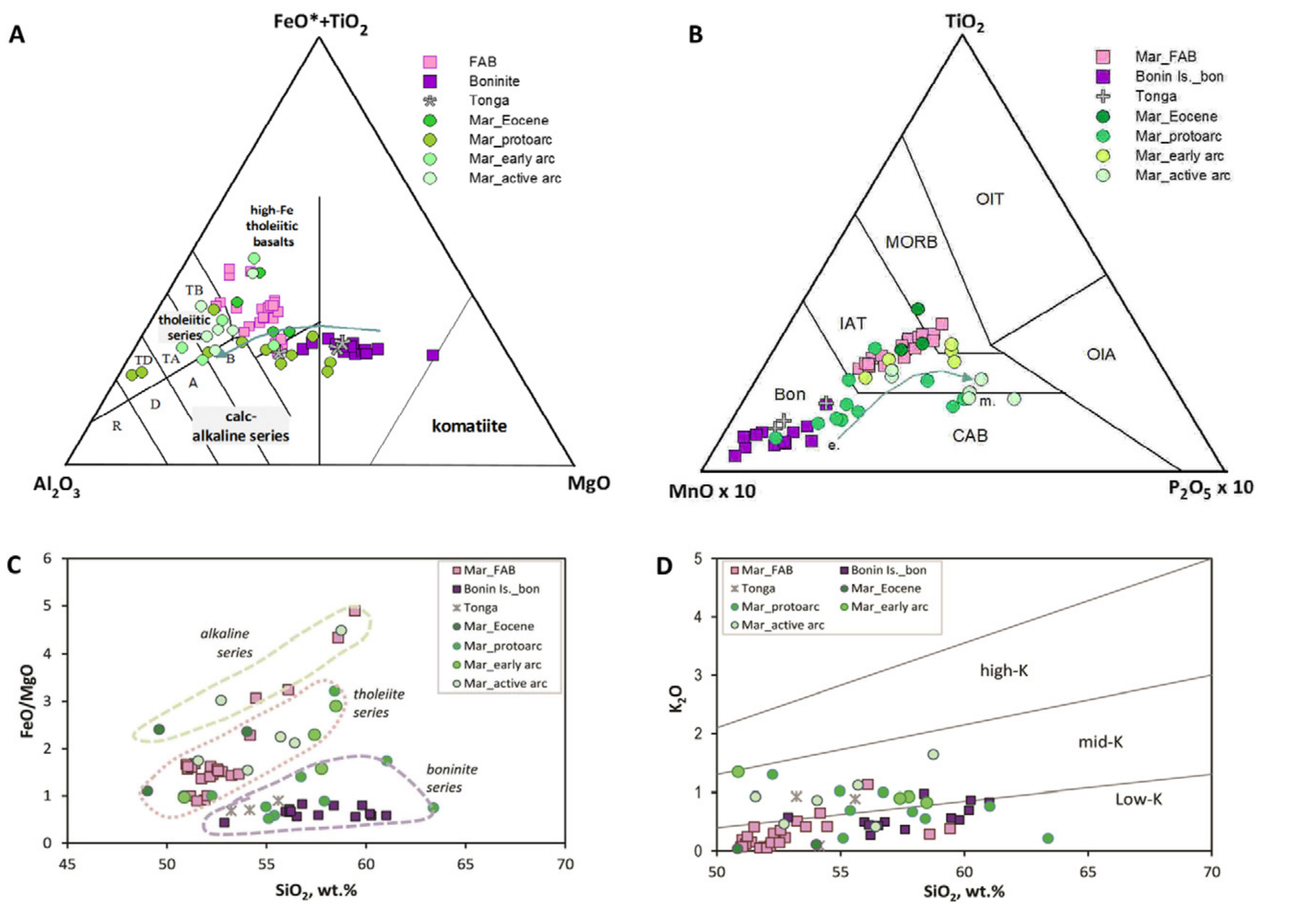

Fig. 3. Typical geochemical characteristics of modern IOA lavas. A - Al_2O_3 – FeO* + TiO_2 – MgO diagram (Jensen, 1976). Tholeiitic series: TA, andesite; TD, dacite; TR, rhyolite. Calc-alkaline series: CB, basalt; CA, andesite; CD, dacite; CR, rhyolite; kom – komatiite. B – tectonic discrimination diagram MnO-TiO_2-P_2O_5 (Mullen, 1983); CAB – calc-alkaline basalt; IAT, island-arc tholeiite; MORB, mid-oceanic ridge basalt; OIA, oceanic island alkaline basalt; OIT, oceanic island tholeiite; bon - boninite. C, D - Harker diagrams for FeO*/MgO, and K2O. Different magma series (C) are after (Reagan and Meijer, 1984). The low-, medium- and high-K subdivisions (D) are those of (Peccerillo and Taylor, 1976). Sources of data: Pearce et al., 1999; Dobson et al., 2006; Falloon et al., 2008; Reagan et al., 2010.

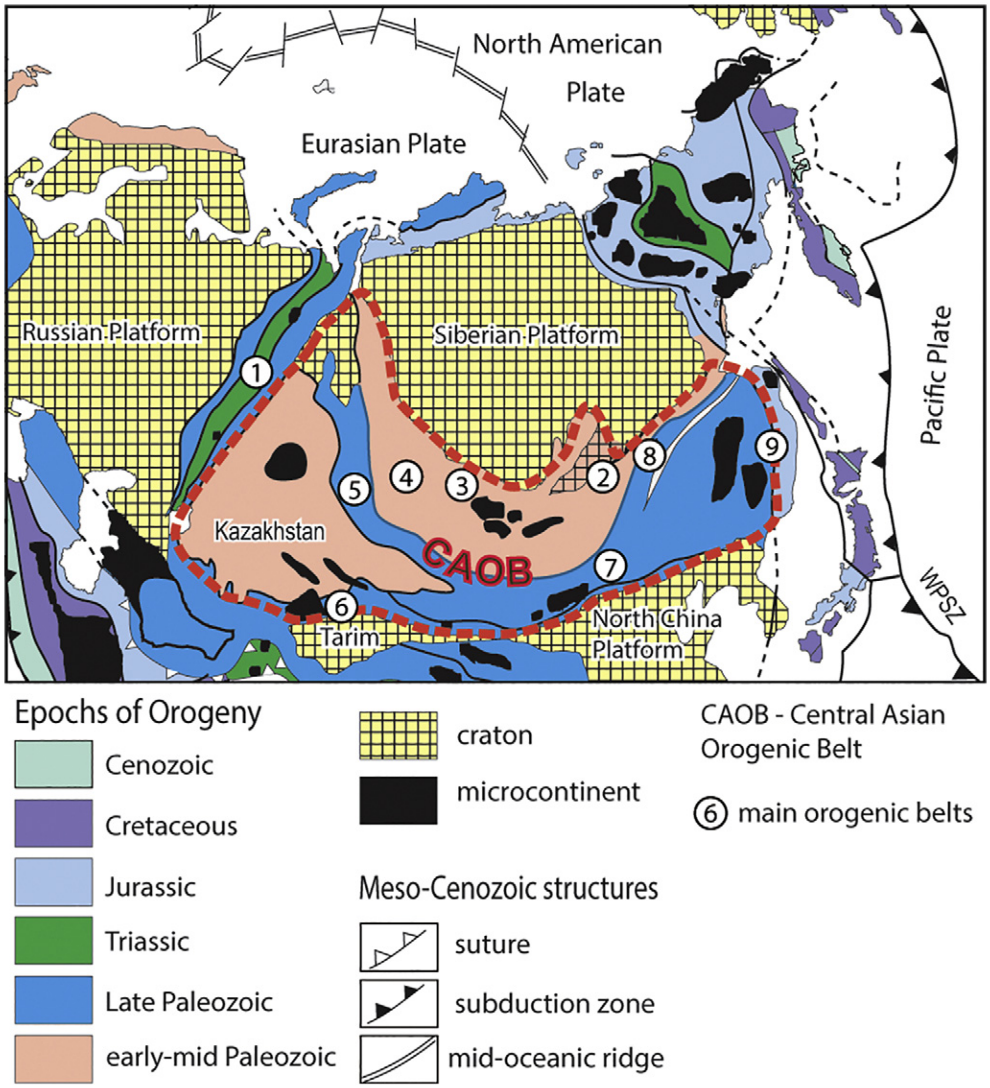

Fig. 4. Tectonic outline of Central Asia showing major continental blocks surrounding the CAOB, its contours and major constituents: orogenic belts and microcontinents (modified from Maruyama and Sakai, 1986; Maruyama et al., 1989; Safonova and Maruyama, 2014). Numbers in circles are for orogenic belts: 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.

magma types, tholeiites, boninites and calc-alkaline basalts, which probably were generated from one subduction zone. The earliest lavas of the Mariana IOA (Fig. 1B), e.g., those erupted on the Kyushu-Palau Ridge and Mariana Fore-arc (e.g., Lin et al., 1989, 1990; Lee et al., 1995; Pearce et al., 2005; Marske et al., 2011; Straub et al., 2015), are island arc tholeiites (IAT) and boninites (Table 1; Fig. 3). These are characteristic of very primitive (young, immature) oceanic island arcs, and are not usually erupted on continents or in the later stages of arc development. IAT have similarities with mid-ocean ridge basalts (MORB) in having depleted rare-earth element (REE) patterns, but are usually more Fe-rich and with low Cr and Ni contents, very low Nb and Ta, higher K contents and high K/Rb ratios (Fig. 3A, C). Boninites are high-Mg lavas, but have high silica contents more typical of andesites; they have high Cr and Ni, but have lower Ti and higher K, Rb, Ba and Sr contents than would normally be expected for high-Mg rocks (Fig. 3A, B). Boninites are thought to result from wet melting of the rather refractory Mg-rich mantle wedge beneath the developing arc - with the wedge being contaminated with elements such as K, Rb, Ba, Sr transported from the subduction zone during dehydration of the hydrous ocean crust (Crawford et al., 1989). IAT could be melts of the more fertile asthenosphere, the magmas then undergoing extensive crystal fractionation en route to the surface. Or they could represent melts of subducted ocean basalt crust, but possible only in the very beginning of subduction when the ocean lithosphere is starting its travel to the deep hot mantle.

Later, after the spreading opened the back-arc basin at 17 Ma, the arc volcanism shifted to what is now the West Mariana Ridge (Fig. 1B) and continued building up the arc for ca. 9 m.y. Those younger lavas were mainly calc-alkaline basalts (CAB) and basaltic andesites, with higher Al and K contents, much higher Sr and Ba contents and light rare-earth enriched rather than depleted REE (rare-earth elements) patterns (Table 1, Fig. 3A, D). These lavas are more similar to calc-alkaline lavas erupted at continental margins, though the latter are typically dominated by andesite rather than basaltic andesites. The West Mariana CAB magmas may have been derived from the mantle wedge enriched in Ba, Sr, light REE (LREE), etc., perhaps as a result of continued fluid transport of these elements into the wedge from the dehydrating subducting slab (e.g., Saunders et al., 1988; Johnson, 2014). Modern lavas erupted at the active Mariana Arc tend to be mainly andesites and basaltic andesites having characteristics in between those of IAT and CAB (Fig. 3A–C). There is some evidence (low εNd values) that a small component (ca. 0.5%) of subducted abyssal sediment is involved in their source regions (Hiekey-Vargas and Reagan, 1987; Lin et al., 1989, 1990; Pearce et al., 1999; Hochstaedter et al., 2000).

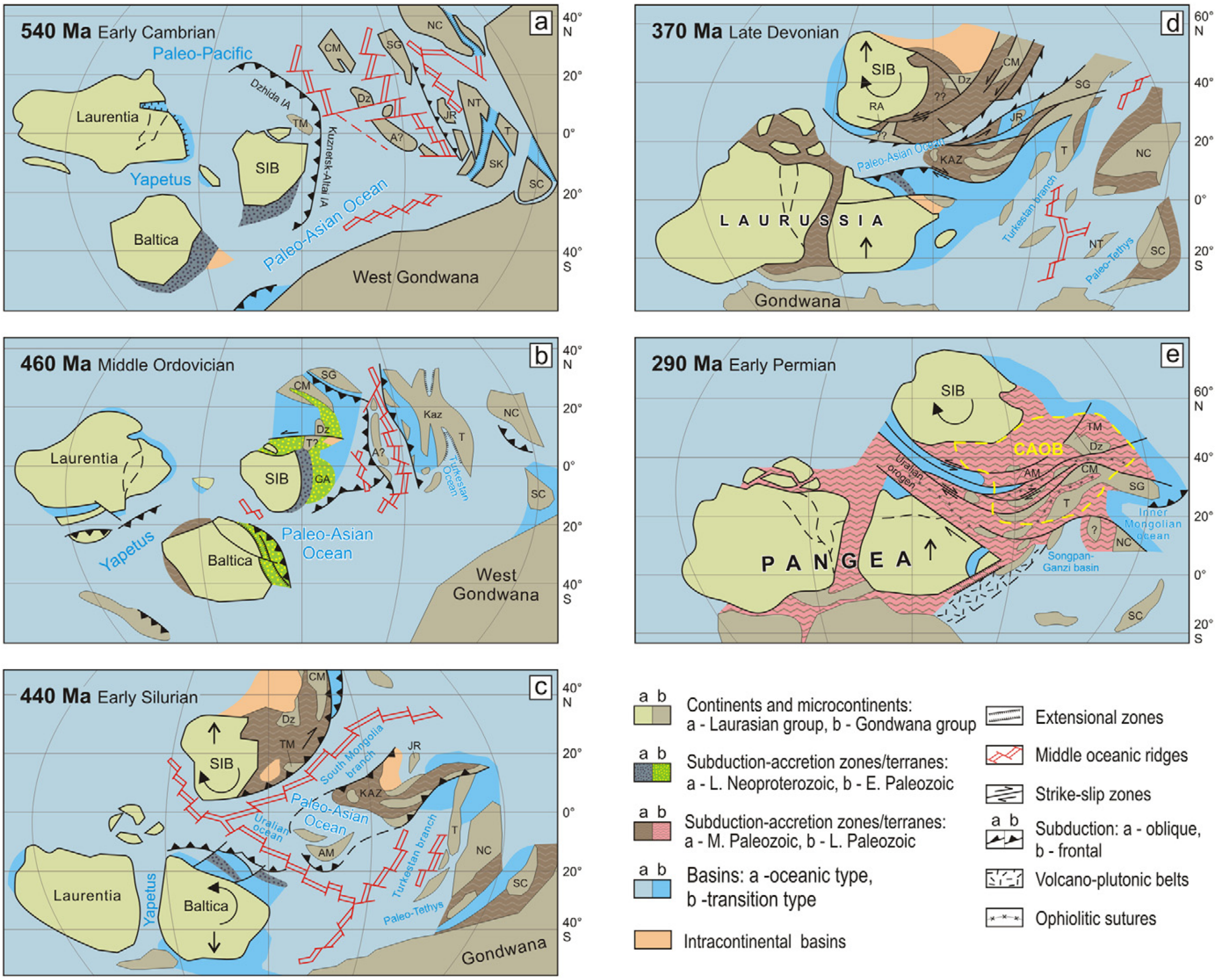

Fig. 5. Geodynamic reconstruction of the Paleo-Asian Ocean based on geological, paleomagnetic and petrologic data (based on reconstructions from Buslov et al., 2001, 2004a; Kurenkov et al., 2002; Dobretsov et al., 2003; Torsvik and Cocks, 2017). Abbreviations for continental blocks: CM, Central Mongolian; Dz, Dzabkhan (Baydrag); JR, Junggar; Kaz — Kazakhstan; NC, North China; NT, North Tienshan; SC, South China, SG, South Gobi; SIB, Siberian Craton; T, Tarim; TM, Tuva-Mongolian; subduction-accretion zones: AM — Altai-Mongolian, GA – Gorny Altai; RA — Rudny Altai; IA — island arc.

The boninites of the Bonin Island are the most primitive rocks plotting away from most of the tholeiitic and other lavas in all diagrams (Fig. 3A–C). The IOA basalts are, in average, different from back-arc basin or marginal basin basalts, which are similar to N-MORB. However, during the early stages of back-arc spreading, when the uprising mantle diapir splits the volcanic arc (Section 8.2), the basalt magmas are derived from the sub-arc mantle. These basalts tend to have an arc-like geochemical signature. Thus their REE patterns may be slightly light REE enriched, they may have higher Ba, Sr, K and Rb (not always though), but low Nb and Ta (Table 1). Moreover they tend to have higher water contents and be vesicular - a consequence of fluids distilled from the subducting slab. These features are useful discriminants in trying to characterize ophiolites as being derived from either obducted ocean floor or back-arc basin crust (e.g., Saunders and Tarney, 1984).

In general, in the Mariana IOA system (Figs. 1, 2), the fore-arc lavas, early arc and active arc lavas are dominated by typical island-arc tholeiites (Fig. 3A–C). Boninites occur in the Tonga and Bonin arcs and in the Mariana protoarc of 50–45 Ma age (Pearce et al., 1999) (Fig. 3B, C). The Mariana early to recent lavas show a clear trend from more primitive to more evolved magmas (Fig. 3B). The boninites and early arc tholeiites are characterized by lower potassium (Fig. 3D). Calc-alkaline lavas occur in the Mariana youngest lavas formed at the active arc and fore-arc (Fig. 3B, C).

3. Paleo-Asian Ocean and Central Asian Orogenic Belt

The CAOB extends over a huge area from Kazakhstan in the west, through Kazakhstan, Kyrgyzstan, Uzbekistan, north-western China to Altai-Sayan and Transbaikalia in Russia, Mongolia, and north-eastern China to the Russian Far East (Fig. 4). The CAOB has evolved during at least 800 m.y., from Neoproterozoic time until the Cenozoic (the time of intracontinental rifting and magmatism), as a result of the evolution and subsequent closure of the Paleo-Asian Ocean and subsequent collisions of the Siberian, Kazakhstan, Tarim and North China continents (Fig. 5) (e.g., Zonenshain et al., 1990; Dobretsov et al., 1995; Windley et al., 2007; Safonova, 2009; Xiao et al., 2010; Donskaya et al., 2013; Xiao and Santosh, 2014; Liu et al., 2017). According to numerous geological, geochemical and isotopic data obtained during the last 15 years the CAOB was a major site of juvenile crustal growth during the Phanerozoic (e.g., Jahn et al., 2000; Kovalenko et al., 2004; Windley et al., 2007; Sun et al., 2008; Xiao et al., 2010; Safonova et al., 2011a; Safonova, 2017). However, it includes not only oceanic, intra-oceanic and continental margin arc terranes, but also

Table 1
Selective characteristics of typical volcanic rocks of the Mariana IOA.

Stage	Early		Late	Modern
Arc type*/age	Protoarc, 50–45 Ma		Younger arc, 36–32 Ma	Active arc
Series	Boninite	IAT	CAB-CAA	High-K
	Primitive		Evolved	
SiO_2	52–62	46–51	45–66	52–60
TiO_2	0.1–0.8	0.6–1.2	0.6–1.0	0.8–1.0
FeO	5–10	7–14	7–10	9–14
MgO	6–20	5–8	3–8	2–5.5
K_2O	0.1–1.3	0.1–0.5	0.5–1.6	0.5–2.7
Al_2O_3	5–17	14–17	13–18	13–17
Cr	100–900	20–450	3–60	1–6
Ni	20–400	10–110	3–20	3–5
Nb	2–6	5–6	0.3–0.6	1.4–2.8
Rb	5–10	1–30	1–5	10–40
Sr	17–45	60–130	170–200	300–650
Ba	70–300	40–300	40–90	150–400
εNd	2–9	8–11		6–7
εHf	12–18	14–19	13–14	14–16

Oxides and trace elements given in wt% and ppm, respectively. IAT – island-arc tholeiite, CAB-CAA, calc-alkaline basalt to andesite. Sources: Hiekey-Vargas and Reagan, 1987; Frolova and Burikova, 1997; Pearce et al., 1999; Reagan et al., 2010. * - according to Pearce et al., 1999.

numerous fragments of Precambrian microcontinents and collisional and post-collisional complexes. The subduction erosion and post-orogenic tectonics both could partly or fully destroy the intra-oceanic arc complexes.

Numerous late Neoproterozoic to late Paleozoic intra-oceanic arcs of the Paleo-Asian Ocean (most of them include boninites) outcrop in eastern and western Tuva-Sayan regions, Mongolia, Russian-Kazakh-Chinese-Mongolian Altai, and western and eastern Junggar (e.g., Simonov et al., 1994; Buslov et al., 1998, 2001; Pfander et al., 2002; Kovalenko et al., 2004; Dijkstra et al., 2006; Zhang et al., 2005; Niu et al., 2006; Gordienko et al., 2007; Ota et al., 2007; Jian et al., 2010; Buriánek et al., 2017) (Table 2). The intra-oceanic arcs of CAOB are related to the evolution of the Paleo-Asian Ocean including its Turkestan (or South Tienshan) and Junggar (or Ob-Zaisan) branches, and active margins of the Siberian, Kazakhstan, and Tarim continents (Fig. 5). The Paleo-Asian Ocean, at different times, separated the Siberian, Kazakhstan and Tarim continents and smaller microcontinents, some probably derived from Gondwana (e.g., Dobretsov et al., 1995; Buslov et al., 2001; Jiang et al., 2011; Safonova and Santosh, 2014).

The intra-oceanic arc terranes of the CAOB formed during a long period from the late Neoproterozoic to the late Paleozoic. The formation of the CAOB started from accretion of intra-oceanic arcs and pieces of oceanic lithosphere to the southern margin of the Siberian Craton during late Neoproterozoic-early Paleozoic time (e.g., Buslov et al., 2002; Pfander et al., 2002; Turkina, 2002; Kuzmichev et al., 2005; Gordienko et al., 2006, 2007) (Table 2; Fig. 5A, B). During the middle and late Paleozoic, the subduction of the Paleo-Asian Ocean continued under the southern active margin of the Siberian Craton and under active margins of the Kazakhstan, Tarim and North China continents (e.g., Kovalenko et al., 2004; Windley et al., 2007; Xiao et al., 2010; Yarmolyuk et al., 2012; Donskaya et al., 2013; Yang et al., 2015b). The closure of the Paleo-Asian Ocean and its Turkestan and Junggar branches ceased the subduction diachronously, during a period from the late Paleozoic to the Mesozoic (Fig. 5D, E). The last stages of Paleo-Asian Ocean evolution are recorded in oceanic and supra-subduction complexes of 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; Yang et al., 2015a) and Far East (Dobretsov et al., 2003; Donskaya, et al., 2013; Yang et al., 2015b).

The suturing of the Paleo-Asian Ocean started by the middle-late Paleozoic approach and collision of the Kazakhstan and Siberian continents (Fig. 5C, D). 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). Closure of the Turkestan Ocean, a southern branch of the Paleo-Asian Ocean, and collisions of the Kazakhstan continent and Tarim Craton and smaller microcontinents in the South Gobi area formed the Tienshan orogen extending across the territories of Kazakhstan, Uzbekistan, Tajikistan,

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

No.	arc name	Location	Age, Ma[a]	Length, km[b]	Bon	$εNd_T$	References[c]
1	Dunzhugur	Eastern Tuva-Sayan - NW Mongolia	812–784	220	Yes	n.d.	Kuzmichev et al., 2001;
2	Shishkhid	Eastern Tuva-Sayan - NW Mongolia	808–790	110	Yes	+6.9	Kuzmichev et al., 2005
3	Ilchir	Eastern Tuva-Sayan - NW Mongolia	ca. 600	90	Yes	n.d.	Dobretsov et al., 1989
4	Agardag	Western Tuva-Sayan, Russia	ca. 570	115		+4.8–8.5	Pfander et al., 2002
5	Shatskii	Western Tuva-Sayan, Russia	ca. 580	330		+8.8	Mongush et al., 2011a
6	Tannu-Ola	Western Tuva-Sayan, Russia	539–518	210		+6.4–+8.4	Mongush et al., 2011b
7	Kurtushibin	Western Tuva-Sayan, Russia	Late Neoproterozoic	440	Yes	n.d.	Dobretsov et al., 1992
8	Dariv	Western Mongolia	573–571	260	Yes	n.d.	Khain et al., 2003; Dijkstra et al., 2006
9	Khan-Taishirin	Western Mongolia	573–568	610	Yes	+6.5–+10.5	Buriánek et al., 2017
10	Dzhida	Mongolia-Transbaikalia	570–510	280	Yes	+6.4–+7.6	Gordienko et al., 2006, 2012
11	Kurai-Ulagan	Russian Altai	647–598	95	Yes	+4–+6	Buslov et al., 1993; Chen et al., 2016
12	Selety-Urumbai	Northern Kazakhstan	Cambrian–early Ordovician	390	Yes	n.d.	Degtyarev, 2011, 2012
13	Bozshakol-Chingiz	East Kazakhstan	Late Cambrian early Ordovician; 501–480	1470	Yes	+4.1–+6.9	Shen et al., 2015; Degtyarev, 2011, 2012
14	Baydaulet-Aqbastau	East Kazakhstan	Floian–Late Ordovician	910		n.d.	Degtyarev, 2012
15	Chatkal-Atbashi	Southern Tienshan, Kyrgyzstan	470–460	230		+0.9 to −2.6	Alexeiev et al., 2016
16	Fan-Karategin	Southern Tienshan, Tajikistan	Late Ordovician–Early Silurian	330		n.d.	Volkova and Budanov, 1999
17	Gurvansayhan-Zoolen	SW Mongolia	Late Silurian–Late Devonian; 421–417	870	Yes	+6–+9	Badarch et al., 2002; Helo et al., 2006
18	Saerbulake	Chinese Altai	Early Devonian	315	Yes	+3–+4.3	Zhang et al., 2005; Niu et al., 2006
19	Zharma-Saur	East Kazakhstan–western Junggar	M. Devonian; 380–356	490		+6–+16[d]	Hong et al., 2017; Li et al., 2016a
20	Char	East Kazakhstan	Middle–Late Devonian	95	Yes		Kurganskaya et al., 2014
21	Bogda	Chinese Tienshan	E. Carboniferous 347–315	600		7.9–9.4	Chen et al., 2013

More see in the text; d – Hf-in-zircon isotopes; n.d. – no data; Bon – found boninites; E. – early, M. – middle.
[a] Present-day/exposed.
[b] Age estimates determined by isotope analysis are given in Ma and those inferred from fossils are given as stratigraphic periods/stages.
[c] These references are selective.

Kyrgyzstan and northwestern China (e.g., Biske, 1996; Bakirov and Maksumova, 2001; Xiao and Kusky, 2009; Charvet et al., 2007; Biske and Seltmann, 2010; Wang et al., 2011). Early-middle Paleozoic intra-oceanic arcs of the Turkestan Ocean occur in the South Tienshan (Volkova and Budanov, 1999; Alexeiev et al., 2016; Dolgopolova et al., 2017). The Junggar Ocean, which once separated the Junggar block and the active margins of the Kazakhstan and Siberian continents, was closed in the late Paleozoic, most probably in late Carboniferous time (Buslov et al., 2001; Safonova et al., 2012; Kuibida et al., 2016).The eastern part of CAOB was assembled during the Late Permian to Triassic approach and further collision of the North China and Siberian Cratons and smaller microcontinents (e.g., Zhao et al., 1990; Xiao et al., 2004; Zhao et al., 2013; Zhou and Wilde, 2013; Yang et al., 2015a, b; Li et al., 2016a, b) (Fig. 5D). Subduction of the Mongol-Okhotsk Ocean (Didenko et al., 1994; Dobretsov et al., 1995; Donskaya et al., 2013; Ruppen et al., 2014) probably produced volcanic arcs in Inner Mongolia and northeast China (e.g., Chen et al., 2000; Dergunov et al., 2001; Miao et al., 2008; Jian et al., 2010). However, there is still no consensus about the intra-oceanic vs continental margin origin of those arcs, neither about the origin of several older active margin terranes of the eastern CAOB (e.g., Eravna and Argun; Yakubchuk, 2017).

The amount of the late Neoproterozoic (ca. 800 to 540 Ma) intra-oceanic arcs is much greater than that of early and middle Paleozoic arcs (Figs. 4, 6; Table 2). The western part of the Paleo-Asian Ocean, including the Turkestan branch, was progressively suturing during the Carboniferous and Permian, whereas its eastern part probably remained opened until the Triassic in Solonker and even middle Jurassic in Mongol-Okhotsk (e.g., Zonenshain et al., 1990; Dobretsov et al., 1995; Buslov et al., 2001; Windley et al., 2007; Biske and Seltmann, 2010; Xiao et al., 2010; Donskaya et al., 2013; Yakubchuk, 2017). In this paper, we review PAO intra-oceanic arcs based on their geological position, i.e. in association with reconstructed back-arc basins, rock chemistry and, if available, isotopes. The tectonic definitions for island-arc domains, e.g., terrane, ophiolite belt, magmatic arc, zone, etc., are given as in original publications (see the references cited in Sections 4–7).

4. Neoproterozoic – early Cambrian IOAs of Siberia and Mongolia

We recognize 11 intra-oceanic arc complexes have been identified in the central and oldest part of the CAOB – in eastern and western Tuva-Sayan, Transbaikalia, Russian Altai, western, north-western and northern Mongolia (Figs. 4–6). They mark an early stage of Paleo-Asian Ocean evolution (Dobretsov et al., 2003). The arcs span a long period from the late Neoproterozoic (ca. 800 Ma) to the earliest Cambrian. Most of them include boninite-bearing volcanic units and are associated with back-arc basin complexes.

4.1. Eastern Tuva-Sayan - NW Mongolia: Dunzhugur, Shishkid, Ilchir

The oldest Dunzhugur IOA occurs in eastern Tuva, Russia (no. 1 in Figs. 4, 6) and represents a part of the active margin of the Precambrian Tuva-Mongolian microcontinent surrounded by the Palaeozoic foldbelts. The rocks of the Tuva-Mongolian microcontinent are exposed at the Gargan Block, which consists of Early Precambrian crystalline basement and Neoproterozoic passive-margin sediments (Fig. 7) (Kozakov et al., 1999). Dunzhugur volcanic rocks are thrust onto the Gargan Block. They show a full ophiolite assemblage including pillow lavas, gabbros and ultra-mafic rocks, sheeted dykes and fore-arc complexes facing the microcontinent. The Dunzhugur arc probably collided the Gargan Block at ca. 800 Ma (Kuzmichev et al., 2005). The OPS sedimentary rocks associated with the Dunzhugur ophiolites show strong deformation by thrusting and intrusion of post-ophiolite sills of tholeiitic dolerite. The OPS sediments, hemipelagic siliceous shale and turbidite, concordantly overlie the IOA lavas. The composition of the turbidites is

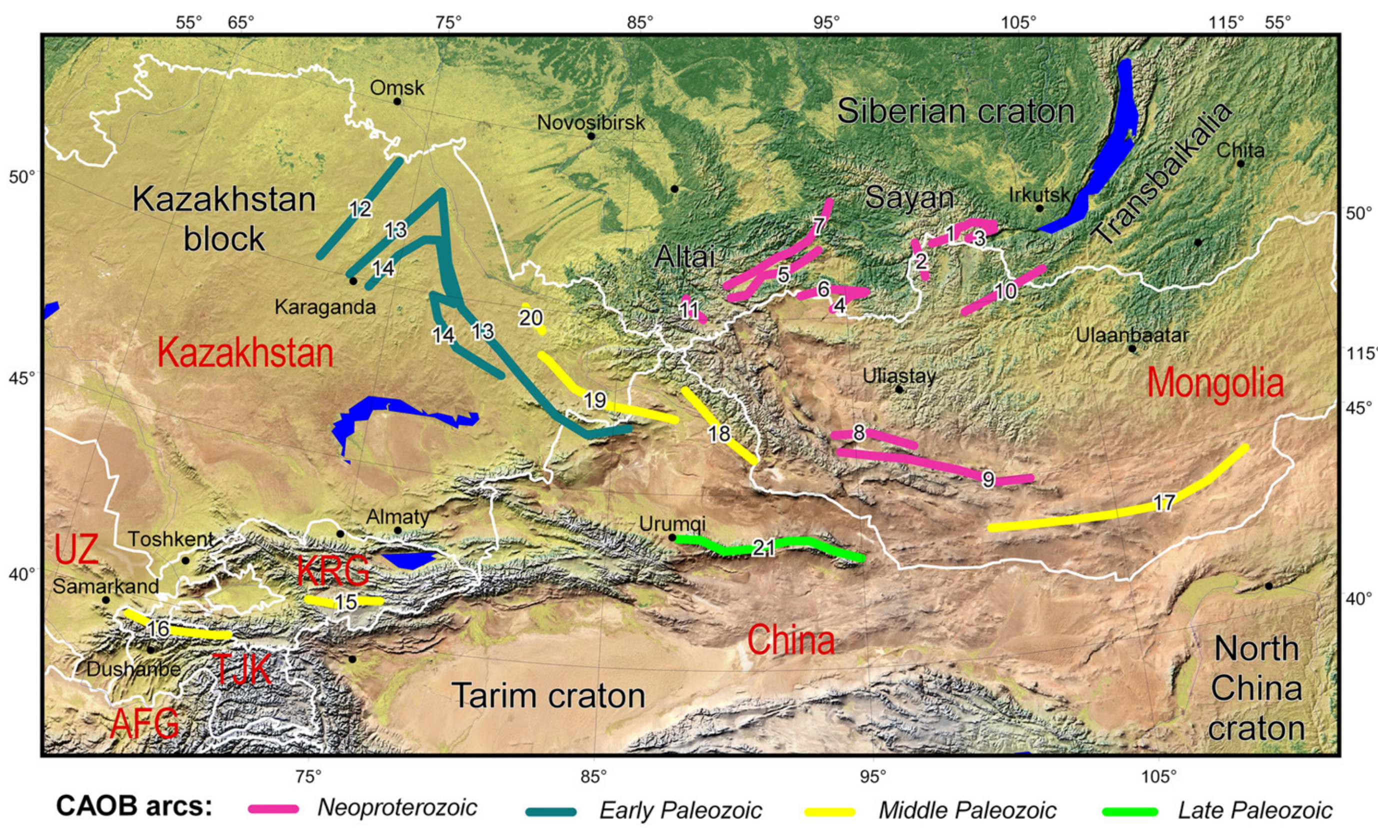

Fig. 6. Fossil IOA systems of the CAOB. Siberian Group: Neoproterozoic to early Cambrian IOAs of Siberia and Mongolia (pink). Kazakhstan Group: early Paleozoic arcs of Kazakhstan (blue). Southern Group: middle Paleozoic arcs of the Tienshan, Chinese Altai and Mongolia (yellow). The lengths of arcs were determined with help of ArcView software. AFG, Afghanistan; KRG, Kyrgyzstan; TJK, Tajikistan; UZ, Uzbekistan.

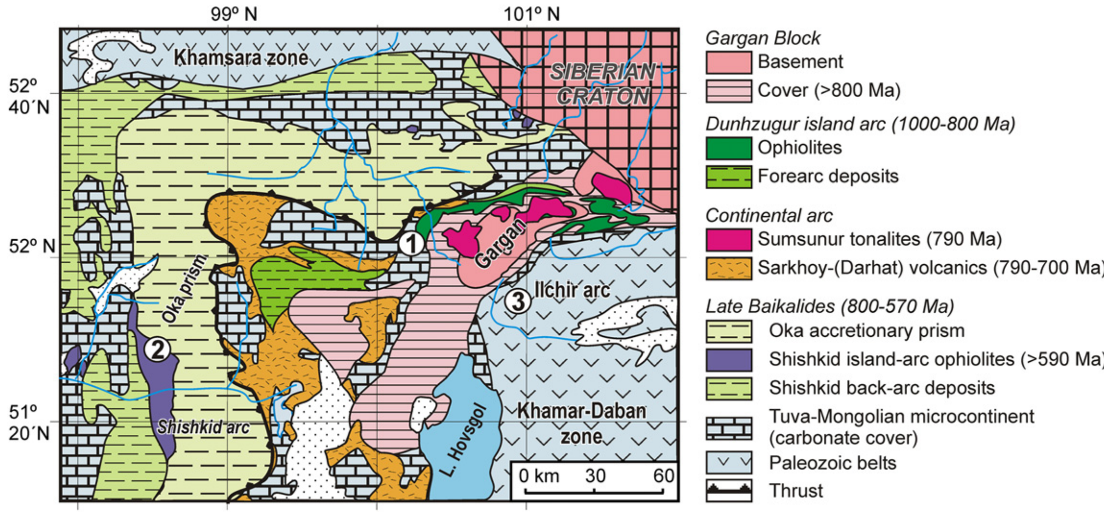

Fig. 7. Major geological units of eastern Tuva-Sayan (modified from Kuzmichev et al., 2005).

similar to that of arc mafic volcanic rocks. The IOA lavas overlain by turbidities suggest an environment similar to that of recent forearc basins (Bloomer et al., 1995) dominated by turbidites with lenses of unsorted coarse-grained volcaniclastics containing serpentinized ultramafic and limestone clasts (Underwood et al., 1995). The presence of the mafic sills is also a common feature of IOA-related fore-arc basins (Taylor et al., 1995; Kuzmichev et al., 2001). The approximate age of the Dunzhugur IOA is 1000–800 Ma as deduced from the 790 Ma age of the Sumsunur tonalities, which intruded the Gargan after the emplacement (obduction?) of the Dunzhugur ophiolites (Kuzmichev et al., 2001; Khain et al., 2002). The Dunzhugur dykes and volcanic rocks possess tholeiitic and boninitic to calc-alkaline composition (Dobretsov et al., 1986; Sklyarov et al., 2016): the boninitic lavas are characterized by increased SiO_2 (56–58 wt%) and MgO (8–10 wt%) and decreased TiO_2 (0.2–0.4 wt%) suggesting a long-lived intra-oceanic arc similar to the Mariana arc (Fig. 8; Suppl. Electr. Mat.).

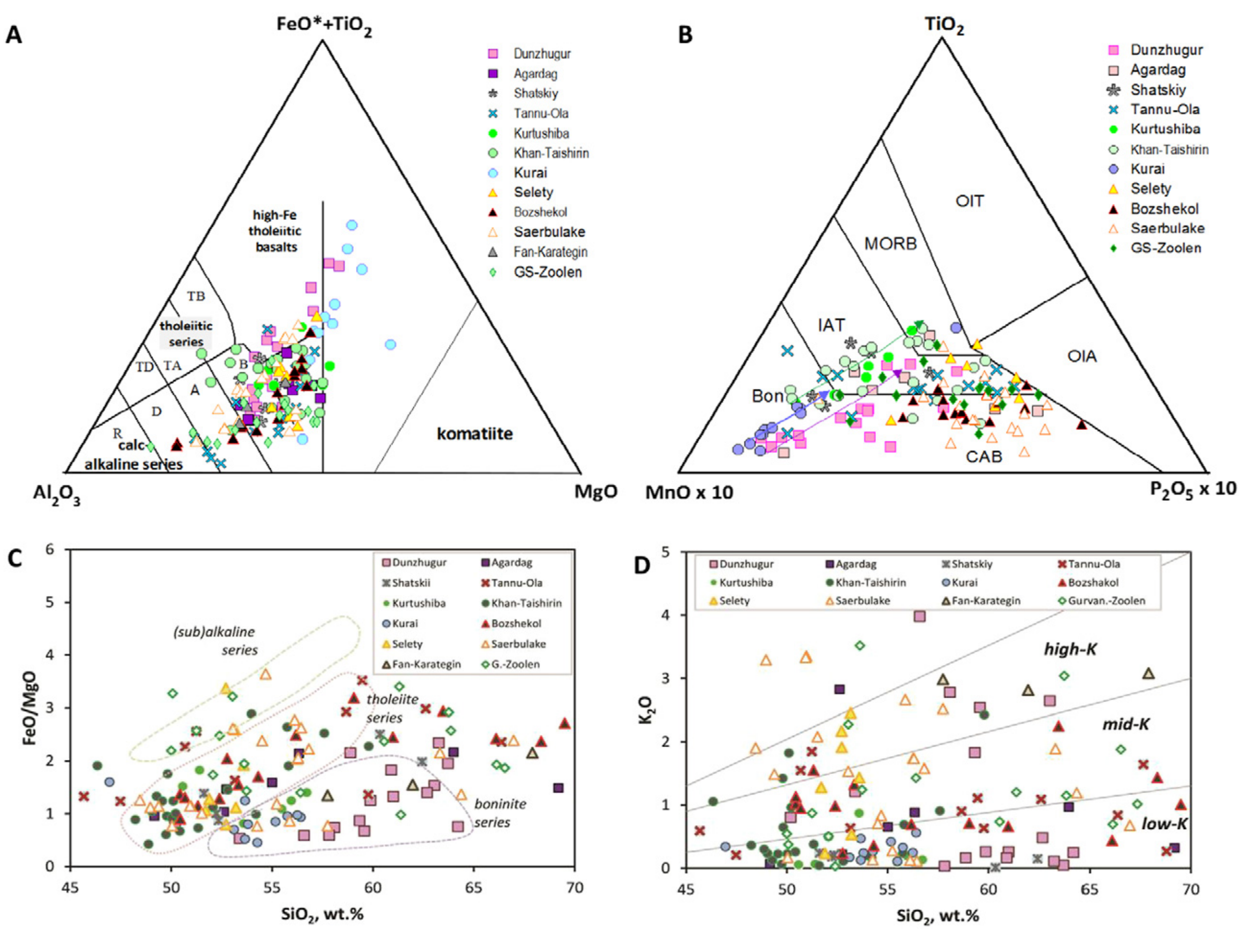

Fig. 8. Geochemical characteristics of selected IOA lavas of the CAOB. Explanations for the plots and symbols as in Fig. 3. Sources: Buslov et al. (1998); Volkova and Budanov (1999); Pfander et al. (2002); Kuzmichev et al. (2005); Helo et al. (2006); Niu et al. (2006); Volkova et al. (2009); Mongush et al. (2011a, b); Degtyarev (2012); Shen et al. (2015); Buriánek et al. (2017).

The Shishkhid arc or ophiolitic terrane is located west of Dunzhugur IOA (Fig. 7). It is a part of the western active margin of the Tuva-Mongolian microcontinent (Kuzmichev et al., 2005). The Shishkhid IOA extends from northern Mongolia to eastern Tuva (no. 2 in Figs. 6, 7) and occupies an area over 70 km long and 15 km wide. The Shishkhid IOA is associated with a late Neoproterozoic back-arc basin, which is located to the west. The neighboring units are the Oka accretionary belt hosting OPS units, both igneous and sedimentary (Kuzmichev et al., 2005) in the north and the Sarkhoi continental arc (Kuzmichev, 2004) in the west (Fig. 7). The Shishkhid arc is thrust onto the Oka belt and is tectonically underlain by a mélange consisting of serpentinite lenses intercalated with Oka accretionary units. The Shishkhid arc consists of (bottom to top) ultramafic rocks, gabbro, sheeted dykes, basalt, rhyolite and andesitic pyroclastic rocks thus representing a 4.5 km thick typical supra-subduction ophiolite (Kuzmichev et al., 2005). The volcanic rocks are overlain by a 3 km-thick sedimentary unit suggesting progressive subsidence of the volcanic edifice after the cessation of volcanism. The Shishkhid igneous units are geochemically very variable. The volcanic rocks range from basalt to rhyolite of the tholeiitic, boninitic and calc-alkaline series and possess many geochemical features typical of modern IOAs (Kuzmichev et al., 2005). Most of the samples, both intrusive and extrusive, belong to the calc-alkaline series though and are characterized by high LILE, show flat to LREE-enriched rare-earth patterns and Nb negative peaks in the multi-elements diagrams. The initial Nd isotopes (εNd) range from +6.9 to 0.0 (Table 2) suggesting both a depleted sub-arc mantle wedge source and subducted sediment component (see Kuzmichev et al., 2005 for details). In general, the composition of the Shishkhid volcanic rocks is similar to that of the "back-arc knolls zone" of the Izu-Bonin arc (e.g., Pearce et al., 1992; Hochstaedter et al., 2000; Tamura and Tatsumi, 2002). SHRIMP U—Pb zircon dating of a rhyolite from the volcanic unit yielded a crystallization age of 800 ± 2.6 Ma (Kuzmichev et al., 2005). This date and the geological position suggest that the Shishkhid arc evolved during most of the late Neoproterozoic and finally collided with the Sarkhoi continental arc at ca. 600 Ma. The Shishkhid IOA can be traced for about 600 km and there are coeval arcs in other terrains of CAOB (see other IOA examples of Section 4) suggesting that during the late Neoproterozoic there were many intra-oceanic arcs in the Paleo-Asian Ocean.

The Ilchir ophiolite terrane with intra-oceanic arc units is located farther to the east of the Dunzhugur and Shishkhid ophiolitic belts (no. 3 in Figs. 6, 7), in East Sayan, southern Siberia. There are two types of ophiolites in the Ilchir terrane (Dobretsov et al., 1985). Type 1 ophiolite section is exposed in the northwestern part of the terrane and includes pyroxenites, Mg-rich gabbro, sheeted dykes, and boninitic pillow lavas and pillow breccias of Mg-rich basaltic andesite (dominant) and boninite. Type 2 ophiolite section consists of deformed harzburgite-dunite and layered ultramafic-gabbro-anorthosite cumulates. The gabbroic part also includes boninitic varieties. The ophiolite sections are overlapped by black shales and calc-alkaline basalt-andesite-dacite volcanic rocks. The parallel sheeted dikes consist of basaltic andesites, in which the concentrations of SiO_2, TiO_2 and MgO and the shape of REE spectra are similar to those of boninites (Dobretsov et al., 1992). Compositionally, the boninitic dykes seem to be older than the sheeted dykes. Thus, the boninitic series is represented by gabbros, dykes and lavas. The ophiolitic sections are cut by younger diabase and gabbro-diabase dykes geochemically and petrographically similar to the OIB-type sills hosted by the turbidite series overlying the Ilchir ophiolites (Safonova and Santosh, 2014). In summary, the geological position and geochemical characteristics of the Ilchir ophiolite suggest their formation in an intra-oceanic arc environment. The age of the Ilchir IOA has not been well constrained, but geologically it is definitely much older than Ordovician and probably close to that of the glaucophane schists of the Oka zone, for which the Rb—Sr isochron dating yielded an age of ca. 600 Ma (Dobretsov et al., 1989), although these data should be verified by more up-to-date techniques.

4.2. Western Tuva–Sayan: Agardag, Shatskii, Tannu-Ola, Kurtushibin

The Agardag (or Agardag-Teschem) ophiolitic terrane is located in western Tuva (no. 4 in Fig. 6) and marks the northwestern border of the Tuva-Mongolian microcontinent (Fig. 9). The late Neoproterozoic age of the ophiolite is constrained by the 569.6 ± 1.7 Ma $^{207}Pb/^{206}Pb$ age of plagiogranite (Pfander and Kroner, 2004; Table 2). The ophiolite consists of three major units separated by thrusts and strike-slip faults: 1) ultramafic unit (serpentinized dunites, harzburgites, wehrlites and pyroxenites) with subordinate gabbro and plagiogranite; 2) intrusive gabbroid unit (hornblende-bearing gabbros, olivine-gabbros, gabbronorites, gabbrodiorites and wehrlites; 3) subvolcanic-volcanic unit of fine-grained gabbro, dolerite and basalt. The last unit is overlain by conglomerates, sandstones and shallow-marine limestones suggesting a marine environment. The entire ophiolite was obducted onto the Tuva-Mongolian microcontinent in early Palaeozoic time and is

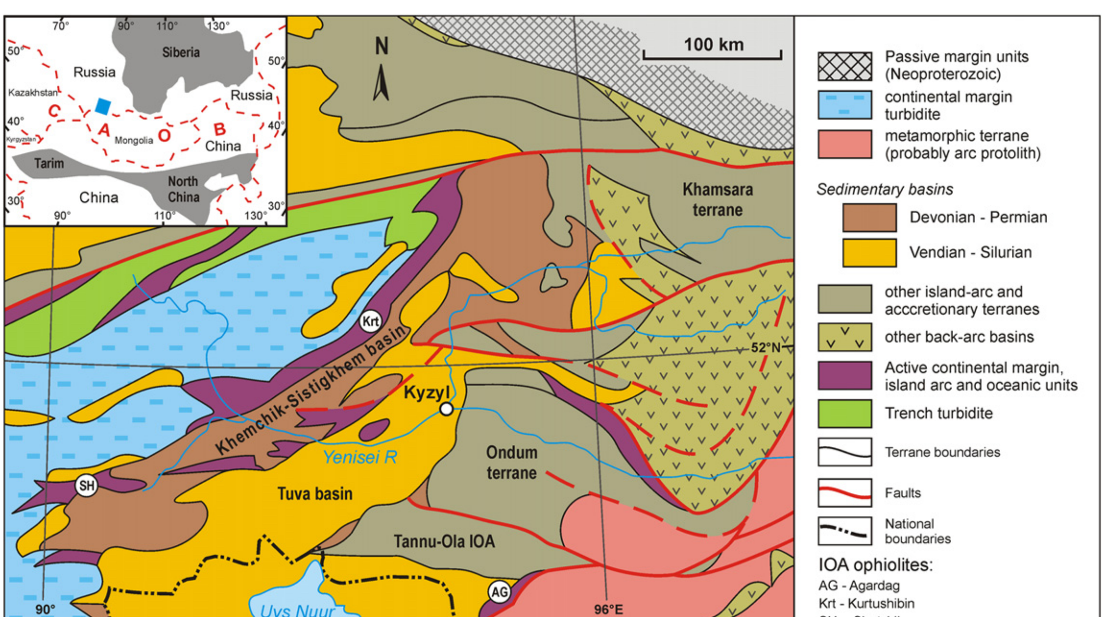

Fig. 9. Major geological units of western Tuva-Sayan arcs (modified from Berzin et al., 1994; Mongush et al., 2011b).

embedded within a tectonic mélange consisting of OPS sediments, limestones, ultramafic rocks, pillow lavas and massive andesibasalt and thus is part of an accretionary wedge. There are two groups of volcanic rocks. One group is dominated by high-alumina basalts and basaltic andesites having island-arc affinities (Pfander et al., 2002; Fig. 8). They were derived from a low melted parental magma by predominantly clinopyroxene fractionation. The island-arc rocks have high abundances of incompatible trace elements ($La/Yb_n = 14.6–5.1$) and negative Nb anomalies ($Nb/La = 0.37–0.62$), but low Zr/Nb ratios (7–14). Initial Nd values are around +5.5. Enrichment of LILE within this group is significant ($Ba/La = 11–130$). The island arc volcanic rocks formed during an early stage of subduction by low-degree melting from a depleted mantle source, containing subducted sediments. Another group of samples are back-arc tholeiites probably formed from a similar depleted mantle source as the island-arc rocks, but by higher degrees of melting (8–15%) without an appreciable influence of the downgoing slab. They have lower abundances of incompatible trace elements, flat rare-earth element patterns ($(La/Yb)n = 0.6–2.4$) and higher Nd values (+7.8 to +8.5). Negative Nb anomalies are absent ($Nb/La = 0.81–1.30$), but Zr/Nb is high (21–48) (Pfander et al., 2002; Suppl. Electr. Mat.). There are also samples compositionally similar to boninites (Fig. 8B, C). The most likely geodynamic environment to produce these characteristics is a young, intra-oceanic island-arc system and an associated back-arc basin. In contrast to the volcanic rocks, the gabbros are strongly depleted in incompatible trace elements but also have negative Nb anomalies and initial εNd values varying between +4.8 and +7.1 (Table 2), which all are consistent with an island arc origin. In general, the mineralogy and geochemistry of the plutonic rocks indicate the presence of boninitic parental melts (Pfander and Kroner, 2004).The ophiolite therefore consists of an association of island arc and back-arc related sequences, which have been amalgamated during subduction-accretion and collisional obduction.

The Shatskii island arc or ophiolitic terrane is located in western Tuva . The rocks of the terrane formed during the late Neoproterozoic to Cambrian and are spatially related to the Khemchik–Systygkhem accretionary complex (no. 5 in Fig. 6, Fig. 9). The terrane consists of numerous tectonic sheets f late Neoproterozoic ophiolites including intraplate oceanic, island arc, and continental arc units and olistostromes (Berzin, 1987; Pfander et al., 2002; Mongush et al., 2011a), all overlain by Ordovician to Devonian strata. The Shatskii arc contains a nearly full set of typical ophiolitic rocks overlain by olistostrome hosting lens-like blocks of seamount (OIB, limestone), pelagic (MORB, chert), hemipelagic (siliceous shale and mudstone) and trench (sandstone) oceanic plate stratigraphy rocks submerged into a sand-clayish terrigenous matrix (Mongush et al., 2011a; Safonova and Santosh, 2014; Safonova et al., 2016b). The Shatskii arc is dominated by harzburgite, gabbro and diabase, gabbro-diabase and andesite “dyke-in-dyke” complexes. The harzburgite, in places, is cut by veins of dunite and orthopyroxenite. The Ar—Ar age of the “upper” gabbro (hornblede) is 578.1 ± 5.6 Ma (Mongush et al., 2011a), which is close to the age of the Agardag arc (Pfander et al., 2002) (Table 2). The Shatskii rocks span a wide range of SiO_2 (40.0 to 77.4 wt%). The mafic varieties are medium to high-Mg (6.7–8.5 wt% MgO) and are characterized by low TiO_2 (0.3–0.7 wt%) and medium concentrations of $Fe_2O_3^*$ (7.9–10.4 wt%), Al_2O_3 (16.2–18.4 wt%) and CaO (10.0–10.4 wt%) (Fig. 8C; Suppl. Electr. Mat.). The rocks have low concentrations of REE and their primitive mantle normalized multi-element spectra show profound negative Nb—Ta anomalies relative to Th and La. The values of εNd(T) range from +8.1 to + 7.3, i.e. close to the depleted mantle (εNd(0.57Ga) = +8.8) (Mongush et al., 2011a). All this suggests derivation of primary melts from a mantle wedge source and their eruption in a supra-subduction setting, most likely at primitive or intra-oceanic island arcs and back-arc basins.

The Tannu-Ola island-arc is located about 150 km south and southeast of the Khemchik–Systygkhem accretionary complex hosting the Shatskii arc, which was characterized above (no. 6 in Fig. 6, Fig. 8). The volcanic and plutonic arc rocks, back-arc terrigenous deposits and OPS units are unconformably overlain by volcanogenic-sedimentary deposits containing Early Cambrian fossils (Vinkman et al., 1980; Mongush et al., 2011b; Safonova and Santosh, 2014). The island-arc complex includes both effusive and intrusive mafic to felsic rocks. The volcanic rocks are dominated by basalt, andesibasalt, andesite, and their tuffs with subordinate rhyodacite and rhyolite and their tuffs, which erupted mostly in underwater conditions. The volcanic rocks belong to the boninitic, tholeiitic and calc-alkaline series (Fig. 8A–C); they also span a wide range of SiO_2 (48–77 wt%, i.e., from basalt to rhyolite). The mafic varieties are characterized by medium TiO_2 (1–1.3 wt%) and MgO (3.5–8.6 wt%) and medium to high Fe_2O_{3t} (11.8–15.4 wt%) and Al_2O_3 (14.3–18.3 wt%). All varieties possess primitive mantle normalized patterns of incompatible elements with negative anomalies at Nb, Ta, and Ti and weakly to moderately fractionated REE patterns. The plutonic rocks are gabbro-norite, gabbro, diorite, tonalite and plagiogranite. Norite, plagiogranite and diorite yielded ages of 539 ± 6 (Ar—Ar), 518 ± 2 U-Pb) and 522 ± 4 (U—Pb) Ma, respectively (Mongush et al., 2011b). Both the Tannu-Ola volcanic and plutonic rocks show high positive εNd(T) values ranging from +6.4 to +8.4 (Table 2). According to the whole set of geochemical and isotopic data, the Tannu-Ola igneous rocks formed in a supra-subduction setting, probably in an intra-oceanic arc because they are separated from an active margin of the Tuva-Mongolian microcontinent by East Tuva back-arc deposits (Figs. 7, 9).

The Kurtushibin ophiolitic terrane in western Sayan extends over almost 500 km from the Khemchik River in the southwest to the Amyl River in the northeast (no. 7 in Fig. 6, Fig. 9). It consists of three zones separated by tectonic contacts: (1) the Koyard ophiolite massif; (2) Kurtushibin mélange, hosting OPS units, and (3) blueschists and greenschists of the Dzhebash Group (Melyakhovetskii and Sklyarov, 1985). The Koyard massif consists of three units (bottom to top): deformed dunite–harzburgite (2 km thick), dunite–clinopyroxenite–gabbro cumulate (840 m thick), and dike complex (700–800 m thick). The dykes are gabrroic (1–5 m) and basaltic (<1 m). The gabbroic dykes formed at a mid-oceanic ridge and the younger basaltic dykes formed during island-arc or/and back-arc rifting (Kurenkov et al., 2002). The ophiolite is overlain by lavas, pillow lavas, basalt breccias and less abundant tuffs of tholeiitic basalts with interbeds and lenses of siliceous schists and metagreywackes of the Verkhnekoyardsky Formation (30–1600 m). The basalts probably erupted in an island-arc setting. The black shales associated with the basalts contain Osagian-type microfossils of late Neoproterozoic age (Dobretsov et al., 1992). The pillow lavas of the Verkhnekoyardsky Formation ($SiO_2 = 47–48.7$, $TiO_2 = 0.5–0.8$, $MgO = 7.7–10.2$ wt%) and the basaltic dykes are characterized by subchondritic values of rare-earth elements (REE), LREE depleted rate-earth patterns with $(La/Sm)_N$ is lower than those typical of N-MORB, 7–9 chondrite values heavy REE and small Eu anomalies, both negative or positive (Volkova et al., 2009). In general, the volcanic assemblage is dominated by boninites and tholeiites with subordinate calc-alkaline basalts. According to the distribution of incompatible HFSE (Hf, Nb and Zr) and Yb, the basalts were derived at high degree melting of depleted spinel peridotite. The primitive mantle normalized multi-element diagrams show negative peaks of Nb and Ti. Dobretsov et al. (1992) also mentioned boninites in the Kurtushibin terrane. In total, the geological and geochemical data suggest the formation of the younger basaltic dykes of the Kurtushibin ophiolites and overlapping pillow basalts in an intra-oceanic arc setting.

4.3. Western Mongolia: Dariv, Khan-Taishir

The Dariv ophiolite terrane is exposed in the Dariv Range of the Mongolian Altai in western Mongolia (no. 8 in Figs. 6 and 10). It contains (from bottom to top) serpentinized harzburgites and dunites, an igneous layered complex consisting of gabbro–norites, websterites, orthopyroxenites and dunites, sheeted dykes and a series of mafic to andesitic lava flows (Dijkstra et al., 2006). The late Neoproterozoic age

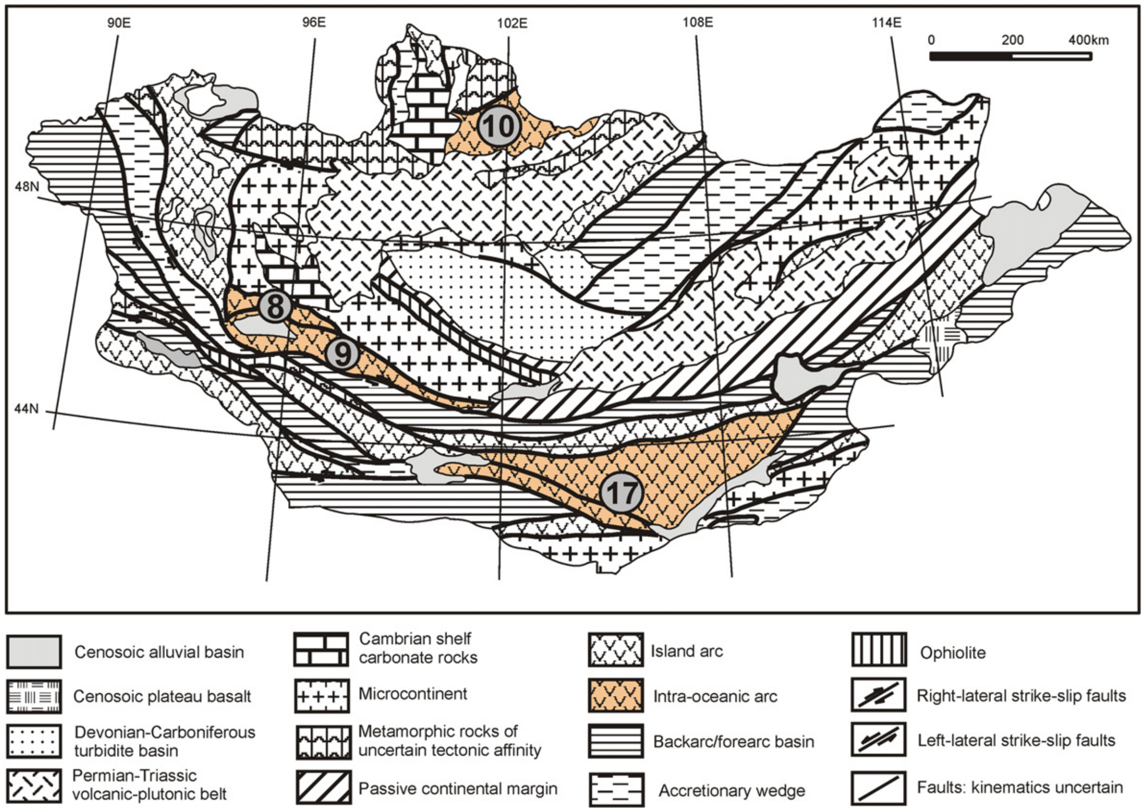

Fig. 10. Main terranes of Mongolia (modified after Badarch et al., 2002). The numbers in circles are for the intra-oceanic arcs and their hosting terranes: 8 – Dariv, 9 – Khan-Taishirin, 10 – Dzhida, 17 – Gurvansayhan-Zoolen. The numbers correspond to those in Table 2.

(573–571 Ma) of the ophiolite is constrained by U—Pb zircon data from plagiogranite (Khain et al., 2003) (Table 2). The lavas are trachybasalts and trachy-andesites. They are enriched in LILE with respect to middle and heavy REE and transition metals in primitive mantle-normalized multi-element diagrams and show clear negative Nb and Ti in respect to the neighboring elements. The sheeted dyke complex consists of metre-scale mutually intrusive dykes ranging in composition from basaltic andesites to rhyolites (Dijkstra et al., 2006). The chemical compositions of the basaltic andesite dykes resemble those of boninites, i.e. high SiO_2, Cr and Ni, and low TiO_2 (0.4%) and Zr (38–45 ppm), but they are too evolved (~7 wt% MgO, no orthopyroxene) to be classified as "true" boninites (Hickey and Frey 1982; Taylor et al., 1994). The multi-element diagrams are similar to those of the lavas, although the concentrations are generally lower in the dykes. Even though the rocks are fractionated, the concentrations of the most compatible elements (e.g., high-field strength elements - HREE) are depleted with respect to N-MORB that is also typical of boninites, indicating derivation from a relatively refractory mantle source (Taylor et al., 1994). Based on the compositions of the crustal units and the crystallization sequences in the mafic and ultramafic cumulates, Dijkstra et al. (2006) show that the Dariv rocks bear a strong resemblance to rocks recovered from the modern Izu–Bonin–Mariana intra-oceanic arc, a fragment of proto-arc oceanic basement, and consequently the Dariv ophiolite is suggested to originate in a similar tectonic setting.

The Khan-Taishir ophiolite also referred to as Hantayshir, Khantaishir or Khan-Taishirin) is part of the Lake (Ozyorny or Ozyornaya) zone or terrane in western Mongolia (no. 9 in Figs. 6 and 10) (Gibsher et al., 2001; Badarch et al., 2002). It tectonically overlies the crystalline basement of the Baydrag or Dzabkhan microcontinent, which is located east of the Lake zone. The Neoproterozoic (Ediacaran) age of the Khan-Taishir ophiolitic sequence is constrained by the ca. 573–568 Ma U—Pb age of zircons from plagiogranite (Gibsher et al., 2001; Jian et al., 2014) (Table 2). Kepezhzinskas et al. (1987) were the first to find boninites in the Khan-Taishir ophiolite spanning SiO_2 from 55 to 61 wt% at $TiO_2 = 0.14–0.23$, $MgO = 8–14.7$, $FeO_t = 6.9–8.9$ and $Al_2O_3 = 10.7–14.9$ wt%. Later, Buriánek et al. (2017) reported tholeiitic basalts and basaltic andesites with subordinate boninites and calc-alkaline basalts and andesibasalts (Fig. 8; Suppl. Electr. Mat.). Chondrite-normalized REE show no Eu anomalies ($Eu/Eu^* = 0.90–1.19$), medium to high total REE contents and flat to fractionated REE patterns ($La_N/Yb_N = 1.1–11.5$). The primitive mantle normalized multi-element plots deep negative Nb and Ti. In the tectonic discrimination diagrams of Meschede (1986), Wood (1980) and Pearce (2008), the Khan-Taishir basaltoids show a clear subduction-related signature (Buriánek et al., 2017). The mafic varieties dominate over felsic rocks. The typically low Zr/Th, Nb/Yb and Th/Yb ratios suggest formation of the Khan-Taishir mafic volcanic assemblages at an intra-oceanic arc. This accords well with a signature of mantle wedge melting in the Nb/Yb vs. TiO_2/Yb diagram of Pearce (2008), which is typical of island arcs. The Hf-in-zircon data from a gabbro-diorite show its derivation from a juvenile magma source. The Nd isotopic compositions of basaltic lavas, metatuff and peridotite samples are all highly radiogenic ($\varepsilon Nd_T = +6.5$ to $+10.5$) (Table 2) suggesting their derivation from depleted mantle. The Khan-Taishir ophiolite also hosts ultramafic rocks and gabbro-diorite ($SiO_2 = 47–52$ wt%) showing a tholeiitic trend in the AFM plot (Buriánek et al., 2017). The REE patterns of plutonic rocks are mostly flat with few samples of gabbro showing elevated LREE contents ($La_N/Yb_N = 3.4$). The whole set of geological, geochronological and geochemical data from the Khan-Taishir ophiolites indicates that a subduction zone was initiated along the NE active margin of Gondwana at around 570–550 Ma (e.g. Gibsher et al., 2001; Buriánek et al., 2017). That giant subduction system explains the origin of supra-subduction ophiolites along the whole margin of Baydrag-Dzabkhan and Tuva-Mongolian microcontinents (Salnikova et al., 2001; Soejono et al., 2016).

4.4. Northern Mongolia - southern Transbaikalia: Dzhida

The Dzhida island arc or terrane occupies a large area in southern Transbaikalia and northern Mongolia (no. 10 in Figs. 4, 6 and 10), approximately 600 to 700 km (Badarch et al., 2002; Gordienko et al., 2007). It is located between the Tuva-Mongolian microcontinent and Khamar-Daban terrane and unites three different types of complexes: relicts of island arc, seamount OPS and flysch of marginal paleo-basins (back-arc basins). The island arc terrane includes a plagiogranite-tonalite-diorite complex and an ophiolitic complex consisting of mafic-ultramafic, boninite-basaltic, rhyolite-andesitic, granitoid, carbonate, and tuff assemblages. The mafic-ultramafic complex consists of serpentinite, pyroxenite, dunite, harzburgite, gabbro, talc-carbonate rocks, and tholeiitic plagiogranites, all typically sheared and deformed (Gordienko and Filimonov, 2005). The mafic-ultramafic complex represents the lower part of the ophiolite. The intra-oceanic arc consists of tholeiitic and calc-alkaline basalts, boninite, and dolerites plus small gabbro-dolerite, microgabbro, and pyroxenite bodies. Basalts are tholeiitic and calc-alkaline (Gordienko and Filimonov, 2005). The tholeiitic basalts are characterized by lower SiO_2, Al_2O_3, Cr and Ni and higher TiO_2 and FeO compared to the calc-alkaline basalts. The tholeiitic and calc-alkaline basaltic series formed on an early stage of intra-oceanic arc development. Evidence for this comes from the presence of boninites (Almukhamedov et al., 2001) in the Bayan-Gol accretionary prism, where they occur as dikes and large blocks of fine-grained volcaniclastics. The large blocks of boninite volcaniclastics contain thin boninite and basalt lava flows suggesting their eruption in a fore-arc submarine environment. The late Neoproterozoic to early Cambrian age of the intra-oceanic arc is constrained by the 800–740 Ma Nd model ages of island-arc gabbro and granitoids also having high positive εNd(T) (+6.4 to +7.6), 570–510 Ma U—Pb ages of island-arc ophiolites and granitoids (Gordienko et al., 2006, 2012) (Table 2), and Early Cambrian fossils in carbonates of the Dzhida seamount located near the island-arc unit (Gordienko et al., 2007). The basement of the Dzhida seamount is intruded by Early Ordovician granitoids (Reznitskii et al., 2005). All these complexes (arc, seamount, back-arc basin) were juxtaposed during the Late Paleozoic collision of the Siberian and North China continents (Buslov et al., 2004; Gordienko and Filimonov, 2005).

In summary, the distribution of late Neoproterozoic to early Cambrian island arcs in Transbaikalia-Mongolia suggests four successive periods of birth and death of island arc systems. The oldest Dunzhugur arc evolved during the middle to late Neoproterozoic (Kuzmichev et al., 2001). The next came the Shishkhid arc of ca. 800 to 600 Ma age (Kuzmichev et al., 2005) and the Ilchir arc of late Neoproterozoic age (Dobretsov et al., 1989). The three Dunzhugur, Shishkhid, and Ilchir arc terranes are now located south-west of Lake Baikal and north of Lake Darkhad (Figs. 6, 7). The Agardag, Shatskii, Tannu-Ola and Kurtushibin arcs are all of late Neoproterozoic to early Cambrian age and possibly represent a single arc system, which now extends from NW Mongolia to western Tuva-Sayan. The Dariv-Khantaishir arc system is located in western Mongolia and formed at ca. 570–500 Ma (Figs. 6, 10). The Dzhida arc is located to the east of all three arc systems and evolved during a long period from ca. 800 Ma to the Cambrian. All these arcs form a kind of orocline (Fig. 6).

4.5. Russian Altai: Kurai-Ulagan

The Altai orogenic belt extends for ca. 2500 km from Russia (SW Siberia), through eastern Kazakhstan and northwestern China to southwestern Mongolia, bounded by the Junggar block in the southwest and the Sayan Mts. in the northeast, and occupies the central-southern portion of the CAOB. The Russian Altai is the northern branch of the Altai orogen in the western CAOB (no. 11 in Fig. 6, Fig. 12) hosting at least one primitive or intra-oceanic arc, which were also referred to as the Kuznetsk-Altai primitive island arc system (Buslov et al., 2001, 2002). The boninite-bearing island arc is structurally and spatially related to the Kurai accretionary complex of the Gorny Altai terrane also hosting late Neoproterozoic-early Cambrian OPS (Buslov et al., 1993, 1998; Safonova et al., 2008, 2011b; Safonova, 2014). The island arc complex is best outcropped northeast of Chagan-Uzun Village (Fig. 11) and extends to the west and north-west to the Ulagan suture-shear zone with ophiolites boninites and blueschists (Buslov et al., 2002; Chen et al., 2016). The island arc consists of volcanogenic-sedimentary rocks, dikes and sills, sheeted dikes, and layered gabbro-pyroxenite consisting of gabbro, clinopyroxenite, wehrlite and serpentinite. The tholeiite–boninite rocks are intruded by dikes of calc-alkaline quartz-diorite and plagiogranite. The dike, sills and sheeted dikes include dolerite, gabbro and boninitic rocks suggesting their relation to a primitive island arc (Buslov et al., 1998, 2002). The late Neoproterozoic to early Cambrian age of the island arc is constrained by the 647 ± 80 Ma age of Chagan-Uzun clinopyroxenite (Dobretsov et al., 1995), the 598 ± 25 Ma age of the limestones associated with Kurai seamount basalt (Uchio et al., 2004), by the Ar—Ar ages of ca. 630–560 Ma of Chagan-Uzun eclogite (Buslov et al., 2002), by the ca. 500–472 Ma U—Pb zircon ages of Ulagan meta-sedimentary rocks (Chen et al., 2016) and by the early Cambrian microfossil age (microphytolites, algae, siliceous sponge spicules) of the sediments of the Katun seamount (Postnikov and Terleev, 2004).

Geochemically, the Chagan-Uzun boninites are similar to those of the western Pacific (Simonov et al., 1994; Buslov et al., 1998): SiO_2 = 53.5–58.7, TiO_2 = 0.16–0.42, MgO = 9.3–16.1, Al_2O_3 = 4.6–12.5, Fe_2O_3 = 8.1–10.8 wt% (Figs. 3, 8; Suppl. Electr. Mat.). The calc-alkaline rocks are less abundant and occur in the lower part of the Chagan-Uzun structure as a sheet of andesitic lava, tuffaceous and sedimentary rocks (siliceous mudstone, limestone, etc.), which is thrust over the ophiolite complex. The volcanogenic-sedimentary sequence is mainly composed of micritic and massive reef limestones interbedded with green-gray chlorite-bearing shales and volcaniclastic sandstones. Boudinated and deformed gabbro, gabbro-diabase, and diabase dykes cut the lower ophiolite sheet and are compositionally close to the calc-alkaline island arc series and probably have early to middle Cambrian age. The metasediments formed after island-arc turbidites possessing positive εNd values (+4 to +6) (Chen et al., 2016). The whole island arc unit consists of five packages of tectonic sheets forming a north-west striking synfold. The lower 1.5 km thick package consists of Lower-Middle Cambrian fore-arc turbidites of a normal, more evolved island arc (Fig. 11). The second is a 1 km thick package of primitive island arc sheets composed of gabbro-diabase, sheeted dyke complex, sills of tuffs and volcanic rocks. The third package includes up to 1 km thick sheets of gabbro-pyroxenites. The fourth is a 1.5–2 km thick package of carbonate-terrigenous black shales and turbidites of the outer slope of the primitive or intra-oceanic arc or marginal sea. The uppermost fifth package is up to 0.5–1 km thick and is compositionally and lithologically similar to Package 2. The total thickness of the tectonic packages of the Kurai zone including the Chagan-Uzun area exceeds 5–6 km. Boninitic to tholeiitic lavas, dike-sill and turbidite complexes belong to an intra-oceanic arc. In general, the temporal and lateral polarity of magmatic rocks is typical of volcanic arcs: older primitive-arc tholeiite-boninite series (Late Neoproterozoic to Early Cambrian) are followed by younger normal-arc tholeiitic and calc-alkaline arc volcanic series. Laterally, various volcanic units are recognized within large fragments of more evolved island arcs ranging in composition from tholeiitic (high-Mg andesite and basalt) in frontal parts, through calc-alkaline in central parts to shoshonitic in back-arc areas (Buslov et al., 1993, 2002; Simonov et al., 1994).

5. Early Paleozoic IOAs of the Kazakh Orocline

Degtyarev (2011, 2012) proposed to identify Early Paleozoic intra-oceanic arc terranes as the Sary-Arka volcanic arc in the Kazakhstan Orocline (Fig. 12). The Saryarka volcanic arc forms a horse-shoe shaped

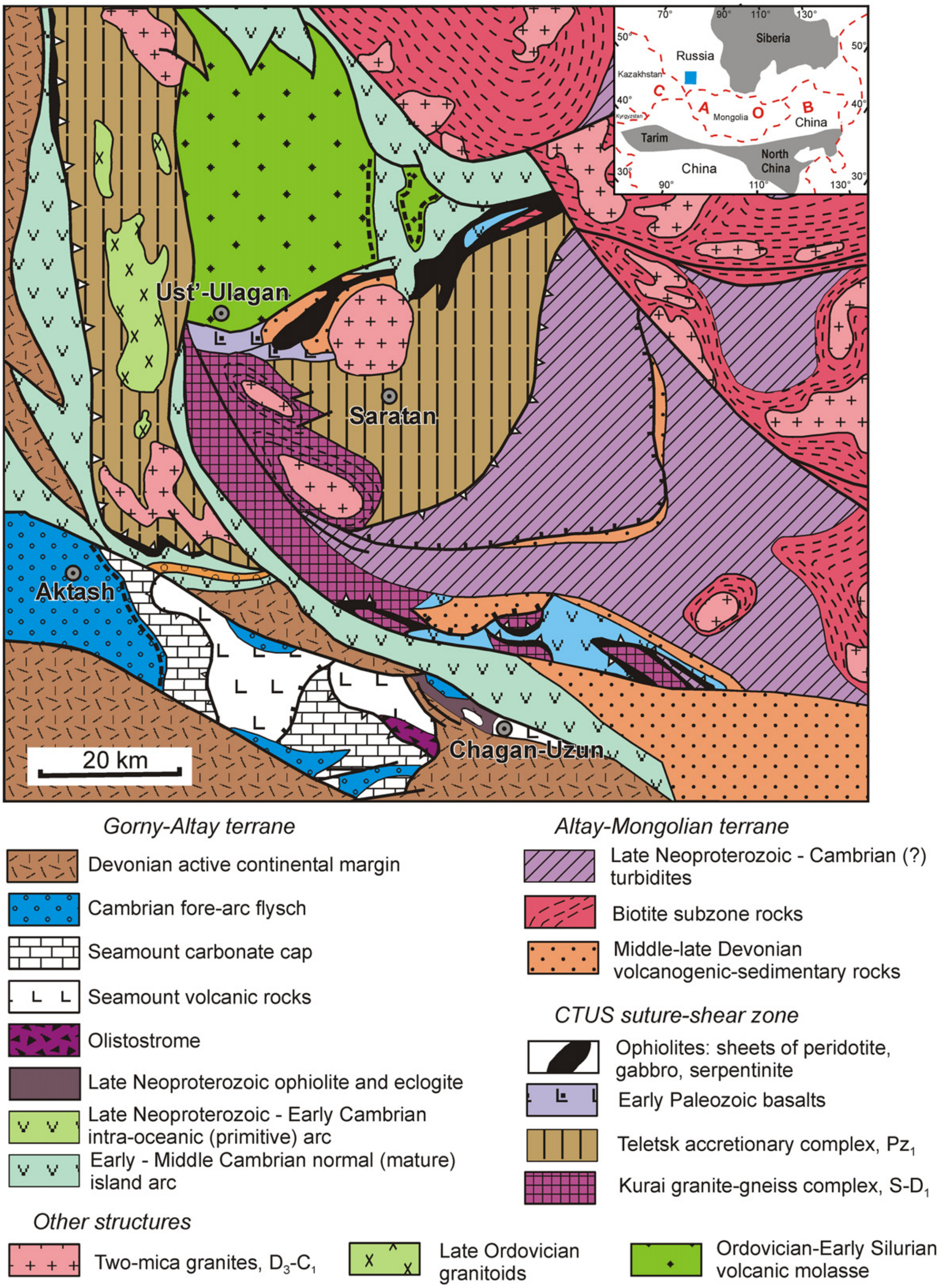

Fig. 11. The location and geological framework of the Kurai-Ulagan island-arc system in the Russian Altai (modified from Buslov and Watanabe, 1996; Chen et al., 2016).

bend extending from the northern Tienshan to the Kokchetav Precambrian massif in the north-west and then to north-eastern and eastern Kazakhstan over a total distance of more than 2000 km. The belt consists of tectonically juxtaposed volcanic and plutonic complexes of Cambrian – early Ordovician intra-oceanic arcs (referred to as ensimatic arcs by Degtyarev, 2011) and OPS units (ultramafic and mafic rocks, basalt, chert and other siliceous sediments). These complexes occur as tectonic sheets. The better preserved segments of intra-oceanic arcs are in Urumbai-Selety (north), Bozshakol-Chingiz (NE and east) and Baydaulet-Aqbastau (Degtyarev, 2011).

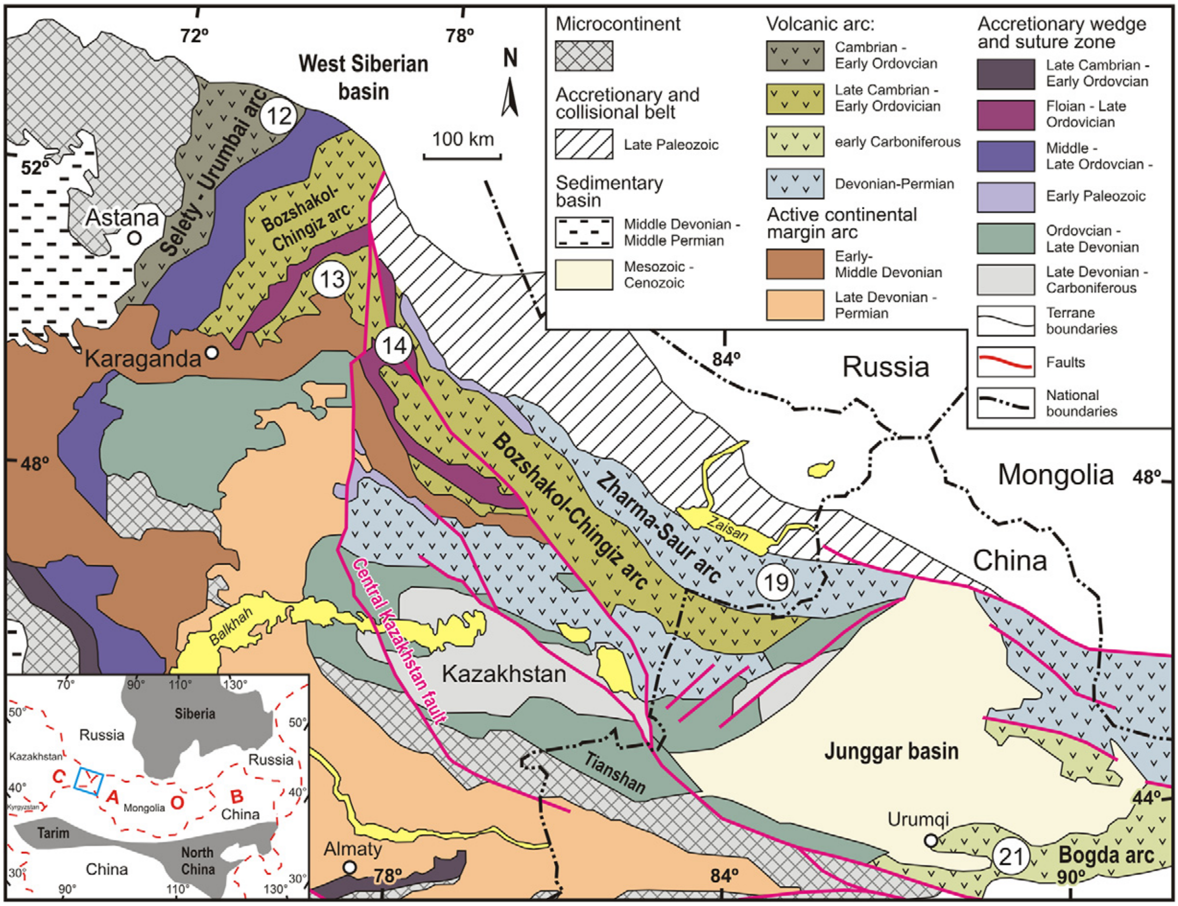

Fig. 12. The tectonic map of the Kazakh orocline and adjacent areas (modified from Windley et al., 2007; Degtyarev, 2012).

5.1. Early to middle Cambrian: Selety-Urumbai

The early-middle Cambrian Selety and Urumbai arcs or a single Selety-Urumbai arc (no. 12 in Figs. 6, 12) are dominated by volcanic and plutonic rocks of an intra-oceanic arc overlapped by carbonate-siliceous sediments and tuffs. The late Cambrian to Early Ordovician units are differentiated volcanic and plutonic igneous rocks of an island-arc origin plus coeval OPS rocks (MORB, chert, and siliceous fine clastics) and Early Ordovician–Darriwilian flysch and siliceous-terrigenous deposits. The early–middle Cambrian island arc complexes are ophiolites including granitoids, bimodal basalt–rhyolite series and boninite-bearing differentiated volcanic series. The ophiolite section consists of peridotites, layered and massive gabbro, sheeted dikes of dolerites and overlying pillow basalts. The gabbros are overlain by tonalities and then by younger plagiogranites, which U—Pb ages range from 525 to 519 Ma (Ryazantsev et al., 2009; Degtyarev et al., 2010). The well-preserved bimodal basalt–rhyolite series probably marks a back-arc rifting near continental margin, which gave rise to an intra-oceanic arc. The mafic volcanics are dominantly basalt and andesibasalt with subordinate calc-alkaline varieties (Fig. 8B, C). Intermediate varieties (andesite, dacite, trachyandesite) are subordinate. The felsic volcanics are rhyolite, rhyodacite, dacite, lavabreccias, and tuffs. The Early Cambrian age of the bimodal volcanic series is constrained by fossils found in the Selety zone (Ivshin et al., 1993). The differentiated volcanic series are early to middle Cambrian, i.e. younger than the basalt–rhyolite series (Degtyarev, 2012). In general, the plutonic and volcanic rocks of the Sary-Arka belt possess geochemical features of supra-subduction magmas, in particular, those of the Izu–Bonin IOA or back-arc basins. In the Urumbai zone (northern and eastern segments), the volcanic rocks (basalt, andesite, dacite) have primitive isotopic signatures (Degtyarev, 2012), some are close to boninites suggesting their formation in an intra-oceanic arc setting. The age of the island-arc volcanic complexes spans the late Cambrian - Tremadocian interval in the southern, southwestern, northern, and northeastern segments and the late Cambrian to late early Ordovician interval in the eastern segment of the Kazakh orocline (Degtyarev, 2011) (Fig. 12). There is no concensus among researchers regarding if there was one Selety-Urumbai arc or there were two Selety and Urumbai arcs, but the Selety and Urumbai volcanic complexes definitely have different ages and are characterized by different sets of dominant lithologies. However, early Cambrian ophiolites with supra-subduction signatures occur in the neighboring Bozshakol zone, although no detailed geochemical or isotopic data have been published yet. Therefore, it is quite probable that all these complexes once represented a single arc, which volcanic front should have migrated, probably from east to west (K. Degtyarev, pers. comm.).

5.2. Late Cambrian to Early Ordovician: Bozshakol-Chingiz

Late Cambrian to Early Ordovician island arc complexes have been identified in all segments of the Sary-Arka belt. They have tectonic contacts with the early–middle Cambrian island-arc units, Cambrian-Ordovician OPS units and Paleozoic granitoid complexes emplaced along the belt (Degtyarev, 2011). The northeastern and eastern segments of the Saryarka belt host a 100 km long Bozshakol-Chingiz arc (Degtyarev and Ryazantsev, 2007; Shen et al., 2015) also referred to as Boschekul-Chingiz (Degtyarev, 2011) or Boshchekul-Chingiz (Windley et al., 2007) consisting of compositionally diverse volcanic series and granodiorite plutons (no. 13 in Figs. 6, 12). The U—Pb zircon ages of the volcanic and plutonic rocks of the Bozshakol-Chingiz arc in the eastern segment (no. 13a in Fig. 6) are 501.8 ± 3.2 and 489.5 ± 3.3 Ma, respectively

(Shen et al., 2015). The Bozshakol tholeiitic to calc-alkaline basalt and calc-alkaline andesite and dacite (Fig. 8) are enriched in light REE with a marked negative Nb anomaly and Th/Yb-enrichment (Suppl. Electr. Mat.). They also have low initial $^{87}Sr/^{86}Sr$ ratios (0.7026–0.7048), high zircon εHf(t) and whole-rock εNd(t) values (+9.7 to +17.0 and +5.4 to +6.7, respectively). The Bozshakol plutonic rocks are fine- and medium-grained medium-K calc-alkaline tonalite porphyries possessing geochemical characteristics similar to normal arc granitoids (low initial $^{87}Sr/^{86}Sr$ ratios = 0.7036–0.7039, high zircon εHf(t) = +10.7 to +17.2 and positive whole-rock εNd(t) = +4.9 to +5.7). The medium-grained tonalite porphyries possess geochemical affinities to adakites. Shen et al. (2015) suggest that both, the volcanic and plutonic rocks, were derived by partial melting of the mantle wedge and subducted slab.

The southern part of the Bozshakol-Chingiz arc is dominated by chert, tuff, tephra and mafic, andesitic and felsic pyroclastics, which probably deposited on island arc slopes and in adjacent marginal basins (Degtyarev, 2011). These volcanogenic-sedimentary sequences overlie the island-arc ophiolites composed of ultramafic rocks, gabbroids, parallel dikes, aphyric basalts and associated OPS (chert). The volcanic rocks of the north-eastern segment (no. 13b in Fig. 6) are subalkaline and alkaline basalts and andesites with SiO_2 = 47.7–63.6, TiO_2 = 0.2–0.9, MgO = 2.1–8.1, Al_2O_3 = 15.1–19.6, Fe_2O_3 = 5.0–13.1 wt% (Suppl. Electr. Mat.). The subalkaline varieties belong to the calc-alkaline series. Their REE patterns show moderate enrichment of light REE and undifferentiated heavy REE. The multi-element diagrams show clear negative peaks at Nb and Ti. Thus, the geochemical compositions of the rocks, both volcanic and plutonic, suggest their formation in a supra-subduction setting. In the northern segment, the late Cambrian–Early Ordovician volcanics of the lower part of the section, i.e. arc basement, possess boninitic characteristics (Fig. 8 B, C). The other segments are dominated by andesitic volcanics with subordinate basalts. In general, the geochemical data suggest significant fractionation of initial melts, i.e. large thickness of the island arc crust and the related formation of intermediate magma chambers. The Sr and Nd isotope ratios obtained from the arc volcanic rocks and granitoids (Table 2) also suggest rather primitive and melanocratic island arc basement. The occurrence of chert–tuff packages overlying the ophiolite section suggest that the late Cambrian–Early Ordovician island arc was separated from a continental margin by a marginal sea (back-arc) basin with oceanic-type crust. The late Cambrian – early Ordovician volcanic rocks are conformably overlain by Tremadocian–Arenigian (northern segments) or upper Arenigian–lower Llanvirnian (eastern segment) siliceous tuffs and siliceous–terrigenous sequences suggesting the cessation of volcanism in early Ordovician time and island arc retreat (Degtyarev and Ryazantsev, 2007; Degtyarev, 2011, 2012).

5.3. Floian to late Ordovician: Baydaulet-Aqbastau

One more island arc has been identified in the Baydaulet–Aqbastau volcanic belt by K. Degtyarev (2011, 2012) although no detailed geochemical and isotope data have been obtained yet. The Baydaulet–Aqbastau volcanic belt is located south of the Selety-Urumbai and Boschekul-Chingiz arcs, more inside the Kazakh orocline, and extends for a distance of ca. 900 km (no. 14 in Figs. 6, 12; Table 2). It is composed of Ordovician volcanic and volcanic–sedimentary complexes conformably overlain by Silurian terrigenous sequences and unconformably by Devonian volcanic and volcanic–sedimentary sequences. The structure and rock composition of the Lower–Middle Ordovician complexes are different from those of the Late Ordovician ones. There are two types of Lower–Middle Ordovician island-arc complexes, which formed over a mafic, probably oceanic crust. The first one is dominated by aphyric pillow basalts and associated chert, intermediate and acid tuffs and tephroids, and tuffs (Satpaev and Balatundyk–Otyzbes zones in the NW and SE segments of the belt, respectively). The second type is dominated by basalt, basaltic andesite, pyroclastic rocks with subordinate cherty siltstone, tuffite, and jasper (North Qaraghandy and Aqbastau–Qosmurun zones in the NW and SE segments of the belt, respectively). The Late Ordovician complexes consist of basalt, basaltic andesite, andesite, dacite, and rhyolite and their tuffs. In the North Qaraghandy and Aqbastau–Qosmurun zones the volcanism was followed by the emplacement of granite–granodiorite intrusions. The mafic rocks of oceanic affinities, mostly basalts, occur at both, northwestern and southeastern, sides of the belt. The Baydaulet–Aqbastau island arc is bounded by the Esqembay–Balqybek, Teqturmas and North Balqash ophiolite zones of the Zhungar–Balqash Region. Thus, the structural position of the island-arc complexes and the composition of their hosted volcanic rocks suggest that the Baydaulet–Aqbastau belt was a large island-arc system, which existed in the western Paleo-Asian Ocean during almost the whole Ordovician. During the Middle Ordovician, the interarc basin separated the Chingiz–North Tien Shan and the Baydaulet–Aqbastau arcs. During the Late Ordovician, the basin started to shrink and fully closed by the end of Ordovician to form the narrow Esqembay–Balqybek ophiolite belt (Degtyarev 2011, 2012).

6. Ordovician to Middle Paleozoic intra-oceanic arcs

Remnants of intra-oceanic arcs have been recognized in the western and southwestern CAOB: in the Tienshan orogenn (Burtman, 2008; Biske and Seltmann, 2010), in East Kazakhstan and East Junggar and in southwestern Mongolia. The Tienshan orogen extends over the territories of Uzbekistan, Tajikistan, Kyrgyzstan and China (Figs. 6, 13), where it is traditionally divided into three Paleozoic tectonic domains, northern, middle and southern (Bakirov and Maksumova, 2001). Some arc or supra-subduciton units extend from Tajikistan to Uzbekistan (Chatkal-Kurama arc) or from Uzbekistan to Kyrgyzstan (from Kyzylkum and Nurata areas in Uzbekistan to Chatkal and Atbashi areas in Kyrgyzstan; Dolgopolova et al., 2017 and Alexeiev et al., 2016, respectively) and farther to NW China making their geographic references confusing. Several authors mentioned a probability of the presence of remnants of intra-oceanic arcs within metamorphic terranes formed after subduction-accretionary complexes, for example, in the Fan-Karategin belt in Tajikistan or Atbashi belt in Kyrgyzstan (Volkova and Budanov, 1999; Alexeiev et al., 2016). Numerous episodes of strong intra-continental deformation and metamorphism accompanied the oceanic suturing in the Turkestan Ocean and/or Junggar oceans (Fig. 5B, C). Those IOAs typically lack boninites or reliably identified coeval back-arc basins. Consequently, there are few intra-oceanic arcs of Silurian-Devonian age, which intra-oceanic origin has been well constrained geologically, geochronologically and geochemically.

6.1. Kyrgyz Tienshan: Chatkal-Atbashi

A new arc system of Middle Ordovician age (470–460 Ma) has been recently recognized south of the Kyrgyz Middle Tienshan microcontinent (Alexeiev et al., 2016). The two fragments of this magmatic arc are located within the northern Atbashi and southern Chatkal ranges. The Chatkal-Atbashi arc (no. 15 in Figs. 6 and 13) consists of three units: volcanic, metamorphic and ophiolitic. The Middle to Late Ordovician volcanic sequence is dominated by rhyolite, dacite and sub-volcanic porphyrites and tuff with subordinate andesitic and mafic varieties. The volcanic rocks are calk-alkaline and sub-alkaline and have whole-rock Nd- and Hf-in-zircon isotopic data implying rather a continental arc (Table 2), possibly similar to that of the modern Japanese Islands (Jahn, 2010): εNd(t) ranges from +0.9 to −2.6 and εHf(t) from +1.8 to −6.0 (Alexeiev et al., 2016). Geological data show that in the north the arc was separated from the microcontinent by a back-arc basin, which is represented by the Karaterek ophiolite belt at the northern slope of the Chatkal Range (Ivanov et al., 2002). The ophiolite consists of pyroxenite, gabbro-amphibolite, metagabbro, metabasalt, and pillow basalt, which are associated with siliceous shale, chert, black siltstone, and turbidite. The major element composition of amphibolite and metabasalt were interpreted as back-arc material (Ivanov et al., 2002):

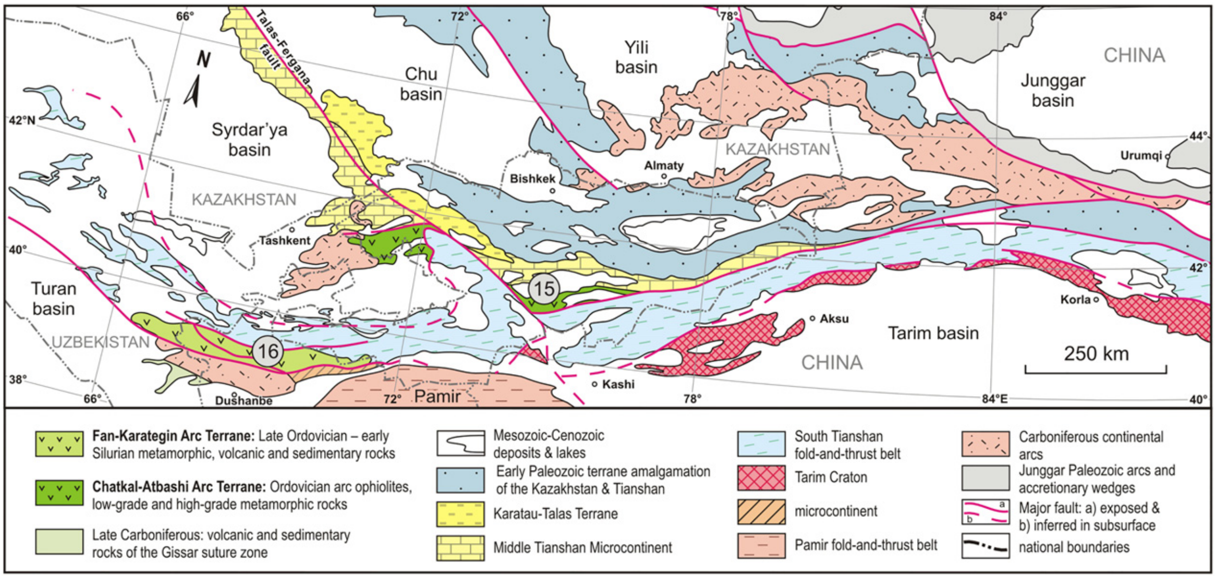

Fig. 13. Simplified geology of the Tian-Shan orogen showing the Fan-Karategin, Kysylkum-Nurata and Chatkal-Atbashi arcs (based on Volkova and Budanov, 1999; Biske and Seltmann, 2010; Alexeiev et al., 2016).

they are similar to MORB and the igneous rocks are associated with deep-marine sediments. The same ophiolites were also considered as an intra-oceanic arc by Burtman (2006). The age of the ophiolites is constrained as Early to Middle Ordovician by conodonts preserved in chert (Ivanov et al., 2002). During the late Ordovician, island arc collided with the microcontinent. The timing of the collision is constrained by the cessation of sedimentation on the Middle Tienshan microcontinent, by the age of an angular unconformity within the Karaterek suture zone, and by the age of syncollisional metamorphism and magmatism of the southern Chatkal Range (Fig. 13) (Alexeiev et al., 2016). The disappearance of a major crustal block with transitional facies on the continental margin and the too short distance between the arc and the adjacent coeval accretionary complex suggest that the oceanic subduction beneath the Chatkal-Atbashi arc and subsequent plate convergence were accompanied by subduction erosion. This can explain the removal of the Ordovician arc as well as the Silurian and Devonian fore-arcs east of the Talas-Ferghana Fault (Alexeiev et al., 2016; Safonova, 2017). Thus, Alexeiev and co-authors (2016) recognized a long-lived early Palaeozoic arc in the southwestern Middle Tienshan suggesting an oceanic domain between the microcontinent and the Tarim craton in the Middle Ordovician. In addition, a younger arc was proposed in the western extension of the Chatkal-Atbashi domain in Uzbekistan (see Section 6.3 below) (Dolgopolova et al., 2017). Both suggest that the Turkestan basin remained quite wide during a long period from the Middle Ordovician to the late Silurian.

6.2. Tajik Tienshan: Fan-Karategin

The Tajik Tienshan represents the southwestern segment of an extensive middle-late Paleozoic domain including oceanic, passive margin and accretionary units and their metamorphic belts of both the South and Middle Tienshans, which extend from western Uzbekistan to the Chinese South Tienshan (Volkova and Budanov, 1999; Konopelko et al., 2017; Fig. 13). The Fan-Karategin low-grade metamorphic belt is located in Tajikistan (no. 16 in Figs. 6 and 13). It isan approximately 300 km long and 40 km wide WE-trending belt bounded by the Zeravshan fault in the north and by Gissar–Karategin fault in the south (Fig. 13). The Fan-Karategin belt probably extends to the South Tianshan orogen in NW China (Gao et al., 1995; Volkova and Budanov, 1999). The Fan-Karategin belt represents a former subduction–accretionary complex and consists of three tectonostratigraphic units of different ages and lithologies: oceanic, island arc and clastic (Volkova and Budanov, 1999). The 1200 m thick oceanic or OPS unit (Gorif Series) is composed of tectonic sheets of OIB-type alkali and tholeiitic metabasalts associated with metachert and metacarbonate (marble) of Middle Ordovician - early Silurian age (Vinogradov and Torshin, 1963). These rocks are regarded as former pelagic and shallow-water carbonate sediments developed upon oceanic islands (Volkova and Budanov, 1999; Safonova and Santosh, 2014). The seamount basalts and pelagic sediments were metamorphosed in the blueschist and greenschist facies. The oceanic unit is tectonically overlain by a 1000 m thick metavolcanic arc-derived unit, Norvat Series (Late Ordovician – early Silurian). The volcanic rocks are basalt, andesite, rhyolite and trachyte possessing geochemical characteristic of supra-subduction basalts. The arc volcanic unit is overlapped by a clastic unit consisting of Late Ordovician to early Silurian metasedimentary rocks of the Yagnob Formation (bottom to top): graywackes containing volcaniclastic arc-derived material, mudstones, tuffs and volcanoclastic sandstones (Volkova and Budanov, 1999). The Fan-Karategin arc has a very complicated structure of numerous tectonically mixed, stacked and overthrust sheets and slices. It still lacks precise geochemical and geochronological data from igneous and sedimentary rocks. The only three available analyses show tholeiitic andesite and dacite (Fig. 8C; Suppl. Electr. Mat.). All this, coupled with its metamorphic nature, makes the verification of its tectonic origin, intra-oceanic vs continental margin arc, hard. However, the overall predominance of mafic supra-subduction volcanics, presence of accreted seamounts and blueschists formed after OIB, MORB and arc basalts, pelagic, hemipelagic and turbidute sedimentary lithologies, and absence of abundant coeval granites nearby suggest rather an intra-oceanic then a continental magmatic arc. In addition, a back-arc basin was proposed in the neighboring Alai range of Kyrgyzstan, i.e. to the east of the Fan-Karategin belt (Alexeiev et al., 2015).

6.3. South Mongolia: Gurvansayhan-Zoolen

The Gurvansayhan-Zoolen arc is located in southern Mongolia and extends for a distance of >800 km (no. 17 in Figs. 6 and 10). According to Badarch et al. (2002) island arc units occur within the Gurvansayhan and Zoolen terranes bounded by the Gobi Altai back-arc/fore-arc terrane in the north. The Gurvansayhan island arc terrane is dominated by Silurian-Devonian island arc tholeiites and andesites and middle Devonian volcaniclastic rocks (Badarch et al., 2002; Helo et al., 2006). The arc units are covered by middle Carboniferous terrigenous sediments and intruded by Silurian to late Permian granitoids (Badarch et al. 2002). The Zoolen terrane was previously regarded as accretionary, but later Helo et al. (2006) showed it was rather an island arc. The Zoolen terrane is ca. 500 km long and up to 80 km wide (Fig. 10). It consists of strongly deformed rocks in the greenschist-facies of metamorphism. Thrust sheets, tectonic slivers, and mélange contain Ordovician-Devonian pillow basalts and andesites (Badarch et al., 2002). The volcanic rocks are associated with clastic sediments and limestones and contain fragments of peridotite, serpentinite, and gabbro. Post-accretionary units are similar to those in the Gurvansayhan island arc and comprise volcanic and sedimentary rocks (Badarch et al., 2002). The U—Pb zircon age of andesidacites of the Zoolen terrane is 421–417 Ma (Helo et al., 2006) (Table 2). These ages are consistent with the late Silurian to early Devonian paleontological ages of the Gurvansayhan and Zoolen sedimentary assemblages (Ruzhentsev et al., 1985). The volcanic rocks of the Gurvansayhan and Zoolen terranes are dominated by andesitic and basaltic varieties of boninitic, tholeiitic and calc-alkaline series (Fig. 8B–D). The basalts have low TiO_2 (0.5–1.3 wt%) and high Mg-numbers (~70), indicating near primary mantle-derived compositions. The andesites are moderately fractionated with Mg-numbers of 50–60. There are andesites with high MgO (7.1 wt%) and SiO_2 (60 wt%), i.e. similar to boninites (Helo et al., 2006). The calc-alkaline voclanics (andesites to rhyolites) are characterized by differentiated REE and concave HREE patterns due to fractionation of hornblende, which is typical of thick and mature island arcs. The tholeiitic basalts have low Th/Nb (~ 0.3) and low-enriched LREE (Helo et al., 2006) suggesting a very small amount of recycled sediments that is typical of immature island arcs (Stern, 2010). So, the Gurvansayhan and Zoolen volcanics are enriched in light REE and depleted in HFSE. Conclusively, the geochemical data in combination with high initial εNd-values of ca. +6 to +9 (Helo et al., 2006; Table 2) are consistent with a juvenile intra-oceanic arc. There are also volcanic varieties compositionally similar to adakites, which, together with boninites, suggest a fore-arc setting. The volcanic rocks from adjacent terranes (Gobi Altai) also have high initial εNd values (+6 to +10) and exhibit calc-alkaline, LREE enriched island arc as well as tholeiitic LREE-depleted back-arc basin signatures of predominantly juvenile composition (Helo et al., 2006). The formation of the Gurvansayhan and Zoolen arc terranes is attributed to the evolution on the South Mongolian branch of the Paleo-Asian Ocean (Zonenshain et al., 1990; Dobretsov et al., 1995; Buslov et al., 2001).

6.4. Chinese Altai: Saerbulake

The Chinese Altai is the central segment of the Altai orogen (see Section 4.5). According to the terminology of Cai et al. (2011) it consists of four domains: i) late Devonian–early Carboniferous island-arc domain (North Altai); ii) middle Ordovician–early Devonian dominantly turbiditic-pyroclastic active continental margin domain (Central Altai); iii) the Qiongkuer domain of late Silurian sedimentary rocks, early Devonian arc-related pyroclastic rocks and middle Devonian turbiditic rocks, and iv) a domain of Devonian sediments and late Carboniferous volcaniclastic rocks (South Altai). Further south, the Chinese Altai is separated from the Junggar basin by the Erqis Fault (Fig. 14), one of the large-scale (1000 km) sinistral strike-slip faults in Central Asia (Xiao et al., 2010).

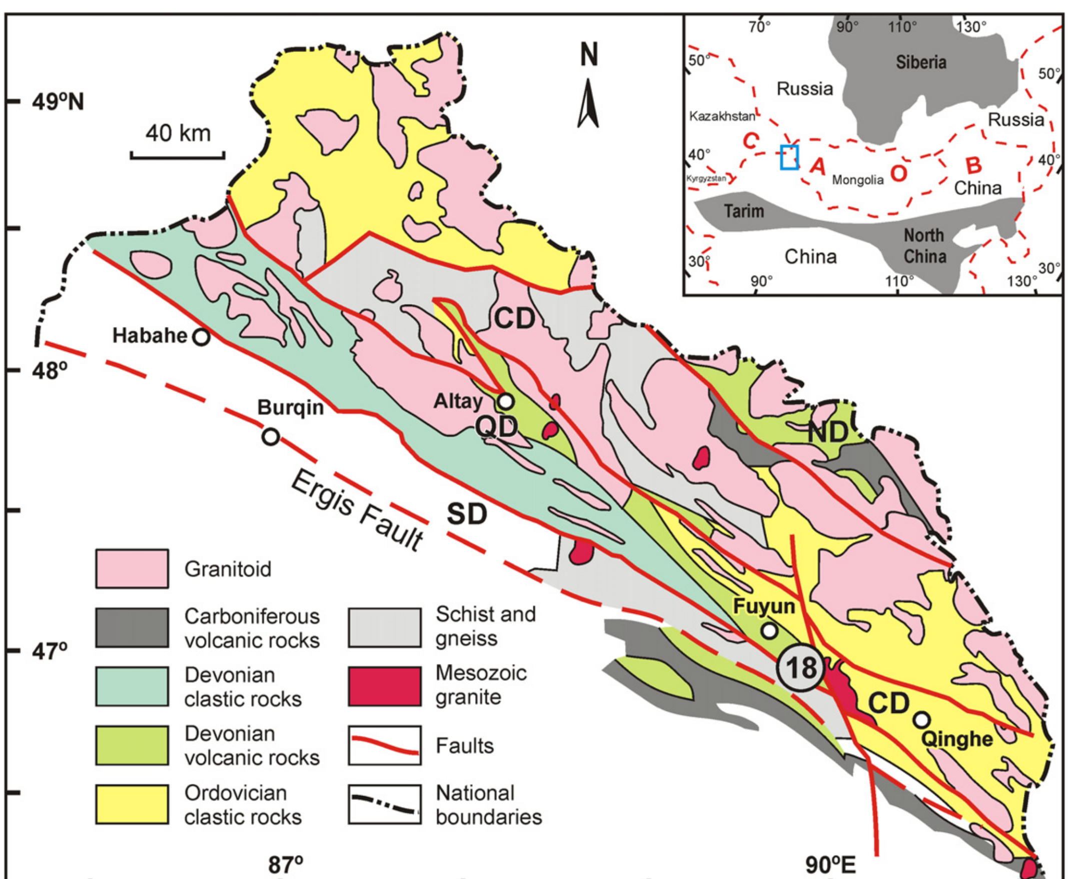

Fig. 14. Geology of the Chinese Altai (modified from Windley et al., 2002; Yuan et al., 2007). Domains (as in Yuan et al., 2007): northern (ND), southern (SD), central (CD) and Qiongkuer (QG). The Saerbulake arc is located within the field of Devonian volcanic, southern and central domains.

The Chinese Altai includes a NW-striking Saerbulake volcanic arc (Zhang et al., 2005; Niu et al., 2006), which is located in the Central and South domains (no. 18 in Fig. 6). The Saerbulake arc extends along the north-eastern side of the East Junggar terrane and consists mainly of Devonian–Carboniferous volcanic rocks: including picrite, boninite, adakite, high-Mg andesite, and Nb-rich basalt (Niu et al., 2006). Back-arc basin units crop out along the southern margin of the Chinese Altai. The Devonian volcanic arc rocks are adakite, boninite, tholeiitic and calc-alkaline basalt to dacite, which occur as tectonic blocks in the Suoerkuduke and Saerbulake areas (Figs. 8, 14). The adakitic hornblende and pyroxene andesites (Suoerkuduke area) are characterized by high Sr/Y (120–136) and Y (14–16 ppm), SiO_2 spanning 56.3–64.3 wt% and very high Ba (739–889 ppm), and Na_2O (4.9 wt%). Their REE patterns show enriched LREE and moderately depleted HREE. The adakites have initial Sr isotope (0.70465–0.70489) and εNd(T) (+3.0 to +4.3) values suggesting their generation by melting of subducted oceanic crust. Boninites occur in the Saerbulake area as pillow lava or volcanic breccia and are associated with high-TiO_2 basalt/gabbro and low-TiO_2 basalt. The boninites are characterized by low and U-shaped REE patterns with variable La/Yb ratios. They have high SiO_2 (54.9–56.8 wt%), CaO (10.2–8.9 wt%) and MgO (8.4–9.4 wt%), Ni (95–142 ppm) and Cr (322–562 ppm) and low TiO_2 (0.26–0.28 wt%) (Suppl. Electr. Mat.) The Saerbulake boninites show La/Yb varying from 1.8 to 8.6 at relatively depleted HREE. The low La/Yb boninites show slightly U-shaped REE patterns with negative Eu anomalies, whereas the high La/Yb boninites lack significant Eu-anomalies. All these characteristics are consistent with high-Ca boninite (Crawford et al., 1989) derived from melting of mildly refractory mantle peridotite fluxed by a slab-derived fluid component under normal mantle potential temperature conditions (Niu et al., 2006). The initial Sr isotopic compositions and εNd(T) are 0.7014–0.7052 and +6.6, respectively (Table 2), i.e. similar to those of the Izu-Bonin arc (Taylor et al., 1994; Cameron et al., 1983). Thus, the Devonian formations of the southern Chinese Altay represent a tectonic juxtaposition of volcanic rocks of various origins, i.e., adakite and boninites associated with high- and low-Ti basalts and/or gabbros, respectively. The basalts, both high-Ti and low-Ti, occur as massive flows, pillowed lavas, tuff breccia, lapilli tuff and blocks in tectonic mélanges. Those mélanges formed by multiple events of accretion and subsequent deformation during the Devonian–Carboniferous suturing of the Paleo-Asian Ocean.

6.5. East Kazakhstan – West Junggar: Zharma-Saur

The Zharma-Saur terrane or, in Soviet literature, structural-formation zone (no. 19 in Fig. 6), is located north of the Bozshakol-Chingiz arc and south of the Char suture (Figs. 12, 15). It extends over a distance more than 450 km from East Kazakhstan to NW China (West Junggar). Many papers mention the Zharma-Saur zone as an arc terrane, although no detailed or up-to-date information about the composition and age of the Zharma part of the Zharma-Saur belt located in Kazakhstan have been published so far. Geological maps and explanatory notes show that the Zharma zone consist of Cambrian, Ordovician, Silurian, Devonian and Carboniferous rocks (Sokratov, 1962). The Cambrian-Ordovician sequence consistrs of metamorphosed and altered sandstone, mudstone, jasper, limestone (with trilobite) with subordinate mafic lavas and tuffs. The Cambrian part is dominated by sedimentary rocks, and the Ordovician part is dominated by volcanic rocks. The Cambrian-Ordovician rocks are overlain by Silurian rocks with an angular unconformity. The Silurian sequence consists of conglomerate, sandstone, mudstone and limestone (with brachiopods and corals) and mafic lavas and volcaniclastics. It is tectonically overlain by Early to Middle Devonian felsic volcanic rocks (albitophyre, quart porphyre), limy sandstone, pyroxene and plagioclase porphyric basalt. The Late Devonian (Frasnian) rocks are dominantly mafic volcanics and sandstones. With an angular or tectonic unconformity the Devonian rocks are overlain by early Carboniferous terrigenous rocks (sandstone, siliceous shahe and mudstone, siltstone), limestone (with brachiopods) and mafic volcanic (subordinate) rocks. Filippova et al. (2001) suggested that the Zharma arc formed on a sialic basement but not isotopic data have been obtained so far.

In NW China, the Saur part of the Zharma-Saur terrane is considered as a late Palaeozoic volcanic arc, which is separated from the early Paleozoic Bozshakol–Chingiz volcanic arc by the Hongguleleng ophiolite belt (Li et al., 2015b). Evidence for an oceanic basin between these two arcs comes from geological data and detrital zircon data. The zircon data show that the late Carboniferous sedimentary rocks of the Zharma-Saur arc are dominated by Devonian to Carboniferous detrital zircon grains (Li et al., 2016a), whereas early Paleozoic detrital zircon grains are abundant in the Boshchekul-Chingiz arc (Choulet et al., 2012). The existence of such an oceanic basin is consistent with a north-dipping subduction zone (Xiao et al., 2004, 2010).The Saur Mountains host late Devonian to early Carboiniferous arc granitoids with juvenile Hf-in-zircon isotopic charactertistics (εHf = +6 − +16) and typical arc-related geochemical signatures (Chen et al., 2010; Hong et al., 2017). In general, the Zharma-Saur arc is characterized by a transition from Devonian I-type to late Carboniferous granitoids, possibly due to a tectonic shift from subduction to a postcollisional environment (Chen et al., 2010; Li et al., 2016b).

6.6. East Kazakhstan: Char

The Char suture-shear zone or ophiolite belt is located in eastern Kazakhstan (no. 20 in Fig. 6, Figs. 12, 15). It is known to host fragments of Devonian-Early Carboniferous oceanic crust, MORB, OPB and OIB, (Ermolov et al., 1981; Safonova et al., 2012). However, recently we have identifiedd island-arc units (Kurganskaya et al., 2014). The Char volcanic and subvolcanic supra-subduction rocks are basalt, microgabbro, dolerite, andesite, tonalite and dacite. The mafic to andesitic volcanics possess low TiO_2 (0.85 wt% av.) and show MgO vs. major elements crystallization trends suggesting two magma series: tholeiitic and calc-alkaline. The tholeiitic varieties are less enriched in incompatible elements then the calc-alkaline ones. There are also high-Mg and low-Ti andesibasalts similar to boninites. The rocks possess moderately LREE enriched rare-earth element patterns and are characterized by negative Nb anomalies present in the multi-element spectra (Nb/La_{pm} = 0.14–0.47; Nb/Th_{pm} = 0.7–1.6).The distribution of rare-earth elements (La/Sm_n = 0.8–2.3, Gd/Yb_n = 0.7–1.9) and the results of geochemical modeling in the Nb—Yb system suggest high degrees of melting of a depleted harzburgite-bearing mantle source at spinel facies depths. Fractional crystallization of clinopyroxene, plagioclase and opaque minerals also affected the final composition of the volcanic rocks (Kurganskaya et al., 2014; Yang et al., 2015a). Clinopyroxene monomineral thermometry indicates crystallization of melts at 1020–1180 °C. The numerical calculations based on the compositions of melt inclusions show that primary melts were derived at 1350–1530 °C and 14–26 kbar and crystallized at 1150–1190 °C. The clinopyroxene-hosted melt inclusions show compositions close to boninites and arc tholeiites (Simonov et al., 2010). All these features are indicative of a supra-subduction origin of rocks, probably, in an intra-oceanic arc. The age of andesite and tonalite is early Carboniferous (ca. 322–336 Ma) and that of the gabbro and dolerite is Devonian (387–395 Ma) (Safonova et al., 2016c). We suggest two volcanic arcs: the Middle Devonian intra-oceanic arc and the Early Carboniferous arc of enigmatic origin. As the Char zone is strongly sheared it is impossible to reconstruct a back-arc basin there, although, in general, there are many slivers composed of deep-marine (chert) and hemipelagic (shales, mudstones, sandstones) sediments (Fig. 15). During the late Carboniferous, the island-arc units were probably accreted to the active margin of the Kazakhstan continent or collisionally squeezed between Kazakhstan and Rudny Altai.

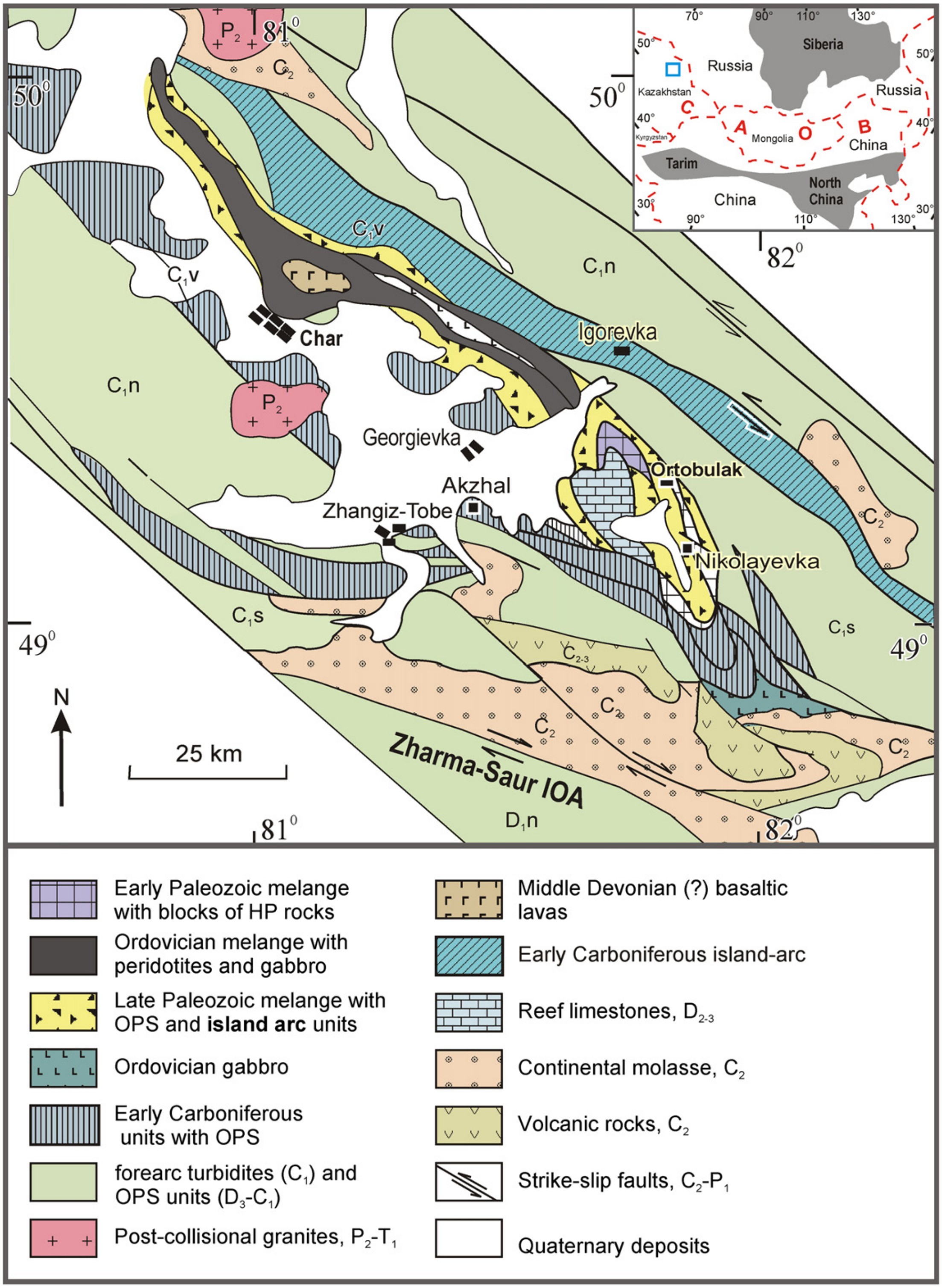

Fig. 15. Geological map of the Char zone (modified from Ermolov et al., 1981). Notably, the IOA units occur as tectonic slivers together with OPS units, all hosted by accretionary rocks with several types of mélange (Safonova et al., 2012).

7. Late Paleozoic arcs

The late Paleozoic was a period of Paleo-Asian Ocean closure and suturing. Several supra-subduction or island-arc terranes have been recognized in the eastern CAOB, an area of the termination of the Paleo-Asian Ocean (Fig. 5D, E). The most well known locality of late Paleozoic ophiolites of the eastern CAOB is the Solonker (or Sulinheer) suture, which is extended along the northern boundary of the North China Craton in Inner Mongolia and then farther along the Mongolia-China boundary. The suture zone separates the Songliao basin and the North China Craton. It includes the Hegenshan ophiolite and its hosting accretionary complex (Xiao et al., 2003; Miao et al., 2008). However, the oceanic vs intra-oceanic arc origin of the Hegenshan ophiolite remains unclear. As far as that time the Paleo-Asian Ocean was shrinking and probably was not very wide, we assess the probability of existence of late Paleozoic intra-oceanic arcs in the eastern Paleo-Asian Ocean as low. A late Paleozoic intra-oceanic arc – Bogda arc – has been identified in the Chinese North Tienshan .

7.1. Bogda arc

The E–W trending Bogda arc is located in the northern part of the Chinese North Tienshan. It extends for a distance of about 600 km and separates the Junggar Basin in the north and the Tu–Ha Basin in the south (no. 21 in Figs. 6, 12). A Devonian–Carboniferous arc terrane was proposed in this region long time ago, however its true tectonic nature (IOA or continental arc) has long time remained debatable. The Bogda arc is bounded by ophiolites. Xie et al. (2016) believe that this is further supported by the occurrence of arc-like early Late Carboniferous high-alimina basalts because those arc lavas occur in modern intra-oceanic arcs (e.g., Crawford et al., 1987). However, such an argumentation looks vague as high-Al basalt may occur in other tectonic settings. The better studied are the Bogda pillow basalts, which have relatively low TiO_2 (1.4–1.65%), FeO_t (9.5–10.7%), Nb/La (0.24–0.32) and high εNd(t) (7.9–9.4) and are considered to have formed in a rear-arc or back-arc setting (Chen et al., 2013; Xie et al., 2016). Bogda granitoids also display island arc geochemical signatures evidenced by enrichment of LILEs and depletion of HFSEs (Li et al., 2015a). Although no detailed geochemical data have been obtained from the Bogda volcanic rocks so far, many authors believe it is a true island arc the Carboniferous rocks outcropped north of the Tianshan orogen are mainly volcanic and volcaniclastic rocks and turbidites closely associated with the northern Tianshan ophiolite belt (Wang et al., 2003). The Carboniferous Bogda island arc is nucleated by mafic-ultramafic complexes and large doleritic hypabyssal dike and sill complexes (Wang et al., 2003). This arc became part of northern Tianshan subduction-accretion complexes in Carboniferous time and was most likely associated with south-dipping subduction of the Paleo-Tianshan Ocean (e.g., Xiao et al., 2004, 2010).

8. Discussion

8.1. Geochemical features of volcanic rocks from fossil intra-oceanic arcs

In this section, we will discuss the major element composition of volcanic rocks from fossil intra-oceanic arcs of the Paleo-Asian Ocean in comparison with those from the modern intra-oceanic arcs in the western Pacific (Figs. 3, 8). In total, 150 analyses of volcanic rocks of fossil arcs in the CAOB including 11 new analyses from the Kurai arc have been included in this comparative study (Fig. 8). These analyses are compared to the previously published data sets obtained from the Izu-Bonin-Mariana-Tonga system (Fig. 3).

The Al_2O_3 – $FeO^* + TiO_2$ – MgO (AFM) classification diagram (Jensen, 1976) (Figs. 3A and 8A) shows different groupings of compositional points of modern and fossils arc volcanics. The modern rocks fall in the tholeiitic and high-Mg (boninitic) fields and, to a much lesser degree, to the calc-alkaline field. They form clear trends from “classic” boninites to older and then younger tholeiities. The fossil arc volcanics plot mostly in the calc-alkaline field and, in lesser amounts, to the high-Fe tholeiitic field. Most of the Kurai volcanics (late Neoproterozoic) and a part of the Dunzhugur (late Neoproterozoic) volcanics plot away from the main cluster, in the high-Mg field (Fig. 8A). The overall difference between the distributions of compositional points of old and modern IOA volcanics in the AFM triangles can be due to several reasons. The modern mantle differs from the Neoproterozoic to middle Paleozoic mantle in composition, temperature, thickness, and degree of metasomatism. Moreover, the older volcanis rocks have experienced stronger post-magmatic secondary alteration. All these factors plus specific conditions at each subduction zone (thickness of oceanic crust, angle of subduction, degree of mantle wedge metasomatism) could affect the degree of mantle melting, composition of parental melts and consequently the proportions between FeO, TiO_2, Al_2O_3 and MgO.

In the MnO-TiO_2-P_2O_5 tectonic discrimination diagram (Mullen, 1983), the difference between the modern and old IOA volcanics is not so significant (Figs. 3B and 8B). Most of the Paleo-Asian Ocean rocks also plot in the boninitic and tholeiitic fields. Similarly to the modern arc, a small part of the Paleo-Asian Ocean analyses plot in the MORB field. However, a part of the Bozshakol (middle Cambrian), Saerbulake (early Devonian) and Gurvansayhan-Zoolen (late Silurian) compositions shift toward the calc-alkaline field, like in the AFM plot (Fig. 8A, B).

In the SiO_2 vs FeO_t/MgO Hacker's diagram, the Paleo-Asian Ocean volcanic rocks also plot in the boninitic, tholeiitic fields and calc-alkaline field with a smaller amount of points in the high-K field (Figs. 3C and 8C). The boninitic field includes the compositional points of the Dunzhugur, Kurtushibin, Kurai, and Searbulake IOAs. In the SiO_2 vs K_2O diagram, the proportion of old IOA compositions in the mid-K and high-K area is much greater than that of modern IOAs. The elevated K abundances are possibly due to alteration. The effect of seafloor alteration was previously shown for the eastern group boninites from north Tonga (Falloon et al., 2008). In summary, the old Paleo-Asian Ocean boninites have whole-rock geochemistry (apart from K) essentially identical to that of the modern, fresh boninites of the Tonga Trench, whereas the tholeiitic and calc-alkaline varieties show much wider variations of major oxides (Fig. 8A, C, D). The variations of incompatible elements are less profound (Fig. 8B). Unlike major oxides, trace elements show no notable differences between the modern and fossil counterparts (see Suppl. Electr. Materials).

8.2. Role of back-arc basins in identification of intra-oceanic arcs

Back-arc basin is a major constituent of IOA systems (Fig. 1). Back-arc basins form by fast (>10 cm per year) or slow (1–2 cm per year) sea-floor spreading (Stern, 2010). Most world IOAs are neighboring back-arc basins with active spreading, e.g. the Mariana, Tonga–Kermadec, New Britain, and Vanuatu (Fig. 2). Several other IOAs, e.g. the Lesser Antilles, Aleutians and Solomon, are associated with fading/shrinking back-arc basins. The spreading results from extension that splits a continental arc, intra-oceanic arc or fore-arc.

Identification of back-arc basins in old accretionary orogens is of special importance because boninites, which are often considered as a key diagnostic feature of IOA, may be absent in both modern and fossil IOAs. Initiation of a new IOA start with the opening of back-arc basin, which typically goes by three ways: active margin rifting, intra-oceanic arc rifting and fore-arc rifting (Stern, 2010) (Fig. 16). The active margin rifting forms an initial IOA and an initial back-arc basin (Fig. 16A) parallel to a still active continental arc. The volcanic front without interruption shifts oceanward and forms an intra-oceanic arc. The Sea of Japan possibly formed by a similar scenario, when the Japanese Islands were rifted off the East Asia active continental margin in early Miocene time, i.e. at ca. 20–15 Ma (Khanchuk et al., 1989; Tatsumi et al., 1990; Jolivet et al., 1994). A similar situation can be reconstructed in the middle Cambrian - early Ordovician island-arc terranes of the Kazakh orocline (Fig. 12; Section 5.1). The intra-oceanic arc rifting (Fig. 16B) forms a new arc parallel to still active initial arc and a new back-arc basin leaving a remnant basin in the rear part of the subduction system. The rifted part of the arc can split from the volcanic front to form a remnant arc that subsides as the spreading continues (Karig, 1972). Such a scenario probably acted in the Mariana IOA system (Fig. 1B) and in the Devonian Gurvansayhan-Zoolen IOA of southern Mongolia (Fig. 10). The fore-arc rifting also forms a new arc whereas the active arc migrates continentward and, with time, becomes a remnant arc (Fig. 16C). The Shishkhid IOA of eastern Tuva-Sayan probably formed by such a scenario (Fig. 7).

Back-arc spreading produces new sea floor which structure is, to a great extent, similar to that of that produced at mid-ocean ridges. Inter-arc rifting may create large basins in between intra-oceanic arcs, active and remnant, but without sea-floor spreading. Typical examples

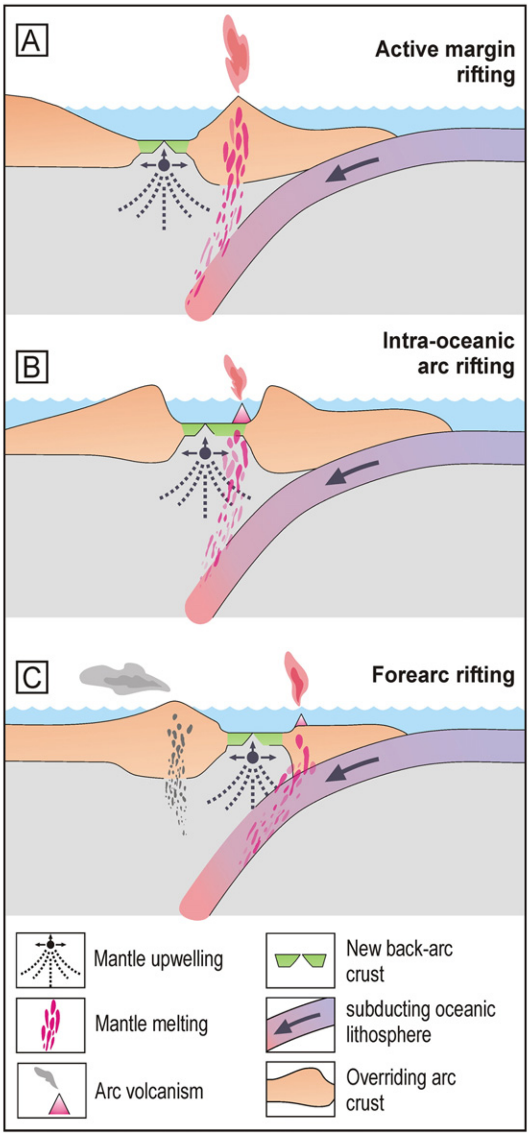

Fig. 16. The models for back-arc rifting and formation of IOA (Stern, 2010).

of inter-arc rift are the Shikoku basin in the back of the Izu-Bonin arc and the Parece Vela Basin in the back of the Mariana arc (Fig. 1B).

8.3. Major periods of intra-oceanic arc construction in the Paleo-Asian Ocean

We divided the IOA complexes of the CAOB into three groups according to their age and relation to active margins of different continental masses, which finally collided to form the CAOB (Figs. 6, 5E). The Neoproterozoic – earliest Cambrian IOAs of the Siberian Group formed in the northern Paleo-Asian Ocean (in present coordinates; Fig. 5), during the early stages of its evolution. The arcs of the Siberian Group now are incorporated into accretionary complexes and adjacent active margin/island-arc terranes of eastern and western Tuva-Sayan regions, Transbaikalia, northern and southwestern Mongolia and Russian Altai (Figs. 6, 7, 9–11). Before accretion the arcs were evolving during ca. 250–270 m.y. (Table 2). Consequently, they probably formed over more than one oceanic plate, i.e. similar to the IOAs of the modern Pacific, which now exist at the zone of subduction of the Pacific plate and, to a much lesser extent, the Philippine and other plates. The Japanese arc, which was formerly an active continental margin, keeps records of already disappeared Farallon and Izanagi oceanic plates. Thus, the Paleo-Pacific to Pacific oceans has been evolving during more than 300 m.y. (Maruyama et al., 1997).

The oldest IOA system of the Paleo-Asian Ocean is represented by the Dunzhugur and Shishkid arcs in the eastern Tuva-Sayan (Figs. 1, 7), which formed at ca. 800 Ma, i.e. at earliest stages of Paleo-Asian Ocean evolution (Fig. 5A). The Agardag and Shatskii arcs (and probably the Kurtushibin arc, whose precise age has not been constrained yet) of the western Tuva-Sayan formed at ca. 580–570 Ma. They evolved almost coevally with the Dariv and Khan-Taishirin arcs of south-western Mongolia and Kurai-Ulagan arc of the Russian Altai (Figs. 1, 9–11). The Tannu-Ola ophiolite represents the youngest arc of the Siberian Group (539–518 Ma). Most of the arcs of this group include both boninites and back-arc deposits and therefore are parts of former intra-oceanic arc systems.

The Kazakhstan Group includes a limited number of intra-oceanic island arcs, which formed and evolved from the middle Cambrian to the early Ordovician (Table 2). The Selety-Urumbai (middle to late Cambrian) and Bozshakol-Chingiz (late Cambrian-early Ordovician) island-arc terranes are parts of the Kazakh Orocline (Fig. 12). Their intra-oceanic arc origin is not as obvious as that of the Siberian Group, however, there are back-arc formations and abundant tholeiitic volcanic rocks with positive Nd and Hf isotope characteristics.

The Southern Group is more diverse, both geographically and lithologically. It includes Middle Paleozoic island-arc terranes of the Tienshan Orogen, Chinese Altai, southern Mongolia and eastern Kazakhstan. The Middle Paleozoic arcs formed in the Turkestan/South Tienshan, Junggar/Ob'-Zaisan and South Mongolian oceans, southern branches of the PAO. No boninitic rocks have been identified in the Tienshan arcs of Fan-Karategin and Chatkal-Atbashi (Fig. 13). However there are back-arc units and island-arc volcanic rocks with positive Nd and Hf isotopic values, and also the island-arc units are associated with back-arc basins. These regions remain understudied though and more reliable identification of intra-oceanic island arcs requires future research. Few, if any, well-defined IOAs of late Paleozoic age have been recognized in the CAOB: the only Bogda arc in the northern Chinese Tienshan hosts granitoids with positive epsilon Nd characteristics while its volcanic rocks remain understudied.

8.4. Parameters of intra-oceanic arcs: modern vs fossil

The major parameters of IOA are geometric length, life length and dominating lithologies with related geochemical signatures. There are 13 modern intra-oceanic arc systems in the world, 10 of which are located in the northern, western and southwestern Pacific (Leat and Larter, 2003). The better known localities are the Izu-Bonin-Mariana, Tonga-Kermadec, Vanuatu, Solomon, New Britain arc, western Aleutian, South Sandwich and Lesser Antilles arcs (Fig. 2). The modern IOAs constitute about 40% of all Pacific-type convergent margins. Compared to continental arcs, they are less studied because most of them are submerged below the sea level. However, many scientists agree that the modern IOAs are lithologically dominated by boninites, tholeiitic basalts and andesites, and less abundant tholeiitic dacite and subalkaline basalt to dacite (see Sections 1 and 2). The IOAs are sites of juvenile magma generation, high seismicity, subduction-related metamorphism and tectonic erosion. All these processes form and transform the geometries and lithologies of IOAs. During suturing of the oceanic lithosphere the arcs or their preserved fragments can be accreted to continental margins and then incorporated into intracontinental orogenic belts, like the Central Asian Orogenic Belt. The major parameters of the Pacific-type convergent margins are: (i) accreting vs. eroding; (ii) intra-oceanic or continental margin types of supra-subduction complex; (iii) length and quantity of island arcs; (iv) time span of continuous subduction; (v) type of accreted OPS (pelagic, hemipelagic, trench, oceanic island/seamount/plateau) (Safonova, 2017; Safonova et al., 2016b). Among those, the length and amount of island arcs and the tiem span of continuous subduction are of special importance as they record the periods and scales of juvenile crust generation.

The total length of IOAs in the Circum-Pacific is more than 20,000 km (Fig. 2). The longest is the Mariana – Izu-Bonin IOA system, which extends for a distance of about 4000 km, spanning about 50 m.y., from the Eocene to the present time. This arc system is the oldest active. The Mariana arc has been evolving over the oldest subducting oceanic crust of 150 Ma age. The length of most of the Pacific arcs exceeds 1000 km compared to a limited number of short arcs (New Britain, Halmahar and Sangihe) around Indonesia, which can be regarded as part of a single arc though.

The IOAs of the Paleo-Asian Ocean have been identified in almost all major orogens or foldbelts of the CAOB. However, the largest number of IOAs was recorded in southern Siberia – northern Mongolia (Fig. 6). We discussed the duration of intra-oceanic arc magmatism in Section 8.3. The total length of the arcs of the Siberian Group (Neoproterozoic to early Cambrian) is around 2700 km (Table 2). The longest arcs, or the longest preserved segments of the fossil arcs, are the Kurtushibin and Khan-Taishirin island-arc terranes, 440 and 610 km, respectively, which may be separated fragments of formerly single arc. The longest arc of the Kazakhstan Group (middle Cambrian - Ordovician) is the Bozshakol-Chingiz terranealso probably representing a single arc together with the Selety-Urumbai. However, it remains unclear if the Bozshakol-Chingiz arc was a true intra-oceanic arc and if it was extended over the whole length of the former subduction zone. The total length of middle Cambrian to Ordovician arcs is about 1500 km. The middle Paleozoic arcs were more numerous and wider geographically distributed. The longest arc of the Southern Group is the Gurvansayhan-Zoolen island-arc terrane in southerm Mongolia extended for more than 850 km.

Thus, the oldest modern arc system is Mariana (ca. 50 Ma), whereas several fossil arcs show much longer life spans (Table 2). This can be explained by the still continuing evolution of the Pacific Ocean or by its smaller size compared to the Paleo-Asian Ocean. The lithologies of the modern and fossil arcs are, in general, similar, although geochemically the fossil arcs contain more calc-alkaline varieties suggesting either their more evolved character, which agrees well with their longer life spans (some evolving longer than 50 m.y.), or by different conditions of magma generation. The total length of the modern arcs is greater than that of each group of fossil IOAs. This is also quite explainable as now we can see only fragments of former arcs, which survived after many stages of subduction-related and collision-related orogeny, metamorphism and intra-continental deformations.

8.5. Preservation of intra-oceanic arcs and tectonic erosion in the CAOB

A very important issue is tectonic erosion of IOAs and/or their direct subduction that may result in partial or complete disappearance of IOAs (Scholl and van Huene, 2007; Clift et al., 2009; Yamamoto et al., 2009; Stern and Scholl, 2010; Safonova et al., 2015). For example, there are at least six intra-oceanic arcs in the Philippine Sea, the largest of which are the Kyshyu-Palau and Izu-Bonin arcs, which, according to P-wave tomographic data, are currently subducting under the Honshu arc down to the mantle with minor to nil accretion instead, thus suggesting ongoing tectonic erosion and/or delamination of arc crust (Yamamoto et al., 2009).

Tectonic erosion at Pacific-type convergent margins can be recognized by the distance between magmatic arc and accretionary complex. In modern arcs, the distance between arc and trench is typically more than 100 km. If an accretionary complex in old accretionary belt crops out too close or adjacent to a coeval arc, we may suggest a loss of crust due to tectonic erosion (e.g., Scholl and van Huene, 2007; Safonova et al., 2015; Safonova, 2017). A case of tectonic erosion has been recently well proved in the Chatkal-Atbashi Ordovician arc of the Kyrgyz Middle Tienshan (Alexeiev et al., 2016). These authors suggested a loss of 50 to 100 km of middle-late Devonian crust in the northern Atbashi Range (Fig. 12). They found that the distance between the arc granitoids, supra-subduction ophiolites and coeval fore-arc deposits and accretionary units appeared much less than should be expected and concluded about an important episode of tectonic erosion in early Carboniferous times (Alexeiev et al., 2016). In Devonian to early Carboniferous times, the Tienshan orogen was probably a site of strong subduction erosion (Safonova, 2017). The early IOA formations, in particular, boninitic and tholeiitic, could be eroded and subducted to the deep mantle without leaving any fragment at the surface.

Therefore, the structural and compositional records of fossil IOA allow us to make only minimal estimations of their geometric length, life length, crust thickness and dominant lithologies. Consequently, while identifying the nature of a supra-subduction terrane, we should take into account the whole integrity of available data: the absence/presence of boninites, the absence/presence of back-arc basin assemblages, the tholeiitic versus calc-alkaline dominant lithology, whole-rock Nd and Hf-in-zircon isotopes, etc.

9. Conclusions

Twenty one (or less if considering some as parts of one arc) intra-oceanic arcs have been identified in the Central Asian Orogenic Belt. They formed during three time intervals in the evolution of the Paleo-Asian Ocean: Neoproterozoic, early Paleozoic and middle Paleozoic. According to their age and geographic position, we divided them into three groups: Siberian, Kazakhstan and Southern.

The Siberian Group includes eleven fragments of Neoproterozoic to earliest Cambrian arcs, most of which with boninites, with a total length of ca. 2700 km. The longest arc fragments are the Kurtushibin and Khan-Taishirin island-arc terranes, 440 and 610 km, respectively. The Neoproterozoic – earliest Cambrian IOAs formed during the early stages of the Paleo-Asian Ocean evolution. The arcs of the Siberian Group are associated with accretionary complexes, back-arc and active margin terranes of eastern and western Tuva-Sayan, Transbaikalia, northern and southwestern Mongolia and Russian Altai.

The Kazakhstan Group includes three fragments middle Cambrian – Ordovician arc systems, which contain few boninites. The middle Cambrian to early Ordovician arcs formed and evolved in the middle Cambrian to the early Ordovician. The Selety, Urumbai (middle-late Cambrian) and Bozshakol-Chingiz (late Cambrian-early Ordovician) island-arc terranes are parts of the Kazakh Orocline. The total length of the middle Cambrian to Ordovician arcs is about 1500 km.

The Southern Group includes six late Ordovician to Devonian arcs, some also with boninites, totally about 2000 km long. The Southern Group is more diverse, both geographically and lithologically and can be split into the Tienshan, East Kazakhstan – West Junggar and South Mongolian subgroups. The Middle Paleozoic arcs of the Southern Group formed in the Turkestan/South Tienshan, Ob'-Zasian/Junggar and South Mongolian oceans, all branches of the Paleo-Asian Ocean. No boninitic rocks have been identified in the Tienshan arcs of Fan-Karategin and Chatkal-Atbashi. Therefore, their intra-oceanic origin remains debatable. The longest arc of this group is the Gurvansayhan-Zoolen island-arc terrane of southern Mongolia, which extends for a distance of >850 km.

The eastern CAOB hosts several volcanic belts which may include fragments of IOAs, however, only one Late Paleozoic IOA – Bogda - has been reliably recognized so far, probably because the late Paleozoic was the time of PAO shrinking and suturing. Therefore, the ocean was not be large enough to produce intra-oceanic arc systems.

The lithologies of the modern and fossil arcs are similar, although geochemically the fossil arcs contain more calc-alkaline varieties suggesting either their more evolved character, which agrees well with their recorded longer life span or there were different conditions of magma generation. In addition, an intraoceanic arc may represent an early stage in the evolution of immature arcs and accordingly they are longer living. The boninites of fossil IOAs of the Paleo-Asian Ocean have essentially identical whole-rock geochemistry (apart from potassium) to the modern fresh boninites of the Tonga Trench, whereas the tholeiitic and calc-alkaline volcanic rocks show much wider variations of major oxides.

Of special importance is the identification of back-arc basins in old accretionary orogens, because boninites, which are often considered as a key diagnostic feature of IOA, may be absent in both modern and fossil IOAs. The formation of back-arc basins might have followed three scenarios: active margin rifting, intra-oceanic arc rifting and fore-arc rifting. All these scenarios can be reconstructed in many fossil intra-oceanic arcs, for example, in the arcs of the Kazakh, Siberian or Southern Groups, respectively.

Intra-oceanic arcs could be tectonically eroded and/or directly subducted into the deep mantle. This may result in partial or complete disappearance of IOAs. Therefore, the structural and compositional records of fossil IOA allow us to make minimal estimations of their length, life span, and crust thickness.

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Appendix A. Supplementary data

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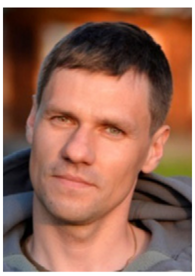

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