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The Altaids of Central Asia: a Tectonic and Evolutionary Innovative Review

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Abstract

The Altaids, one of the largest and long-lived accretionary orogens in the world, developed from ca. 600 Ma to 250 Ma by the multiple accretions of terranes of different origin, chiefly microcontinents and island arcs. Considerable geological information supported by geochemical, radiometric and isotopic data suggest that modern geodynamic processes such as seamounts/plateau accretion, ridge-trench interaction, the formation of supra-subduction ridges and back-arc basins, arc-arc collisions and oroclinal bending were responsible for the evolution of the Altaid archipelagos. Because of the paucity of palaeomagnetic and radiogenic data it is still not possible to present a definitive palaeo-reconstruction of the Altaids. Nevertheless, considering the voluminous literature appearing today on the Altaids, it is timely and appropriate to present a review of current understanding of the many inherent tectonic problems, some of which are controversial.

The Altaids began its development in Vendian (610-570 Ma)-Early Palaeozoic oceans between three approaching cratons, Siberia, Gondwana and Tarim-North China, where it continually evolved during the Middle-Late Palaeozoic. The peri-Siberian part of the orogen formed around the microcontinents of Tuva-Mongolia and Altai-Mongolia through the multiple accretion of exotic Izu-Bonin-type island arcs (e.g. Uimen-Lebed, Lake-Khamsara), and oceanic islands/seamounts/plateaus (e.g. Kurai, Dzhida, Bayanhongor), and by the formation of back-arc basins (i.e. Altai-Sayan, Barguzin). These multiple accretion-collision events led to the formation of major peri-Siberian sutures by the end of the Early Palaeozoic (e.g. Bayanhongor, Dariv-Agardagh, Borus, Kurtushiba, Dzhida, Olkhon). The Mongol-Okhotsk Ocean opened within this new accreted continent in the Early-Middle Palaeozoic.

The Kazakhstan Continent formed mostly by the Early Silurian in Eastern Gondwana by the accretion-collision of several ribbon-microcontinents (e.g. Chatkal-Karatau, Chu-Yili, Aktau-Junggar) and island arc-type terranes (e.g. Boshchekul-Chingiz, Baidaulet-Akbastau). Most Kazakhstan microcontinents originated in Gondwana from which they were detached through two probable stages of stretching in the Vendian and Amgaian (Middle Cambrian). Kazakhstan was finally created by formation of the Kumdykol, Kyrgyz-Terskey, Dzhalair-Naiman sutures in the Arenigian (Lower Ordovician), and by formation of the Maikain-Kyzyltas, Yili-Erementau sutures in the Hirnantian-Rhuddanian (Lower Silurian). The completed Kazakhstanian Continent moved westward toward Siberia and Tarim-North China in the Middle-Late Palaeozoic.

The Tarim-North China craton(s) was likely located to the north of Eastern Gondwana during the Vendian-Early Paleozoic. The tectonic evolution of the northern margin of Tarim-North China in the Early-Middle Palaeozoic mostly took place by island arc accretion (i.e. Tulinkai island arc), active margin accretion (i.e. Bainaimiao arc and Ondor Sum wedge) and by the opening of back-arc basins, which led to separation of the Central Tianshan-Hanshan Microcontinent.

From the mid-Paleozoic, Siberia, Tarim-North China and Kazakhstania began to mutually interact. The new plate tectonic arrangements led to the oroclinal bending and large-scale rotation of Kazakhstania during the Carboniferous, and to the main terminal sutures of the Altaids (i.e. South Tianshan, Turkestan, Uralian, Chara, Junggar-Balkash and Solonker) by the Permo-Triassic. Following the completion of the Altaids, only the Mongol-Okhotsk remained opened until the Jurassic-Cretaceous.

During our synthesis we discuss alternative plate tectonic hypotheses, and we propose new models, which may provide potential perspectives for future investigations.

Keywords

Altaids, tectonic, Palaeozoic, cratons, terranes

Contents

1. Introduction

Accretionary orogens are currently of considerable interest, because they provide the best available evidence of the structural, metamorphic and magmatic events and processes that take place during subduction, exhumation and accretion (Cawood et al., 2009, Schulmann and Paterson, 2011). On our planet extant accretionary orogens are best displayed along the western Pacific, where on-going processes are taking place in Alaska (Sisson et al., 2003), the

Philippines and Indonesia (Hall, 2002, 2009), and Japan (Isozaki et al., 2010). Importantly, the accretionary processes in Japan can be tracked back to their beginning at about 500 Ma (Isozaki et al., 2010).

The Altaids in Central Asia (Sengör et al., 1993; Sengör and Natal'in, 1996) is the largest accretionary orogen worldwide, which evolved from about 600 Ma to 250 Ma (Sengör et al., 1993), and thus overlaps in time with the accretionary orogen of the Japanese Islands. The Altaids have been variably termed the "Ural-Amurian", "Ural-Mongolian" or "Central Asian" orogenic belts, and its re-definition, and at times casual use, have caused some confusion, as well explained by Sengör and Natal'in (2007). Note that the Central Asian Orogenic Belt (CAOB) has an age range from 1 Ga to 250 Ma (Windley et al., 2007), because it includes the Neoproterozoic Uralides and Baikalides, which in the original definition of Sengör et al. (1993) were situated on the north-western and northern sides of the Altaids, respectively. In this paper we follow the original definition of Sengör et al. (1993).

Several palaeotectonic/palaeogeographic models have been presented for different regions and time-periods of the Altaids. Some of the most important English-language overviews are: Zonenshain et al. (1990), Mossakovsky et al. (1994), Gusev and Khain (1996), Sengör and Natal'in (1996), Filippova et al. (2001), Badarch et al (2002), Keraskova et al. (2003), Khain et al. (2003), Yakubchuk (2004), Cocks and Torsvik (2007), Dobrestov and Buslov (2007), and Xiao et al. (2010a) from which two main models emerge to account for the evolution of the Altaids: (1) tectonic duplication and amalgamation of one to three major arcs of common origin: (2) multiple amalgamation of many microcontinents and island arcs of different origins. From our re-evaluation we suggest that current data lead to a model that involves several major ribbon-terranes of different origin that were accreted at different times. Sengör and Natal'in (1996) presented the first detailed, systematic description of each "puzzle piece", correlationed them across the whole Altaids, and integrated them into a palaeotectonic model for the Palaeozoic. Our synthesis and tectonic interpretations are based on a similar methodology.

In spite of considerable research in the last half-century, there is still much uncertainty about the relative roles of accretionary versus collisional tectonics within the overall development of the Altaids, and inevitably there are considerable disputes about many key tectonic processes and events such as the time of formation/exhumation of eclogites and collision in the western Tianshan, the Kazakhstan Superterrane and orocline, the history of the Mongol-Okhotsk Ocean, the putative Tarim plume, and the timing of terminal collision along the main ca. 5500-long suture zone.

The purpose of this paper is to provide an overview of current knowledge of the main components, timing and evolution of the Altaids in order to outline the principal differences of interpretation, as a means of better understanding the accretionary evolution of this major orogen. We use the time-scale of Gradstein et al. (2004): e.g. Neoproterozoic, 1000-542 Ma; Cryogenian 850-630 Ma; Ediacaran 630-542 Ma), and the commonly used Vendian at 610-542 Ma.

2. The Altaids: components and terms

The Altaids of Central Asia (Sengör et al., 1993) extend from the Ural mountains (Puchkov, 1997) eastwards (present coordinates) to the Pacific coast (Wilde et al., 2010), and from near the border of the Siberia craton (Vernikovsky et al., 2004) southwards to the main Altaid suture that extends from the Pacific margin (Wilde et al., 2010), westwards via the Solonker Suture north of Beijing (Xiao et al., 2003), through Beishan (Xiao et al., 2010b), to the southern Tianshan in western China (Xiao et al., 2008), and farther westwards in Kyrgyzstan (Biske and Seltmann, 2010) and Uzbekistan where it finally joins the Uralian Suture (Alexeiev et al., 2009).

Sengör et al. (1993) defined 26 first-order Altaid units and attributed them to independent orogens such as the Altai-Sayan, Mongol-Okhotsk, Kazakhstania, Tianshan and Uralian. We generally follow a similar approach, but define more regions (Fig. 1). Our re-evaluation suggests the following key regions: (1) Northern Kazakhstan, (2) Altai-Sayan, (3) Baikal, (4) Siberian Altai, (5) Chinese Altai, (6) Northern Mongolia, (7) Southern Kazakhstan, (8) Junggar, (9) Southern Mongolia, (10) Kyrgyz Tianshan, (11) Chinese Tianshan, (12) Beishan, (13) Inner Mongolia (Fig. 1) divided into 107 zones (see Figs. 2 and 9).

The terminology used in palaeotectonic/palaeogeographic models to define continental fragments in orogenic belts (e.g. tectonic units, microcontinents, terranes, blocks) have often led to misinterpretations and confusion. For example, although Sengör and Dewey (1990) pointed out the general unsuitability of the term "terrane", it was later used by Badarch et al. (2002) and Cocks and Torsvik (2007) in an Altaid context. To avoid confusion with existing terminology, we use the non-genetic term "zone", as on tectonic maps by Sengör and Natal'in (1996) and Naumova et al. (2006) for the Altaids, by Badarch et al. (2002) for Mongolia, and by Windley et al. (2007) for Kazakhstan. Zones contain rocks with distinctive geology and tectonic features, and they are delimited by tectonic boundaries such as suture zones or major faults. We also use the term "terrane" for a collection of "zones", which have common affinities that enable short or long mutual correlations (see Figs. 2 and 9).

3. Vendian to Devonian accretionary growth of the peri-Siberian Continent

The Pre-Silurian growth of the peri-Siberian Continent was complex and different issues related to the number, nature, origin, interactions and evolution of terranes involved in its formation are far from resolved. However, some interesting situations reside. Of the different microcontinents (ss. 3.3) some were considered to have a Gondwanan origin (Zonenshain et al., 1990; Mossakovsky et al., 1994; Ruzhentsev and Mossakovskiy, 1996; Kheraskova et al., 2003;

Buslov et al., 2004a; Kheraskova et al., 2010), and others a Siberian origin (Sengör and Natal'in, 1996; Yakubchuk, 2004; Kuzmichev et al., 2005). Island arcs formed in the Siberian Altai, SW Mongolia, and Transbaikalia (Buslov et al., 2002; Dijkstra et al., 2006; Gordienko et al., 2007; Ota et al., 2007; Kovach et al., 2011; Yarmolyuk et al., 2011), and some back-arcs in Siberia (e.g. Belichenko et al., 2006; Zorin et al., 2009) (ss. 3.2). Several island arcs of Vendian-Cambrian age are only preserved as relicts in subduction-accretion complexes. Only two major, well-studied island arcs are presented in detail in this synthesis (ss. 3.4), and lesser-known small relics in their subduction-accretion complexes (ss. 3.5). Some, still poorly known, but large arcs are considered in their subduction-accretion zones (Borus-Kurtushiba and Dzhida-Bayangol zones); future investigations may lead to their definition as new terranes (Fig. 2).

3.1. The Neoproterozoic Siberian margin

It is generally accepted that the Siberian Craton was a part of Rodinia and detached from Northern Laurentia at ca. 780-750Ma (Cocks and Torsvik, 2007). The Rodinian rifting was localized on the western Siberian margin at Yenisei, Birusa and Sharizhalgay (Fig. 2), and it extended eastwards to the north of Lake Baikal as far as the Patom area - Fig. 2 (Smethurst et al., 1998; Vernikovsky et al., 2003; Vernikovsky et al., 2004; Gladkochub et al., 2006, 2007; Metelkin et al., 2007; Sovetov et al., 2007; Stanevich et al., 2007; Li et al., 2008; Vernikovsky et al., 2009).

The Siberian margin underwent multiple accretion in the Cryogenian (850-630 Ma) (Vernikovsky et al., 2003, 2004, 2009), following which it became a passive margin at ca. 600Ma when it was obducted by ca. 700-630 Ma island arcs (i.e. Yenisei, Kan, Baikal-Muya; Vernikovsky et al., 2003, 2004; Turkina et al., 2007). Subsequently, the southern and western margins of the Siberian craton were fringed by an active margin and occupied by a peripheral foreland basin until ca. 540 Ma (Khomentovskii and Postnikov, 2001; Vernikovsky et al., 2004; Sovetov et al., 2007).

The conclusion of Sovetov et al. (2007) that this active margin underwent collision at ca. 550 Ma was based on their interpretation of the closure of remnant marginal basins (i.e. interruption of marine deposition) and establishment of a wide orogen (i.e. molasse deposits). This compressional event was closely followed by important extension: Early Cambrian (ca. 542-510 Ma) rift sediments were deposited in a collapsed orogen (Sovetov et al., 2007). This evidence may also reflect the collapse of a Cordillera when lower plate buoyancy changes and leads to back-arc stretching (Stampfli and Borel, 2002). According to Zorin et al. (2009) Siberia was surrounded by back-arc basins in the Vendian-Cambrian.

3.2. Siberian back-arc basins

3.2.1. Barguzin

In the Baikal area, the controversial nature and origin of the Ikat-Barguzin/Baikal-Muya zones (Fig. 2) have been interpreted by various palaeotectonic models, either as a displaced or exotic microcontinent and/or as an island arc accreted to Siberia in the Neoproterozoic or Palaeozoic (Khomentovskii and Postnikov, 2001; Khain et al., 2003; Vernikovsky et al., 2004; Fedorovsky et al., 2005; Belichenko et al., 2006; Dobretsov and Buslov, 2007; Makrygina et al., 2007; Rytsk et al., 2007; Gladkochub et al., 2008; Zorin et al., 2009; Rytsk et al., 2011). Although their precise timing remains unclear, there were probably two main accretionary/collisional events: (1) in the Vendian the obduction of an island arc (i.e. Baikal-Muya) onto the Siberian passive margin and (2) a later collision between the Barguzin "block" and Siberia.

Although the nature of the Barguzin block is subject to controversy (Belichenko et al., 2006) the partial formation of the Ikat-Barguzin Zone by back-arc processes seems today most likely, given the data available (Khomentovskii and Postnikov, 2001; Belichenko et al., 2006; Makrygina et al., 2007; Zorin et al., 2009). From detailed geochemical data Makrygina et al. (2007) identified a continental arc-back arc within the Barguzin and Hamar-Davaa zones (at Talanchan, Olkhon and Slyudyanka, Fig. 3). This is consistent with the presence of a Cambrian Siberian back-arc basin in the Altai-Sayan area (ss. 3.2.2) and supported by ca. 500 Ma metamorphism identified along the craton margin (Olkhon, Slyudyanka, Derba; Gladkochub et al., 2008 and references therein) as well as by coeval folding in a high deformation zone (Rytsk et al., 2011), which likely marks the suture of a closed peri-Siberian back-arc basin (Fig. 5). Rytsk et al. (2011) presented an interesting discussion on alternative versions of palaeogeodynamic reconstructions for the Baikal-Muya Belt during the Neoproterozoic.

3.2.2. Altai-Sayan

Vendian to Middle Cambrian ophiolites, back-arc basins and island arcs were identified in the Kizir-Kazir area (Altai-Sayan back-arc basin, Telesk and North Sayan zones on Fig. 2) (Naumova et al. 2006). The Altai-Sayan back-arc basin opened in the Late Cambrian. Sovetov et al. (2007) underlined that the Siberian active margin was affected by extension (i.e. collapsed orogen, and rift sediments) in the Cambrian (ca. 443-410 Ma). This is consistent with coeval post-collisional intrusions and metamorphism in the accreted Kan island arc (ss. 3.1), which likely reflect exhumation. However, that metamorphism did not affect the Derba passive margin located in front of the accreted island arc (Turkina et al., 2007), which implies that the passive margin was probably not Neoproterozoic as previously thought (Naumova et al., 2006) but certainly Cambrian, having formed by the opening of the Altai-Sayan back-arc basin. A volcanic island arc source reported in the Derba metasediments (Dimitrivea et al. 2006 in Turkina et al. 2007) likely reflects the weathering of the accreted arc.

Although precise geochronological and geochemical studies are still lacking, it is most likely that the maximum age of back-arc sediments was 520 to 500 Ma (Naumova et al. 2006), when a metamorphic-magmatic event affected the Derba passive margin sediments at ca. 500 Ma (Turkina et al., 2007) and syn/post-collisional granitoids intruded the Siberian margin, the back-arc sediments and the arc assemblages (i.e. Tannuola Belt, Naumova et al., 2006) in the Middle Cambrian-Early Ordovician (ca. 520-470 Ma) (Figs. 4).

3.3. Microcontinents

3.3.1. Mongolian and Siberian microcontinents: one or several terranes?

Within the Mongolian and Siberian Altaids Precambrian microcontinents have been identified at Hamar-Davaa, Tuva-Mongolia, Baidrag-Dzabkhan, Khangai (Tarvagatai), Idermeg and Argunsky (Badarch et al., 2002; Kovalenko et al., 2004; Naumova et al., 2006; Kozakov et al., 2007b; Kröner et al., 2007; Windley et al., 2007; Yarmolyuk et al., 2008; Demoux et al., 2009a, 2009c; Kozakov et al., 2011; Rojas-Agramonte et al., 2011) (Fig. 2). Several authors have proposed that the Tuva-Mongolian (Tuva-Mongolian and Hamar-Davaa zones) and Central Mongolian microcontinents (Baidrag-Dzabkhan, Khangai, Idermeg, Argunsky zones) formed a single terrane (Sengör and Natal'in, 1996; Khain et al., 2003; Yakubchuk, 2004; Kuzmichev et al., 2005; Windley et al., 2007). This correlation is especially well constrained by the Vendian Dariv-Agardagh ophiolitic belt that fringes the western side of the Baidrag-Dzabkhan and Tuva-Mongolian microcontinents (Pfänder et al., 2004; Dijkstra et al., 2006) (ss. 3.4.1). The Tuva-Mongolian Microcontinent has been considered either as exotic (i.e. Gondwana, Zonenshain et al., 1990; Mossakovsky et al., 1994; Ruzhentsev and Mossakovskiy, 1996; Dobretsov et al., 2003; Dobretsov and Buslov, 2007, Levashova et al. 2011)) or originated from Siberia (Sengör and Natal'in, 1996; Kuzmichev et al., 2001, 2005; Khain et al., 2002; Yakubchuk, 2004); the second interpretation is more consistent with recent palaeomagnetic data, according to which the southeastern margin of Siberia was located in the equatorial zone (Cocks and Torsvik, 2007).

This palaeo-geographic position is supported by carbonate sediments (Letnikova and Geletii, 2005) and palaeomagnetic data (Kravchinsky et al., 2001), which indicate that the Tuva-Mongolian Microcontinent was located in the peri-Siberian tropics in the Late Vendian-Cambrian. And a Siberian origin is supported by affinities of tonalites and island arcs in the Siberian margin and Tuva-Mongolia (Kuzmichev et al., 2001), who suggested that an elongate chain of Precambrian massifs including Tuva-Mongolia had separated from Siberia in the late Neoproterozoic.

In the Tuva-Mongolian basement a Neoproterozoic tectono-magmatic event can be correlated with one in the Baikal-Muya Belt on the Siberian margin (Kuzmichev et al., 2001, 2005; Khain et al., 2002, 2003; Zorin et al., 2009). The Baikalian Orogeny has been interpreted as a result of accretion of a Neoproterozoic (ca. 825-700Ma) island arc (Shishkhid/Baikal-Muya) to an older Gargan/Barguzin Microcontinent between 650-560 Ma (Gusev and Khain, 1996; Khomentovskii and Postnikov, 2001; Kuzmichev et al., 2001, 2005). Similar isotopic ages are found in the Baidrag-Dzabkhan (Yarmolyuk et al., 2008) and Idermeg (Kröner et al., 2007) microcontinents, which may also have undergone the Baikalian Orogeny. The correlation of Tuva-Mongolia with Barguzin/Baikal-Muya is also supported by the occurrence of a similar Vendian-Cambrian sedimentary cover overlying the two basements (Huvsgol-Bokson basin and Upper Angaran Basin; Mossakovsky et al., 1994; Badarch et al., 2002; Kuzmichev et al., 2005; Letnikova and Geletii, 2005; Naumova et al., 2006; Ruzhentsev and Nekrasov, 2009; Zorin et al., 2009) (Fig. 2-3).

The above information suggests that Central Mongolia, Tuva-Mongolia and Barguzin formed one single ribbon-microcontinent in the Vendian-Cambrian that was attached or located close to Siberia. Considering that the Altai-Sayan and Ikat-Barguzin zones may correspond to a Siberian back-arc basin opened after the Late Vendian (ca. 600 Ma) accretion of the potentially single Yenisei/Kan/Baikal-Muya/Shishkhid island arc (ss. 3.1-3.2), the Mongolian microcontinents, surrounded by Vendian-Cambrian oceans, was probably not detached from this part of the Siberian margin, but was derived from the West Stanovoy area by back-arc opening; this is compatible in a larger tectoic context as proposed by Kuzmichev et al. (2001). The existence around it of major, long-lived oceans (Dariv-Agardagh, Bayanhongor, and Dzhida-Bayangol, 3.4.1, 3.5.1-2) implies that it was isolated during Vendian and Cambrian time. It could also have been a continental promontory of the Siberian Craton, formed, for example, as a result of Rodinian break-up, which then moved westwards along the Southern Siberian margin during the Cambrian. Therefore, the Tuva-Mongolian and Central-Mongolian microcontinents may have evolved together as an independent ribbon-terrane (Fig. 5a).

The location between the Baidrag-Dzabkhan and Khangai microcontinents of the Bayanhongor Ocean from the Vendian through Early Ordovician (3.5.1) suggests the existence of another terrane: the Khangai-Argunsky Ribbon-Microcontinent. Although poorly documented, the Herlen Zone is considered here as the westward equivalent of the Bayanhongor Suture Zone because Vendian-Early Cambrian ophiolites occur there (Badarch et al., 2002). However, as discussed in section 3.5.1, the Bayanhongor oceanic fragments may also have originated in the Dzhida-Bayangol Ocean and thus, the Baidrag-Dzabkhan and Khangai Precambrian rocks may have formed parts of the same microcontinent (Fig. 5a).

3.3.2. The Altai-Mongolian Microcontinent: a possible extension in Inner Mongolia

The Altai-Mongolian Microcontinent is located to the southwest of the Vendian-Cambrian arc systems (Gorny-Altai, Lake) that were accreted to Siberia or Peri-Siberian terranes by the Late Cambrian (Fig. 2 and ss. 3.4). During the Cambrian, the Siberian accretionary wedge prograded northwards in the Charysh-Terekta and Hovd zones and remained in the area at least till the Early-Middle Ordovician (e.g. Buslov et al., 2002; Xiao et al., 2004a). This implies that the Altai-Mongolian Microcontinent was exotic and was accreted to Siberia after the Early

Ordovician. The Gondwanan origin of this microcontinent is generally accepted by the scientific community (Zonenshain et al., 1990; Mossakovsky et al., 1994; Ruzhentsev and Mossakovskiy, 1996; Dobretsov et al., 2003; Buslov et al., 2004a, b; Dobretsov and Buslov, 2007; Glorie et al., 2011; Yang et al., 2011), but its exact location remains enigmatic. A Tarim-North China origin was also speculated on by Wilhem (2010) based on kinematic and spatial constraints required by plate tectonics principles.

The presence of Precambrian detrital zircons in metasedimentary rocks and of xenocrystic zircons in granitoids suggests to Glorie et al. (2011) and Jiang et al. (2011) that the Altai-Mongolian Microcontinent had a Precambrian basement. However, Cai et al. (2011) reported that the so-called Precambrian basement rocks in the Chinese Altai have zircon U-Pb ages in the range 528 to 488 Ma, similar to that of the main low-grade marine quartzo-feldspathic turbidites interpreted as an active margin; Buslov et al. (2001) had interpreted these as a passive margin shelf. Unconformable unmetamorphosed marine sediments and andesitic volcanics, interpreted as a continental fore-arc basin, have a Late Ordovician age (Windley et al., 2002). However, this interpretation was recently challenged because of the occurrence of ca. 540-450 Ma detrital zircons of arc origin within the metasediments (the Habahe flysch), which were reinterpreted as a Middle Palaeozoic subduction-accretion complex (Long et al., 2010; Sun et al., 2008). Thus, the existence of a passive margin fringing the Altai-Mongolian Microcontinent during the Early Palaeozoic is not certain, but an active margin setting is evident according to ca. 540-440 detrital zircons, ca. 500 rhyodacite (Windley et al., 2002), ca. 460-430 Ma arc granitoids (e.g. Wang, T. et al., 2006; Briggs et al., 2007) as well as ca. 450-440 Ma rhyolite and ca. 410 Ma dacite-rhyolite suites (Wang, Y. et al. 2011). From Hf isotopic compositions of zircons in the granites and of detrital zircons in the sediments, Cai et al. (2011) calculated that as much as 84% of the Chinese Altai is probably made up of juvenile Palaeozoic materials, and accordingly there is little Precambrian basement.

This Cambrian-Ordovician arc-related magmatism was generally attributed to the southern active margin of the Altai-Mongolian Microcontinent (Windley et al., 2002; Wang, T et al., 2006; Long et al., 2010; Sun et al., 2008; Xiao et al., 2009b). However, the Altai-Mongolian Microcontinent is narrow (ca. 200km) and this arc magmatism could be also partly related to the northern margin. Glorie et al. (2011) recently underlined the occurrence of Middle-Late Ordovician (ca. 470-450 Ma) arc magmatism on the northern margin of the Altai-Mongolian Microcontinent in the Siberian Altai. The existence of a southern active margin since the Middle Ordovician is well constrained by intrusions in the Altai-Mongolian and Kalba-Narym zones and associated coeval accretionary wedge (i.e. Chara Zone, see below), but the pre-Middle Ordovician detrital zircons could also have been transported from a probable northward active margin towards a potential southern passive margin (Fig. 3).

It is probable that the well-documented, extensive Ordovician-Silurian southern active margin of the Altai-Mongolian Microcontinent extended as a plate boundary along strike. Relics from coeval continental arc and accretionary wedges occur farther north in the Kazakhstan and southeastward in the Southern and Inner Mongolia (Fig. 2). The Late Palaeozoic Chara strike-slip zone extends from Kazakhstan to the Chinese Altai (Buslov et al., 2004b). Although mainly composed of Late Palaeozoic rocks, the Chara Zone contains relics of Ordovician-Silurian subduction-accretion complexes mainly represented by ophiolitic mélanges with Ordovician oceanic crust and ca. 450-430 Ma HP complexes (tholeiite basalt, OIB and E-MORB protoliths) (Buslov et al., 2004a, b; Safonova et al., 2004; Volkova and Sklyarov, 2007; Xiao et al., 2009b).

In southern Mongolia, tholeiitic pillow basalt (MORB) of supposed Ordovician age, tuff, chert and clastics occur in the Bidz area (Badarch et al. 2002) and Precambrian continental crust was identified in the East Tseel Zone (ca. 1.5 Ga; Helo et al., 2006). The Precambrian basement seems to extend eastwards into the South Gobi and West Mandalovoo zones (Helo et al., 2006; Kröner et al., 2010). A thick intensively deformed assemblage of clastics, chert, tuff, shallow-marine limestone, and ultramafic and gabbroic bodies was metamorphosed to amphibolite facies at ca. 450-430 Ma (Naumova et al. 2006), and Silurian (ca. 435-425 Ma) arc-related volcanic rocks also occur in the West Mandalovoo Zone) (Helo et al. 2006). To the east, the presence of Precambrian basement and Ordovician-Silurian greenschist-facies sediments in the Gurvansayhan-Zoolen Zone (Badarch et al., 2002; Helo et al., 2006; Kröner et al., 2007) supports the extension of the Chinese Altai continental active margin as far as southern Mongolia (Fig. 6).

Ordovician-Silurian continental are assemblages are also present in Inner Mongolia (Nuhetdavaa-Enshoo and Baolidao zones on Fig. 2). It is well accepted that the Nuhetdavaa-Enshoo Zone corresponds to the Early-Late Palaeozoic active margin of Siberia (Wang, Q. and Liu, 1986; Hsü et al., 1991a; Yue et al., 2001; Badarch et al., 2002; Xiao et al., 2003; Miao et al., 2007b). In this area, the Precambrian continental basement was covered by Cambrian to Early Ordovician passive margin limestones, cherts and turbidites. A Middle Ordovician-Silurian sedimentary-volcanic arc assemblage succeeded the passive margin sediments and indicates a major change in the geodynamic evolution of the continental margin (Yue et al. 2001; Xiao et al. 2003). To the South, the Baolidao Microcontinent was interpreted as a terrane detached from the Mongolian active margin (Nuhetdavaa-Enshoo Zone) in the Late Palaeozoic (Miao et al., 2007b) and thus can be considered as the Early-Middle Palaeozoic southern margin of the Nuhetdavaa-Enshoo Microcontinent (Fig. 2). In this area, an Ordovician accretionary wedge is characterized by the Xilinhot and Sunidzuoqi metamorphic complexes, which were interpreted as the relics of a forearc basin (Shi et al., 2003; Chen et al., 2009) and/or accretionary wedge (Xiao et al., 2003; Miao et al., 2007b; Jian et al., 2008) associated with a continental arc (Li et al., 2011). The Xilin Hot metasediments (i.e. turbidite protoliths) yield ages ranging from Archean to Carboniferous (2900-2200 Ma, 1800-1500 Ma, 950-780 Ma and 536-302 Ma; Shi et al. 2003, Chen et al. 2009). The Sunidzuoqi complex was intruded by the ca. 490-310 Baolidao magmatic suite, which yield similar inherited zircon ages (Chen et al., 2000, 2009; Jian et al., 2008). However, the Early

Ordovician Baolidao arc magmatism (ca. 490-470 Ma; Chen et al., 2000; Jian et al., 2008) and Cambrian detrital zircons (Shi et al. 2003, Chen et al. 2009) contradict the existence of a passive margin at this time. For this reason, some authors proposed that the northern continental active margin started already in the Early Cambrian (Chen et al., 2009) or Late Cambrian (Miao et al., 2007b). However, Jian et al. (2008) underlined the island arc characteristics (i.e. near-trench intrusions in forearc sediments, and juvenile arc crust) of the Late Cambrian-Early Ordovician arc assemblage (Sunidzuoqi-Xilinhot). Thus, the Cambrian-Early Ordovician arc-related rocks are thus most probably remnants of an island arc accreted to a passive margin of the Nuhetdavaa-Enshoo Microcontinent. This event may have caused the onset of subduction (Cloetingh et al., 1984; Stern, 2004). The existence of an active margin of Middle Ordovician-Silurian age was well established in the Baolidao and Nuhetdavaa-Enshoo zones (Hsü et al., 1991a; Xiao et al., 2003; Miao et al., 2007b; Jian et al., 2008; Chen et al., 2009) (Fig. 6).

As proposed by some authors (Badarch et al., 2002 and references therein), the occurrence of Precambrian basement, arc-related magmatism and subduction-accretion complexes in Kazakhstan, Mongolia and Inner Mongolia strongly suggests the existence of a single active margin during Late Ordovician and Silurian time (Fig. 2-6). These spatial correlations are also supported by the continuation of accretion processes throughout the Middle Palaeozoic (ss. 3.6). This would imply that the Altai-Mongolian Microcontinent may have accreted to Siberia before the Late Ordovician (ss. 3.5.4). The pre-Middle Ordovician tectonic history of the margin is less clear, and the published models developed for the Chinese Altai and Inner Mongolia seem to contradict each other: Cambrian active margin versus Cambrian passive margin. However, as discussed above, the Cambrian-Early Ordovician arc-related rocks of the Chinese-Altai could also be attributed to a potential northern active margin, and thus a passive margin stage may have existed on its southern margin during the Cambrian to Early-Middle Ordovician (Fig. 3). The latter interpretation is consistent with the coeval onset of an active margin in Inner Mongolia. This leads to the hypothesis that the Altai-Mongolian Microcontinent could have been a ribbon-terrane that extended as far as northeastern China; its accretion to Siberia may have caused the onset of an active margin on its southern side. However, the poor Early Palaeozoic data available from western Mongolia and northeastern China prevent the testing of such a model. The recognition of Vendian-Cambrian subduction-accretion complexes (i.e. potential extension of the Lake Zone) between the Idermeg/Argunsky and Nuhetdavaa-Enshoo microcontinents would imply the exotic origin of the latter and thus its probable correlation with the Altai-Mongolian Microcontinent (Fig. 5b). At the present state of knowledge, the Nuhetdavaa-Enshoo Zone could also correspond to the southern extension of the Idermeg Microcontinent considered as part of the Siberian Tuva-Mongolian terrane (ss. 3.3.1). In this hypothesis, the Sunidzuoqi-Xilinhot island arc could be the eventual extension of the Cambrian-Early Ordovician Altai-Mongolian southward subduction zone.

3.4. Island arcs

3.4.1. The Dariv-Agardagh ophiolitic belt and the Lake-Khamsara arc

The external boundary of the Tuva-Mongolian Ribbon-Terrane is well defined by the Dariv-Agardagh ophiolitic belt. The ophiolites are located on the western margin of the Baidrag-Dzabkhan and Tuva-Mongolia microcontinents (Badarch et al., 2002) (Fig. 2). The correlation of the Khantaishir, Dariv and Agardagh ophiolites is strongly supported by their similar geochemistry (IAT, BABB and bonninites, Pfänder et al., 2002, 2004; Khain et al., 2003; Matsumoto and Tomurtogoo, 2003; Dijkstra et al., 2006), formation ages (ca. 570 Ma; U-Pb zircon; Pfänder et al., 1998; Kozakov et al., 2002; Khain et al., 2003) and structural position (i.e. thrusted onto a microcontinent passive margin; Kozakov et al., 2002; Pfänder et al., 2004; Dijkstra et al., 2006; Kröner et al., 2010; Stipska et al., 2010). Geochemical data suggest that the Dariv-Agardagh ophiolitic rocks formed in a proto-intra oceanic arc associated with forearc and back-arc basins (Pfänder et al., 2004; Dijkstra et al., 2006). Stipska et al. (2010) recently demonstrated from detailed geochronological and metamorphic studies that the Baidrag-

Dzabkhan passive margin was obducted in the Early Cambrian, earlier than previously thought (ca. 520-490 Ma, Pfänder et al., 2004; Dijkstra et al., 2006) (Fig. 3). This Early Cambrian age is well consistent with the ca. 540 Ma metamorphic event defined in the Tuva-Mongolian Massif (Moren, Fig. 2-5b, Salnikova et al., 2001).

The ophiolitic belt is juxtaposed to the Lakes-Khamsara Zone (West and East Lake, Hovd, Tannuola and Khamsara zones, Fig. 2), which consists of Vendian-Cambrian ophiolitic mélanges, island arc, back-arc, fore-arc and accretionary rocks (Kovalenko et al., 1996a, 1996b; Badarch et al., 2002; Kozakov et al., 2002; Naumova et al., 2006). It is generally well accepted that the Lake Zone extends northward into the Tannuola-Khamsara Zone (Mossakovsky et al., 1994; Ruzhentsev and Mossakovskiy, 1996; Sengör and Natal'in, 1996; Badarch et al., 2002; Naumova et al., 2006; Dobretsov and Buslov, 2007; Windley et al., 2007). Recent detailed geochemistry and geochronology of magmatic rocks and associated siliceous-terrigenous sediments support the previous hypothesis and confirm that the Lake Zone formed in the Vendian-Cambrian in an intra-oceanic arc environment (Kovach et al., 2011; Yarmolyuk et al., 2011). Kröner et al. (2010) underlined the existence of a Proterozoic basement in the Lake Zone and demonstrated with zircon dates its affinity with the northern Baidrag-Dzabkhan Microcontinent. They also proposed that the Khantaishir accretionary complex most probably originated in an ocean located to the south of the Lake Zone (Fig. 5a), and they identified a Late Cambrian thermal event, which is well consistent with a coeval tectono-magmatic event identified in the Tuva-Mongolian massif (Salnikova et al., 2001), the Dariv Range (Dijkstra et al., 2006) and the Baga Bogd Massif (i.e. West Gobi-Altai Zone; Demoux et al., 2009c). Although this event was previously attributed to the accretion of the arc, it was most likely related to isostatic and thermal relaxation in a post-collisional context (Kröner et al., 2010) (Fig. 3).

To the west of the Lake Zone, the Hovd and Ulgey Zones (Fig. 2), consisting mainly of Middle Cambrian-Ordovician limestones, clastic, volcaniclastic and sedimentary rocks metamorphosed in the greenschist facies was interpreted as a northern accretionary wedge associated with the Lake arc (Badarch et al., 2002) which was likely emplaced after the obduction of the ophiolite (i.e. subduction reversal) (Fig. 3). Nevertheless, as suggested by Kröner et al. (2010), the interpretation of the Hovd Zone as an accretionary wedge may be challenged, because no Late Cambrian-Early Ordovician arc-related rocks have been recognized in southern Mongolia. The continental margin may also have remained passive following the obduction; the Hovd volcaniclastics could have been derived from the erosion of this new high-standing arc.

3.4.2. The Uimen-Lebed arc and its Gorny-Altai accretionary wedge

In the Siberian Altai, Vendian to Late Cambrian island arc assemblages and associated accretionary complexes occur in the Uimen-Lebed, Gorny-Altai and Charysh-Terekta zones (Fig. 2, Dobretsov et al., 2004a; Ota et al., 2007). The evolution of the Uimen-Lebed island arc and its related Gorny-Altai accretionary wedge was well studied in the Salair, Katum and Kurai areas with detailed field, metamorphic, geochemical and geochronological data (Buslov et al., 2000, 2002; Ota et al., 2002; Dobretsov et al., 2004a, b; Safonova et al., 2004, 2008, 2011; Ota et al., 2007; Utsunomiya et al., 2009; Glorie et al., 2011; Safonova, 2008). These authors emphasized the occurrence of oceanic islands and the westward progradation of the arc system (Fig. 4). The Uimen-Lebed arc (Uimen-Lebed Zone) was generally interpreted as a Vendian-Cambrian island arc similar to the Izu-Bonin-Mariana arc system located close to the Siberian margin (Buslov et al., 2002; Dobretsov et al., 2004b; Ota et al., 2007; Glorie et al., 2011). The coeval existence of the Altai-Sayan back-arc basin (ss. 3.2.2 and Fig. 4) implies that at least two arcs existed in the area during the Vendian-Cambrian time (Fig. 5). The peri-Siberian plate tectonic framework was most likely complex during the Vendian and Cambrian; it probably looked like the Cenozoic SE Asian plate tectonic archipelago - see the reconstructions of Hall (2002).

3.5. Subduction-accretion zones

3.5.1. The Bayanhongor Zone: relics of at least two oceans

The Bayanhongor Zone is located between the Baidrag-Dzabkhan and Khangai (Tarvagatai) microcontinents (Badarch et al., 2002) (Fig. 2). It consists of various tectonic units that trend NW-SE and dip southwest and consist from south to north of a Neoproterozoic mélange, an Early Ordovician terrigenous volcano-clastic sequence, a sedimentary-dominated mélange, an ophiolitic mélange and a sedimentary-dominated mélange with seamount relics and schists (see details in Buchan et al., 2001 and Osozawa et al., 2008). The age of oceanic crust, which was dated as ca. 570 Ma (Sm-Nd; Kepezhinskas et al., 1991) and ca. 660 Ma (U-Pb; Kovach et al., 2005), led to some confusion (Windley et al., 2007), but a ca. 660-640 Ma age (U-Pb) was recently corroborated by Jian et al. (2010a). The age range and the oceanic plateau affinity of some ophiolitic rocks (Kovach et al., 2005) may imply that the Bayanhongor Ocean was wide (Windley et al., 2007).

The Baidrag-Dzabkhan Microcontinent and the Bayanhongor Ophiolite mélanges were affected by a Vendian-Early Cambrian (ca. 580-520 Ma) tectono-magmatic event (Buchan et al., 2002; Kozakov et al., 2006, 2008; Demoux et al., 2009a; Jian et al., 2010a; Kröner et al. 2011). This long tectono-magmatic event likely reflects more than one accretionary event. Jian et al. (2010a) proposed that an island arc and seamounts were successively accreted within the Bayanhongor Zone during Vendian-Early Cambrian time. Two distinct major accretionary/collisional events emerge from the palaeotectonic models of the authors: (1) the accretion of the Bayanhongor Ophiolite and (2) the collision between the Baidrag-Dzabkhan and Khangai microcontinents (Buchan et al., 2002; Kovalenko et al., 2005; Osozawa et al., 2008;

Demoux et al., 2009a). From the presence of Middle Cambrian-Early Ordovician flysch, Early Ordovician structures, Early-Middle Ordovician syn/post-collisional granitoids, and Late Ordovician greenschist metamorphism, it is generally accepted that the Bayanhongor Ocean finally disappeared during the Ordovician (Voznesenskaya, 1996; Buchan et al., 2001, 2002; Kovalenko et al., 2005; Osozawa et al., 2008) (Fig. 3).

The presence of post-Ordovician rocks and deformation indicates later tectonic activity in the Bayanhongor Zone (Dergunov, 2001). Devonian thrust stacking was attrributed by Osozawa et al. (2008) to the collision of Kazakhstan with the peri-Siberian margin (ss. 6.3). The occurrence of fossiliferous Carboniferous shallow-marine limestones and the seamount-bearing sedimentary-dominated mélange strongly suggests the extension of the Mongol-Okhotsk Ocean in the Bayanhongor area. The undated mélange shows a strong affinity with rocks in the Devonian-Carboniferous Khangai-Khentey basin, well accepted as the Mongol-Okhotsk Ocean (Parfenov et al., 1999; Zorin, 1999; Kelty et al., 2008; Kurihara et al., 2009; Ruzhentsev and Nekrasov, 2009). Jian et al. (2010a) recently documented the occurrence in the Bayanhongor area of Permo-Triassic ophiolitic gabbros and lavas (ca. 298-210 Ma) that also contain Precambrian zircon xenocrysts. This means that there are two ophiolites in the Bayanhongor Zone of 660-640 Ma and 298-210 Ma age. The first formed during the evolution of the Altaids, and the second most likely formed in a narrow rift (thus explaining the presence of Precambrian zircons) ocean at the western end of the Mongol-Okhotsk Ocean, which has long been regarded as a triangular-shaped ocean that closed progressively eastwards in a scissor-like fashion (Zhao et al., 1990; Zonenshain et al., 1990; Tomurtogoo et al., 2004). Bayanhongor is situated on the western end of the Mongol-Okhotsk Suture outlined by Zorin (1999). Closure by subduction of the Mongol-Okhotsk Ocean would accordingly explain the presence of the 220-200 Ma calc-alkaline Khentei batholith (Kovalenko et al., 2004), and farther east of the accreted Onon island arc (ss. 3.5.5). Obduction by thrusting, probably at the end-Triassic, placed the second ophiolite against the first ophiolite that had already been emplaced by thrusting in the Neoproterozoic, thus making this the longest and widest ophiolite zone in Central Asia. However, at present the boundaries and extent of the first and second ophiolites are still unknown (more mapping is required), and none of the first or second generation of thrusts has been dated.

The close affinities between the Dzhida-Bayangol and Bayanhongor zones (i.e. age and nature of oceanic materials, timing of accretion) strongly suggest their origin in a single ocean that finally closed in the Early Ordovician (Fig. 3 and ss. 3.5.2). Such a hypothesis would imply, as proposed by Kovalenko et al. (2005), the imbrication and duplication of the Khangai and Baidrag-Dzabkhan microcontinents. However the existence of strike-slip deformation within the Bayanhongor Zone was rejected by Osozawa et al. (2008). Alternatively, the Baidrag passive margin may have extended as far as the Tarvagatai Precambrian basement, and the northerly obducted Bayanhongor Ophiolite originated in the Dzhida-Bayangol Ocean (Fig. 5a,b).

3.5.2. The Dzhida-Bayangol Zone: multiple accretionary events within a Vendian-Cambrian ocean

In the middle of the Sibero-Mongolian orocline, outlined by the above-described microcontinents (Tuva-Mongolian, Khangai, Baidrag-Dzabkhan, Argunsky and Idermeg) and by Siberia is the Dzhida-Bayangol Zone (Dzhida, Eravna and Bayangol Zones on Fig. 2, Naumova et al., 2006). This zone is mainly characterized by Vendian-Cambrian arcs and subduction-accretion complexes (Badarch et al., 2002; Naumova et al., 2006; Gordienko et al., 2007; Windley et al., 2007; Zorin et al., 2009). Geochemichal and isotopic studies on Phanerozoic granitoids in west-central Mongolia suggest the massive production of juvenile crust with limited influence of old microcontinents (Jahn et al., 2004). The Dzhida-Bayangol Zone was studied in detail in the Dzhida and Bayangol regions where there are subduction-accretion complexes including island arc rocks (i.e. boninites and basalts of supposed Vendian age, Early-Middle Cambrian rhyolites and andesites, ca. 506-505 Ma plagiogranite-diorite complexes and related mélanges), oceanic islands (i.e. ophiolitic mélanges with oceanic crust and Early Cambrian limestones), mélanges, flysch, and metamorphic-magmatic complexes (Badarch et al., 2002; Gordienko and Filimonov, 2005; Gordienko et al., 2007; Safonova, 2009) (Fig. 3). As proposed by Zorin et al. (2009), the Cambrian Dzhida island arc likely extends eastwards into the Eravna Zone where there are also Vendian-Cambrian island arc volcano-sedimentary assemblages (Naumova et al., 2006). In the southern part of the Barguzin Zone, a dismembered ophiolite represented by metamorphosed gabbro, diorite, plagiogranite and ultrabasic rocks has a Sm-Nd age of ca. 545+/-13 Ma (Belichenko et al., 2006). A metamorphosed volcano-sedimentary complex containing Vendian ophiolites (i.e. N-MORB, E-MORB and OIB with ca. 580 Ma Sm-Nd model ages) was reported in the Onon Zone (Naumova et al., 2006); these rocks occur in the eastern extension of the Vendian-Cambrian accretionary complexes. The above information implies that a major ocean existed between Siberia and the Tuva-Mongolian Terrane in the Vendian and Cambrian and that numerous accretionary events probably affected its margins; we call this the Dzhida-Bayangol Ocean (Fig. 5a,b).

Although the Dzhida-Bayangol Zone is poorly documented, the available data below support the closure of this ocean by the Early Ordovician. An Ordovician collisional event documented by flysch deposits intruded by ca. 490-470 Ma collisional granitic rocks and by regional metamorphism occurs in the Dzhida area (Gordienko et al., 2007). These ages are consistent with the period of Ordovician regional metamorphism identified in the Baikal area (Gordienko et al., 2007 and references therein), and with Late Cambrian-Early Ordovician flysch deposits in the northern Khentey Zone (Haraa locality, Badarch et al., 2002; Naumova et al., 2006; Kelty et al., 2008). The Haraa flysch sediments contain zircons with ages ranging from ca. 2.6-2.4, 1.95-1.8, 9.0-8.0 Ga to 605-504 Ma, which likely come from the Mongolian microcontinents (Kelty et al., 2008). Forearc basalts, andesites and dacites interlayered with volcanic-derived sandstones, siltstones and shales in the Haraa area (Badarch et al., 2002) contain detrital zircons with ages ranging from 483 to 2568 Ma (Kröner et al., 2007); these ages imply that the arc activity ended by the Early Ordovician (Fig. 3).

The end of arc activity in the Haraa area is consistent with the time of Early Ordovician collision identified in the Dzhida, Baikal and Bayanhongor regions and with the general absence of Middle-Late Ordovician deposits in Central Mongolia and Transbaikalia (Naumova et al., 2006) (Fig. 3). However, Kröner et al. (2007) reported brief arc magmatic activity between 460 and 435 Ma in the Haraa and Herlen areas; this contradicts the Early Ordovician closure of the Dzhida-Bayangol Ocean, but the Late Ordovician-Silurian arc activity could be related to the northward subduction of the Palaeo-Asian Ocean (i.e. peri-Siberian margin, ss. 3.6 and Fig. 6). The pre-Silurian accretion of the Mongolian microcontinents and other Vendian-Cambrian island arcs to Siberia (i.e. “Tuva-Mongolian Orogeny”, Fig. 3) is strongly supported by the existence of a peri-Siberian margin located south of the Main Mongolian Lineament (Badarch et al., 2002; Windley et al., 2007).

3.5.3. The Borus-Kurtushiba Zone: an arc-arc collisional zone?

The Borus-Kurtushiba Zone corresponds to the West Sayan Zone, which is delimited to the south and north by the Kurtushiba and Borus ophiolitic and HP metamorphic belts, respectively (Fig. 2, Naumova et al., 2006). It mainly consists of Vendian-Cambrian ophiolites and volcano-sedimentary accretionary rocks, thick Cambrian-Ordovician sands and shales interpreted as flysch deposits, and of Silurian shallow-water terrigenous carbonate shelf sediments (Naumova et al., 2006). HP and ophiolitic rocks are associated with the volcano-sedimentary accretionary rocks in both belts (Dobretsov and Buslov, 2004; Naumova et al., 2006 and references therein). The different geochemistry of ophiolites (MORB and OIB) and of HP ages (at Kurtushiba: 600-520 Ma, Dobretsov and Sklyarov, 1989 in Volkova and Sklyarov, 2007; and at Borus: 540-440 Ma; Dobretsov and Buslov, 2004) suggest different accretionary events.

The formation time of the Borus-Kurtushiba "Suture Zone" is suggested by Cambrian-Ordovician flysch and Ordovician shallow-water molasse deposits occurring in West Sayan and Tannuola/Khamsara (the Khemchik-Sistigkhem Basin on Fig. 2, Naumova et al. 2006). The younger age of the flysch is not precisely known, but the dated younger arc activity suggests ocean closure was in the Middle Cambrian. The particular structure of the Borus-Kurtushiba Zone (i.e. accretionary and flysch rocks surrounded by HP-ophiolitic belts and shallow marine cover) could possibly be attributed to facing "soft" arc-arc collisions (Hall, 1999; Hall and Smyth, 2008). Similar structures in the Blue Mountains were attributed to facing arc-arc collisions (Dorsey and LaMaskin, 2007). The Borus and Kurtushiba belts may be accretionary wedges associated with the Sayan and Khamsara arcs, respectively. However, the Western Sayan regional geology is still poorly known and detailed fieldwork together with geochronology and geochemistry are required to test this hypothesis (Fig. 5b).

3.5.4. The Charysh-Terekta Zone: the accretion of the Altai-Mongolian Microcontinent

The Charysh-Terekta dextral strike-slip zone (Fig. 2) was interpreted as a Late Devonian-Early Carboniferous suture resulting from the oblique collision between the Altai-Mongolian Terrane and Siberia (Buslov et al, 2000, 2004a, b; Safonova et al., 2004; Dobretsov and Buslov, 2007). The strike-slip zone is marked by mélange-greenschist facies shear zones containing fragments of the following: a Cambrian-Ordovician Gorny-Altai arc system (i.e. Middle-Late Cambrian Anui-Chuya forearc sediments, Early-Middle Ordovician forearc sediments with pelagic fauna and Cambrian-Early Ordovician clasts of magmatic arc origin), an ocean (i.e. ophiolitic mélanges with Late Tremadocian-Early Arenigian cherts), an Early Palaeozoic accreted terrane (the Uimon HP metamorphosed volcano-sedimentary complex with E-MORB, N-MORB and OIB, ca. 490-480 Ma blueschists; Volkova and Sklyarov, 2007), an Ordovician-Silurian marginal basin (Ordovician clastics with benthic and pelagic fauna and an Early-Middle Ordovician volcanogenic component, a Late Ordovician through Devonian carbonate platform, Yolkin et al., 1994), a Devonian arc and related basin (an Early-Middle Devonian sedimentary-volcanic arc assemblage, an Early Devonian limestone with crinoids and corals), and finally, clastic fragments of the Altai-Mongolian Terrane (see details in Buslov et al., 2004b) (Fig. 4).

In the Gorny-Altai the arc activity ended by the Early-Middle Ordovician; a deep to shallow-marine basin progressively developed from the east (Uimen-Lebed) in the Tremadocian to the west (Charysh-Terekta) in the Arenigian and it lasted until the Early Devonian (Yolkin et al. 1994). The presence of ca. 490-480 Ma blueschists in the Uimon ophiolitic complex is consistent with a Tremadocian deformation phase (metamorphism, magmatism and folding) identified in the Gorny-Altai and Uimen-Lebed areas (Buslov et al., 2002; Dobretsov et al., 2004b; Glorie et al., 2011). This accretion-collision event is also supported by Early-Middle Ordovician flysch sediments, which replaced Late Cambrian forearc-type deposits in the Gorny-Altai area (Yolkin et al., 1994) (Fig. 4).

The Early Ordovician accretion of the Uimon Terrane (island arc system?) within the Gorny-Altai accretionary wedge likely caused the end of arc magmatism and the westward progradation of subduction in the Chara Zone (i.e. Late Ordovician-Silurian northward accretionary wedge, ss. 3.3.2). This implies that a coeval arc was located in the Kalba-Narym and/or Rudny Altai zones where no Early Palaeozoic rocks crop out. The Ordovician-Devonian deep-shallow marine sediments were not deposited on a passive margin, as sometimes proposed, but most probably in a back-arc according to Yolkin et al. (1994). The lack of a magmatic component may be explained by the existence of a deep basin to the west that trapped the sedimentology input. The poorly documented metamorphosed sedimentary-volcanogenic rocks associated with an Ordovician-Silurian (?) oceanic basement in the Rudny-Altai Zone (Buslov et al., 2004a, b) could possibly be the relic of a back-arc basin (Fig. 4).

The time of oblique accretion of the Altai-Mongolian Microcontinent to Siberia was well constrained in the Siberian Altai as Late Devonian (Charysh-Terekta Shear Zone, Buslov et al., 2000, 2004a, b). Glorie et al. (2011) recently reported that voluminous granitoids predated (Middle-Late Devonian) the large-scale strike-slip displacements. Buslov et al. (2000, 2004b) proposed that the Altai-Mongolian Microcontinent was part of the palaeo-Kazakhstan continent, which was still isolated in the Palaeo-Asian Ocean during the Middle Palaeozoic, and which started to collide with Siberia in the Middle-Late Devonian. Nevertheless, the probable existence of a Middle-Late Ordovician through Carboniferous peri-Siberian active margin led to the hypothesis that the Altai-Mongolian Terrane may have already been accreted to Siberia before, and thus became the new Siberian active margin after the Late Ordovician (ss. 3.3.2, Fig. 6). This potential Middle Ordovician accretion-collision event is consistent with the nearly coeval accretion of the Uimon Terrane in the Siberian Altai (Fig. 5b). For these reasons, we suggest that the Altai-Mongolian Microcontinent was already accreted southwards in the Mongolian part of Siberia in the Late Ordovician and, as proposed by Buslov et al. (2000), it later moved along the Siberian margin in the Late Devonian (ss. 6.3). This last assumption prevents the attribution of the Altai-Mongolian Microcontinent to the Kazakhstan Terrane that was located at this time in eastern Gondwana (ss. 4.2), but no available data seem to contradict such a distinct origin and evolution of both terranes. This potential model (Fig. 5b) seems problematic because a microcontinent-continent collision should have widely affected the Siberian margin and no such major tectonic event has been documented. Nevertheless, Lamb and Badarch (2000) suspected a Middle-Late Ordovician compressional event from the occurrence of Ordovician angular unconformities and of variable degrees of deformation and metamorphism at different localities in southern Mongolia (i.e. unconformable shallow-marine deposits). These observations were recently corroborated by the detailed lithostratigraphic studies of Kröner et al. (2010). These unconformities may be explained by soft collision between the Altai-Mongolian Terrane and Siberia.

We have seen in section 3.3.2 that the Altai-Mongolian Terrane possibly extended to Inner Mongolia. In southwestern Mongolia, the attribution of a Precambrian basement to the Altai-Mongolian Terrane is unclear. As suggested by Kröner et al. (2010), the South Gobi Precambrian basement may also be a part of the Baidrag-Dzabkhan Microcontinent. As shown by the significant thinning of tectonic units (see map of Naumova et al., 2006) and by detailed metamorphic and structural studies (Kozakov et al, 2007a, Lehmann et al., 2010), southwestern Mongolia underwent intense strike-slip deformation in the Middle Palaeozoic (ss. 6.3). This makes spatial correlations difficult and suggests that the units were extensively stretched. From palaeomagnetic data Buslov et al. (2000) argued that the Altai-Mongolian Microcontinent migrated after the Emsian over a distance of more than 2000km along the Siberian margin. Such a distance seems to be consistent with the potential extension of the Altai-Mongolian Microcontinent and the Nuhetdavaa-Enshoo Microcontinent as one independent terrane, which underwent Ordovician accretion in Mongolia (Fig. 5b). Southwestern Mongolian Precambrian rocks may be fragments of this terrane (ss. 3.3.2)

3.5.5. The Onon Zone of the Mongol-Okhotsk Ocean: a Vendian or Silurian opening time?

The Mesozoic Onon Suture Zone (or Aga Zone) occupies a central position within the Altaids (Naumova et al., 2006, Fig. 2). It is generally accepted as the main suture of the Mongol-Okhotsk Ocean (e.g. Sengör and Natal'in, 1996; Zorin, 1999; Windley et al., 2007). The Mongol-Okhotsk Orogenic Belt was formed through the Jurassic-Cretaceous following a diachronous oceanic closure from Mongolia to the Russian Pacific margin (Gusev and Khain, 1996; Zorin, 1999; Kravchinsky et al., 2002; Yakubchuk, 2004, 2008; Cogné et al., 2005; Tomurtogoo et al., 2005; Ruzhentsev and Nekrasov, 2009). The opening time of the Mongol-Okhotsk Ocean has been variably interpreted in the last decades;Vendian (Gordienko, 1994; Gusev and Khain, 1996; Sengör and Natal'in, 1996; Tomurtogoo et al., 2005), Silurian (Zorin, 1999; Badarch et al., 2002;

Kröner et al., 2007; Windley et al., 2007; Ruzhentsev and Nekrasov, 2009; Bussien et al., 2011), or Permian (Zonenshain et al., 1990). Although the most recent geological data preclude a Permian opening, both Vendian and Silurian times are acceptable by the scientific community.

In Mongolia and Transbaikalia relics of the Mongol-Okhotsk Ocean are represented by the Onon Suture Zone and the Khangai-Khentey basin (Fig. 2) (Zorin, 1999; Ruzhentsev and Nekrasov, 2009). The Khangai-Khentey basin was generally interpreted as a Devonian-Carboniferous accretionary wedge of the northern Siberian (Parfenov et al., 1999; Zorin, 1999; Kurihara et al., 2009; Bussien et al., 2011) or southern Mongolian margins (Kelty et al., 2008). It consists mainly of thick turbidites with OIB and associated Pridolian to Frasnian cherts (Kelty et al., 2008; Kurihara et al., 2009). The Onon Suture Zone mainly consists of polymict mélanges containing Visean-Tournaisian island arc rocks (quartz diorite, granite, basalt, dolerite, dacite, tuffs and clastics), Visean-Serpukhovian back-arc sediments, Late Ordovician-Early Silurian (ca. 450-430 Ma) ophiolitic rocks (layered gabbro, gabbrodiorite, plagiogranite and dolerite dykes with a MORB chemical affinity), Late Silurian-Early Devonian oceanic crust, Early Carboniferous ophiolites (ca. 325 Ma; Tomurtogoo et al., 2005) and other Devonian-Triassic marine clastics (see details in Ruzhentsev and Nekrasov, 2009).

The various interpretations of the time of opening of the Mongol-Okhotsk Ocean mainly come from the spatial correlations between the different oceanic relics. A Vendian opening time was commonly proposed because of the occurrence in the Dzhida-Bayangol and Bayanhongor zones of Vendian-Cambrian oceanic relics attributed to the Mongol-Okhotsk Ocean (Gusev and Khain, 1996; Sengör and Natal'in, 1996; Tomurtogoo et al., 2005). Although the simplicity of this model (i.e. a single ocean from the Vendian through the Jurassic) is attractive, the oceanic relics may also have formed within different oceanic domains. We have seen above that the Bayanhongor and Dzhida-Bayangol oceans without doubt existed in the Vendian, and that the "Tuva-Mongolian Orogeny" most probably formed by the Early-Middle Ordovician (ss. 3.5.1-2 and Fig. 3). Although many investigations are still required in the Dzhida-Bayangol area, present knowledge favours a Silurian time of opening for the Mongol-Okhotsk Ocean. The Onon Suture Zone and Khangai-Khentey Basin are exclusively composed of Silurian through Mesozoic oceanic materials (Ruzhentsev and Nekrasov, 2009), which are consistent with the occurrence of Late Ordovician-Early Silurian (ca. 450-430 Ma) extensional magmatism (Ruzhentsev and Nekrasov, 2009) that most likely reflects the first rifting of the Mongol-Okhotsk Ocean. This is also supported by the general lack of Ordovician deposits in Transbaikalia and northern Mongolia (Naumova et al., 2006) and the return of shallow to deep marine sediments during the Silurian (i.e. passive margin and slope sediments unconformably overlie older basement in the Argunsky area and Vendian-Cambrian rocks in the Khentey area, respectively (Badarch et al., 2002; Ruzhentsev and Nekrasov, 2009) (Fig. 3).

A Late Ordovician-Silurian opening is most likely from a plate tectonic point of view. The presence of a Silurian-Carboniferous peri-Siberian active margin (Fig. 6) implies that the southern Mongolian margin of the Mongol-Okhotsk Ocean remained attached to Siberia. This also supported by the occurrence of Silurian Tuvaella fauna within clastics in the Argunsky area (Badarch et al., 2002) and is consistent with palaeomagnetic data (Kravchinsky et al., 2002). Sengör and Natal'in (1996) proposed an interesting palaeotectonic model that took account of the above considerations, according to which the extension of the Siberian active margin in Mongolia was located on the external side of the Vendian-Jurassic Khangai-Khentey Ocean (Mongol-Okhotsk Ocean). However, the existence of such a long-lived ca. 500 Ma ocean, subducting under Siberia and Mongolia for most of its life span, is hardly viable in terms of modern plate tectonics. Although no geological data clearly contradict the opening of the Mongol-Okhotsk during the Vendian, the model of a Silurian scissor-like opening (Zhao et al., 1990; Zonenshain et al., 1990; Badarch et al. 2002; Tomurtogoo et al., 2004) seems most likley in the light of current available data (Fig. 14-15-16-17 and 18).

3.6. *The Devonian to Carboniferous peri-Siberian active margin*

The pre-Late Ordovician Siberian accretion of the Barguzin and Altai-Sayan back-arcs (ss. 3.2), the Tuva-Mongolian and Altai-Mongolian microcontinents (ss. 3.3), and the Lake-Khamsara and Uimen-Lebed island arcs (ss. 3.4), and of numerous other relicts within the Bayanhongor, Dzhida-Bayangol and Charysh-Terekta suture zones (ss. 3.5) is strongly supported by the remnants of Devonian-Carboniferous continental arcs, subduction-accretion complexes as well as back-arc basins located all along the Siberian margin (Siberian Altai, Chinese Altai, western Mongolia, southern Mongolia and Inner Mongolia, Figs. 6-9).

In the Siberian Altai a Late Carboniferous-Early Permian mélange (Silurian-Carboniferous oceanic crust and fragments of Early Palaeozoic mélanges) and Late Devonian-Early Carboniferous sedimentary-volcanic rocks (Upper Devonian basalts and cherts, Early Carboniferous island arc rocks associated with siliceous sediments with radiolaria, Visean-Namurian olistostromes containing Middle Devonian-Early Carboniferous oceanic crust probably originated from oceanic islands/seamounts as well as Namurian turbidites and tuffs) are associated with Early Palaeozoic mélanges within the Chara Zone (Buslov et al., 2004a, 2004b; Safonova et al., 2004, 2012). Subduction-accretion complexes containing pelagic and slope deposits and oceanic crustal fragments ranging in age from Ordovician to Visean are also reported in the Chinese Altai (Keksanto, Qiaoxiahala and Qinghe ophiolitic mélanges, Xiao et al., 2009b and references therein). Arc-type volcano-sedimentary rocks and numerous ca. 440-360Ma arc-related granitoids in the Chinese Altai were attributed to the Late Ordovician-Carboniferous active margin of the Altai-Mongolian Microcontinent (Windley et al., 2002; Wang,T. et al., 2006; Briggs et al., 2007; Yuan et al., 2007; Sun et al., 2008; Chai et al., 2009; Cai et al., 2011) (Fig. 6).

In section 3.5.4 we suggested that the Ordovician-Silurian Siberian arc was most probably located in the Rudny-Altai and/or Kalba-Narym Zones of the Siberian Altai and that a shallow to deep basin existed in the back-arc region (Gorny-Altai, Charysh-Terekta and Rudny-Altai Zones) during the Silurian and Early Devonian (Fig. 4). The basin was likely closed in the Emsian when deep trench clastics were deposited on metamorphosed rocks in the Rudny-Altai (Buslov et al., 2004a) and an arc assemblage was developed on the Gorny-Altai and Charysh-Terekta carbonate platform (Yolkin et al., 1994). The active margin was again affected by extension as indicated by Middle-Late Devonian extension-related magmatism and Late Devonian-Early Carboniferous back-arc rocks (Yolkin et al., 1994; Naumova et al., 2006); the arc was displaced westwards in the Rudny-Altai Zone and associated fore-arc sediments were deposited in the Kalba-Narym Zone (Yolkin et al., 1994) (Figs. 6-13).

Cai et al. (2011) summarized data from the Chinese Altai that are indicative of a typical subduction-accretion orogen largely of Silurian-Devonian age, such as thrusted turbidites with marine flysch rhythms (Habahe Group), Devonian pillow basalts, andesites, dacites, pyroclastic rocks, boninites and adakites generated in a subduction-related setting (Xu et al., 2003; Niu et al., 2006), and 440-360 Ma (the main period of arc activity) calc-alkaline granitic rocks with positive $\varepsilon_{Hf}(t)$ zircons suggesting juvenile mantle origin (Wang, T. et al., 2006). In the Chinese Altai, a Devonian back-arc basin is documented by the ca. 370 Ma BABB Kuerti Ophiolite (Xu et al., 2003; Zhang, H. et al., 2003), and possibly by Middle Devonian turbidites in the Kalba-Narym Zone, and by thick metamorphosed Middle Devonian to Early Carboniferous cherty clastics in the Altai-Mongolian Zone (see Fig. 9, Windley et al. 2002). The opening time of this back-arc basin is indicated by ca. 415-400 Ma extensional continental active margin magmatism in the Kalba-Narym and Altai-Mongolian zones (Wang, T. et al., 2006; Yuan et al., 2007). Also the ages of granitoids in the Chinese Altai (Windley et al., 2002; Wang, T. et al., 2006; Briggs et al., 2007; Yuan et al., 2007; Sun et al., 2008; Wang, W. et al., 2009) support the opening of a back-arc basin in the Early-Middle Devonian. The intrusions stopped between ca. 400 and 375

Ma in the Altai-Mongolian Zone and continued until 380 Ma in the Kalba-Narym Zone; this most likely reflects the displacement of the Kalba-Narym arc with respect to the new Altai-Mongolian passive margin (Fig. 6-13).

In Mongolia, the Tseel Devonian arc assemblage attributed to a transitional arc-back-arc setting may represent the extension of the Kuerti back-arc basin in Western Mongolia (Helo et al., 2006; Demoux et al., 2009b). Ordovician through Early Carboniferous subduction-accretionary complexes occur in the Bidz and Zoolen Zones (southern part of the Gurvansayhan-Zoolen Zone, Badarch et al., 2002; Helo et al., 2006). The associated Silurian through Mississippian continental arc documented by magmatism and volcano-sedimentary rocks is located farther north in the northern part of the Gurvansayhan-Zoolen zones and in the West and Central Mandalovoo zones (Lamb and Badarch, 2000, 2001; Badarch et al., 2002). The arc assemblages were interpreted as a Japan-type magmatic arc according to zircon ages, metamorphic petrology and geochemical data (Kozakov et al., 2002, 2007a; Helo et al., 2006; Demoux et al., 2009b; Kröner et al., 2010). Continental Devonian through Early Carboniferous shallow to deeper water volcano-sedimentary back-arc rocks are in the West and Central Mandalovoo and the West and Central Gobi-Altai Zones (Lamb and Badarch, 2000, 2001; Badarch et al., 2002; Blight et al., 2008). In Southern Mongolia, back-arc extension started in the earliest Devonian (Lamb and Badarch, 2000). Recently, Kröner et al. (2010) showed that a first phase of continental stretching affected the Gobi-Altai and Mandalovoo Zones in the Late Silurian, giving rise to a carbonate platform (Fig. 6-13).

In Inner Mongolia, an Ordovician through Carboniferous Siberian continental arc margin (Fig. 9) is mainly documented by ca. 490-310 Ma Baolidao arc magmatism (Chen et al., 2000; Jian et al., 2008; Chen et al., 2009), by ca. 536-302 Ma detrital zircons within the Xilinhot accretionary wedge or/and forearc metamorphosed complex (Shi et al., 2003; Chen et al., 2009), by the Early Carboniferous Jiaoqier Ophiolite (Miao et al., 2007b) and other ophiolites along the

Jiaoqier-Xilinhot Fault (Wang, Q. and Liu, 1986; Xiao et al., 2003), and by the Sunidzuoqi subduction-accretion complex (Jian et al. 2008). The Devonian Mongolian back-arc basin likely extended into the Nuhetdavaa-Enshoo area where there is a thick and complete shallow-marine, arc-related sedimentary series (Yue et al., 2001; Badarch et al., 2002) (Fig. 6-9). The occurrence of uppermost Silurian turbidites (Yue et al., 2001) is also consistent with coeval continental stretching recognized in Mongolia. Although the back-arc basin was probably not ‘oceanised’ in Mongolia, the nature of the basin is less clear in China because of variable ages attributed to the Hegenshan ophiolites (Devonian and/or Permian; Hsü et al., 1991a; Robinson et al., 1999; Xiao et al., 2003; Miao et al., 2007a; or Carboniferous and Cretaceous; Jian et al., in press). Two back-arc basins probably existed in Inner Mongolia: (1) a Devonian oceanic/continental basin and (2) a Permian oceanic basin (Miao et al., 2007a) (Fig. 13).

4. Early Palaeozoic formation of the Kazakhstan Continent

The time of formation of the Kazakhstan continent in the Palaeozoic has been widely discussed (Zonenshain et al., 1990; Mossakovsky et al., 1994; Sengör and Natal'in, 1996; Kheraskova et al., 2003; Yakubchuk, 2004; Degtyarev and Ryazantsev, 2007; Kröner et al., 2008; Windley et al., 2007; Abrajevitch et al., 2008; Burtman, 2008; Levashova et al., 2009; Alexeiev et al., 2010; Biske and Seltmann, 2010; Alexeiev, 2011), and from this discussion two main ideas emerge: (1) The Kazakhstan continent was created by Middle-Late Palaeozoic thrust duplication and bending of one to three island arcs that originated in the Baltica and Siberia continents (Sengör and Natal'in, 1996; Yakubchuk, 2004), and (2) the Kazakhstan continent was formed by the successive amalgamations of island arcs and microcontinents that originated in eastern Gondwana by the Silurian, and the derivative continent was then bent into an orocline in the Late Palaeozoic (Windley et al., 2007; Abrajevitch et al., 2008; Alexeiev et al., 2010; Biske and Seltmann, 2010). As pointed out by the latter authors, substantial geochemical, geochronological, palaeomagnetic and geological data, published since Sengör and Natal'in

(1996), have corroborated pre-Silurian, multiple amalgamation (ss. 4.1) in eastern Gondwana (ss. 4.2).

4.1. Major stages of formation and potential correlations

The mains issues that inhibit the construction of a well-constrained palaeotectonic model for continental growth of Kazakhstan in the Early-Middle Palaeozoic mainly reside in the definition and potential spatial correlations of sutures and terranes (Alexeiev et al., 2010) (Fig. 2). Although more fieldwork and geochronologic/geochemical studies are still necessary in order to better constrain the pre-Silurian amalgamation of Kazakhstan, some important considerations arise from the available data. The continental growth of Kazakhstan was aided by four major tectonic events in: (1) the Terreneuvian (ca. 540-530Ma), (2) the Amgan (ca. 520-510 Ma), the Arenigian (ca. 480-470 Ma) and the Hirnantian-Rhuddanian (ca. 450-440 Ma) (Fig. 7).

The Terreneuvian event has been recognized mainly in northern Kazakhstan in the Kokchetav Massif (the Kumdykol Suture, Degtyarev and Ryazantsev, 2007; Windley et al., 2007; Alexeiev, 2011), which created a diamond/coesite-bearing UHP tectono-metamorphic mélange in a Cambrian-Early Ordovician accretionary complex (Dobretsov et al., 2005b). Maruyama and Parkinson (2000) presented new data and an impressive overview of the geology, petrology and tectonic framework of the UHP belt. The metamorphic rocks that have continental and oceanic protoliths yielded a ca. 540-530 Ma UHP metamorphic peak and a ca. 530-520 Ma exhumation trajectory (Dobretsov and Buslov, 2004 and references therein). The Early Cambrian metamorphism resulted from the northward subduction of a microcontinental (the West Teniz Zone on Fig. 2) passive margin under a Vendian-Cambrian island arc (Dobretsov et al., 2005a; Degtyarev and Ryazantsev, 2007). Following the obduction, subduction started under the microcontinent (Dobretsov et al., 2005a, 2005b). No coeval UHP metamorphic complex has been identified in Kazakhstan. However, Vendian-Early Cambrian oceanic and ophiolitic rocks occur in the Erementau Mountains (seamounts, Yakubchuk, 1990), in Central Chingiz (Early Cambrian ophiolites, Yakubchuk and Degtyarev, 1994), in the Aktyuz Massif (ca. 530 Ma metagabbro, Kröner et al. 2012), in the Kyrgyz and Terskey Ranges (Early Cambrian IAT/MORB, Lomize et al., 1997), and in western Junggar (Tangbale, ca. 540-520 plagiogranites with E-MORB signature and gabbro, Buckman and Aitchison, 2004; Jian et al., 2005); all these relict oceanic rocks may be derived from the Kumdykol Ocean (Fig. 7).

The Amgan event is characterized by a stratigraphic unconformity, which crops out in the East Teniz (Urumbai), Selety-Erementau (Selety), West and East Boshchekul-Chingiz Zones (Boshchekul and Chingiz Ranges) (Fig. 2). At the Amgan unconformity arc-related rocks are unconformably overlain by shallow to deep marine terrigenous-carbonate sediments (Degtyarev and Ryazantsev, 2007). Cambrian oceanic rocks and deep-marine sediments are generally associated Amgan unconformity (Alexeiev et al., 2010; Degtyarev and Ryazantsev, 2007; Kröner et al., 2012). Amgan to Arenigian deep-marine sediments occur in the Central Chingiz, Selety, Urumbai and Dzhalair-Naiman Zones (Yakubchuk, 1990; Yakubchuk and Degtyarev, 1994; Degtyarev and Ryazantsev, 2007; Tolmacheva et al., 2008; Alexeiev et al., 2009), and Middle-Late Cambrian cherty terrigenous sediments are in the Kendyktas Massif (Issyk-Kul Zone) (Degtyarev and Ryazantsev, 2007). In the Dzhalair-Naiman Suture Zone there are strongly deformed sandstones, which conformably overlie red jasper cherts and basaltic pillow lavas underlain by gabbro cut by diabase dykes with lenses of plagiogranite dated at ca. 512 Ma (Kröner et al., 2008). A similar ca. 516 Ma, SHRIMP U-Pb zircon age of basalts in the southern Tian Shan in China (Xiate locality on Fig. 2, Qian et al., 2009) suggests the eastward extension of the suture. Also, there are ca. 520 Ma (TIMS U-Pb) plagiogranites in a sheeted dyke complex in an ophiolite in the Dzhalair-Naiman Zone (Andassai and Dulankara massifs, Ryazantsev et al., 2009), and Cambrian-Early Ordovician ophiolitic rocks in the Kyrgyz-Terskey Ranges (pillow basalts associated with Cambrian-Earliest Ordovician cherts in an ophiolitic mélanges, Lomize et al., 1997) and in the Arkalyk Suture Zone (mélange with Late Cambrian-Lower Ordovician cherts and basalts, Yakubchuk and Degtyarev, 1994). Therefore, the Urumbai, Dzhalair-Naiman, Kyrgyz-Terskey and Arkalyk sutures (Windley et al., 2007 and Fig. 2) may contain the relics of a single ocean (Fig. 7). In Central Chingiz, the Amgan-through-Arenigian deep-marine sediments are associated with a coeval arc assemblage, interpreted to have formed in a back-arc (Yakubchuk and Degtyarev, 1994; Tolmacheva et al., 2008). The correlation of the Boshchekul and Chingiz Ranges is strongly supported by detailed stratigraphy, which shows Early Cambrian through Latest Ordovician arc assemblages (Yakubchuk, 1990; Yakubchuk and Degtyarev, 1994; Degtyarev and Ryazantsev, 2007; Windley et al., 2007). Early Cambrian (ca. 508 Ma granodiorites, ca. 534 Ma metadacites) and Early Ordovician (ca. 480 Ma metadacites, granodiorites and granitoids) arc rocks were reported in the Chu-Yili (Kröner et al., 2008; Gao et al., 2009; Alexeiev et al., 2010) and Issyk-Kul (a Cambrian to Arenigian arc assemblage, Lomize et al., 1997) microcontinents. Thus evidence suggests that a Cambrian through Early Ordovician arc was associated with a back-arc basin (Fig. 7).

The Arenigian event is locally characterized by a cessation of arc activity, stratigraphic unconformities, olistostromes, flysch deposits, HP metamorphism, folding, and granitic intrusions, which together clearly reflect a significant accretion-collision event (Dobretsov et al., 2005a; Degtyarev and Ryazantsev, 2007; Kröner et al., 2007, 2012; Alexeiev et al., 2010; Biske and Seltmann, 2010; Alexeiev, 2011). Most of the Kazakhstan sutures formed in the Arenigian, implying that they evolved from the closure of a single ocean. The Arenigian event was recognized or suspected in the Kumdykol Suture Zone (Arenigian olistostrome and thrusting, Dobretsov et al., 2005a, b), in the Kyrgyz-Terskey Suture Zone (Arenigian ophiolitic mélanges and olistostromes, Lomize et al., 1997, and Arenigian maximum protolith rock age, thrusting and granitic intrusions, Bazhenov et al., 2003; Alexyutin et al., 2005), in the Dzhalair-Naiman Suture Zone (ca. 490 Ma maximum age of detrital zircons within deformed turbidites, Kröner et al., 2008; ca. 470 Ma metamorphism, Kröner et al., 2012; Arenigian fresh overlap assemblages, Alexeiev et al., 2010; and ca. 470 syn-collisional adakitic diorite, Qian et al., 2009), and in the

Urumbai Suture Zone (Middle Cambrian to Arenigian cherty-terrigenous sediments thrusted with coeval volcaniclastics, Degtyarev and Ryazantsev, 2007). The formation time of the Arkalyk Suture Zone is uncertain, but it may be attributed to the Arenigian because of the Early Ordovician maximum age of cherts and basalts, and the occurrence of Amgan limestone lenses within flysch sediments (Yakubchuk and Degtyarev, 1994; Degtyarev and Ryazantsev, 2007). This timing is supported by Arenigian deformation, granitic intrusions and fresh overlap sediments, identified in the Chingiz Range (Yakubchuk, 1990; Yakubchuk and Degtyarev, 1994; Collins et al., 2003; Tolmacheva et al., 2008) (Fig. 7).

The Arenigian accretion/collision event seems to have systematically affected the Amgan-Arenigian marine sediments. Following this event, the arc activity continued and lasted till the Latest Ordovician-Early Silurian. Middle-Late Ordovician arc-related rocks occur in the Stepnyak Massif (Kröner et al., 2008; Degtyarev et al., 2008), Kyrgyz Range (Konopelko et al., 2008), and in the Chingiz and Boshchekul Ranges (Yakubchuk, 1990; Yakubchuk and Degtyarev, 1994). The Erementau Zone, which extends from northern Kazakhstan (eastern Selety Zone) to western Balkash (Yili-Erementau Zone) is well accepted as the accretionary wedge of the Middle-Late Ordovician continental arc (Yakubchuk, 1990; Windley et al., 2007; Alexeiev et al., 2010). The Arenigian event was also recorded in the accretionary wedge, but it does not mark the end of the accretion (ophiolitic mélanges are unconformably overlain by Middle-Late Ordovician flysch and olistostromes interpreted as forearc deposits, Alexyutin et al., 2005 and references therein) (Fig. 7).

The Hirnantian-Rhuddanian event has been interpreted as the principal formation time of the Kazakhstan Continent caused by collision between the Aktau-Junggar and Arenigian-formed terranes following closure of the Erementau-Yili Ocean (Yakubchuk, 1990; Filippova et al., 2001; Bykadorov et al., 2003; Kröner et al., 2008; Windley et al., 2007; Alexeiev et al., 2010; Kheraskova et al., 2010; Alexeiev, 2011). This collisional event is recognized in the Stepnyak area (folds intruded by ca. 450-440 Ma granitic rocks, Kröner et al., 2008), in the Kendyktas Range (folds intruded by ca. 466-438 Ma granitic rocks, Alexyutin et al., 2005), in the Kyrgyz and Terskey Ranges (folds and granitoids, Lomize et al., 1997), in the Aktau-Junggar Microcontinent (ca. 450-440 Ma granitoids, Degtyarev et al., 2006), in the Central Chingiz and Boshchekul Ranges (end of arc activity, red bed overlap and granitic intrusions, Collins et al., 2003), in the Baidaulet and Akbastau Ranges (end of arc magmatism,Yakubchuk, 1990) and possibly in the Chinese Borohoro Range (basal conglomerate, Gao et al., 1998) and western Junggar (Late Ordovician magmatic rocks unconformably overlain by Lower Silurian slope deposits, Buckman and Aitchison, 2004). The formation time of the Maikain-Kyzyltas Suture is well constrained as Latest Ordovician-Early Silurian, and the suture likely contains relics of a main ocean (Arenigian island arc, Yakubchuk, 1990) (Fig. 7). The Erementau-Yili accretionary wedge, which also shows cessation of sedimentation at this time (Alexeiev et al., 2010) (Fig. 7), probably borders the same ocean.

Present-day knowledge suggests that at least four major tectonic events contributed to the Early Palaeozoic continental growth of Kazakhstan. Degtyarev and Ryazantsev (2007) proposed that the Amgan unconformity is related to the Terreneuvian island arc obduction. Although we do not contest this model here, we propose an alternative according to which the Amgan and Terreneuvian events were different. As suggested byYakubchuk and Degtyarev (1994) and Tolmacheva et al. (2008), the Amgan unconformity may also reflect the opening of a back-arc basin. The recognition of these different events throughout Kazakhstan leads to some potential correlations (Fig. 7). Although the Tar-Muromtsev Zone is poorly known (Sengör and Natal'in, 1996), we assume it was equivalent to the Boschekul-Chingiz arc of Windley et al. (2007). Considering the close affinities between the Chu-Yili and Issyk-Kul microcontinents (and assuming the absence of extensive nappes), the possibility that they belonged to a single terrane would imply major tectonic duplication (Fig. 2). This may be explained either by complex oblique collision or by later oroclinal bending. Important strike-slip faults were probably associated with the oroclinal bending process (Sengör and Natal'in, 1996; Johnston, 2004).

4.2. The probable Gondwanan origin of the Kazakhstan microcontinents

Two different Vendian origins have been proposed for the Kazakhstan continental fragments. (1) Baltica-Siberia (Sengör and Natal'in, 1996; Yakubchuk, 2004) and (2) East Gondwana (Mossakovsky et al., 1994; Kheraskova et al., 2003; Windley et al., 2007; Biske and Seltmann, 2010). The location of the Kazakhstan fragments in the equatorial zone of eastern Gondwana is currently strongly favoured by their marked lithostratigraphic affinities (Kheraskova et al., 2003; Degtyarev and Ryazantsev, 2007; Windley et al., 2007; Biske and Seltmann, 2010), by palaeomagnetic data (Bazhenov et al., 2003; Collins et al., 2003; Alexyutin et al., 2005, Levashova et al., 2011), by coeval orogenic events (see below), and by the East Gondwanan affinity of Early Palaeozoic Kazakhstan faunas (Popov et al., 2009).

In the Vendian and Early Palaeozoic East Gondwana (in the equatorial zone) was probably partly formed by South China (Yangtze craton), and the Lesser Himalaya and Iran (Burrett et al., 1990; Nie, 1991; Van der Voo, 1993; Cocks and Torsvik, 2002; Jiang et al., 2003; Torsvik and Cocks, 2009; Stampfli et al., 2011) (Fig. 14). The lithostratigraphic affinities of Vendian-Late Ordovician passive margin sediments (Vendian rift sediments associated with glacial deposits followed by shallow-marine terrigenous carbonates with phosphorites) between Lesser Himalaya, South China (Jiang et al., 2003) and the Ishim-Naryn Zone of Kazakhstan (Degtyarev and Ryazantsev, 2007) suggest that the western Kazakhstan microcontinents (Chatkal-Karatau, Tourgai, Naryn, West Chu Sarysu and West Teniz in Fig. 2, Windley et al., 2007) were included in East Gondwana (Fig. 8). This is supported by the Latest Ordovician final amalgamation age of the Kazakhstan microcontinents (Kröner et al., 2008; Windley et al., 2007; Alexeiev, 2011) and by a coeval orogenic event that affected South China (Chen et al., 1997; Su et al., 2009; Charvet et al., 2010).

As indicated by the palaeotectonic models of Kazakhstan by Degtyarev and Ryazantsev (2007) and of South China by Charvet et al. (2010), the Vendian-Ordovician rocks in South China-Ishim-Naryn were probably not deposited on a "passive margin", but on the margin of a deep marginal continental basin. Other Kazakhstan microcontinents (Issyk-Kul and Chu-Yili) were also situated in the equatorial zone of East Gondwana according to palaeobiogeographic and palaeomagnetic data (Bazhenov et al., 2003; Alexyutin et al., 2005; Popov et al., 2009), and they may have formed on the conjugate margins of Gondwana (Degtyarev and Ryazantsev, 2007). This model is also compatible with latest Ordovician deformation that affected the external parts of Kazakhstan (Stepnyak, Kröner et al., 2008; and Kyrgyz-Terskey, Lomize et al., 1997): the collision of the Aktau-Junggar Microcontinent with the internal margin of Kazakhstan likely caused the inversion of the South China-Ishim-Naryn basin in the Latest Ordovician. The origin of the Aktau-Junggar Microcontinent and its potentially associated Baidaulet-Abkbastau arc (Fig. 7), separated from the rest of Kazakhstan by the main Yili-Maikain-Kyzyltas Ocean (Early Cambrian OIB, Middle-Late Ordovician island arc relics; Yakubchuk, 1990), remains enigmatic. Based on these considerations and on the hypothesis outlined in sections 4.1 and 4.2, we propose (see Figure 8) a tentative and preliminary palaeotectonic configuration for the Early Palaeozoic continental growth of Kazakhstan. Numerous pluridisiplinary studies are required in Kazakhstan in order to reach a viable understanding of its apparently complex Early Palaeozoic tectonic evolution.

5. The northern Tarim-North China margin during the Early and Middle Palaeozoic

From their relevant palaeotectonic-palaeogeographic models the North China and Tarim blocks were considered either as a single continent (Mossakovsky et al., 1994; Heubeck, 2001; Metcalfe, 2006; Yakubchuk, 2008) or two different continents (Zonenshain et al., 1990; Cocks and Torsvik, 2002; Fortey and Cocks, 2003; Torsvik and Cocks, 2004). Palaeozoic palaeobiogeographic, palaeomagnetic and palaeoclimatic data from both blocks all indicate low palaeolatitudes (Burrett et al., 1990; Nie, 1991; Zhao et al., 1993, 1996; Chen et al., 1999; Huang et al., 1999, 2000; Yang et al., 2002; Fortey and Cocks, 2003), and thus they do not contest a common evolution. The attribution of the North China and Tarim cratons to a single continent during the Paleozoic requires, not a similar, but a correlative tectonic evolution of their different margins. The Palaeozoic tectonic environments recognized in the Kunlun, Qilian and Qinling show strong affinities, which lead to a coherent plate tectonic evolution (Bian et al., 2001; Stampfli et al., 2011, and our observations). The tectonic evolution of the northern Tarim margin was subject to hot debate in the last decades (ss. 6.2). Although different issues still need to be resolved, the present state of knowledge indicates marked affinities between the Tarim and North China margins (ss. 6.1), which in turn suggests a common palaeotectonic setting (ss. 6.3.).

5.1. The North China margin: the onset of an Ordovician active margin

It is well accepted that the North China margin was passive in the Cambrian-Early Ordovician and became active in the Middle-Late Ordovician (Wang, Q. and Liu, 1986; Hsü et al., 1991a; Yue et al., 2001; Xiao et al., 2003; Miao et al., 2007b). In Inner Mongolia, the Bainaimiao and Ondor Sum zones contain a Middle Ordovician to Middle Silurian continental arc and an associated accretionary wedge on the margin of the North China Craton (Wang; Q. and Liu, 1986; Shao, 1989; Hsü et al., 1991a; Yue et al., 2001; Xiao et al., 2003) (Fig. 2). The southdirected subduction-accretion complex contains Early Ordovician (ca. 490-470 Ma) MORB and IAT ophiolitic rocks (the Tulinkai Ophiolite at Ondor Sum, Xiao et al., 2003; Miao et al., 2007b; Jian et al., 2008). According to geochemical data (SSZ and boninitic trondhjemite), the Tulinkai Ophiolite is a relic of the basement of an Ordovician intra-oceanic island arc (Jian et al. 2008). In the Late Ordovician a metamorphic-magmatic event affected the Ondor Sum complex (ca. 450 Ma blueschist matrix, ca. 451-436 amphibolitic sole of the Tulinkai Ophiolite and ca. 460-450 Ma addakite, de Jong et al., 2006; Miao et al., 2007b; Jian et al., 2008) and arc-related rocks in the Bater area (ca. 450-440 Ma adakite and ca. 450 Ma diorite, Jian et al. 2008). The adakitic magmatism and amphibolite sole strongly imply ridge subduction at ca. 450 Ma (Jian et al. 2008). Another tectonic event affected the northern China margin in the Late Silurian. The Ordovician-Silurian Bainaimiao arc and Ondor Sum accretionary complex are unconformably overlain by shallow marine clastics (Wang, Q. and Liu, 1986; Shao, 1989; Hsü et al., 1991a; Yue et al., 2001; Xiao et al., 2003), and the Bater arc was intruded by ca. 419-415 Ma low-K tonalities (Jian et al. 2008). These events caused the end of arc activity in the Bainaimiao and Ondor Sum Zones (Fig. 10). However, Devonian-Permian arc-related intrusions in the northern margin of the North China Craton (Inner Mongolia Paleo-Uplift, Zhang, S. H. et al., 2006, 2007a, 2007b, 2009a, 2009b) and ca. 400-360 Ma magmatic zircons in Upper Carboniferous-Permian clastics (at Daqinshan; Cope et al., 2005) indicate that subduction was only temporarily interrupted.

According to Hsü et al. (1991a) and Yue et al., (2001), the Late Silurian event may have resulted from the collision between the Sunid (or Xilinhot) Microcontinent (i.e. Baolidao Zone) and North China. However, the Permian-age Solonker Suture Zone, well accepted as the main suture between North China and Siberia (Wang, Q. and Liu, 1986; Sengör and Natal'in, 1996; Xiao et al., 2003; Li, 2006; Miao et al., 2007a, 2007b), and this challenges such a palaeotectonic model (Fig. 2). The Sunid Microcontinent was most likely detached from the Siberian margin by back-arc opening in the Late Palaeozoic (Miao et al., 2007a). Jian et al. (2008) proposed that the

Tulinkai island arc was accreted to a microcontinent located in the Bater area in the Late Silurian (Precambrian xenocrysts in ca. 419-415 Ma tonalities). Nevertheless, the existence of a microcontinent in the area is uncertain, and according to most interpretations the Bater Precambrian rocks more likely belong to the North China Craton. The island arc may have been already accreted to North China in the Middle-Late Ordovician (ca. 460 Ma) prior to the subduction of the ridge at ca. 450 Ma. The ridge-subduction and arc-accretion processes were most probably mutually related (supra-subduction ridge associated with the ca. 480-460 Tulinkai island arc). This palaeotectonic model, consistent with the important geochemichal/geochronological data presented by Jian et al. (2008), explains why the passive margin was transformed into an active margin (Fig. 12D). Subduction inception in a passive margin setting is unlikely (there are no modern analogues); the most likely process for such a transition is the accretion/collision of a terrane obducted onto the passive margin followed by subduction reversal (Cloetingh et al., 1984; Stern, 2004). Xiao et al. (2003) already proposed such a scenario for the onset of an active margin in Inner Mongolia. The end of the Sino-Korean platform development in the Middle-Late Ordovician (Meng et al., 1997; Meng and Ge, 2003) also supports the onset of an active margin at this time. Based on the above considerations, neither the Tulinkai arc nor the Sunid Microcontinent can be related to Late Silurian collision; the wider plate tectonic context suggests the involvement of the Kazakhstan Continent (Fig. 11, ss. 6.2.1).

5.2. The Tarim margin: two passive margin stages?

The tectonic evolution of the northern margin of Tarim during the Palaeozoic is subject to hot debate (Wang, Q. et al., 2010). The Tarim margin has commonly been considered to be a continuous passive margin from the Sinian through the Late Palaeozoic that was fringed by the South Tianshan Ocean (Windley et al., 1990; Gao et al., 1998; Chen et al., 1999; Carroll et al., 2001; Xiao et al., 2004b, 2009a). However, other authors have proposed that the Tarim margin was temporarily active during the Ordovician before Silurian back-arc opening of the South Tianshan Ocean and detachment of the Central Tianshan Microcontinent (Shu et al., 2002, 2004; Charvet et al., 2007; Wang, B. et al., 2008; Lin et al., 2009). This key controversy about the Altaid evolution mainly arises from the spatial correlations of the sutures in the Chinese Tianshan (ss. 5.2.1), Beishan (ss. 5.2.2), and Kyrgyz Tianshan (ss. 5.2.4) (see; ss. 5.2.3-5.2.5). The widely investigated Chinese Tianshan is a good point of departure for discussing the potential extension of the sutures.

5.2.1. The Chinese Tianshan

The back-arc model for the Chinese Tianshan is well constrained by stratigraphic, structural, palaeontologic, metamorphic, magmatic, geochemical and geochronological data as well as by kinematic constraints (Shu et al., 1999; Laurent-Charvet et al., 2002, 2003; Shu et al., 2002, 2004; Charvet et al., 2007, 2011; Wang, B. et al., 2008; Lin et al., 2009). These authors defined the Central Tianshan Microcontinent according to the position of the South Tianshan Suture, which extends along the Tarim margin, and to that of the Central Tianshan Suture that is located in the Nalati and North Tianshan Faults (Fig. 9). The South Tianshan Suture separates the Tarim Craton and the Central Tianshan Microcontinent (Yili or Yili-Central Tianshan; Windley et al., 1990; Gao et al., 1998; Chen et al., 1999; Carroll et al., 2001; Zhou et al., 2001; Shu et al., 2002; Xiao et al., 2004b, 2011; Charvet et al., 2007, 2011; Wang, B. et al., 2008; Gao et al., 2009; Lin et al., 2009). The suture is defined in the western to eastern Tianshan by numerous ophiolitic mélanges that contain HP blueschists, Ordovician-Silurian limestones, Silurian to Lower-Middle Devonian cherts, and Middle Silurian to Late Devonian oceanic crust (ca. 425-380 isotopic ages) in a Late Devonian to Early Carboniferous matrix (the Heiyinshan, Kule, Yushugou, Tonghuashan-Liuhuangshan and Kawabulak ophiolites or ophiolitic mélanges; Chen et al., 1999; Shu et al., 2002; Xiao et al., 2004b, 2008; Charvet et al., 2007). The Central Tianshan Suture was not always recognized to separate the Central Tianshan and Yili zones (Fig. 9); for example, the two zones were considered as a single microcontinent (Gao et al., 1998; Chen et al., 1999; Zhou et al., 2001). However, it is commonly accepted today that the Nalati Fault is a suture between the Yili and Central Tianshan microcontinents, because of their different Early-Middle Palaeozoic geology (Wang, B. et al., 2008; Gao et al., 2009; Qian et al., 2009). The Central Tianshan Suture is especially well documented along the North Tianshan Fault by the Gangou-Mishigou ophiolitic mélange that contains Ordovician through Silurian (ca. 470-420 Ma) oceanic relics (Charvet et al., 2007). The Aqqiikkudug-Weiya Fault was interpreted as its eastward extension, because of similar Ordovician ophiolitic mélanges with HP relics in the Central Tianshan and Xingxingxia microcontinents (Shu et al., 2002; Xiao et al., 2004b; Charvet et al., 2007). The Central Tianshan Microcontinent (Central Tianshan and Xingxingxia zones, Fig. 9) is characterized by a Precambrian basement and a Middle Ordovician-Devonian arc (Gao et al., 1998; Chen et al., 1999; Shu et al., 2002; Xiao et al., 2004b, 2011; Charvet et al., 2007). The opening of the South Tianshan Ocean as a back-arc basin in the Early Silurian is consistent with the occurrence of post-Ordovician-Early Carboniferous oceanic relics within the well-documented South Tianshan Suture Zone, and of an Ordovician-Silurian continental arc in the Central Tianshan Microcontinent, and with the existence of an older Central Tianshan Ocean to the north (Fig. 10). Other ideas, such as Silurian syn-rift deposition, south-dipping subduction of the Central Tianshan Ocean, and the closure time of oceans were also proposed within the back-arc model. Confirmation of this model has to be found in its relevance to other key regions in a larger plate tectonic context (Beishan, Kyrgyz Tianshan, Tarim).

5.2.2. Beishan

In Beishan the Hanshan Microcontinent, which is located between the Hongshishan and Xiaohuangshan suture zones (Fig. 9), is characterized by Precambrian basement, a Cambrian to Early Ordovician passive margin and a Middle Ordovician to Late Silurian arc (Hsü et al., 1991b; Zuo et al., 1991; Yue et al., 2001; Xiao et al., 2010b). The northern boundary of the

Hanshan Microcontinent is not clear, but Zuo et al. (1991) defined its continental margin based on Ordovician turbidites located to the north of the Middle Ordovician-Silurian Yuanbaoshan continental arc (active margin of the Hanshan Microcontinent). The Hongshishan ophiolitic mélange has a Carboniferous age, but the time of formation of oceanic crust needs to be better constrained (Xiao et al., 2010b). There are Ordovician and Silurian clastics, chert, quartzite and pillow basalts metamorphosed in the greenschist facies in the Hashaat area (Badarch et al., 2002). The location of the Tsagaan Uul Precambrian Terrane to the south also supports the extension of the Hanshan Microcontinent in southern Mongolia. The geology of this microcontinent and the occurrence of Ordovician oceanic relics to the north strongly suggest the eastward extension of the Central Tianshan Microcontinent and of the Central Tianshan Ocean through Beishan as far as southern Mongolia (Fig. 9-10).

Based on these considerations, the Xiaohuangshan Zone located between the Hanshan Microcontinent and Tarim craton (Dunhuang) most probably represents the westward extension of the South Tianshan Suture. The Xiaohuangshan Zone is bound by two belts of ophiolites (Hongliuhe-Xichangjing and Xingxingxia-Shibanjing, Xiao et al., 2010b). Various ophiolitic mélanges, Middle-Late Ordovician-Silurian arc assemblages, Precambrian rocks, and a Cambrian-Ordovician passive margin are found in the Xiaohuangshan Zone (Zuo et al., 1991; Xiao et al., 2010b). Except for a ca. 426 Ma gabbro in the Hongliuhe ophiolitic mélange, Ordovician cherts occur in a complete ophiolite (Xichangjing), and Ordovician-Silurian turbidites are associated with the ophiolitic rocks in the Hongliuhe-Xichangjing belt (Hsü et al., 1991b; Zuo et al., 1991; Yue et al., 2001; Xiao et al., 2010b). There are Ordovician-Silurian fossils in highly deformed ophiolitic mélanges in the Xingxingxia-Shibanjing belt (Xiao et al., 2010b). As previously proposed, the Silurian oceanic crust in westernmost Hongliuhe is most probably a relict of the South Tianshan Ocean (Charvet et al., 2007; Xiao et al., 2008, 2010b) (Fig. 10). The correlation of the Hongliuhe ophiolitic mélange with the Ordovician Xichangjing oceanic relics, as well as the presence of the Ordovician Xingxingxia-Shibanjing ophiolitic belt

(Xiao et al., 2010b) that probably originated in one or two main oceans located to the south of the Hanshan Microcontinent (Hsü et al., 1991b; Zuo et al., 1991; Yue et al., 2001; Xiao et al., 2010b) challenge the Tianshan back-arc model in Beishan.

5.2.3. Problematic correlations between Beishan and its neighbouring regions

Although the Xiaohuangshan oceanic remnants were generally considered to be derived from the main Early Palaeozoic Ocean that separated the Hanshan and Tarim continental blocks, the opening of a Silurian back-arc ocean is suspected from other relevant geological data. A Sinian to Lower Ordovician passive margin in the Dunhuang, Xiaohuangshan and Hanshan zones (Zuo et al., 1991; Yue et al., 2001; Xiao et al., 2010b) is most probably a remnant of a single passive margin, which would imply a Tarim origin for the Hanshan Microcontinent. In the northern Dunhuang Zone (south of Hongliuhe), a Silurian olistostrome containing relics of a passive margin (Sinian red chert-bearing marble, Lower Cambrian black chert and sandstone, and Middle Cambrian limestone) was interpreted by Charvet et al. (2007) to have formed in an early rift of the South Tianshan Ocean. In the same region, Zuo et al. (1991) described a Silurian slope deposit, which unconformably covers a Cambrian basement, and suggested it formed in the early syn-rift of the Hongliuhe back-arc ocean. The only well-dated oceanic crust is Silurian in age. Finally, as discussed above, the closed affinities between the Central Tianshan and Hanshan microcontinents strongly support their common origin and evolution, and the application of the back-arc model in Beishan (Fig. 10).

The different tectonic units in Beishan may have been generated in two (Zuo et al., 1991; Yue et al., 2001) or multiple accretionary-collisional processes (Xiao et al., 2010b). In their review of Beishan geology the last authors presented an innovative palaeotectonic model for the whole Palaeozoic according to which another microcontinent (Shuangyinshan-Huaniushan) existed between the Hanshan and Dunhuang continental blocks, and is characterized by Precambrian to

Ordovician shelf carbonates and clastics as well as by slope deposits on which an Ordovician to Early Silurian arc developed (Xiao et al., 2010b). The Shuangyinshan-Huaniushan Microcontinent was defined by a Carboniferous-Permian ophiolitic mélange (i.e. Liuyuan) located on its southern side (Mao et al., 2012a). Because of the occurrence of ca. 460 Ma eclogites (Qu et al., 2011) and ca. 450-420Ma adakites and Nb-enriched basalts (Mao et al., 2011b), the Liuyuan mélange was interpreted as a relic of the Cambrian-Permian Paleo-Asian Ocean (Xiao et al., 2010b). Nevertheless, a possible common palaeogeographic origin and tectonic evolution for the Shuangyinshan-Huaniushan and Dunhuang (part of the Tarim block) continental blocks is suspected because of their close affinities with Inner Mongolian geology (a Cambrian-Ordovician passive margin followed by an Ordovician-Early Silurian active margin). This correlative hypothesis is supported by the geochemical similarities of the magmatic rocks (MORB, IAT, adakites) contained within the Liuyuan mélange (Mao et al., 2012a, 2012b) and the Ondor Sum mélange (Jian et al., 2008), as well as the Late Silurian unconformity in the Shuangyinshan-Huaniushan arc (Xiao et al., 2010b) and the Ondor Sum complex (Xiao et al., 2003). Thus the Dunhuang block may be the equivalent of the North China block, and the Shuangyinshan-Huaniushan arc of the continental Bainaimiao arc (Fig. 10). This would imply that the Early Palaeozoic Inner Mongolian scenario (ss. 5.1, Fig. 12D) could also be viable for the Beishan, and that the Ordovician ophiolites may have originated from an island arc obducted onto the Dunhuang-Shuangyinshan-Huaniushan passive margin. The Carboniferous-Permian rocks contained in the Liuyuan mélange (Mao et al., 2012a) may also contain the record of one or more later distinct events (e.g. continental rifting, collision, back-arc opening, and subduction-accretion (see discussion in Guo et al., 2012).

Detailed structural, stratigraphic, geochemical/geochronological studies are still required in Beishan in order to test and constrain the correlations between the different tectonic units. A common palaeogeographic origin for some units is suspected by their common affinities (see review of Xiao et al., 2010b). Nappe emplacement (klippen and window) and distinctive weathering in the Chinese Tianshan (Charvet et al., 2007) and Kyrgyz Tianshan (Biske and Seltmann, 2010) may also explain some tectonic duplications in Beishan; and thus a simpler model than the one presented by Xiao et al. (2010b) may also be viable. According to available data the interpretation of the Xiaohuangshan Suture Zone as a relict of a single Silurian-Devonian Ocean is unlikely because, according to their current location, the Ordovician oceanic remnants were interpreted as derived from an ocean located to the south of the Hanshan Microcontinent. Nevertheless, northward obduction of the Tarim margin (Dunhuang-Shuangyinshan-Huaniushan-Hanshan) by an Ordovician island arc developed in a northern oceanic domain (Central Tianshan-Hashaat) remain possible and should be tested (Fig. 12D-C). Because of relevant close affinities with and between surrounding regions (Chinese Tianshan, Tarim and Inner Mongolia, Fig. 10) and of consistency within a larger plate tectonic framework (ss. 5.3), we consider that the second back-arc model is the more likely and applicable to Beishan, as illustrated in our plate tectonic reconstructions (see model on Fig. 11-12-13-14).

5.2.4. The Kyrgyz Tianshan

In the Kyrgyz Tianshan, the Turkestan Suture Zone is well accepted as the principal suture between the Tarim and Kazakhstan continents (Biske, 1995; Sengör and Natal'in, 1996; Burtman, 2006, 2008; Windley et al., 2007; Pickering et al., 2008; Biske and Seltmann, 2010). The suture zone is mainly represented by southward allochtons of oceanic origin in the Bukuntau-Kokshaal belt (Turkestan-Alay and Kokshaal Zones, Fig. 9). The uppermost units consist of Early Ordovician ophiolites (e.g. Sartale) associated with clastic breccias that underlie clastic and volcanic rocks metamorphosed in the greenschist facies with HP-UHP relics (Biske and Seltmann, 2010). Other oceanic remains in the upper allochtons, but not connected with the Ordovician ophiolites, consist of basalts of probable seamount origin associated with Upper Silurian-Devonian open-sea sediments (Pickering et al., 2008; Biske and Seltmann, 2010). The lower allochtons contain Silurian to Moscovian pelagic sediments, Silurian to Devonian shallow-marine carbonates and Middle-Devonian to Visean turbidites that are interpreted as the passive margin of the Tarim-Alay continent (Burtman, 2008; Biske and Seltmann, 2010). With detailed stratigraphic and structural data, the last authors described and correlated the continuity of the lower allochtons from Uzbekistan eastwards as far as the Chinese Tianshan. The opening of the Turkestan Ocean was dated as Sinian by rift and passive margin sediments (Biske and Shilov, 1998; Carroll et al., 2001).

5.2.5. Problematic correlations between the Chinese Tianshan and Kyrgyz Tianshan

The correlations between the Chinese and Kyrgyz sutures are uncertain because of the westward thinning and/or disappearance of the Chinese Central Tianshan Microcontinent in the Far East Chinese Tianshan (e.g. Biske and Seltmann, 2010; Charvet et al., 2011) (Fig. 9). Some authors proposed that the Central Tianshan Microcontinent extends in the Naryn Zone (i.e. Middle Tianshan; Gao et al., 2009; Qian et al., 2009), but this correlation was negated by Biske and Seltmann (2010), because of their different geology; the Naryn Zone belongs to the Kazakhstan Domain (Windley et al., 2007; Biske and Seltmann, 2010) (Fig. 9). We should expect that the extension of the Central Tianshan-Hanshan active margin or equivalent arc should be in the Kyrgyz Tianshan. The major HP-LT belt that extends from the Western Chinese Tianshan (Changawuzi) through the Kyrgyz Tianshan (Atbashi and Alay) as far as Uzbekistan (Biske and Seltmann, 2010; Hegner et al., 2010; Gao et al., 2011) may represent the relics of this plate boundary. The belt is composed of blueschist-, eclogite- and greenschist-facies metasedimentary rocks and mafic metavolcanic rocks (with N-MORB, E-MORB, OIB and arc basalt affinities, Gao and Klemd 2003). Ordovician oceanic crust seems to be commonly associated with the HP rocks (Alekseev et al., 2007; Biske and Seltmann, 2010). The HP-LT belt is generally considered to belong to the Turkestan-South Tianshan Suture (Gao et al., 1998; Chen et al., 1999; Zhou et al., 2001; Biske and Seltmann, 2010; Gao et al., 2011). However, some authors have suggested that it has affinities with the Central Tianshan Suture Zone (e.g.

lithostratigraphy, structural data, age of oceanic crust, Charvet et al., 2007, 2010; Wang, B. et al., 2008; Lin et al., 2009).

The double Chinese affinities of the Turkestan Suture would imply that the Turkestan Suture Zone contains relics of both the South and Central Tianshan Oceans (Fig. 10). Although the Ordovician-Silurian arc was not recognized in Kyrgyz Tianshan, the opening of a South Tianshan back-arc type of ocean within the Tarim margin as far as Kyrgyzstan is possible; this arc may have disappeared by subduction processes. The Silurian opening of an ocean in the Kyrgyz Tianshan is supported by the Tarim stratigraphic record. A major Late Ordovician hiatus and Early Silurian unconformity occurs in the Tarim passive margin (Kalpin and Bachu Uplifts; Carroll et al., 2001). The Tarim basin was affected by rapid subsidence during the Early Silurian, which decreased from the Middle Silurian through the Devonian (Carroll et al., 2001). Such a subsidence trend is characteristic of an isostatic adjustment of the lithosphere, which undergoes mechanical stretching (i.e. tectonic unloading) and successive thermal subsidence (i.e. reequibration of the lithosphere-asthenosphere and sediment loading) (Heidlauf et al., 1986; Bott, 1992; Ziegler and Cloetingh, 2004). A period of rifting in the Silurian is supported by the occurrence of syn-rift deposits (coarse pebble conglomerates, breccias and felsic igneous rocks alternating with banded limestones) within the Tarim sediments of the Kokshaal Range (Biske and Shilov, 1998).

Reconciling the available data with a consistent palaeotectonic model, we propose that the Tarim margin underwent two stages of passive margin subsidence: (1) in the Sinian-Middle Ordovician during opening of the Turkestan1-Central Tianshan Ocean; (2) in the Silurian-Carboniferous associated with opening of the Turkestan 2-South Tianshan Ocean (Fig. 10, 12A-B).

5.3. The northern margins of Tarim and North China: a correlative tectonic evolution

During the Cambrian-Ordovician, the Tarim-North China continent was likely isolated and fringed by a passive margin on its northern side (Stampfli et al., 2011). Relics of these passive margin sediments occur in the Kyrgyz Tianshan, Tarim, Beishan and Inner Mongolia (Fig. 10). The passive margin changed to an active margin after the probable accretion of the Tulinkai island arc in the Middle Ordovician (Fig. 12D). This accretionary event was not recognized along the Tarim margin, but is suspected in the Beishan according to the close affinities between the Dunhuang and North China blocks, between the Shuangyinshan-Huaniushan and Bainaimiao continental arcs, as well as between the Ondor Sum and Liuyuan ophiolitic mélanges (Fig. 10). Thus, the Ordovician-Silurian active margin of North China most probably extended during the coeval southward subduction of the Early Palaeozoic Central Tianshan Ocean under the Central Tianshan-Hanshan Microcontinent, and probably it continued as far as Uzbekistan (Fig. 11). After the Silurian, the Tarim-North China active margin had a distinctive and coherent tectonic evolution: a back-arc basin most probably opened along the Tarim segment in the Early Silurian (Fig. 12A-B-C) when the Inner Mongolian segment was affected by collision/accretion at the end of the Early Silurian (Fig. 12D), which was likely related to the motion of the Kazakhstan Continent (Fig. 11, see details in ss. 6.2). Following the Silurian event, magmatic activity in the North Chinese margin was probably only temporarily interrupted and it resumed in the Devonian (Fig. 12D). The onset of the active margin and subsequent Silurian back-arc opening is recorded within the stratigraphic sequence of the Sino-Korean and Tarim platforms (i.e.a Late Ordovician hiatus and a Silurian unconformity, Carroll et al., 2001; Meng and Ge, 2003). The back-arc opening allowed marine incursions into the Tarim basin (Early Silurian marine clastics, Carroll et al., 2001), while the Sino-Korean platform apparently remained emergent during most of the Palaeozoic (a major hiatus, Meng and Ge, 2003).

6. Middle and Late Palaeozoic interactions between Kazakhstania, Siberia and Tarim-North China and an alternative plate tectonic model

It is well accepted that the Kazakhstan superterrane was located between Tarim and Siberia during the Middle and Late Palaeozoic and that it was affected by oroclinal bending (Sengör and Natal'in, 1996; Filippova et al., 2001; Levashova et al., 2003; Yakubchuk, 2004; Abrajevitch et al., 2007, 2008; Levashova et al., 2007, 2009; Windley et al., 2007; Xiao et al., 2010a). The oroclinal bending of Kazakhstan is demonstrated by the oroclinal shape of tectonic units (Windley et al., 2007) and by palaeomagnetic data, which constrain the main rotations from the Late Devonian through the Carboniferous (Abrajevitch et al., 2007, 2008; Levashova et al., 2009). The pre-Silurian amalgamation of the Kazakhstan microcontinents in East Gondwana (ss. 4.1-2) implies that the Kazakhstan continent moved westward toward its present location within the global context constrained by palaeomagnetic data (i.e. Kazakhstania, Abrajevitch et al., 2008; Baltica, Cocks and Torsvik, 2005; Siberia, Cocks and Torsvik, 2007; Tarim, Van der Voo, 1993, Fig. 14). As previously proposed (Sengör and Natal'in, 1996; Abrajevitch et al., 2008; Xiao et al., 2010a), the oroclinal bending of Kazakhstan was evidently related to the relative motions of Baltica, Siberia and Tarim. We demonstrate below that the Middle-Late Palaeozoic motion and bending of the Kazakhstan Superterrane (ss. 6.1) is supported and refined by diachronous tectonic events identified along the Tarim-North China (6.2) and Siberian (ss. 6.3) margins. The progressive closure of the surrounding oceans finally led to the final amalgamation of Kazakhstania in the Permian (6.4).

6.1. The Kazakhstan Continent and its potential extensions

Following its amalgamation in the Latest Ordovician-Early Silurian, the Kazakhstan Continent was mostly emergent during the Silurian and Early Devonian, as demonstrated by continental deposits and subaerial lavas (Daukeev et al., 2002). Early Silurian to Middle

Devonian and Carboniferous arc assemblages were recognized on its western and southwestern margins in Kazakhstan, Uzbekistan, Kyrgyzstan and China (Sengör and Natal'in, 1996; Filippova et al., 2001; Daukeev et al., 2002; Bykadorov et al., 2003; Windley et al., 2007; Gao et al., 2009; Biske and Seltmann, 2010; Alexeiev, 2011). After the Middle-Late Devonian, the sea invaded the western part of Kazakhstan and a terrigenous-carbonate shelf started to develop in the Tourgai, Chatkal-Karatau and Naryn zones, and epi-continental shallow marine basins developed in the Teniz and Chu-Sarysu (Fig. 9). The marine sediments evolved into continental deposits within both basins in the Middle-Late Carboniferous, but the Tourgai, Chatkal-Karatau and Naryn zones remained a continental shelf until the Latest Carboniferous (Daukeev et al., 2002; Windley et al., 2007). The Givetian to Bashkirian marine sediments were interpreted as a passive margin (Windley et al., 2007; Biske and Seltmann, 2010; Alexeiev, 2011). This succession of tectonic environments suggests that the passive margin formed by back-arc opening. As partly shown in some palaeotectonic models (Filippova et al., 2001; Bykadorov et al., 2003), no oceanic crust seems to have formed in the back-arc basin; the related continental arc (the Valerianov and South Tianshan arcs in the Valerianov and Chatkal-Karatau zones on Fig. 9) most probably remained attached to the Kazakhstan continent. The Middle-Late Palaeozoic oroclinal bending and the westward motion of the continent require an external as well as an internal subduction zone.

The Kazakhstan Superterrane was fringed by an internal active margin from the Late Silurian to the Late Carboniferous (Filippova et al., 2001; Bykadorov et al., 2003; Windley et al., 2007; Alexeiev, 2011). Arc assemblages are present all along the Kazakhstan orocline (the Chu-Yili, Aktau-Junggar, East Teniz, Baidaulet-Akbastau, Zhaman-Sarysu and North-Balkash zones, Fig. 9). The importance of accretionary processes is demonstrated by the seaward progradation onto the Zhaman-Sarysu accretionary wedge (Windley et al., 2007). The Ordovician-Devonian Zhaman-Sarysu and Devonian-Carboniferous Junggar-Balkash subduction-accretion zones most likely formed by the multiple accretion of terranes, but their nature and plate tectonic evolution still remain enigmatic (Fig. 9).

The southern branch of the Kazakhstan continent most likely extends into the Turpan-Bogda and Atasbogd zones where there are Devonian-Carboniferous arc assemblages (Lamb and Badarch, 2000, 2001; Badarch et al., 2002; Xiao et al., 2004b, 2010b; Naumova et al., 2006; Charvet et al., 2007) (Fig. 9). The existence of a Precambrian continental basement under the Turpan Basin is subject to controversy (Xiao et al., 2004b; Charvet et al., 2007) and has been only assumed in southern Mongolia (Kröner et al., 2010). Early Palaeozoic rocks are poorly known, but some relics of Ordovician-Silurian volcano-sedimentary arc assemblages occur on the southern edge of the Turpan Basin (the Dananhu arc) and in Beishan (the Queershan arc), which have been interpreted as island arc systems (Xiao et al., 2004b, 2010b). However, these arcs assemblages have also been attributed to the southern Central Tianshan (Xingxingxia Zone) or Hanshan microcontinents by e.g. Zuo et al. (1991), Shu et al. (2002) and Charvet et al. (2007). The enigmatic nature and age of Middle-Late Palaeozoic arc basement suggest, as pointed out by Xiao et al. (2004b), that the Turpan-Bogda Zone may have formed mainly, or maybe only, in the Devonian-Carboniferous by the accretion of island arcs. The Kazakhstan active margin (mainly East-Aktau-Junggar, Chu-Yili and Central Tianshan zones) may have extended in the Chinese Harlik-Dananhu island arc system of Xiao et al. (2004a), see the Turpan-Bogda Zone on Fig. 9.

The Devonian and Carboniferous continental arc of the Kazakhstan northern branch is recognized in the East Baidaulet-Akbastau, North Balkash and East Boshchekul-Chingiz zones (Windley et al., 2007). The Zharma-Saur Zone, interpreted as a Devonian-Carboniferous arc (Windley et al., 2007), may be part of this arc system or another accreted island arc. The subduction zone likely extended in northeastern Junggar where there are coeval arc assemblages (Badarch et al., 2002; Xiao et al., 2004a, 2008, 2009b). This correlation is strongly supported by the northern main Chara-Erqis Suture Zone located to the north (Buslov et al., 2004a; Xiao et al., 2009b). The northeastern Junggar arcs (Junggar-Yamaquan and Dulate-Baytag zones in Fig. 9) may belong to a single intra-oceanic arc system (see Xiao et al., 2009b for details). The existence of Precambrian basement under the Junggar Basin is still not proven, but the juvenile nature of the basement is suggested by geochemical signatures (see discussion in Xiao et al., 2008). Interesting palaeotectonic models showing the continental growth of the Junggar region by multiple accretion of island arcs were presented by Xiao et al. (2008, 2009b). Although the connections between the different islands arcs within the Junggar-Balkash Ocean remain unresolved, an archipelago-type palaeotectonic environment is widely favoured today (see Xiao et al., 2010a and references therein).

6.2. Insights from the Tarim-North China margin

The westward motion of the Kazakhstan continent is recorded in the tectonic evolution of the Tarim-North China margin. The interactions between both continents likely started in the Late Silurian in Inner Mongolia (ss. 6.2.1), evolved during Devonian-Carboniferous time from Beishan (ss. 6.2.2) through the Tianshan (6.2.4-6.2.5) and ended by their final amalgamation in the Late Carboniferous in Uzbekistan (the Ural-Tianshan formation, Alexeiev et al., 2009). There are important current issues on the final time of formation and location of the sutures in Beishan (6.2.3) and Tianshan (ss. 6.2.6).

6.2.1. Inner Mongolia

We have seen in section 5.1 that a collisional/accretionary event temporarily interrupted the development of the Bainaimiao-Ondor Sum active margin in the Late Silurian-Early Devonian (Fig. 10-12D). Spatial and kinetic constraints from a global plate tectonic perspective suggest the proximity of Kazakhstan and North China during the Silurian (Fig. 13). We propose that the cessation of activity in the Ondor Sum accretionary wedge was caused by the oblique collision of Kazakhstan with North China (Fig. 11-12D).

6.2.2. Beishan

A Late Silurian-Early Devonian accretionary-collisional event is recognized in the Beishan (Hsü et al., 1991b; Yue et al., 2001, Zuo, 1991; Xiao et al., 2010b). This Orogeny has generally been attributed to the final closure of the Xiaohuangshan Ocean located between the Hanshan Microcontinent and the Tarim Block (Hsü et al., 1991b; Zuo et al., 1991; Yue et al., 2001). According to Zuo et al. (1991), the Dunhuang margin and the Hanshan Microcontinent were deformed, emerged, intruded and surrounded by foreland basins in the Early Devonian (Fig. 9). However, Xiao et al. (2010b) recently interpreted this event as the collision between two composite arcs (the Shuangyinshan-Huaniushan and Mazongshan-Hanshan arcs) following the closure of the Hongliuhe-Xichangjing Ocean, and they proposed that the "Hanshan" composite terrane was only accreted to Tarim in the Permian (the Permo-Carboniferous Liuyuan mélange, Mao et al., 2012a). The northern Hongshishan Suture (Fig. 9) was considered to be the relict of a main ocean that existed during the whole Palaeozoic, because of the occurrence of Ordovician turbidites to the north of the Hanshan Microcontinent (Zuo et al., 1991) and because of the Lower Carboniferous Hongshishan ophiolitic mélanges (Yue et al., 2001; Xiao et al., 2010b) (ss. 5.2.2-3, Fig. 10). Xiao et al. (2010b) proposed that the Hongshishan Suture formed in the Permian by the collision between the Heiyingshan-Hanshan and Queershan island arcs (the Hanshan and Atasbogd zones on Fig. 9). Conversely, Zuo et al. (1991) and Yue et al. (2001) argued that the Ordovician-Silurian Hanshan continental arc (their Yuanbaoshan arc) ended by the early Devonian, when it was overlapped by molasse sediments.

6.2.3. Location and timing of the final suture in Beishan

The highly conflicting observations and interpretations of many authors summarized above show that much still needs to be learnt in Beishan before an agreed viable model can be constructed. Nevertheless, some considerations can be addressed, which aim to reconcile the available data within a consistent larger plate tectonic framework.

No Devonian-Carboniferous oceanic crust or deep marine deposits have been identified to the south of the Hongshishan Suture; Devonian and Carboniferous rocks mainly consist of arc assemblages and related shallow marine sediments (Hsü et al., 1991b; Zuo et al., 1991; Yue et al., 2001; Xiao et al., 2010b). This arc may correspond to the Dunhuang-Hanshan active margin formed by Late Silurian-Early Devonian subduction of the Hongshishan Ocean and the Hongshishan ophiolitic mélanges the associated accretionary wedge (Fig. 12C). Coeval arc assemblages occur in the Atasbogd Zone located to the north of the Hongshishan Suture (Lamb and Badarch, 2000; Badarch et al., 2002; Naumova et al., 2006; Xiao et al., 2010b) (Fig. 9). This Zone is poorly documented (Lamb and Badarch, 2000; Badarch et al., 2002; Naumova et al., 2006), but was most probably formed by three different accretion-collision events from the Devonian to the Permian: (1) in the Latest Silurian Hanshan (basal conglomerate, Zuo et al., 1991; Yue et al., 2001); (2) in the Carboniferous (Early?) Hongshishan (emplacement of ophiolites, Xiao et al., 2010b); and (3) the Late Carboniferous Altan Uul event (southward ophiolite emplacement, Rippington et al., 2008). This demonstrates the northward continental growth of the Tarim margin in Beishan and southern Mongolia.

Although, Xiao et al. (2010b) consider that an oceanic domain existed between the Dunhuang and Hanshan blocks during the whole Palaeozoic, we follow here the interpretation of Zuo et al. (1991) who argued in favour of a Beishan Orogeny and the disappearance of all oceans in Beishan by the Early Devonian. We assume that no ocean south of the Hanshan Microcontinent remainded until the Permian (see ss. 5.2.3-5.3, Fig. 12C). The oblique collision between the Kazakhstan continent and the Tarim margin in the Late Silurian (Fig. 11) may have caused the activity cessation of the Hanshan arc (i.e. unconformity), the closure of the Hongliuhe Ocean (i.e. ridge failure, ca. 420 Ma Hongliuhe Ophiolite) and the amalgamation of the Hanshan

Microcontinent to Tarim (i.e. Beishan Orogeny). Following the amalgamation, the Xiaohuangshan and Hanshan zones then became the site of a new Devonian continental arc system (i.e. arc and back-arc) formed by the southward subduction of the Hongshishan Ocean (Fig. 12C).

6.2.4. The Chinese Tianshan

In the eastern Chinese Tian Shan, there was pre-Visean accretion implying that the South Tianshan and Central Tianshan sutures can be recognized in the so-called Eo-Tianshan Range (Carroll et al., 1995; Charvet et al., 2007) (Fig. 9). Major deformation, characterized by northward emplacement of ophiolitic mélanges, folding and thrusting, predated unconformable Visean deposits. Some ophiolitic mélanges in the South Tianshan Suture Zone contain radiolaria as young as Tournaisian and thus the suture likely formed in the Tournaisian-Visean. The formation age of the Central Tianshan Suture is constrained as Middle Devonian according to the Devonian matrix in the Gangou-Mishigou ophiolitic mélange and to Middle Devonian intrusions (see details in Charvet et al., 2007). In the easternmost Chinese Tianshan (at Weiya), Latest Silurian-Early Devonian ductile deformation is recorded in ophiolitic mélanges in the Central Tianshan Suture (Shu et al., 1999, 2002, 2004). This is consistent with the coeval time of orogeny in Beishan.

Various geological data imply that the South and Central Tianshan sutures formed before the Visean in eastern Tianshan; see details in Charvet et al. (2007), who proposed that the collision between the Yili (Kazakhstan, ss. 6.1) and Central Tianshan terranes (formation of the Central Tianshan Suture) in the Devonian caused the closure of the South Tianshan Ocean. This well-documented palaeotectonic model is mainly followed here (Fig. 12B). The location and formation of sutures in the western Chinese Tianshan are less certain and thus have been subject to various interpretations; see discussion in section 6.2.6.

6.2.5. *The Kyrgyz Tianshan*

In the Kyrgyz Tianshan, the final closure of the Turkestan Ocean that was responsible for the collision between the Tarim and Kazakhstan continents was in the Middle-Late Carboniferous according to detailed stratigraphic and structural data (Biske, 1995; Biske and Shilov, 1998; Burtman, 2006, 2008; Biske and Seltmann, 2010) (Fig. 12A). These authors described lower nappes composed of passive margin rocks from the Tarim and upper nappes of oceanic crust and accretionary prism rocks that originated in the Kazakhstan margin. The continental subduction of the Tarim passive margin (flysch transgression on the Tarim carbonate platform) and formation of the South Tianshan nappes are documented from Uzbekistan to the Chinese Tianshan (Turkestan-Alay and South Tianshan Zones on Fig. 9, Burtman, 2008; Biske and Seltmann, 2010). This is consistent with ca. 320-300Ma eclogites from the Atbashi Ridge (Hegner et al., 2010) and ca. 300-320 Ma post-collisional magmatism (Konopelko et al., 2007; Biske and Seltmann, 2010; Seltmann et al., 2011 and references therein). The Carboniferous age of the Turkestan Suture in Kyrgyzstan is a well-established time in the Altaids (Fig. 12A).

6.2.6. *Location and timing of final sutures in the Tianshan*

As previously pointed out by Wang, Q. et al. (2010) and discussed in section 5.2.5, the westward extension of the Central and South Tianshan sutures in the Kyrgyz Tianshan is subject to much controversy, which has led to different palaeotectonic models that aim to explain the formation of the Tianshan Mountains (Gao et al., 1998; Chen et al., 1999; Carroll et al., 2001; Zhou et al., 2001; Xiao et al., 2004b, 2009a, 2012; Charvet et al., 2007; Wang, B. et al., 2008; Gao et al., 2009; Lin et al., 2009; Biske and Seltmann, 2010). The main controversies related to the tectonic evolution of the Tianshan concern the subduction vergence of the South Tianshan Ocean and the final closure time of the ocean (see discussions in Xiao et al., 2009a; Wang, Q. et al., 2010; Han et al., 2011).

In the westernmost Chinese Tianshan, peak HP metamorphism took place at ca. 350-340 Ma, related to major Tournaisian-Visean subduction (Gao and Klemd, 2003). However, the Tianshan HP-LT rocks also contain several metamorphic ages (Early Devonian, Early Carboniferous, Middle-Late Carboniferous and Triassic ages), which have led to different suture times (Gao et al., 1995, 2011; Zhang, L. et al., 2002, 2003, 2005, 2007; Gao and Klemd, 2003; Lin and Enami, 2006; Lin et al., 2009; Wang, Q. et al., 2010, Gao et al., 2011). Most authors do not agree on the origin or provenance of the HP rocks, which have been attributed to the Central (Wang, B. et al., 2008; Lin et al., 2009) or South Chinese oceans (Gao et al., 1998; Chen et al., 1999; Zhou et al., 2001; Gao and Klemd, 2003). As proposed by Lin et al. (2009), the rocks most likely record different stages in exhumation. The Central Tianshan Microcontinent, which separates both Chinese sutures, becomes thinner and disappears westwards (Biske and Seltmann, 2010). This may have enabled the HP rocks to record both collisional events. Moreover, the various metamorphic ages recognized along the belt may also reflect the diachronous oceanic closure along the Tarim margin (see below). Nevertheless, other geological information allows us to constrain the time of formation of sutures in the western Tianshan. Han et al. (2011) in their important review of western Tianshan geology, confirmed the Late Carboniferous collision between the Tarim and Kazakhstan-Yili terranes (Fig. 12A).

The occurrence of a carbonate platform as young as the Kungurian associated with basaltic flows in the northern Tarim Basin (at Kunkelaqi; Chen and Shi, 2003) has led to a controversy about the final closure time of the ocean between Tarim and Kazakhstan (see discussions in Xiao et al., 2009a; Biske and Seltmann, 2010). Xiao et al. (2009a), who interpreted the Permian carbonate platform as a passive margin, suggested that another ocean should have existed to the south of the Turkestan-South Tianshan Ocean (the North Tarim Ocean). In a different way, Burtman (2008) interpreted post-Moscovian shallow marine deposits in the Kyrgyz Tianshan and northern Tarim as a syn-collisional basin (the Turkestan Sea), which followed the closure of the Turkestan Ocean (a soft collisional process). Xiao et al. (2009a) also interpreted Permian calkalkaline rocks in the Kuluketaq massif (the northern margin of Tarim) as a continental arc (see references and discussion in Xiao et al., 2009a). However, Pirajno et al. (2008) suggested there was a mantle plume in the Tarim region during the Permian. This interpretation was recently supported by detailed geochemical data from basaltic flows associated with continental clastics in the Kalpin area (Yu et al. 2011). Early Permian extension is also suggested by rapid subsidence the affected the Tarim Basin at this time (Carroll et al., 2001). A major Permian mantle plume does not really negate the coeval existence of an ocean to the north. However, further field and geochemical/geochronological investigations are required in the Kuluketaq area to corroborate the existence of another ocean. If it existed, this ocean may have extended into Beishan (Liuyuan, Xiao et al., 2010b) and Solonker (Balengshan, Xiao et al., 2009a), where there are Permian gabbros. Although well accepted in Inner Mongolia, the occurrence of a Permian ocean in northern Tarim and Beishan remains poorly constrained at the present time; this palaeotectonic model was recently strongly contested by Han et al. (2011). We have not included this ocean in our plate tectonic model in Fig. 17-18), but its potential existence is one of the most interesting issues of the Altaids.

Most geoscientists working in the Tianshan consider that the South Tianshan Ocean was subducted northwards under Kazakhstan (Windley et al., 1990; Chen et al., 1999; Carroll et al., 2001; Zhou et al., 2001; Xiao et al., 2004b; Biske and Seltmann, 2010). This idea is mainly based on the Tarim stratigraphy interpreted as a passive margin (Biske and Shilov, 1998; Carroll et al., 2001), However, according to structural observations, other authors have concluded that the South Tianshan nappes were emplaced northwards onto the Central Tianshan Microcontinent, interpreted as a result of southward subduction of the South Tianshan Ocean (Shu et al., 2002; Charvet et al., 2007; Wang, B. et al., 2008; Lin et al., 2009). Although we do not challenge these structural observations, this interpretation is contradicted by other structural and stratigraphic relations in the Kyrgyz Tianshan. In the Kokshaal Range, Biske and Shilov (1998) documented a Tarim passive margin and its northward continental subduction (flysch transgression on a Kasimovian carbonate platform). To reconcile these observations, Charvet et al. (2011) proposed that the Tarim margin was active in China and passive in Kyrgyzstan, and that the southward Chinese subduction was linked to the northward Kazakhstan subduction by a transform-type plate boundary. However, Burtman (2008) reported the southward emplacement of oceanic allochtons onto the Tarim margin as far as the Chinese Tianshan. A simpler palaeotectonic model that shows the Tarim blocks and associated microcontinents (Alay Microcontinent, Burtman, 2008; Biske and Seltmann, 2010) fringed by a passive margin all along their northern side (i.e. from Beishan to Uzbekistan), seems today the most likely model that respects the available data and plate tectonic configurations (Fig. 11-12-14-15-16). The northward thrusting identified in the Chinese Tianshan may reflect back-thrusting as demonstrated in the Kyrgyz Tianshan (Burtman, 2008; Biske and Seltmann, 2010).

The oblique collision between the Tarim and Yili-Central Tianshan microcontinents, as proposed by e.g. (Carroll et al., 2001, Chen et al., 1999; Zhou et al., 2001; Han et al., 2011), fits well within a larger plate tectonic framework (Fig. 12-14-15-16). The pre-Visean orogenic event is well constrained in the eastern Chinese Tianshan, implying closure of the Central Tianshan Ocean in the Devonian and of the South Tianshan Ocean in the Early Carboniferous (Chen et al., 1999; Charvet et al., 2007). A Visean-Early Carboniferous unconformity is also recorded in the stratigraphic record of the Tarim Basin and Kokshaal Range (Carroll et al., 1995, 2001; Biske and Shilov, 1998), but in the Kyrgyz Tianshan and farther west, the South Tianshan Ocean remained open throughout the Carboniferous (a passive margin, Burtman, 2006; Biske and Seltmann, 2010). This demonstrates that the Visean eastern Tianshan orogenic event induced flexural loading as far as Kyrgyzstan and formed a flexural fore-bulge all along the Tarim margin (Carroll et al., 2001) (Fig. 12A-B). The extension of the Visean orogeny in western Tianshan is not clear, and is still more confusing in westernmost Chinese Tianshan, where Early or Late Carboniferous ages have been proposed (Gao et al., 1998, 2009; Chen et al., 1999; Zhou et al., 2001; Wang, B. et al., 2008). Nevertheless, Han et al. (2011) recently clarified many issues in their review of western Chinese Tianshan geology, which strongly argues in favour of a Late Carboniferous age. They also corroborated the westward diachronous closure of the South Tianshan Ocean. It is important to note that the South Tianshan Suture Zone was later affected by dextral strike-slip faulting in the Permian, which locally modified the original post-collisional framework (Laurent-Charvet et al., 2003). The diachronous model for the Tianshan is consistent with Late Silurian–Early Devonian tectonism that affected the North China margin in Inner Mongolia and the coeval Beishan Orogeny (Fig. 12). Moreover, it is compatiable with a larger plate tectonic context of the Altaids, accounting for the westward motion of the Kazakhstan continent (Fig. 14-15-16-17).

6.3. Insights from the peri-Siberian margin

The interaction between the Kazakhstan and Siberian continents most likely started in the Late Devonian. This is supported by accretionary-collisional events, back-arc closures and strike-slip deformation documented along the peri-Siberian margin and more precisely in Inner Mongolia, western Mongolia, the Chinese Altai and Siberian Altai (ss. 6.3.1). The final amalgamation between the two continents took place in the Carboniferous-Early Permian in the Siberian and Chinese Altai (ss. 6.3.2). Figure 13 summarizes the major tectonic events that affected the Siberian margin.

6.3.1. Middle-Late Devonian interactions

We discussed in section 3.6 the fact that an oceanic back-arc basin existed during the Middle Devonian in the Chinese Altai (the Kuerti back-arc basin) and that it likely extended to western Mongolia (Tseel Zone). Following the opening of the back-arc basin in the beginning of the Middle Devonian, the Altai-Mongolian Microcontinent was fringed by a passive margin (clastics and cherts, Windley et al., 2002) (Fig. 6). Unconformable Carboniferous volcaniclastic metasediments mark the return of volcanic activity in the area at this time. Wang, T. et al. (2006)

emphasized the syn-collisional characteristics of ca. 375 Ma granitoids and attributed them to the closure of the Kuerti back-arc basin. This is consistent with the occurrence of ca. 370-360 Ma syn-tectonic granitoids (Kröner et al., 2007) and ca. 390-380 Ma high-grade metamorphism (Kozakov et al., 2007a) in the Tseel complex (Fig. 13).

The Kuerti Ocean most probably extended northwards in the Siberian continental back-arc basin (Fig. 9), the existence of which is documented by Middle-Late seaward displacement of the arc (from the Charysh-Terekta to Rudny-Altai zones) and by back-arc rocks in the Rudny-Altai (Yolkin et al., 1994) (Fig. 13). The closure time of this ocean is not clear, probably because of important strike-slip deformation that affected the back-arc region in the Late Devonian-Early Carboniferous (Charysh-Terektas Shear Zone). Coeval strike-slip deformation took place on the Kuznetz-Telesk Fault, which extends along the Uimen-Lebed Zone (Buslov et al., 2004b). Middle-Late Devonian syn-collisional granitoids in the Gorny-Altai predate strike-slip deformation and mark the beginning of oblique interaction between Siberia and the Altai-Mongolian Microcontinent (Glorie et al., 2011).

In southwestern Mongolia, Kröner et al. (2010) recently recorded a major Late Devonian-Early Carboniferous tectonic event with zircon ages from flysch deposits, deep crustal growth of gneisses and granitic domes, and intramontane basins that unconformably overlie Early Palaeozoic rocks in the Lake and Gobi-Altai Zones (Fig. 9). This event was corroborated by Lehmann et al. (2010) who demonstrated from structural observations the syn-convergent emplacement of an arc system during E-W shortening in the Gobi-Altai Zone, as well as E-W folding and Silurian ophiolite emplacement in the Trans-Altai Zone (the Mongolian part of the Dulate-Baytag Zone in Fig. 9). The Late Devonian age of arc assemblages in the Lake, Gobi-Altai, Mandalovoo and Trans-Altai Zones implies the amalgamation of these different zones by this time (Lehmann et al., 2010). This tectonic event was considered to reflect the accretion of an island arc (their Trans-Altai Zone) to the Siberian margin (in the Lake, West Gobi and West

Mandalovoo Zones) (Kröner et al., 2010; Lehmann et al., 2010). This intra-oceanic arc was likely accreted in the Gurvansayhan-Zoolen accretionary wedge in which there are intra-oceanic relics (Lamb and Badarch, 2001; Helo et al., 2006) (Fig. 13).

The above information, added to the ca. 380 Ma HP blueschists and unconformity in the Sunidzuoqi subduction-accretion complex in Inner Mongolia (Xu et al., 2001), demonstrate that the peri-Siberian margin was affected by a major accretionary-collisional event in the Middle-Late Devonian (Fig. 13), which also affected more internal areas as at Uimen-Lebed (the dextral Kuznetz-Telesk Fault, Buslov et al., 2004b) and also the Bayanhongor Zone (Devonian thrust stacking, Osozawa et al., 2008). From palaeomagnetic data Abrajevitch et al. (2008) showed that the northern tail of the Kazakhstan Terrane was probably located close to the Siberian margin in the Late Devonian. Buslov et al. (2000, 2004a, 2004b) already demonstrated Late Devonian interaction between the Altai-Mongolian Microcontinent, the Kazakhstan continent and the Siberian margin (i.e. formation of the Charysh-Terekta Shear Zone). We propose in our model that the Mongol-Okhotsk ridge temporarily joined the Kuerti ridge (Fig. 14). The Kazakhstan active margin may have extended to an island arc that was accreted in the Middle-Late Devonian Ma in Inner, southern and southwestern Mongolia. The oblique collision between the Kazakhstan continent and Siberia probably caused the closure of the Kuerti back-arc basin and the northward motion of the Altai-Mongolian Microcontinent along the Siberian margin (Fig. 15-16).

6.3.2. Late Carboniferous-Early Permian interactions

The oblique collision between Kazakhstan and Siberia prevented their amalgamation in the Late Devonian-Early Carboniferous; their final amalgamation took place along the Chara sinistral shear zone in the Late Carboniferous-Early Permian (the Chara, Irtysh, northeastern Bashchelkak and Kuznetz-Telesk Faults, Buslov et al., 2000, 2004a b). The offset of the sinistral

Erqis Shear Zone, which is the southward extension of the main Chara fault zone, was estimated to be 1000 km (Sengör et al., 1993; Laurent-Charvet et al., 2002, 2003). Strike-slip deformation in the Erqis Zone took place in the Early Permian according to isotopic data from metamorphic and magmatic rocks, but sinistral/dextral motion and northwestward folding continued until the Late Permian (Laurent-Charvet et al., 2002, 2003; Briggs et al., 2007). In the Siberian Altai, ophiolitic mélanges in the Chara Zone are sealed by Middle-Late Carboniferous volcaniclastics and Late Carboniferous molasse deposits (Buslov et al., 2004a, 2004b). In the Chinese Altai, accretionary processes in the Erqis Zone and arc activity in Altai-Mongolia continued at least until the Latest Carboniferous (Xiao et al., 2009b; Wan et al., 2011). The end of arc activity was likely caused by the oblique accretion of the northeastern Junggar arc system (the Dulate-Baytag, Junggar-Yamaquan zones in Fig. 9) that is partly considered here as a possible extension of the Kazakhstan Continent (ss. 6.1 and Fig. 16-17). The sinistral motion of the Chara-Erqis Shear Zone is consistent with the probable southeastward diachronous oceanic closures along the Siberian margin (Fig. 13) and with the larger plate tectonic context (ss. 6.5).

6.4. The oroclinal bending of Kazakhstan

During the Middle-Late Palaeozoic the Altaid plate tectonic framework was dominated by oroclinal bending and large-scale rotations (see Xiao et al., 2010a). The timing of oroclinal bending in Kazakhstan is constrained by palaeomagnetic data (Levashova et al., 2003, 2007; Abrajevitch et al., 2007, 2008), and is strongly supported by the geological record along the Siberian (6.3) and Tarim-North China margins (6.2). The latter observations added to the present juxtaposition of the Kazakhstan terranes and their potential extensions in China (6.1) lead us to propose an alternative plate tectonic model for this time-period.

Following its amalgamation in eastern Gondwana, the Kazakhstan Continent was detached from the Gondwanan margin (Fig. 8-14) and moved along the North China margin (Fig. 11).

Since the Early Devonian, the southern branch of Kazakhstan was probably attached to the Tarim-North China continent in the Beishan area (Beishan Orogeny, Fig. 12C); thus a single subduction zone likely extended along the northern side of North China and the eastern side of Kazakhstan (Fig. 15). In the Middle-Late Devonian, the northern Kazakhstan branch started to collide with the Siberian margin (Fig. 16): the island arc tail of the Kazakhstan Continent accreted in Inner and southwestern Mongolia, and consequently oblique collision started between Kazakhstan and the Siberian continents in the Chinese and Siberian Altai. This Middle-Late Devonian plate tectonic framework favoured and initiated or enabled the oroclinal bending and major rotation processes.

The Early Palaeozoic Chingiz and Akbastau arcs, accreted to Kazakhstan microcontinents by the end of the Ordovician (see 4.1 and Fig. 2), are today partly juxtaposed against the Late Palaeozoic Chara Suture Zone (Buslov et al., 2004a; Windley et al., 2007) (Fig. 9). This implies that the northern branch of the Kazakhstan continent (North Balkash, East Baidaulet-Abkastau, Boshchekul-Chingiz and maybe the Zharma-Saur Zone on Fig. 9, ss. 6.1) were separated from the main Kazakhstan continent between the Silurian and Carboniferous. A possible and viable palaeotectonic model that respects the available data would consider the northern branch of Kazakhstan as detached by oblique back-arc opening in the Middle-Late Devonian (i.e. the beginning of major clockwise rotation, Abrajevitch et al., 2008); this would be similar to the present-day plate tectonic situation of the North Fiji Basin (Johnston, 2004). In the Altaids, the opening of the back-arc basin may have been triggered by the Kazakhstan-Siberia oblique interaction (Fig. 16). Assuming this palaeotectonic model to be correct, only the continental southwestern branch of Kazakhstan would have bent in the orocline; accordingly, the tectonic evolution of the northern branch can be explained by large-scale rotation (Fig. 17).

6.5. The final formation of the Altaids in the Permian: the Junggar-Balkash and Solonker sutures

The timing of interactions between Siberia, Kazakhstania and Tarim-North China and their evolution within a global plate tectonic framework is constrained by the final formation of the main Altaid suture zones (Fig. 9): Uralian (not discussed here, Puchkov, 1997), Turkestan (ss. 6.2.5), South Tianshan (ss. 6.2.4-6), Chara (ss. 6.3.2), Junggar-Balkash, and Solonker (see below).

Such a huge volume of crust was formed by multiple subduction-accretion processes in the Junggar-Balkash region (ss. 6.1) that a suture trace becomes difficult to define, so the whole region (i.e. Zhaman-Sarysu, Junggar-Balkash, Junggar, Junggar-Yamaquan, Dulate-Baytag, Turpan-Bogda and Atasbogd zones) might eventually be considered as a "suture zone" (Fig. 9). Nevertheless, we assume that the "main" internal subduction of the Kazakhstan orocline during the Carboniferous is represented by the Junggar-Balkash Suture Zone in Kazakhstan. The southern branch extended to the North Tianshan Suture in China (Charvet et al., 2007) and the Altan-Uul Suture in Mongolia (Rippington et al., 2008), and the northern branch to the Karamai (Buckman and Aitchison, 2004) and Kelameili (Xiao et al., 2009b) sutures in China (Fig. 9). These correlations are proposed according to the similar vergence and age of their accretionary belts (see details in the latter works). The final closure of the internal Junggar-Balkash Ocean was determined by field geology in Chinese Tianshan, western and northeastern Junggar.

In the northern Chinese Tianshan, Middle Permian molasse deposits and ca. 280-265 Ma granitic intrusions post-date north-vergent thrusting and folding of Carboniferous arc-related rocks (Charvet et al., 2007). In western Junggar, Late Carboniferous accretionary rocks are sealed by Lower-Middle Permian volcanic, volcaniclastic and clastic rocks interpreted as the initial in-fill of the Junggar Basin (Buckman and Aitchison, 2004). This idea was recently supported by Chen et al. (2010) who underlined that the western Junggar accretionary complexes were intruded by ca. 304-263Ma post-collisional plutons coeval with those emplaced in the Chara Suture Zone and Zharma-Saur arc. These field observations are consistent with the lithologic-palaeogeographic maps of Daukeev et al. (2002), which illustrate the passage from Late Carboniferous slope deposits to Early Permian coastal plain sediments in the Balkash and Junggar regions. After the latest Early Permian, the sediments were exclusively non-marine in the Junggar and Turpan Basins; a large lacustral plain developed in the Late Permian (Carroll et al., 1990; Wartes et al., 2002). Subsidence analysis shows that the Turpan Basin underwent extension in the Early Permian, while the Junggar Basin remained more or less stable (Carroll et al., 1990; Wartes et al., 2002). This extension was associated with transcurrent deformation and major strike-slip displacement along the North Tianshan Fault (Laurent-Charvet et al., 2002, 2003; Wang, B. et al., 2009). The Chinese Tianshan was extensively intruded by a wide variety of magmatic rocks (e.g. granitoids, mafics, ultramafics, adakites, shoshonites) during the Permian. Although several tectonic environments have been proposed to account for the magmatism (e.g. rifting, syn-post-collisional extension, transcurrent faulting), most geochemical signatures suggest a continental environment (Liu and Fei, 2006; Pirajno et al., 2008; Zhao et al., 2008; Wang, B. et al., 2009; Han et al., 2010; Shu et al., 2010; Chen et al., 2011). As pointed out by several authors (e.g. Carroll et al., 1995; Charvet et al., 2007; Han et al., 2011), the above relations strongly support the complete amalgamation of the Chinese Tianshan and Junggar region by the Latest Carboniferous-Early Permian.

From multidisciplinary data Xiao et al. (2009b) argued that the Jiangjun subduction-accretion complex in northeastern Junggar (Kelameili in Fig. 9), interpreted as the northern accretionary wedge of to the Dulate-Yamaquan island arc system, persisted during the Permian. This northeastern Junggar island arc system seems to have accreted to the Siberian active margin by the latest Carboniferous (the Erqis Suture, Wan et al., 2011) (ss. 6.3.2). The northeastern Junggar and Chinese Altai regions were finally juxtaposed in the Permian by sinistral motion in the

Chara-Erqis Zone (Wan et al., 2011). Following that accretion, northward subduction continued under the Siberian margin until at least the Early Permian (Xiao et al., 2009b; Wan et al., 2011). However, other authors considered that the Junggar region was already amalgamated in the Permian, because it was stitched by post-collisional plutons (see references and discussion in Han et al., 2011). According to Wan et al. (2011), the youngest Alaskan-type hydrous mafic-ultramafic complexes in northeastern Junggar are Early Permian in age. Some of these Alaskan-type complexes have been incorrectly, in our opinion, interpreted as a result of the Permian Tarim plume (Piajno et al., 2008); hydrous amphibole-bearing magmas are not an expected consequence of plume activity. The final suture of the Junggar-Balkash region is located under the Junggar Basin and thus makes the interpretation of timing difficult. Moreover, the region was traversed by and enveloped within prominent strike-slip faults (North Tianshan and Erqis faults, Laurent-Charvet et al., 2002). However, the above available data lead to different considerations that are mutually consistent in a larger plate tectonic context. Much of the Junggar region was probably amalgamated by the latest Carboniferous (i.e. northern Tianshan, western Junggar and Junggar Basin), but the Alaskan-type magmatism indicates that subduction persisted in the Early Permian under the new Siberian margin (northeastern Junggar). This suggests that the Junggar Ocean was progressively closed by oblique motion in latest Carboniferous to Early Permian (Fig. 18), and that the Junggar and Chinese Altai region was completely amalgamated by the end of the Permian when major strike-slip faulting ended.

During the Late Palaeozoic, the Junggar-Balkash Ocean extended to Inner Mongolia where there are numerous remnants of the Permian ocean (e.g. Hegenshan, Solon-Obo, Balengshan, Kedanshan, Banlashan ophiolites, Miao et al., 2007a, 2007b; Jian et al., 2008, 2010b; and references in Xiao et al., 2009a). Although authors have not always agreed on the exact time of suturing, it is now widely accepted that the main Solonker Ocean that separated Siberia and North China disappeared in the Permian (Wang, Q. and Liu, 1986; Hsü et al., 1991a; Sengör and Natal'in, 1996; Yue et al., 2001; Xiao et al., 2003; Li, 2006; Shen et al., 2006; Miao et al., 2007a, 2007b; Jian et al., 2010b). Nevertheless, Xiao et al. (2009a) proposed that accretionary processes lasted to the Early-Middle Triassic mainly because of the occurrence of Late Permian basic rocks (Banlashan and Balengshan). Jian et al. (2010b) recently presented new Carboniferous-Permian geochronological data and detailed geochemistry of magmatic rocks, which help to clarify the tectonic interpretation of the Solonker Zone. In their palaeotectonic model for the evolution of North China Jian et al. (2010b) accounted for the varied magmatic activity by different tectonic events and mechanisms: (1) ca. 294-280 Ma subduction initiation, (2) ca. 281-273 Ma ridge subduction, (3) ca. 271-260 Ma forearc-continent collision, and (4) ca. 255-248 Ma post-collisional slab break-off. Miao et al. (2007a) presented detailed geochronological and geochemical data for the Mongolian Altaids and suggested that a back-arc basin (Hegenshan) probably opened in the latest Carboniferous (Fig. 18) and closed coevally with the Solonker Ocean in the latest Permian. However, this ignores or negates the data in Hsü et al. (1991a) that the Hegenshan Ophiolite is now sitting on a thrust above Jurasic red beds, and thus the ophiolite cannot define a suture. Miao et al. (2007a) and Jian et al. (2010b) supported and refined the palaeotectonic model of Xiao et al. (2003) by interpreting the formation of the Solonker zone/suture by facing arc-arc collisions between Mongolia and North China (Fig. 18). In northeastern China, the Solonker Suture probably continues as the Jilin and Cheongjin sutures that formed in Late Permian-Early Triassic time (Jia et al., 2004; Lin et al., 2008)

The probable southeastward diachronous closure of the Junggar-Balkash-Solonker Ocean during the Late Carboniferous through Permian-Early Triassic is consistent with the Early Permian sinistral strike-slip movements on the Chara-Erqis and North Tianshan Faults (Laurent-Charvet et al., 2002), the Middle-Late Carboniferous Tien-Shan-Urals Orogeny (Alexeiev et al., 2009), and the northward diachronous closure of the Uralian-Khanty-Mansi Ocean through the Permian (Puchkov, 1997; Cocks and Torsvik, 2007) (Fig. 18). Important issues currently exist on the potential existence of a Permian-Triassic Ocean in Beishan (ss. 6.2.3) and in Tianshan (6.2.6), and so they were not considered in our models. With the present available knowledge we assume that only the Khanty-Mansi, Junggar-Solonker and Mongol-Okhotsk Oceans remained opened during the Permian. Following the amalgamation of Baltica-Siberia and North China-Mongolia, only the Mongol-Okhotsk Ocean remained open for most of the Mesozoic (Zorin, 1991). In the Far East of the Altaids, the Heilongjiang Ocean probably opened in the Permian and the Khanka-Jiamusi-Bureya Terrane formed; the latter was already re-accreted in the Triassic (Zhou et al., 2009).

The Middle-Late Palaeozoic reconstructions presented on Figures 14-18 are from a global 600-0 Ma plate tectonic model constructed at ca. 20-10 intervals (Neftex Petroleum Consultant Ltd.), which was refined by spatial and kinematic constraints by Wilhem (2010). These reconstructions only aim to explain the interactions between the three major continents (i.e. Siberia, Tarim-North China and Kazakhstan) that led to the formation of the Altaids of Central Asia. In reality, the tectonic framework of the involved oceans, and particularly the Junggar-Balkash Ocean, was probably more complex than presented here, and more like the present plate tectonic framework of the Circum-Pacific (see Xiao et al., 2008, 2009b, 2010a for the Altaids, and Hall, 2002 for the Circum-Pacific).

7. Final Discusssion and Conclusions

The Altaids was formed by multiple accretion/collision processes through the whole Palaeozoic within distinct Early Palaeozoic palaeogeographic domains, which mutually interacted during the Middle-Late Palaeozoic. The major terranes involved in the continental growth of the Altaids were island arcs, and microcontinents commonly with continental margin arcs. Present knowledge suggests that they mostly originated in three Vendian-Cambrian palaeogeographic domains: (1) peri-Siberia, (2) Eastern Gondwana (Kazakhstan continent) and (3) Tarim-North China.

The peri-Siberian Domain mainly grew during Vendian to Ordovician time by the formation of Siberian back-arc basins such as the Altai-Sayan and Barguzin back-arc terranes, by the accretion of microcontinents of Siberian (Tuva-Mongolian) and exotic (Altai-Mongolian) origin as well as by the accretion of island arcs formed within the palaeo-Asian Ocean (e.g. Lake-Khamsara and Uimen-Lebed island arcs). In addition to these major terranes, wide subduction-accretion complexes (i.e. Dzhida-Bayangol, Borus-Kurtushiba), in places with ocean plate stratigraphy, but often still poorly known, were created by accretion in trenches and contributed to significant crustal growth. Seamounts, oceanic islands and plateaus also widely participated in the growth of accretionary wedges in the Vendian–Early Palaeozoic (e.g. Gorny-Altai, Bayanhongor, Lake, Dzhida). Following the major Ordovician phase of continental growth, continental and oceanic back-arc basins developed all along the new Siberian margin. In the latest Ordovician-Early Silurian the triangular Mongol-Okhotsk Ocean opened.

South China and part of Kazakhstan (western microcontinents) were derived from Eastern Gondwana during the Vendian-Early Palaeozoic. Most of the other Kazakhstan microcontinents probably originated in Far Eastern Gondwana, were detached and finally re-accreted to it at the end of the Early Palaeozoic. Ribbon-shape terranes likely formed in two major periods of extension in the Vendian and Amgan. The Kazakhstan Continent mainly grew as a result of two major periods of collision: (1) the Arenigian and (2) the Hirnantian-Rhuddanian. Island arc accretion played a major contribution in the continental growth of Kazakhstan (i.e. Kokchetav, Baidaulet-Akbastau, and Boshchekul-Chingiz island arcs).

The Tarim-North Chinese Domain was likely isolated and close to Eastern Gondwana during the Vendian-Early Palaeozoic. An island arc was accreted to its northern margin in the Ordovician (i.e. Tulinkai) and most probably caused the onset of active margin growth (i.e. Bainaimiao arc and Ondor Sum accretionary wedge). In the Silurian, the Central Tianshan-

Hanshan Terrane probably formed by the back-arc opening of the South Tianshan-Turkestan 2 Ocean.

Since the Middle Palaeozoic the peri-Siberian, Tarim-North Chinese and Kazakhstan continents mutually interacted. This new plate tectonic arrangement led to the oroclinal bending and large-scale rotation of Kazakhstan during the Carboniferous and to the final amalgamation of the Altaids in the Permian.

Although available knowledge has provided interesting compilations and configurations for the general tectonic evolution of the Altaids, many multidisciplinary studies are still required to refine the scenario and to resolve the many controversies, which still persist in the scientific community. Some relevant polemic issues are presented below:

- The extension of the Tuva-Mongolian Ribbon-Microcontinent: Was the Khangai-Argunsky Microcontinent part of it or another exotic independent terrane? Does the Bayanhongor Zone represent the suture trace between the Tuva-Mongolian and Khangai-Argunsky terranes or just relics derived from the northern Dzhida-Bayangol Ocean?

- The extension of the Altai-Mongolian Microcontinent-type terrane: Did it form an exotic terrane with the Nuhetdavaa-Enshoo Microcontinent? Does the Vendian-Cambrian Lake-Khamsara island arc extend under the eastern Mongolian Middle-Late Palaeozoic cover, implying the exotic origin of the Nuhetdavaa-Enshoo Microcontinent?

- The possible accretion of the Altai-Mongolian Terrane to Siberia in the Ordovician needs to be tested by detailed field studies in southern Mongolia.

- The western extension of the Mongol-Okhotsk Ocean and suture: Did the Bayanhongor and Onon oceanic relics originate from a single or two oceans? At present we know that the

Bayanhongor Zone contains a 660-640 Ma ophiolite and a 298-210 Ma ophiolite), but their extent and occurrence are not known. It is probable that the Permo-Triassic ophiolite is far the biggest. The whole, beautifully exposed, Bayanhongor Zone needs to be remapped in order to work out the distribution, extent, and boundaries of the two ophiolites, new trace element and isotopic studies are required to better understand their ages, characteristics and differences, and the two generations of thrusts should be dated.

- What does the discovery, particularly by Alfred Kröner, mean that ostensibly juvenile Palaeozoic island arc rocks commonly contain Precambrian zircon xenocrysts? Does this mean these are all continental margin arcs? But geological relations often indicate an absence of associated Precambrian rocks. So how much of the Altaids is juvenile?

- The correlation of oceanic domains along the Tarim margin. Detailed stratigraphic and structural fieldwork should be undertaken in the Beishan in order to test and constrain the palaeogeographic origins of the different tectonic units, and thus establish their correlations with those in the Chinese Tianshan and Inner Mongolia. Although the potential Ordovician-Silurian continental arc, well identified in the Chinese Tianshan, it has not found in the Kyrgyz Tianshan; so, does the Turkestan Suture contain relics of this potential disappeared plate boundary, and thus the relics of both Chinese oceans? This would test the back-arc model all along the Tarim margin and constrain the Tarim-origin of the Central Tianshan-Hanshan Microcontinent.

- The location and timing of main sutures leading to the final amalgamation of the Altaids. The Junggar-Balkash-Solonker oceanic relics probably come from the last Palaeozoic ocean of the Altaids, but the precise time of its closure still needs to be clarified and refined (i.e. Carboniferous, Permian or Trias). The available data and interpretations from Xingjian and Inner Mongolia are in conflict, and so a detailed re-investigation is required, which should obtain a better understanding of the field relations and structural geology, which should form a basis for understanding the geochronology and metamorphism. A better palaeotectonic model showing a possible eastward diachronous closure of the ocean from Junggar-Balkash to Solonker and even the Pacific margin is required; and anyway was there a single simple ocean or small pockets of oceanc relicts?

- The potential existence of a Permian Ocean along the Tarim margin. This innovative idea was recently contested by some authors based on a detailed regional review of the Tianshan. However, recent data from Beishan (Permian gabbros and turbidites) and the presence of Alaskan-type hydrous mafic-ultramafic complexes rather than plutons derived from a mantle plume in Kuluketaq and Beishan suggest the existence of an ocean in the Permian. Is the Permian plume magmatism confined to the Tarim plateau basalts, and does not include intrusive plutons and complexes?

So far, all studies of the Altaids have been devoted to understanding what is there. But, in view of the recent realisation that in accretionary orogens as much crust has been lost by subduction as has been accreted (Stern and Scholl, 2010), and in the light of the recent realisation that in the arc-dominated orogen of the Japanese Islands at least one major island arc has gone, presumably subducted (Isozaki et al., 2010), should we also be considering how much or what parts of our island arcs are missing? Many so-called arcs in the Altaids are only a few hundred or a few kilometres thick, and yet modern extant island arcs in the Caribbean are of the order of 30 km thick. This knowledge would help understand how much the chemistry of the mantle has been modified by all the crustal additions.

We personally feel that far more fieldwork and far better understanding of structural relations are required throughout the Altaids, followed by geochemical and isotopic studies, in order to produce the next major advances in this incredible accretionary orogen.

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Figure captions

Figure 1: General map of the Altaids with regions and localities discussed in the text. The localities are relevant geological localities and may correspond to ranges, town or villages. The regions are for the Northern Altaids: (1) Northern Kazakhstan: ss. 4.1, 6.1, 6.4, (2) Altai-Sayan: ss. 3.1, 3.2.2, 3.4.2, 3.5.3, (3) Baikal: ss. 3.1, 3.2.1, (4) Siberian Altai: ss. 3.5.4, 3.6, 6.3, (5) Chinese Altai: ss. 3.3.2, 3.6, 6.3 and (6) Northern Mongolia: ss. 3.3.1, 3.5.2, 3.5.5; Southern Altaids: (7) Southern Kazakhstan: ss. 3.4.1, 3.5.1, 4.1, 6.1, (8) Junggar: ss. 6.5 , (9) Southern Mongolia: ss. 3.3.2, 3.6, 6.3, (10) Kyrgyz Tianshan: ss. 5.2.4-5, 6.2.5-6, (11) Chinese Tianshan: ss. 5.2.1-3, 6.2.4-6, (12) Beishan: ss. 5.2.2-3, 6.2.2-3 and (13) Inner Mongolia: ss. 3.3.2, 3.6, 5.1, 6.5.

Figure 2: Map showing the major terranes involved in the Early Palaeozoic formation of the Altaids; most zones (regular and bold text) were adapted from Sengör and Natal'in (1996), Badarch et al. (2002), Naumova et al. (2006) and Windley et al. (2007); boundaries of zones are precisely defined after Wilhem (2010). See names of white zones on Fig. 9. Capital letters: Major basins: A: West Siberian, B: Kuznetsk, C: Altai-Sayan, D: Khemchik-Sistigkhem, E: Huvsgol-Bokson, F: Upper Angaran; G: Songliao, H: Ordos, I: Tarim, J: Turfan, K: Junggar, L: Chu-Sarysu, M: Teniz. Black lines: Stratigraphic transects presented on Fig. 3 and 4. This figure is in part derivative from the Neftex Geodynamic Earth Model. © Neftex Petroleum Consultants Ltd.

Figure 3: Siberia-Chinese Altai stratigraphic transect (see section on Fig. 2). See legend on Fig. 4; data compiled from the works cited in the text. Time scale from Gradstein et al. (2004).

Figure 4: Siberia-Siberian Altai stratigraphic transect (see section on Fig. 2). Data compiled from the works cited in the text. Time scale from Gradstein et al. (2004).

Figure 5: A tentative tectonic sketch ofthe Vendian-Cambrian continental growth of Siberia (see Fig. 2 for zones and localities): a) Early Cambrian: two potential scenarios (1 and 2) for the formation of the Tuva-Mongolian Ribbon-microcontinent and the origin of the Bayanhongor Ocean; seamounts: KU: Kurai, BA: Bayanhongor, DZ: Dzhida, LA: Lake; b) Late Cambrian: Assuming scenario 1 (i.e. Khangai-Argunsky microcontinent as a part of the Tuva-Mongolian terrane, obduction of the Bayanhongor/Dzhida ophiolites on the microcontinent passive margin).

Figure 6: Correlations of the peri-Siberian active margin from the Late Ordovician through the Carboniferous. After the Ordovician accretion of the Altai-Mongolian microcontinent to Siberia (Fig. 5b), the peri-Siberian subduction zone prograded southward.

Figure 7: Potential correlations of major Early Palaeozoic Kazakhstan Terranes mainly based on the identification of the following tectonic events: (1) Terreneuvian island arc accretion; (2) Amgan back-arc opening; (3) Arenigian collision (i.e. back-arc closure); (4) Hirnantian-Rhuddanian collision (i.e. final formation of the Kazakhstan Continent). See legend on Figure 4; data compiled from works cited in text. Time scale from Gradstein et al. (2004).

Figure 8: A preliminary and tentative sketch of the Early Palaeozoic formation of the Kazakhstan Continent in Eastern Gondwana (see map on Fig. 2 for zones and localities); Major potential tectonic stages (Fig. 7): (1) Terreneuvian accretion of a Vendian-Early Cambrian island arc; (2) Amgan back-arc opening: formation of a new terrane; (3) Arenigian back-arc closure following ridge failure; (4) Hirnantian-Rhuddanian major collision: Accretion of the Aktau-Junggar Ribbon-microcontinent, and its possible Baidaulet-Akbastau continuation (i.e. it could also be independent in the Yili-Maikain-Kyzyltas Ocean) to the Kazakhstan continent, which in turn caused the inversion of the South China-Naryn intra-continental basin; (5) Future Early Silurian detachment of Kazakhstan from Gondwana. Some relevant localities: KU: Kumdykol,

ST: Stepnyak, U: Urumbai, AR: Arkalyk, AN: Anrakhai, DN: Dzhalair-Naiman, AK: Aktyuz, KT: Kyrgyz-Terskey.

Figure 9: Map showing the major terranes involved in the Middle-Late Palaeozoic formation of the Altaids; the zones (regular and bold text) are mainly after the maps of Sengör and Natal'in (1996), Badarch et al. (2002), Naumova et al. (2006) and Windley et al. (2007); precise zone boundaries are after Wilhem (2010). See name of zones forming the Kazakhstan, Peri-Siberian and Mongolian continents on Fig. 2. Capital letters: Some major basins (see legend of Fig. 2). This figure is in part derivative from the Neftex Geodynamic Earth Model. © Neftex Petroleum Consultants Ltd.

Figure 10: Diagram showing affinities between the Early-Middle Palaeozoic tectonic environments and events along the Tarim-North China margin: Environments and events related to: A) the Ondor Sum-Central Tianshan-Turkestan 1 Ocean, (B) the South Tianshan-Turkestan 2 Ocean. See Figure 12 for a potential palaeotectonic scenario.

Figure 11: Tectonic sketch of the Tarim-North China margin in the Late Silurian (Fig. 9)

Figure 12: Potential correlations along the Tarim-North China margin with some relevant geological data from papers cited in the text. Diachronous oceanic closure related to the westward motion of Kazakhstan.

Figure 13: Correlation of major tectonic environments and events affecting the Siberian margin (see also Fig. 6 for main compiled data and map on Fig. 9).

Figure 14: Early Silurian (442 Ma) global plate tectonic reconstruction (orthographic projection, fixed Africa). Some major tectonic events for the Altaids: 1) Rifting that gave rise to the Mongol-Okhotsk Ocean after the Ordovician “Tuva-Mongolian Orogeny” (Fig. 3 and 4); 2)

Rudny Altai back-arc basin (Fig. 4 and 12); 3) Strike-slip detachment of the Kazakhstan Continent from Gondwana following the final major Kazakhstan Orogeny (Fig. 8) and 4). Initial breakup of the South Tianshan Ocean (Fig. 12). Localities (zones on Fig. 9): Gondwana: Ca: Cathay, Ir: Iran and Ya: Yangtze; Hunia: Ku: Kunlun, Qa: Qaidam and Qi: Qinling; Kazakhstan: BA: Baidaulet-Abkastau, BC, Boshchekul-Chingiz, ChK: Chatkal-Karatau, CY: Chu-Yili and Tu: Tourgai; peri-Siberia: AM: Altai-Mongolia, Ba: Baolidao, GA: Gorny Altai, GuZo: Gurvansayhan-Zoolen, Ts: Tseel; North China-Tarim: Al: Alay, CT: Central Tianshan, Ha: Hanshan, On: Ondor Sum. See Stampfli et al. (2011) for information about the global tectonic framework (i.e. Hunia and Rheic Ocean). This figure is in part derivative from the Neftex Geodynamic Earth Model. © Neftex Petroleum Consultants Ltd.

Figure 15: Frasnian (382 Ma) plate tectonic reconstruction (Longitude/Latitude WGS 1984, fixed Europe, legend on Fig. 14). Tectonic events: 1) Mongolian-Inner Mongolian continental back-arc basin (Fig. 13); 2) accretion of an island arc (i.e. Sunidzuoqi 380 Ma blueschist facies, Gurvansayhan island arc relics, Fig. 13); 3) Kuerti back-arc basin (Fig. 13), 4) Siberian Altai continental back-arc basin (Fig. 13); 5) Kazakhstan continental back-arc basin (Chu-Sarysu and Teniz carbonate platform), 6) Island arc accretion inferred from the model (i.e. partial formation of the Zhaman-Sarysu subduction-accretion complexes), 7) inferred microcontinent formed by the opening of the South Tianshan-Turkestan 2 Ocean (Fig. 12A-B); 8) collision between the Central Tianshan microcontinent and Kazakhstan Continent and formation of the Central Tianshan Suture Zone (Fig. 12B) and 9): Beishan back-arc basin following the Late Silurian-Early Devonian Beishan Orogeny (i.e. the Late Silurian oblique collision between the Kazakhstan and Hanshan microcontinents caused the closure of the Hongliuhe Ocean and the accretion of the Hanshan microcontinent to Tarim, Fig. 12C). Localities (zones on Fig. 9): Kazakhstan: BC: Boshchekul-Chingiz, ChSa: Chu-Sarysu, CT: Central Tianshan, CY: Chu-Yili, Na: Naryn, Te: Teniz and ZS: Zhaman-Sarysu; peri-Siberia: AM: Altai-Mongolia, Ar: Argunsky, Ba: Bayanhongor, Ch: Chara, ChT: Charysh-Terekta, GA: Gorny Altai, GuZo: Gurvansayhan-

Zoolen, Ku: Kuerti, RA: Rudny Altai, Su: Sunidzuoqi, Ts: Tseel, St: Stanovoy; Tarim-North China: Al: Alai, Ha: Hanshan, Ho: Hongliuhe, On: Ondor Sum. This figure is in part derivative from the Neftex Geodynamic Earth Model. © Neftex Petroleum Consultants Ltd.

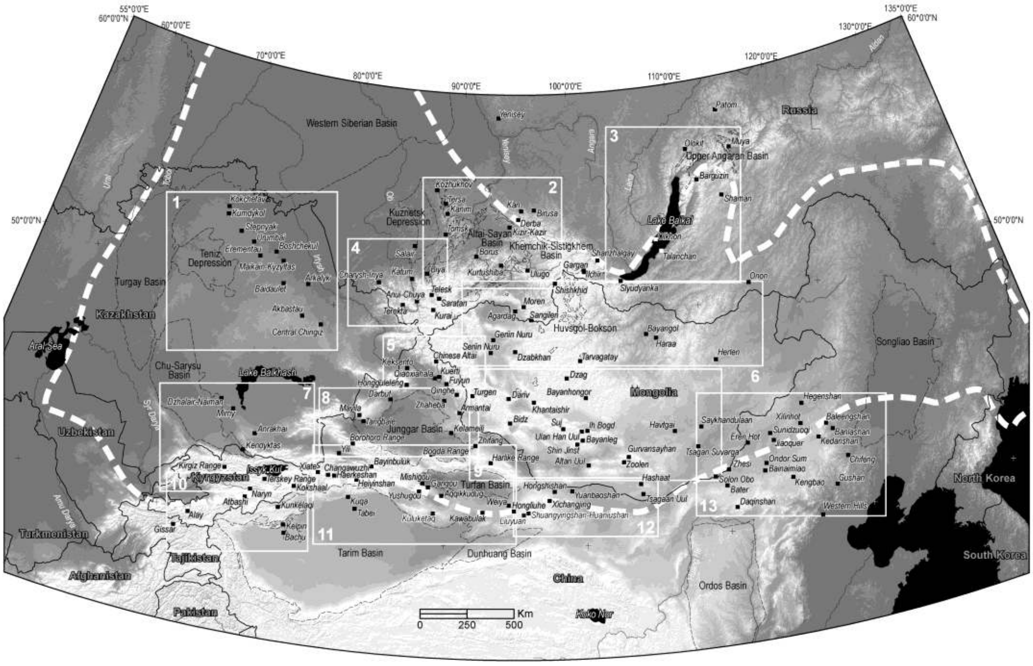
Figure 16: Famennian (370 Ma) plate tectonic reconstruction (Longitude/Latitude WGS 1984, fixed Europe, legend on Fig. 14). Tectonic events: 1) Oblique collision between Kazakhstania and Siberia (Fig. 12), a: accretion of the island arc "tail" of the Kazakhstan Continent (i.e. southwestern Mongolian deformation), b: Bayanhongor stacking, c: closure of the Kuerti back-arc basin and d: beginning of strike-slip deformation within the Charysh-Terekta (ChT) and Kuznetz-Telesk (KuT) zones (i.e. movement of the Altai-Mongolian microcontinent along the Siberian margin); 2) Detachment of the northern "branch" of the Kazakhstan Continent (TM: Tar-Muromtsev) by oblique back-arc opening and 3) Accretion of the inferred Kyrgyz microcontinent (closure of the Turkestan 1 Ocean). Localities: Internal Kazakhstan arc and related accretionary wedge (Fig. 9): Ha: Hanshan, CY: Chu-Yili, NBa: DuBa: Dulate-Baytag, North Balkash, TB: Turpan-Bogda, ZS: Zhaman-Sarysu, ZSa: Zharma-Saur. See Stampfli et al. (2011) for information about the global tectonic framework (i.e. Galatian Superterrane and Rheic Ocean). This figure is in part derivative from the Neftex Geodynamic Earth Model. © Neftex Petroleum Consultants Ltd.

Figure 17: Visean (330 Ma) plate tectonic reconstruction (Longitude/Latitude WGS 1984, fixed Europe, legend on Fig. 14). Tectonic events: 1) Accretion of the Khangai-Khentey Terrane (KK) detached from the Southern Mongolian margin (not discussed in this paper, see Wilhem 2010); 2) Rotation of the northern branch of Kazakhstan (ZhS: Zharma-Saur, DuBa: Dulate-Baytag, Fig. 9) and 3) Formation of the South Tianshan (ST) Suture (Fig. 12B). Localities: Siberian active margin (Fig. 13): ChA: Chinese Altai, IM: Inner Mongolia, SA: Siberian Altai and SM: southern Mongolia. Tarim passive margin (Fig. 12A): Ko: Kokshaal, Al: Alay. Kazakhstan internal active margin (Fig. 9: AJ: Aktau-Junggar, At: Atasbogd, CY: Chu-Yili, NT:

North Tianshan and JuBa: Junggar-Balkash, Kazakhstan external margin (Fig. 9): ChK: Chatkal-Karatau, Va: Valerianov. See Stampfli et al. (2011) for information about the global tectonic framework (i.e. Galatian Superterrane and Rheic Ocean). This figure is in part derivative from the Neftex Geodynamic Earth Model. © Neftex Petroleum Consultants Ltd.

Figure 18: Gzhelian-Asselian (300 Ma) plate tectonic reconstruction (Longitude/Latitude WGS 1984, fixed Europe, legend on Fig. 14). Tectonic events: 1) Chara Zone: Middle-Late Carboniferous formation of the suture in Siberian Altai (SA) and latest Carboniferous-Early Permian strike-slip motion (Fig. 13), 2) Middle-Late Carboniferous formation of the Uralian suture, 3) Late Carboniferous formation of the Turkestan suture (Fig. 12A), 4) Latest Carboniferous partial formation of the Junggar-Balkash suture zone (Bal: Balkash, JuB: Junggar Basin NT: North Tianshan, WJu: Western Junggar), 5) Accretion of the northeastern Junggar arc (NEJu) to the Siberian margin (i.e. Erqis Suture): end of arc activity in the Chinese Altai (ChA) (Fig. 13) and progradation of the northward accretionary wedge (i.e. Kelameili-Karamai Ophiolitic Belt, see Fig. 9), (6) Opening of the Hegenshan back-arc ocean (Ba: Baolidao, Fig. 13), and 7) Eastward diachronous closure of the Junggar-Solonker Ocean (Ocean margins: At: Atasbogd, GuZo: Gurvansayhan-Zoolen) during the Permian. This figure is in part derivative from the Neftex Geodynamic Earth Model. © Neftex Petroleum Consultants Ltd.

Fig. 1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Western Siberian Basin
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Kokchetav
Kumdykol
Stepnyak
Uyimbai
Boshchekul
Ermentau
Teniz Depression
Maikain-Kyzyltas
Arkalyk
Baidaulet
Akbastau
Central Chingiz
Turgay Basin
Kazakhstan
Aral Sea
Chu-Sarysu Basin
Lake Balkhash
Uzbekistan
Turkmenistan
Afghanistan
Tajikistan
Pakistan
Kozhukhov
Tersa
Kanim
Tomsk
Kan
Birusa
Derba
Kizir-Kazir
Kuznetsk Depression
Altai-Sayan Basin
Khemchik-Sistigkhem Basin
Borus
Kurtushiba
Ulugo
Salair
Katum
Charysh-Inya
Anui-Chuya
Telesk
Saratan
Terekta
Kurai
Olokit
Muya
Upper Angaran Basin
Barguzin
Shaman
Lake Baikal
Olkhon
Talanchan
Shanzhalgay
Gargan
Ilchir
Slyudyanka
Shishkhid
Moren
Agardag
Sangilen
Huvsgol-Bokson
Genin Nuru
Senin Nuru
Dzabkhan
Tarvagatay
Bayangol
Haraa
Herlen
Onon
Songliao Basin
Mongolia
Dzag
Dariv
Bayanhongor
Khantaishir
Bidz
Sul
Ih Bogd
Havtgai
Ulan Han Uul
Bayanleg
Shin Jinst
Gurvansayhan
Zoolen
Altan Uul
Harlike Range
Hashaat
Tsagaan Uul
Hongshishan
Yuanbaoshan
Xichanging
Hongliuhe
Weiya
Shuangyingshan-Huaniushan
Liuyuan
Chinese Altai
Keksentо
Qiaoxiahala
Kuerti
Fuyun
Hongguleleng
Qinghe
Turgen
Darbut
Zhaheba
Armantai
Mayila
Tangbale
Junggar Basin
Kelameili
Borohoro Range
Zhifang
Bogda Range
Yili
Bayinbuluk
Changawuzhi
Haerkeshan
Heiyinshan
Mishigou
Gangou
Turfan Basin
Yushugou
Aqqikkuduq
Kuluketaq
Kawabulak
Kuqa
Tabei
Tarim Basin
Dunhuang Basin
Dzhalair-Naiman
Mirny
Anrakhai
Kendyktas
Kirgiz Range
Issyk Kul
Kyrgyzstan
Terskey Range
Xiate
Kokshaal
Naryn
Atbashi
Kunkelaqi
Alay
Gissar
Keipin
Bachu
Syr Darya
Amu Darya
Hegenshan
Saykhandulaan
Xilinhot
Baleengshan
Sunidzuoqi
Baiiashan
Erenhot
Jiaoquer
Kedanshan
Tsagan Suvarga
Zhesi
Ondor Sum
Chifeng
Bainaimiao
Solon Obo
Bater
Kengbao
Gushan
Daqinshan
Western Hills
North Korea
South Korea
China
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Fig. 2

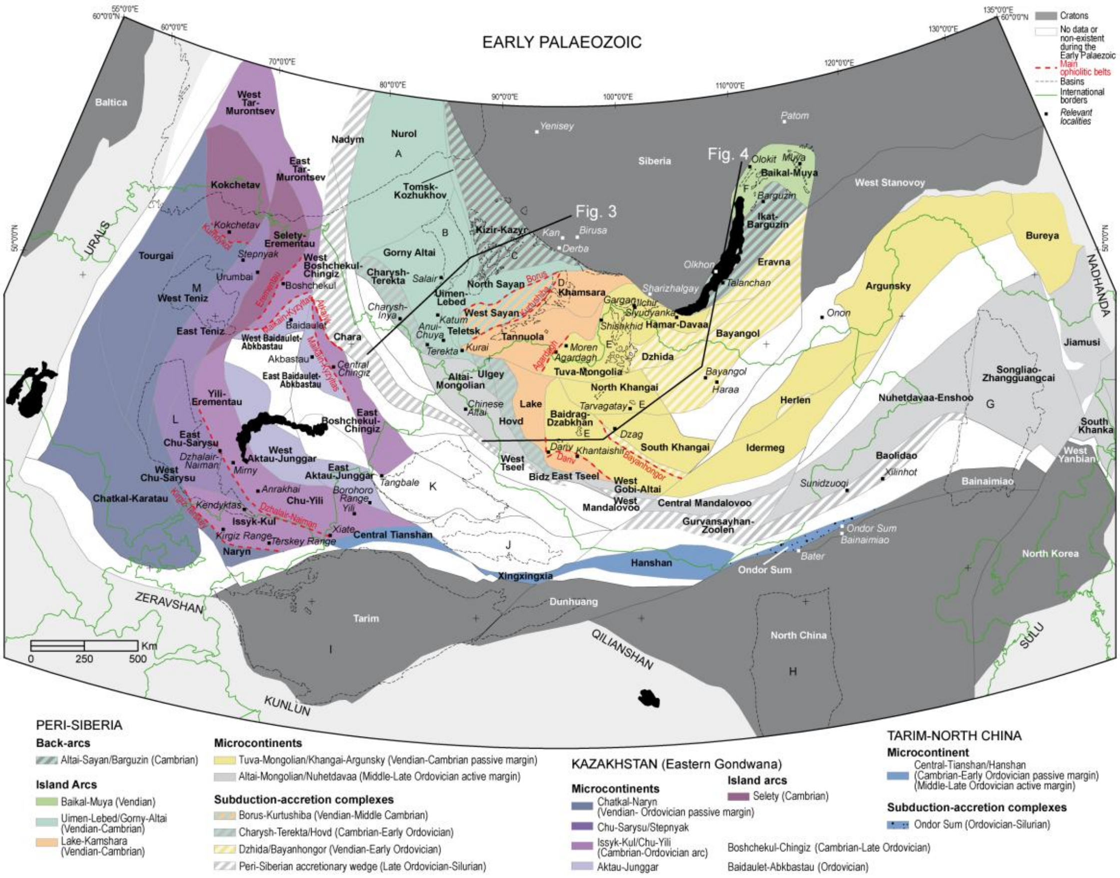

EARLY PALAEOZOIC
Cratons
No data or non-existent during the Early Palaeozoic
Main ophiolitic belts
Basins
International borders
Relevant localities
Fig. 3
Fig. 4
Baltica
Siberia
Tarim
North China
West Stanovoy
URALS
ZERAVSHAN
KUNLUN
QILIANSHAN
NADHANDA
SULU
Nurol
Nadym
Tomsk-Kozhukhov
Gorny Altai
Kizir-Kazyr
North Sayan
West Sayan
Khamsara
Tannuola
Tuva-Mongolia
North Khangai
Baidrag-Dzabkhan
South Khangai
Hamar-Davaa
Dzhida
Bayangol
Eravna
Ikat-Barguzin
Baikal-Muya
Argunsky
Bureya
Jiamusi
Songliao-Zhangguangcai
Nuhetdavaa-Enshoo
Herlen
Idermeg
Baolidao
Bainaimiao
North Korea
South Khanka
West Yanbian
Central Mandalovoo
Gurvansayhan-Zoolen
West Mandalovoo
West Gobi-Altai
East Tseel
West Tseel
Hovd
Lake
Altai-Mongolian
Ulgey
Teletsk
Uimen-Lebed
Charysh-Terekta
Chara
Hanshan
Xingxingxia
Dunhuang
Central Tianshan
Ondor Sum
West Tar-Murontsev
East Tar-Murontsev
Kokchetav
Selety-Erementau
West Boshchekul-Chingiz
East Boshchekul-Chingiz
Tourgai
West Teniz
East Teniz
West Baidaulet-Abkbastau
East Baidaulet-Abkbastau
Yili-Erementau
East Chu-Sarysu
West Chu-Sarysu
West Aktau-Junggar
East Aktau-Junggar
Chu-Yili
Issyk-Kul
Chatkal-Karatau
Naryn
0 250 500 Km
PERI-SIBERIA
Back-arcs
Altai-Sayan/Barguzin (Cambrian)
Island Arcs
Baikal-Muya (Vendian)
Uimen-Lebed/Gorny-Altai (Vendian-Cambrian)
Lake-Kamshara (Vendian-Cambrian)
Microcontinents
Tuva-Mongolian/Khangai-Argunsky (Vendian-Cambrian passive margin)
Altai-Mongolian/Nuhetdavaa (Middle-Late Ordovician active margin)
Subduction-accretion complexes
Borus-Kurtushiba (Vendian-Middle Cambrian)
Charysh-Terekta/Hovd (Cambrian-Early Ordovician)
Dzhida/Bayanhongor (Vendian-Early Ordovician)
Peri-Siberian accretionary wedge (Late Ordovician-Silurian)
KAZAKHSTAN (Eastern Gondwana)
Microcontinents
Chatkal-Naryn (Vendian- Ordovician passive margin)
Chu-Sarysu/Stepnyak
Issyk-Kul/Chu-Yili (Cambrian-Ordovician arc)
Aktau-Junggar
Island arcs
Selety (Cambrian)
Boshchekul-Chingiz (Cambrian-Late Ordovician)
Baidaulet-Abkbastau (Ordovician)
TARIM-NORTH CHINA
Microcontinent
Central-Tianshan/Hanshan (Cambrian-Early Ordovician passive margin) (Middle-Late Ordovician active margin)
Subduction-accretion complexes
Ondor Sum (Ordovician-Silurian)

Fig. 3

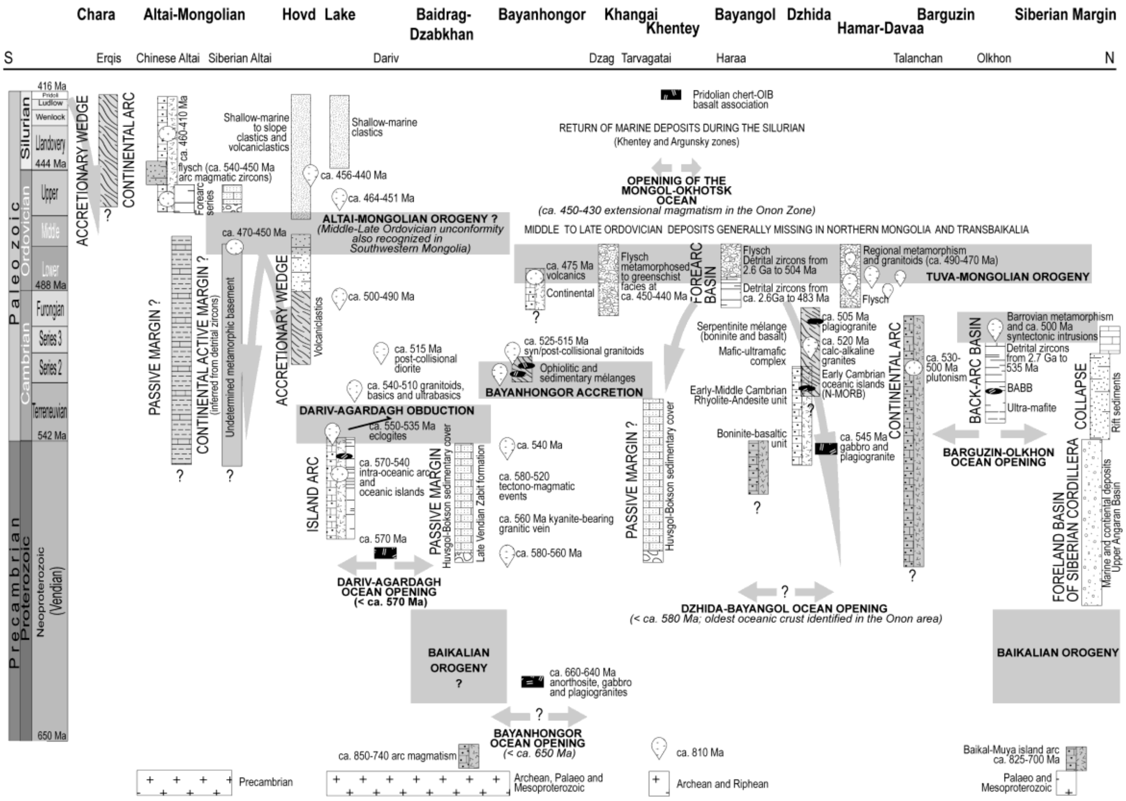

Chara
Altai-Mongolian
Hovd Lake
Baidrag-Dzabkhan
Bayanhongor
Khangai Khentey
Bayangol
Dzhida
Hamar-Davaa
Barguzin
Siberian Margin
S
Erqis
Chinese Altai
Siberian Altai
Dariv
Dzag
Tarvagatai
Haraa
Talanchan
Olkhon
N
416 Ma
444 Ma
488 Ma
542 Ma
650 Ma
Silurian
Ordovician
Cambrian
Paleozoic
Precambrian
Proterozoic
Neoproterozoic (Vendian)
Ludlow
Wenlock
Llandovery
Upper
Middle
Lower
Furongian
Series 3
Series 2
Terreneuvian
ACCRETIONARY WEDGE
CONTINENTAL ARC
PASSIVE MARGIN ?
CONTINENTAL ACTIVE MARGIN ?
(inferred from detrital zircons)
Undetermined metamorphic basement
ACCRETIONARY WEDGE
Volcaniclastics
Shallow-marine to slope clastics and volcaniclastics
Shallow-marine clastics
flysch (ca. 540-450 Ma arc magmatic zircons)
Forearc series
ca. 460-410 Ma
ca. 456-440 Ma
ca. 464-451 Ma
ca. 470-450 Ma
ALTAI-MONGOLIAN OROGENY ?
(Middle-Late Ordovician unconformity also recognized in Southwestern Mongolia)
ca. 500-490 Ma
ca. 515 Ma post-collisional diorite
ca. 540-510 granitoids, basics and ultrabasics
DARIV-AGARDAGH OBDUCTION
ca. 550-535 Ma eclogites
ISLAND ARC
ca. 570-540 intra-oceanic arc and oceanic islands
ca. 570 Ma
DARIV-AGARDAGH OCEAN OPENING (< ca. 570 Ma)
PASSIVE MARGIN
Huvsgol-Bokson sedimentary cover
Late Vendian Zabit formation
BAIKALIAN OROGENY ?
ca. 540 Ma
ca. 580-520 tectono-magmatic events
ca. 560 Ma kyanite-bearing granitic vein
ca. 580-560 Ma
ca. 660-640 Ma anorthosite, gabbro and plagiogranites
BAYANHONGOR OCEAN OPENING (< ca. 650 Ma)
ca. 525-515 Ma syn/post-collisional granitoids
Ophiolitic and sedimentary mélanges
BAYANHONGOR ACCRETION
ca. 475 Ma volcanics
Continental
Flysch metamorphosed to greenschist facies at ca. 450-440 Ma
FOREARC BASIN
Pridolian chert-OIB basalt association
RETURN OF MARINE DEPOSITS DURING THE SILURIAN
(Khentey and Argunsky zones)
OPENINIG OF THE MONGOL-OKHOTSK OCEAN
(ca. 450-430 extensional magmatism in the Onon Zone)
MIDDLE TO LATE ORDOVICIAN DEPOSITS GENERALLY MISSING IN NORTHERN MONGOLIA AND TRANSBAIKALIA
PASSIVE MARGIN ?
Huvsgol-Bokson sedimentary cover
Flysch Detrital zircons from 2.6 Ga to 504 Ma
Detrital zircons from ca. 2.6Ga to 483 Ma
Serpentinite mélange (boninite and basalt)
Mafic-ultramafic complex
Early-Middle Cambrian Rhyolite-Andesite unit
Boninite-basaltic unit
ca. 505 Ma plagiogranite
ca. 520 Ma calc-alkaline granites
Early Cambrian oceanic islands (N-MORB)
ca. 545 Ma gabbro and plagiogranite
DZHIDA-BAYANGOL OCEAN OPENING
(< ca. 580 Ma; oldest oceanic crust identified in the Onon area)
Regional metamorphism and granitoids (ca. 490-470 Ma)
TUVA-MONGOLIAN OROGENY
Flysch
CONTINENTAL ARC
ca. 530-500 Ma plutonism
BACK-ARC BASIN
Barrovian metamorphism and ca. 500 Ma syntectonic intrusions
Detrital zircons from 2.7 Ga to 535 Ma
BABB
Ultra-mafite
COLLAPSE
Rift sediments
BARGUZIN-OLKHON OCEAN OPENING
FORELAND BASIN OF SIBERIAN CORDILLERA
Marine and continental deposits Upper Angaran Basin
BAIKALIAN OROGENY
ca. 850-740 arc magmatism
ca. 810 Ma
Baikal-Muya island arc ca. 825-700 Ma
Precambrian
Archean, Palaeo and Mesoproterozoic
Archean and Riphean
Palaeo and Mesoproterozoic

Fig. 4

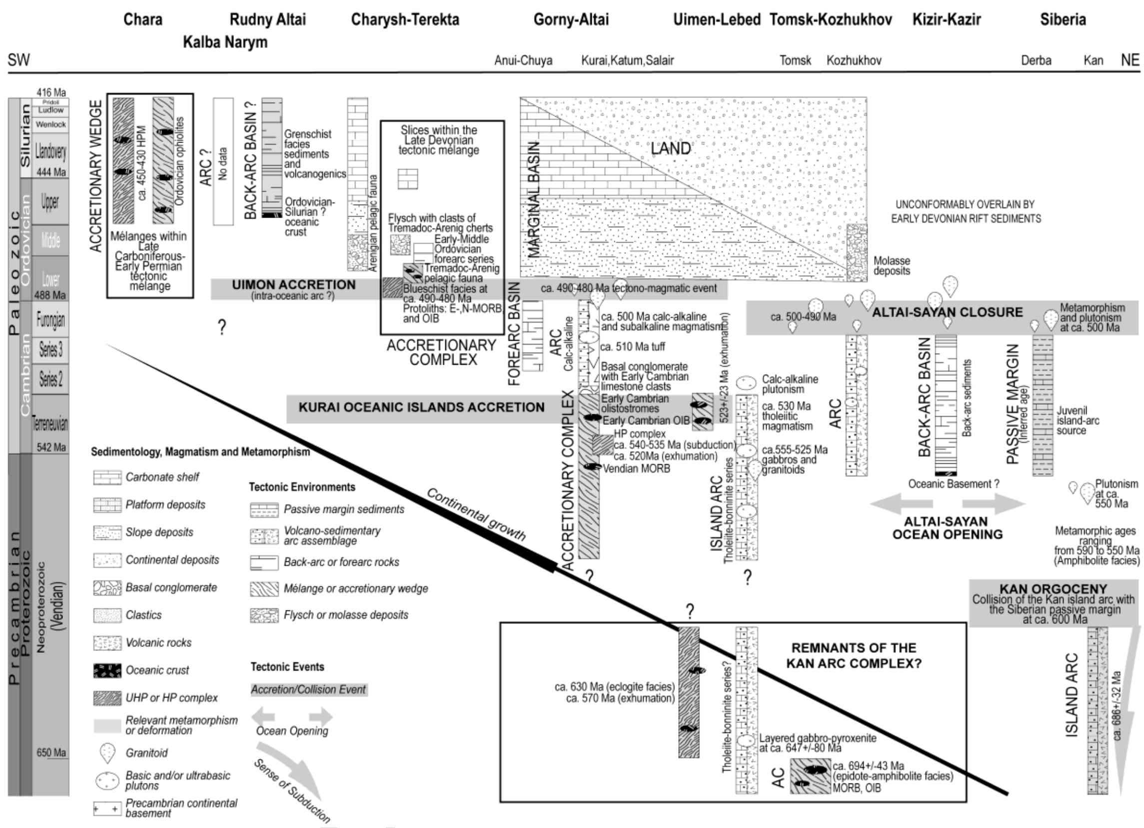

Chara
Rudny Altai
Kalba Narym
Charysh-Terekta
Gorny-Altai
Uimen-Lebed
Tomsk-Kozhukhov
Kizir-Kazir
Siberia
SW
NE
ACCRETIONARY WEDGE
UIMON ACCRETION
KURAI OCEANIC ISLANDS ACCRETION
ALTAI-SAYAN CLOSURE
ALTAI-SAYAN OCEAN OPENING
KAN ORGOCENY
REMNANTS OF THE KAN ARC COMPLEX?
ACCRETIONARY COMPLEX
MARGINAL BASIN
LAND
BACK-ARC BASIN
PASSIVE MARGIN
ISLAND ARC
Continental growth
Sense of Subduction

Fig. 5a

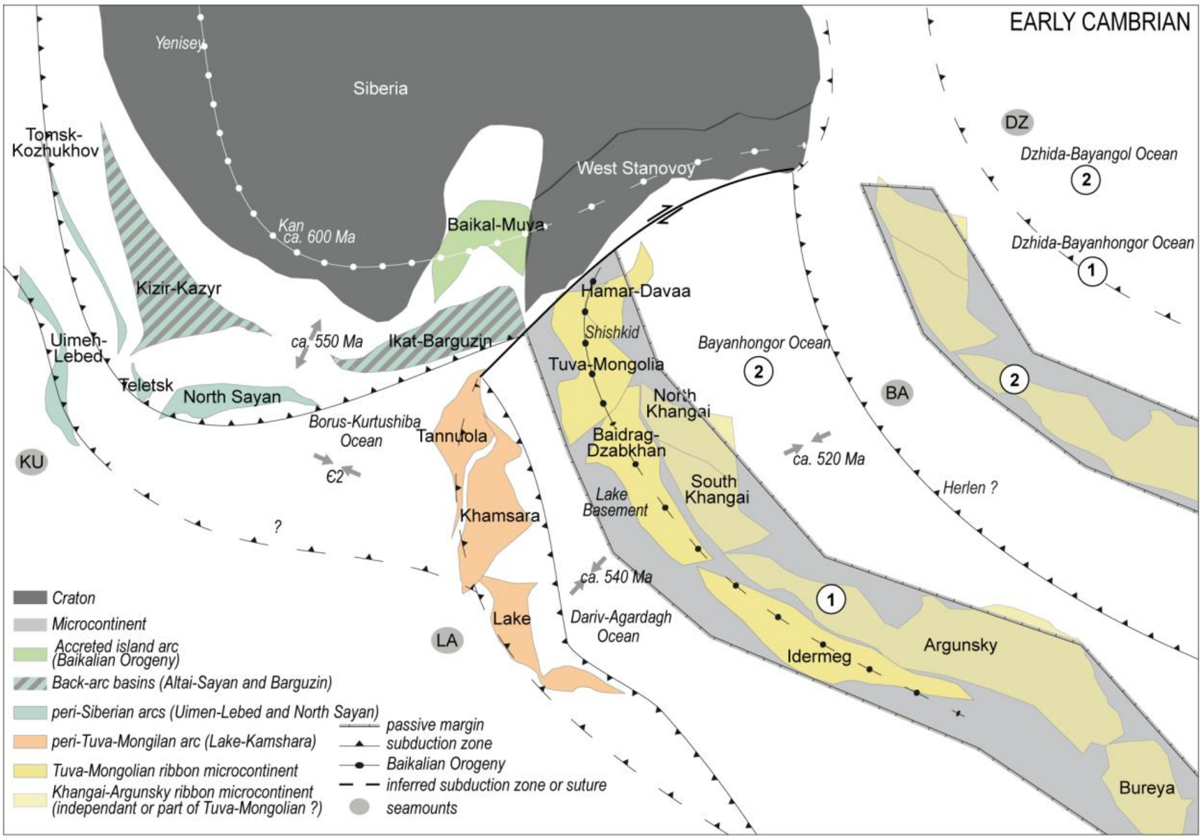

EARLY CAMBRIAN
Siberia
Yenisey
Tomsk-Kozhukhov
West Stanovoy
Baikal-Muya
Kan
ca. 600 Ma
Kizir-Kazyr
Uimen-Lebed
Teletsk
North Sayan
ca. 550 Ma
Ikat-Barguzin
Borus-Kurtushiba Ocean
Tannuola
Khamsara
Lake
KU
LA
€2
?
Hamar-Davaa
Shishkid
Tuva-Mongolia
North Khangai
Baidrag-Dzabkhan
Lake Basement
South Khangai
ca. 540 Ma
Dariv-Agardagh Ocean
Bayanhongor Ocean
ca. 520 Ma
BA
Herlen ?
Idermeg
Argunsky
Bureya
DZ
Dzhida-Bayangol Ocean
Dzhida-Bayanhongor Ocean
1
2
Craton
Microcontinent
Accreted island arc (Baikalian Orogeny)
Back-arc basins (Altai-Sayan and Barguzin)
peri-Siberian arcs (Uimen-Lebed and North Sayan)
peri-Tuva-Mongilan arc (Lake-Kamshara)
Tuva-Mongolian ribbon microcontinent
Khangai-Argunsky ribbon microcontinent (independant or part of Tuva-Mongolian ?)
passive margin
subduction zone
Baikalian Orogeny
inferred subduction zone or suture
seamounts

Fig. 5b

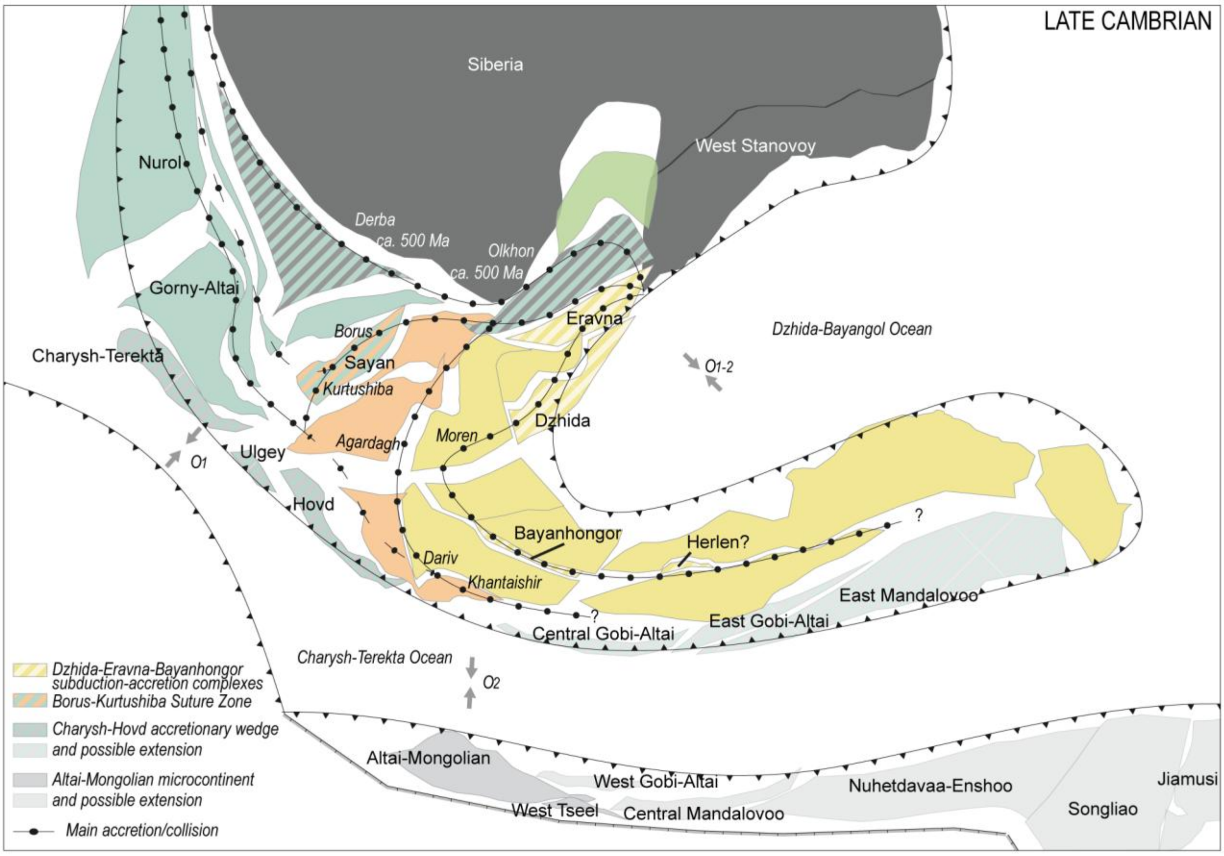

LATE CAMBRIAN
Siberia
West Stanovoy
Nurol
Derba
ca. 500 Ma
Olkhon
ca. 500 Ma
Gorny-Altai
Eravna
Dzhida-Bayangol Ocean
Borus
Sayan
Kurtushiba
Charysh-Terekta
O1-2
Dzhida
Moren
Agardagh
Ulgey
O1
Hovd
Bayanhongor
Herlen?
Dariv
Khantaishir
East Mandalovoo
East Gobi-Altai
Central Gobi-Altai
Charysh-Terekta Ocean
O2
Dzhida-Eravna-Bayanhongor subduction-accretion complexes
Borus-Kurtushiba Suture Zone
Charysh-Hovd accretionary wedge and possible extension
Altai-Mongolian microcontinent and possible extension
Main accretion/collision
Altai-Mongolian
West Gobi-Altai
Nuhetdavaa-Enshoo
Jiamusi
West Tseel
Central Mandalovoo
Songliao

Fig. 6

PERI-SIBERIAN MARGIN

		SIBERIAN ALTAI	CHINESE ALTAI	W — MONGOLIA	MONGOLIA	MONGOLIA — E	INNER MONGOLIA
O1-2		**UIMON ACCRETION (island arc ?)**		**ALTAI-MONGOLIAN ACCRETION**			**NUHETDAVAA-ENSHOO ACCRETION ?**
LATE ORDOVICIAN TO SILURIAN ACTIVE MARGIN	Accretionary wedge	**Chara** Ophiolitic mélanges with ca. 450-430 Ma HP complexes (Buslov et al., 2004a,b; Safonova et al., 2004; Volkova et al., 2007)	**Chara (Erqis)** Ophiolitic mélanges (Xiao et al., 2009b)	**Bidz** Ophiolitic mélanges (Badarch et al., 2002)	**Mandalovoo** Ophiolitic mélanges with ca., 450-430 Ma amphibolites (Naumova et al., 2006)	**Zoolen** Ordovician-Silurian greenschist-facies sediments (Bardach et al., 2002; Helo et al., 2006)	**Baolidao** Sundizuoqi-Xilinhot accretionary complex (Shi et al., 2003; Jian et al., 2008; Chen et al., 2009)
	Continental arc	**Kalba-Narym ?**	**Altai-Mongolian** Continental arc magmatism (Windley et al., 2002; Wang, T. et al., 2006; Briggs et al.,2007; Wang, Y. et al., 2011)		**Mandalovoo** ca.435-425 Ma arc-related rocks (Helo et al., 2006)		**Nuhetdavaa-Enshoo/Baolidao** Volcano-sedimentary arc assemblages (Yue et al., 2001; Xiao et al., 2003) Continental arc plutonism (Chen et al., 2000; Li et al., 2011)
O3-S1				**MONGOL OKHOTSK OPENING**			
BACK-ARC OPENING		**D2-3**	**D1-2**		**(S3)D1**		**(S3)D1**
DEVONIAN TO CARBONIFEROUS ACTIVE MARGIN	Accretionnary wedge	**Chara** Accretionary complexes (Buslov et al., 2004a,b; Safonova et al., 2004)	**Chara (Erqis)** Ophiolitic mélanges (Xiao et al., 2009b)	**Bidz** Accretionary complexes (Badarch et al., 2002)	**Zoolen** Accretionary complexes (Badarch et al., 2002; Helo et al., 2006)		**Baolidao** Sundizuoqi-Xilinhot accretionary complex (Wang et al., 1986; Xiao et al., 2003; Shi et al., 2003; Miao et al., 2007; Chen et al., 2009)
	Continental arc	**Rudny Altai** Arc assemblages (Yolkin et al., 1994)	**Kalba-Narym/Altai-Mongolian** Arc-related magmatism and volcano-sedimentary assemblages (Windley et al., 2002; Wang, T. et al., 2006; Briggs et al., 2007; Yuan et al., 2007)		**Mandalovoo/Gurvansayhan** Continental arc volcano-sedimentary arc assemblages (Lamb et al., 2000, 2001; Badarch et al., 2002; Kozakov et al., 2002, 2007; Kröner et al., 2010;)		**Baolidao** Continental arc magmatism (Chen et al., 2000; Jian et al., 2008; Li et al., 2011)
	Back-arc basin	**Rudny Altai/Gorny Altai** Back-arc sediments (Yolkin et al., 1994)	**Kalba-Narym/Altai-Mongolian** ca. 415-400Ma extensive arc-related magmatism (Wang et al., 2006; Yuan et al., 2007) ca. 370 Ma Kuerti BABB ophiolite (Xu et al., 2003; Zhang, H. et al., 2003) Metamorphosed clastics and chert (Windley et al., 2002)	**Tseel** Transitional arc/back-arc assemblages (Helo et al., 2006; Demoux et al., 2009b)	**Gobi-Altai/Mandalovoo** Continental stretching (Kröner et al., 2010) Back-arc sediments (Lamb et al., 2000, 2001; Bardach et al., 2002, Blight et al., 2008,)		**Nuhetdavaa-Enshoo** Back-arc sediments (Yue et al., 2001; Badarch et al., 2002)

Fig. 7

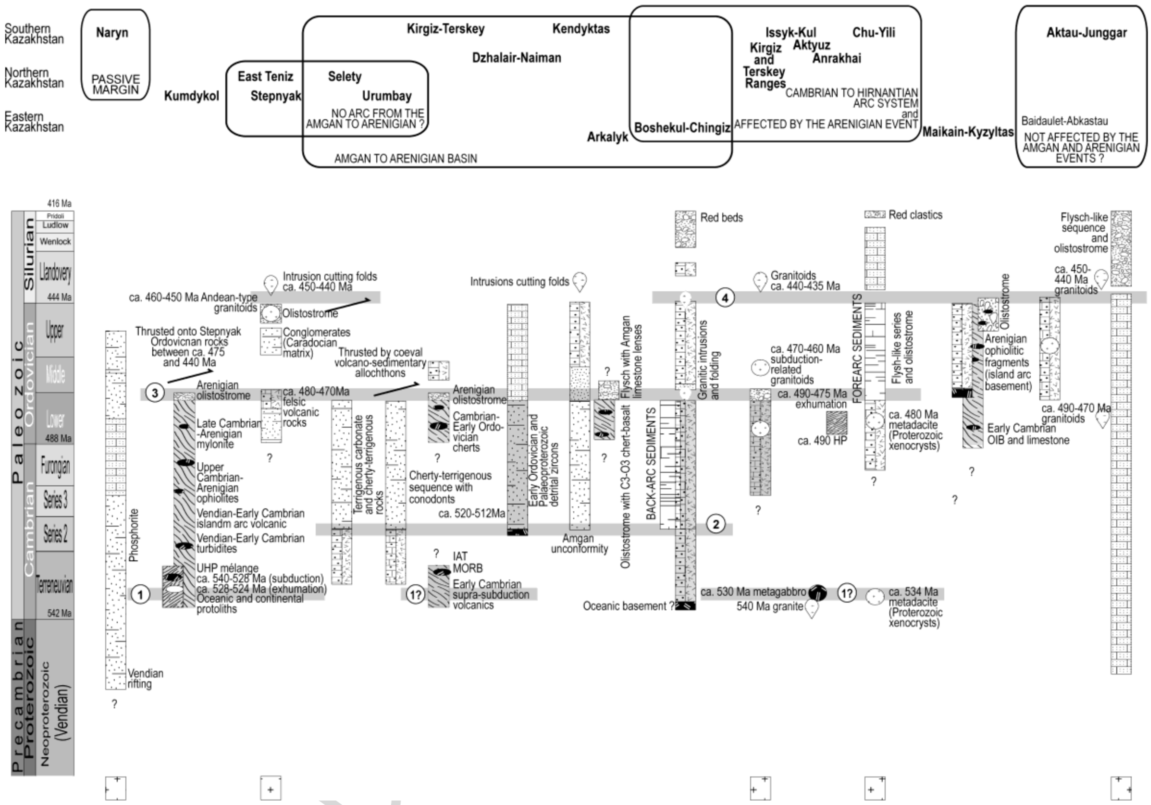
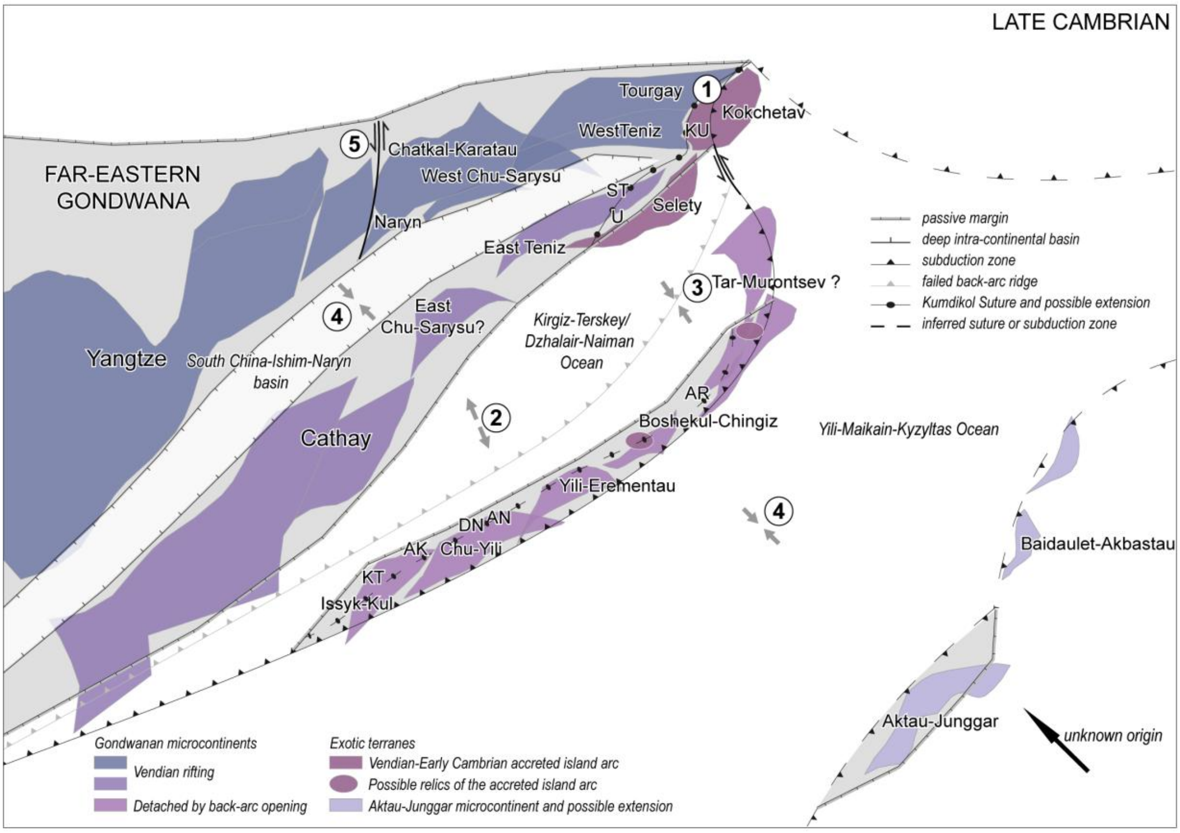

Southern Kazakhstan
Northern Kazakhstan
Eastern Kazakhstan
Naryn
PASSIVE MARGIN
Kumdykol
East Teniz
Stepnyak
Selety
Urumbay
NO ARC FROM THE AMGAN TO ARENIGIAN ?
Kirgiz-Terskey
Kendyktas
Dzhalair-Naiman
AMGAN TO ARENIGIAN BASIN
Arkalyk
Boshekul-Chingiz
Issyk-Kul
Kirgiz and Terskey Ranges
Aktyuz
Anrakhai
Chu-Yili
CAMBRIAN TO HIRNANTIAN ARC SYSTEM and AFFECTED BY THE ARENIGIAN EVENT
Maikain-Kyzyltas
Aktau-Junggar
Baidaulet-Abkastau
NOT AFFECTED BY THE AMGAN AND ARENIGIAN EVENTS ?
416 Ma
Pridoli
Ludlow
Wenlock
Llandovery
444 Ma
Silurian
Upper
Middle
Lower
488 Ma
Ordovician
Furongian
Series 3
Series 2
Terreneuvian
542 Ma
Cambrian
Paleozoic
Precambrian
Proterozoic
Neoproterozoic (Vendian)
Phosphorite
Vendian rifting
Thrusted onto Stepnyak Ordovicnan rocks between ca. 475 and 440 Ma
ca. 460-450 Ma Andean-type granitoids
Intrusion cutting folds ca. 450-440 Ma
Olistostrome
Conglomerates (Caradocian matrix)
Arenigian olistostrome
ca. 480-470Ma felsic volcanic rocks
Late Cambrian -Arenigian mylonite
Upper Cambrian-Arenigian ophiolites
Vendian-Early Cambrian islandm arc volcanic
Vendian-Early Cambrian turbidites
UHP mélange
ca. 540-528 Ma (subduction)
ca. 528-524 Ma (exhumation)
Oceanic and continental protoliths
Thrusted by coeval volcano-sedimentary allochthons
Terrigenous carbonate and cherty-terrigenous rocks
Cherty-terrigenous sequence with conodonts
ca. 520-512Ma
Arenigian olistostrome
Cambrian-Early Ordovician cherts
IAT
MORB
Early Cambrian supra-subduction volcanics
Intrusions cutting folds
Early Ordovician and Palaeoproterozoic detrital zircons
Amgan unconformity
Flysch with Amgan limestone lenses
Olistostrome with C3-O3 chert-basalt
BACK-ARC SEDIMENTS
Oceanic basement ?
Red beds
Granitic intrusions and folding
Granitoids ca. 440-435 Ma
ca. 470-460 Ma subduction-related granitoids
ca. 490-475 Ma exhumation
ca. 490 HP
ca. 530 Ma metagabbro
540 Ma granite
Red clastics
FOREARC SEDIMENTS
Flysh-like series and olistostrome
ca. 480 Ma metadacite (Proterozoic xenocrysts)
ca. 534 Ma metadacite (Proterozoic xenocrysts)
Olistostrome
Arenigian ophiolitic fragments (island arc basement)
Early Cambrian OIB and limestone
Flysch-like sequence and olistostrome
ca. 450-440 Ma granitoids
ca. 490-470 Ma granitoids
1
1?
2
3
4

Fig. 8



LATE CAMBRIAN
FAR-EASTERN
GONDWANA
Yangtze
South China-Ishim-Naryn
basin
Cathay
Tourgay
Kokchetav
WestTeniz
KU
Chatkal-Karatau
West Chu-Sarysu
ST
U
Selety
Naryn
East Teniz
East
Chu-Sarysu?
Kirgiz-Terskey/
Dzhalair-Naiman
Ocean
Tar-Murontsev ?
AR
Boshekul-Chingiz
Yili-Maikain-Kyzyltas Ocean
Yili-Erementau
DN
AN
AK
Chu-Yili
KT
Issyk-Kul
Baidaulet-Akbastau
Aktau-Junggar
unknown origin
passive margin
deep intra-continental basin
subduction zone
failed back-arc ridge
Kumdikol Suture and possible extension
inferred suture or subduction zone
Gondwanan microcontinents
Vendian rifting
Detached by back-arc opening
Exotic terranes
Vendian-Early Cambrian accreted island arc
Possible relics of the accreted island arc
Aktau-Junggar microcontinent and possible extension

Fig. 9

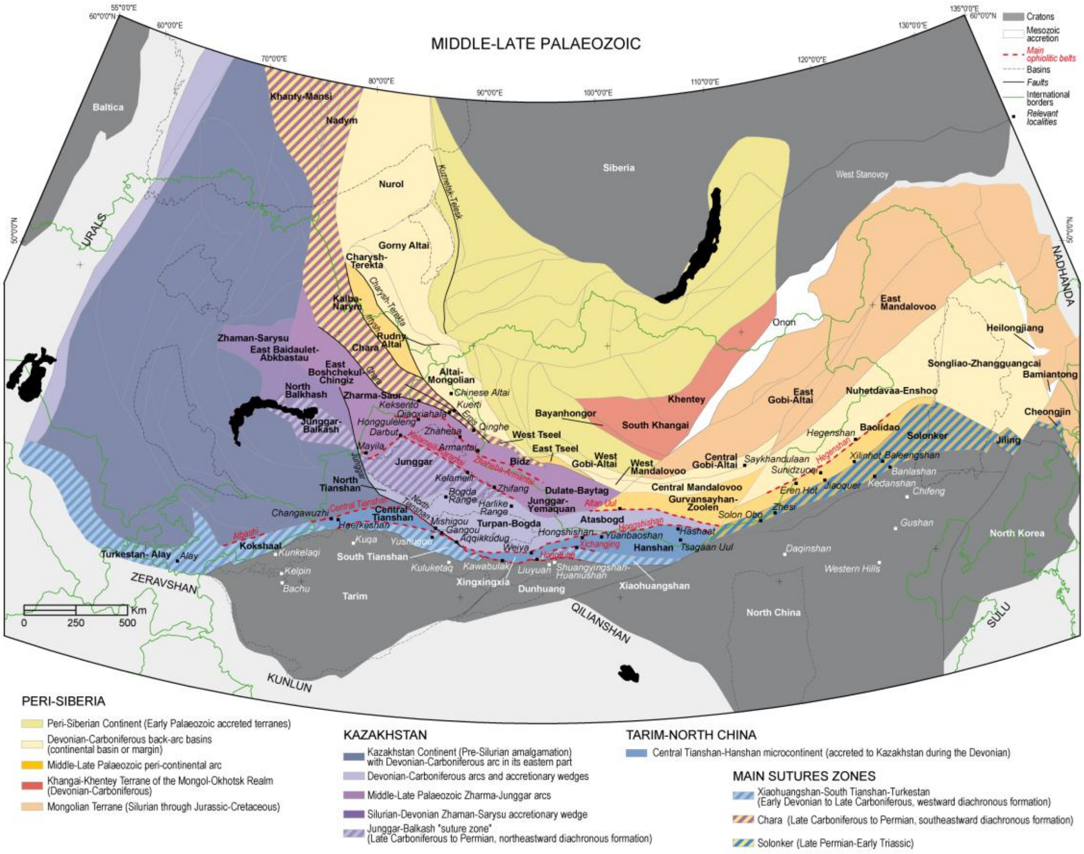

MIDDLE-LATE PALAEOZOIC
Cratons
Mesozoic accretion
Main ophiolitic belts
Basins
Faults
International borders
Relevant localities
Baltica
Siberia
West Stanovoy
URALS
NADHANDA
Khanty-Mansi
Nadym
Nurol
Gorny Altai
Charysh-Terekta
Kalba-Narym
Rudny Altai
Chara
Zhaman-Sarysu
East Baidaulet-Abkbastau
East Boshchekul-Chingiz
North Balkhash
Zharma-Saur
Junggar-Balkash
Altai-Mongolian
Bayanhongor
South Khangai
Khentey
Onon
East Mandalovoo
Heilongjiang
Songliao-Zhangguangcai
Bamiantong
East Gobi-Altai
Nuhetdavaa-Enshoo
Cheongjin
Baolidao
Solonker
Jiling
West Tseel
East Tseel
Bidz
West Gobi-Altai
West Mandalovoo
Central Gobi-Altai
Central Mandalovoo
Gurvansayhan-Zoolen
Junggar
North Tianshan
Central Tianshan
Dulate-Baytag
Junggar-Yemaquan
Turpan-Bogda
Atasbogd
Hanshan
Kokshaal
Turkestan-Alay
South Tianshan
Xingxingxia
Dunhuang
Xiaohuangshan
Tarim
North China
North Korea
ZERAVSHAN
QILIANSHAN
KUNLUN
SULU
0 250 500 Km
PERI-SIBERIA
Peri-Siberian Continent (Early Palaeozoic accreted terranes)
Devonian-Carboniferous back-arc basins (continental basin or margin)
Middle-Late Palaeozoic peri-continental arc
Khangai-Khentey Terrane of the Mongol-Okhotsk Realm (Devonian-Carboniferous)
Mongolian Terrane (Silurian through Jurassic-Cretaceous)
KAZAKHSTAN
Kazakhstan Continent (Pre-Silurian amalgamation) with Devonian-Carboniferous arc in its eastern part
Devonian-Carboniferous arcs and accretionary wedges
Middle-Late Palaeozoic Zharma-Junggar arcs
Silurian-Devonian Zhaman-Sarysu accretionary wedge
Junggar-Balkash "suture zone" (Late Carboniferous to Permian, northeastward diachronous formation)
TARIM-NORTH CHINA
Central Tianshan-Hanshan microcontinent (accreted to Kazakhstan during the Devonian)
MAIN SUTURES ZONES
Xiaohuangshan-South Tianshan-Turkestan (Early Devonian to Late Carboniferous, westward diachronous formation)
Chara (Late Carboniferous to Permian, southeastward diachronous formation)
Solonker (Late Permian-Early Triassic)

Fig. 10

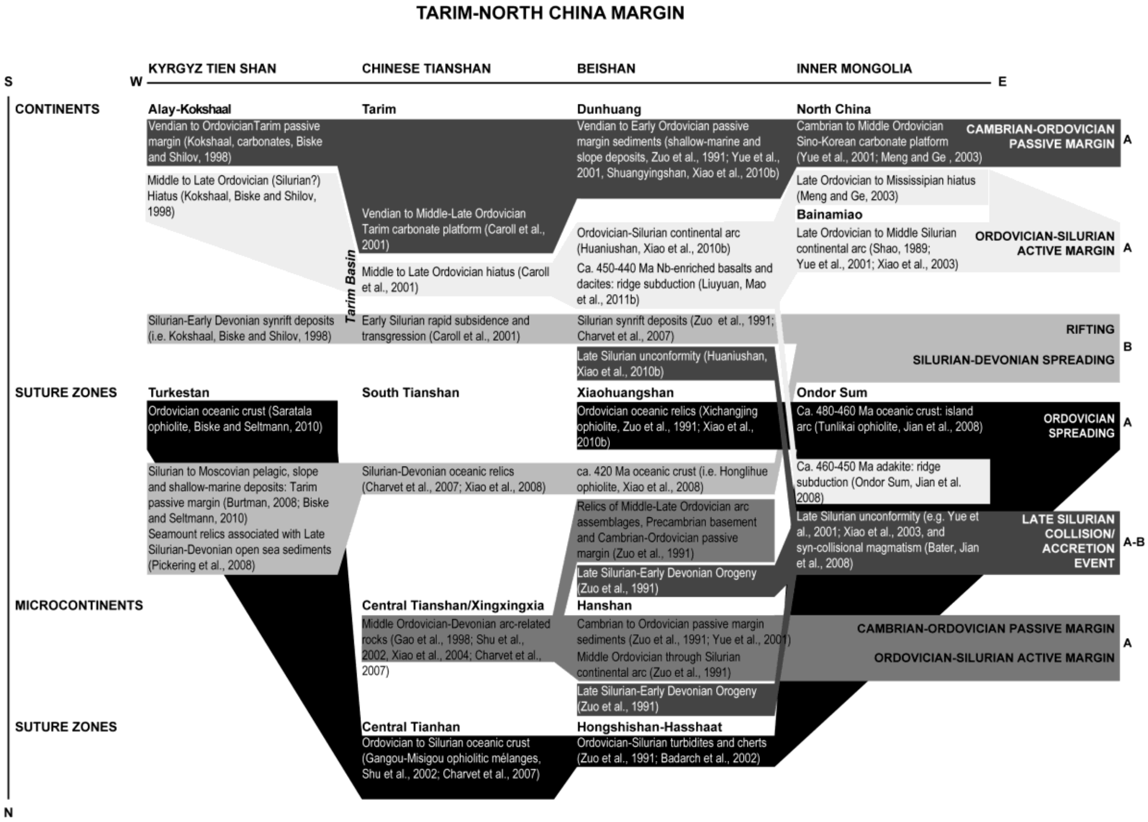

TARIM-NORTH CHINA MARGIN
S
W
KYRGYZ TIEN SHAN
CHINESE TIANSHAN
BEISHAN
INNER MONGOLIA
E
CONTINENTS
Alay-Kokshaal
Vendian to OrdovicianTarim passive margin (Kokshaal, carbonates, Biske and Shilov, 1998)
Middle to Late Ordovician (Silurian?) Hiatus (Kokshaal, Biske and Shilov, 1998)
Silurian-Early Devonian synrift deposits (i.e. Kokshaal, Biske and Shilov, 1998)
Tarim
Vendian to Middle-Late Ordovician Tarim carbonate platform (Caroll et al., 2001)
Middle to Late Ordovician hiatus (Caroll et al., 2001)
Tarim Basin
Early Silurian rapid subsidence and transgression (Caroll et al., 2001)
Dunhuang
Vendian to Early Ordovician passive margin sediments (shallow-marine and slope deposits, Zuo et al., 1991; Yue et al., 2001, Shuangyingshan, Xiao et al., 2010b)
Ordovician-Silurian continental arc (Huaniushan, Xiao et al., 2010b)
Ca. 450-440 Ma Nb-enriched basalts and dacites: ridge subduction (Liuyuan, Mao et al., 2011b)
Silurian synrift deposits (Zuo et al., 1991; Charvet et al., 2007)
Late Silurian unconformity (Huaniushan, Xiao et al., 2010b)
North China
Cambrian to Middle Ordovician Sino-Korean carbonate platform (Yue et al., 2001; Meng and Ge , 2003)
CAMBRIAN-ORDOVICIAN PASSIVE MARGIN
A
Late Ordovician to Mississipian hiatus (Meng and Ge, 2003)
Bainamiao
Late Ordovician to Middle Silurian continental arc (Shao, 1989; Yue et al., 2001; Xiao et al., 2003)
ORDOVICIAN-SILURIAN ACTIVE MARGIN
A
RIFTING
B
SILURIAN-DEVONIAN SPREADING
SUTURE ZONES
Turkestan
Ordovician oceanic crust (Saratala ophiolite, Biske and Seltmann, 2010)
Silurian to Moscovian pelagic, slope and shallow-marine deposits: Tarim passive margin (Burtman, 2008; Biske and Seltmann, 2010)
Seamount relics associated with Late Silurian-Devonian open sea sediments (Pickering et al., 2008)
South Tianshan
Silurian-Devonian oceanic relics (Charvet et al., 2007; Xiao et al., 2008)
Xiaohuangshan
Ordovician oceanic relics (Xichangjing ophiolite, Zuo et al., 1991; Xiao et al., 2010b)
ca. 420 Ma oceanic crust (i.e. Honglihue ophiolite, Xiao et al., 2008)
Relics of Middle-Late Ordovician arc assemblages, Precambrian basement and Cambrian-Ordovician passive margin (Zuo et al., 1991)
Late Silurian-Early Devonian Orogeny (Zuo et al., 1991)
Ondor Sum
Ca. 480-460 Ma oceanic crust: island arc (Tunlikai ophiolite, Jian et al., 2008)
ORDOVICIAN SPREADING
A
Ca. 460-450 Ma adakite: ridge subduction (Ondor Sum, Jian et al. 2008)
Late Silurian unconformity (e.g. Yue et al., 2001; Xiao et al., 2003, and syn-collisional magmatism (Bater, Jian et al., 2008)
LATE SILURIAN COLLISION/ ACCRETION EVENT
A-B
MICROCONTINENTS
Central Tianshan/Xingxingxia
Middle Ordovician-Devonian arc-related rocks (Gao et al., 1998; Shu et al., 2002, Xiao et al., 2004; Charvet et al., 2007)
Hanshan
Cambrian to Ordovician passive margin sediments (Zuo et al., 1991; Yue et al., 2001)
Middle Ordovician through Silurian continental arc (Zuo et al., 1991)
Late Silurian-Early Devonian Orogeny (Zuo et al., 1991)
CAMBRIAN-ORDOVICIAN PASSIVE MARGIN
ORDOVICIAN-SILURIAN ACTIVE MARGIN
A
SUTURE ZONES
Central Tianhan
Ordovician to Silurian oceanic crust (Gangou-Misigou ophiolitic mélanges, Shu et al., 2002; Charvet et al., 2007)
Hongshishan-Hasshaat
Ordovician-Silurian turbidites and cherts (Zuo et al., 1991; Badarch et al., 2002)
N

Fig.11

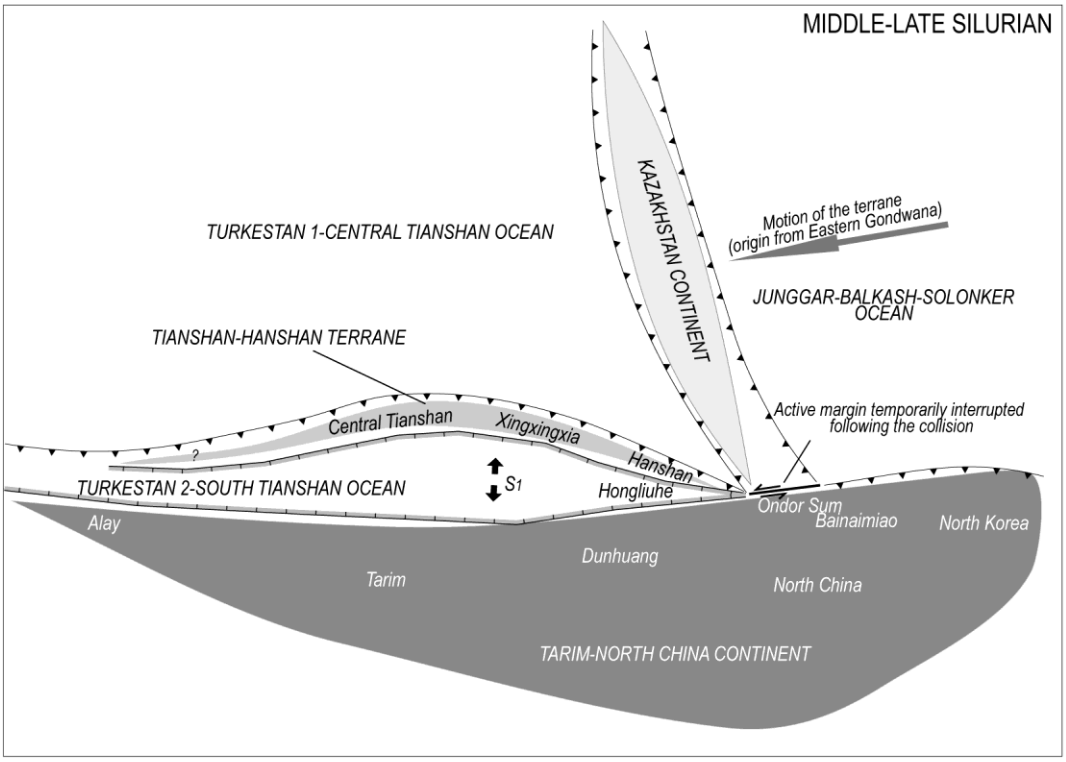

MIDDLE-LATE SILURIAN
KAZAKHSTAN CONTINENT
Motion of the terrane
(origin from Eastern Gondwana)
TURKESTAN 1-CENTRAL TIANSHAN OCEAN
JUNGGAR-BALKASH-SOLONKER
OCEAN
TIANSHAN-HANSHAN TERRANE
Active margin temporarily interrupted
following the collision
Central Tianshan
Xingxingxia
?
Hanshan
S1
TURKESTAN 2-SOUTH TIANSHAN OCEAN
Hongliuhe
Ondor Sum
Bainaimiao
North Korea
Alay
Dunhuang
Tarim
North China
TARIM-NORTH CHINA CONTINENT

Fig. 12

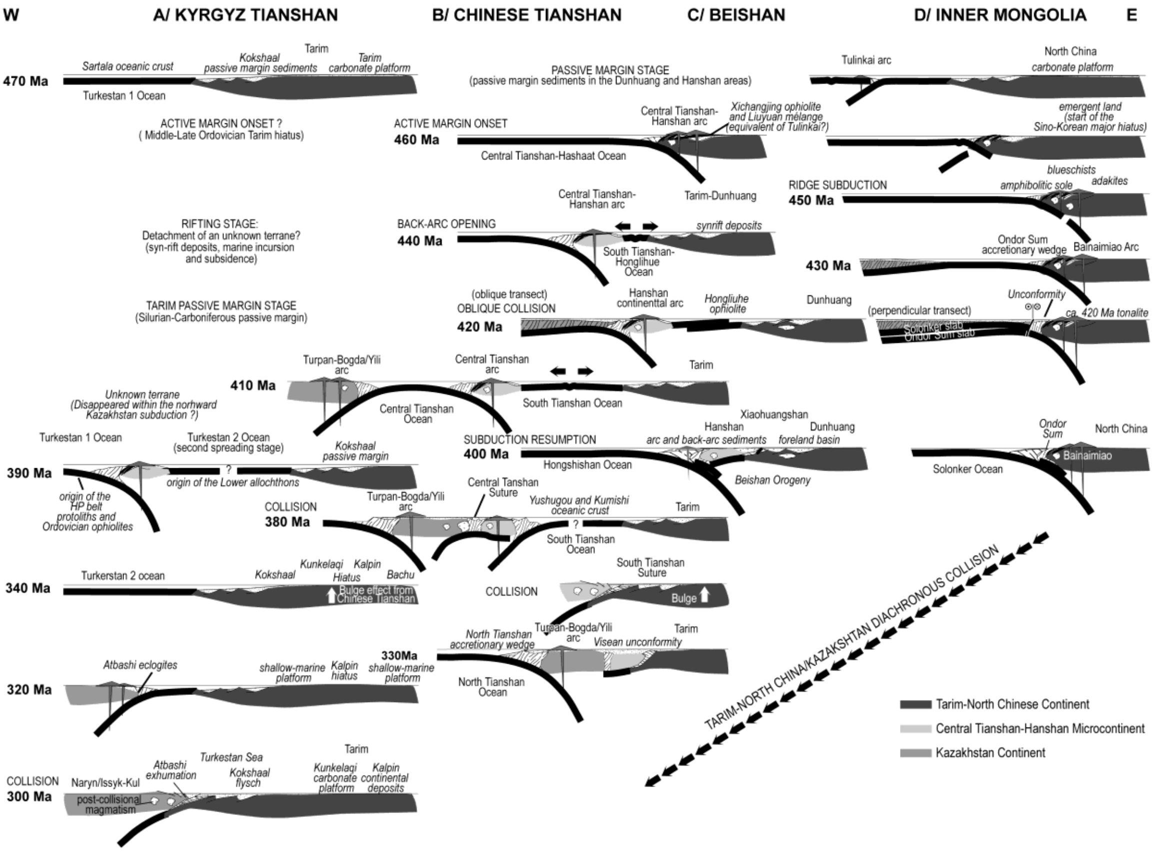

W
A/ KYRGYZ TIANSHAN
B/ CHINESE TIANSHAN
C/ BEISHAN
D/ INNER MONGOLIA
E
470 Ma
460 Ma
450 Ma
440 Ma
430 Ma
420 Ma
410 Ma
400 Ma
390 Ma
380 Ma
340 Ma
330Ma
320 Ma
300 Ma
TARIM-NORTH CHINA/KAZAKHSTAN DIACHRONOUS COLLISION
Tarim-North Chinese Continent
Central Tianshan-Hanshan Microcontinent
Kazakhstan Continent

Fig. 13

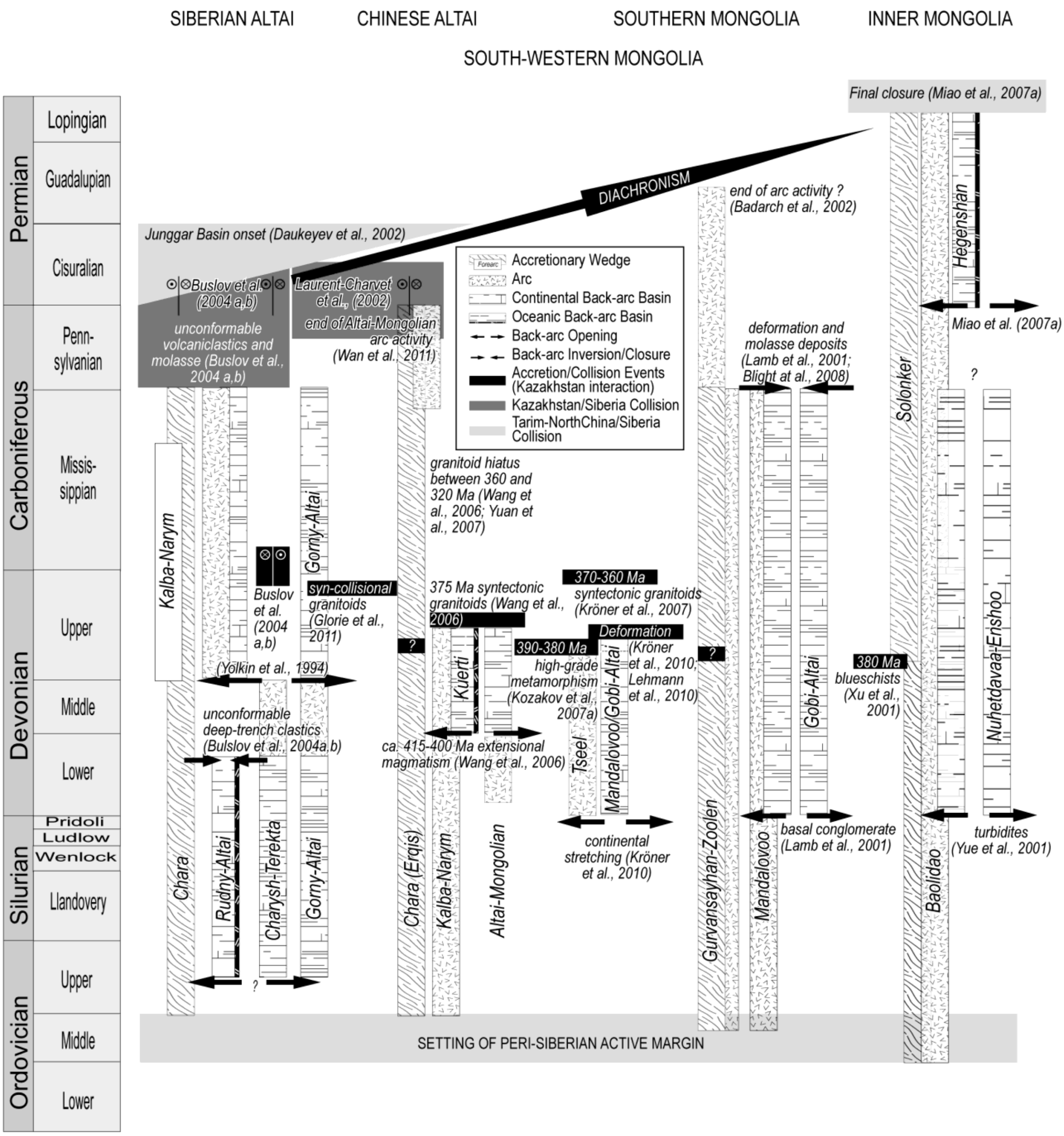

SIBERIAN ALTAI
CHINESE ALTAI
SOUTHERN MONGOLIA
INNER MONGOLIA
SOUTH-WESTERN MONGOLIA
Permian
Lopingian
Guadalupian
Cisuralian
Carboniferous
Penn-sylvanian
Missis-sippian
Devonian
Upper
Middle
Lower
Silurian
Pridoli
Ludlow
Wenlock
Llandovery
Ordovician
Upper
Middle
Lower
Final closure (Miao et al., 2007a)
DIACHRONISM
Junggar Basin onset (Daukeyev et al., 2002)
Buslov et al. (2004 a,b)
unconformable volcaniclastics and molasse (Buslov et al., 2004 a,b)
Laurent-Charvet et al., (2002)
end of Altai-Mongolian arc activity (Wan et al., 2011)
Accretionary Wedge
Forearc
Arc
Continental Back-arc Basin
Oceanic Back-arc Basin
Back-arc Opening
Back-arc Inversion/Closure
Accretion/Collision Events (Kazakhstan interaction)
Kazakhstan/Siberia Collision
Tarim-NorthChina/Siberia Collision
end of arc activity ? (Badarch et al., 2002)
Hegenshan
Miao et al. (2007a)
deformation and molasse deposits (Lamb et al., 2001; Blight at al., 2008)
Solonker
granitoid hiatus between 360 and 320 Ma (Wang et al., 2006; Yuan et al., 2007)
Kalba-Narym
Gorny-Altai
Buslov et al. (2004 a,b)
syn-collisional granitoids (Glorie et al., 2011)
375 Ma syntectonic granitoids (Wang et al., 2006)
370-360 Ma syntectonic granitoids (Kröner et al., 2007)
Deformation (Kröner et al., 2010; Lehmann et al., 2010)
390-380 Ma high-grade metamorphism (Kozakov et al., 2007a)
380 Ma blueschists (Xu et al., 2001)
Kuerti
Mandalovoo/Gobi-Altai
Gobi-Altai
Nuhetdavaa-Ershoo
(Yolkin et al., 1994)
unconformable deep-trench clastics (Bulslov et al., 2004a,b)
ca. 415-400 Ma extensional magmatism (Wang et al., 2006)
Tseel
continental stretching (Kröner et al., 2010)
basal conglomerate (Lamb et al., 2001)
turbidites (Yue et al., 2001)
Chara
Rudny-Altai
Charysh-Terekta
Gorny-Altai
Chara (Erqis)
Kalba-Narym
Altai-Mongolian
Gurvansayhan-Zoolen
Mandalovoo
Baolidao
SETTING OF PERI-SIBERIAN ACTIVE MARGIN

Fig.14

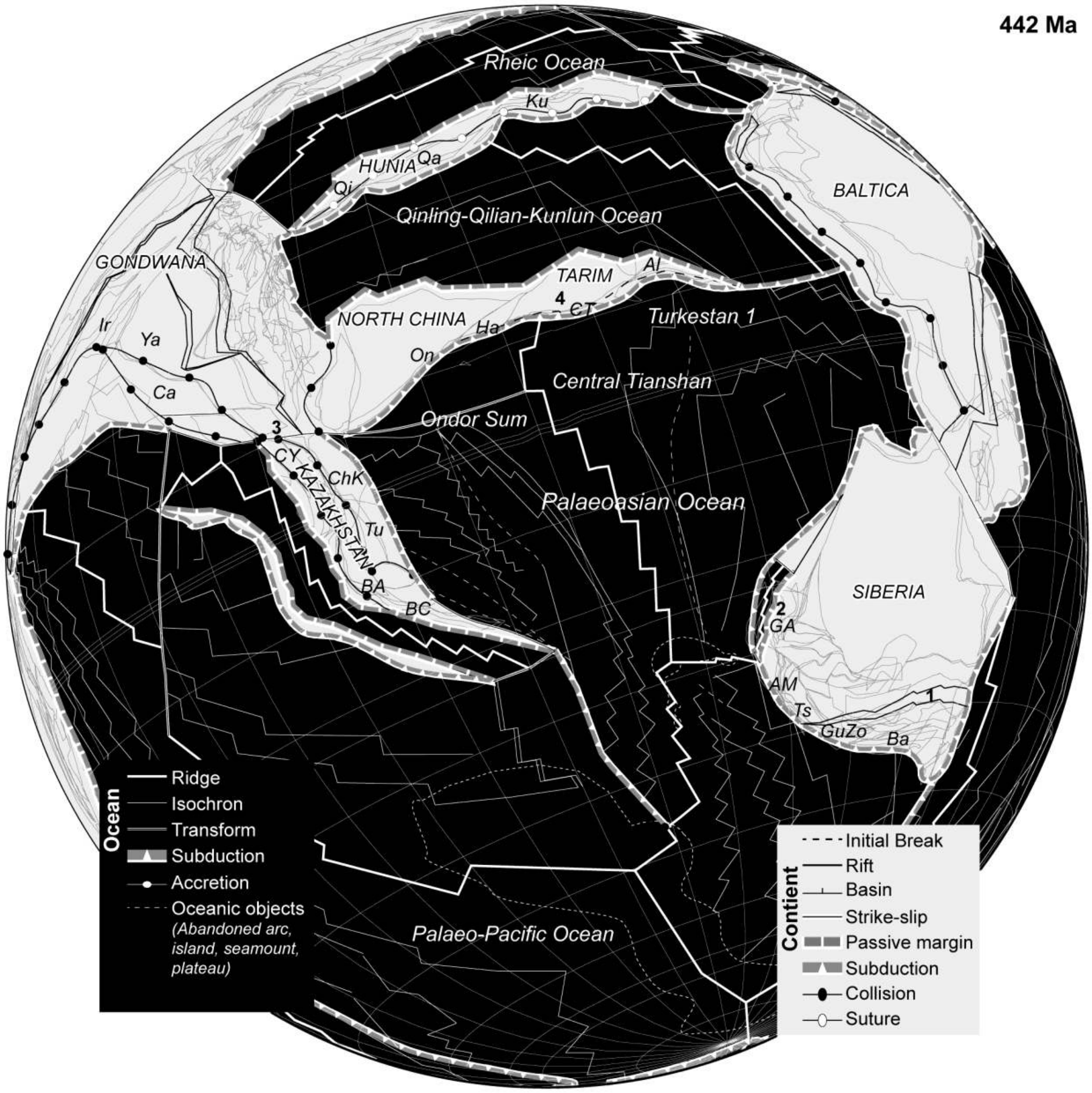

442 Ma
Rheic Ocean
Ku
HUNIA
Qa
Qi
Qinling-Qilian-Kunlun Ocean
BALTICA
GONDWANA
TARIM
Al
NORTH CHINA
Ha
4
ЄT
Turkestan 1
Ir
Ya
On
Central Tianshan
Ca
Ondor Sum
3
CY
ChK
KAZAKHSTAN
Palaeoasian Ocean
Tu
BA
BC
2
GA
SIBERIA
AM
Ts
GuZo
Ba
1
Ocean
Ridge
Isochron
Transform
Subduction
Accretion
Oceanic objects
(Abandoned arc, island, seamount, plateau)
Palaeo-Pacific Ocean
Contient
Initial Break
Rift
Basin
Strike-slip
Passive margin
Subduction
Collision
Suture

Fig.15

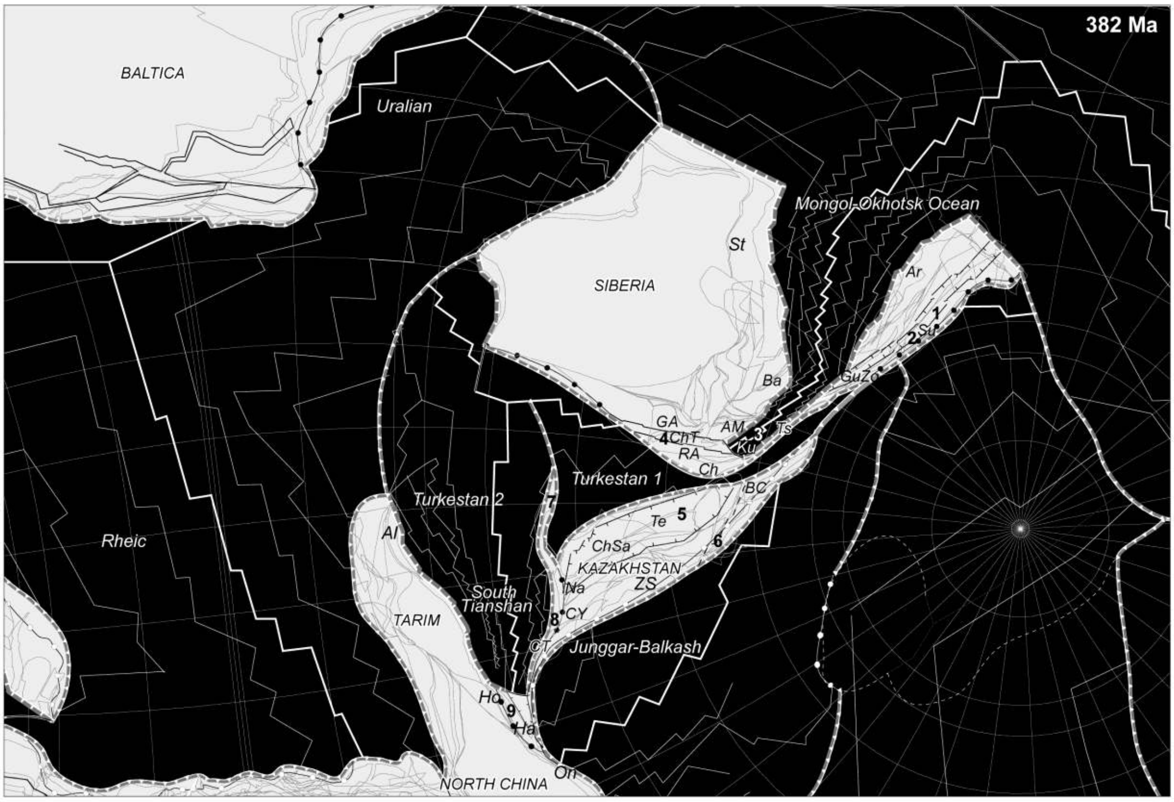

382 Ma
BALTICA
Uralian
Mongol-Okhotsk Ocean
St
SIBERIA
Ar
1
2
Su
GuZo
Ba
GA
AM
3
Ts
4
ChT
RA
Ku
Ch
Turkestan 1
BC
Turkestan 2
7
Te
5
Al
Rheic
ChSa
6
KAZAKHSTAN
ZS
Na
South
Tianshan
CY
8
TARIM
CT
Junggar-Balkash
Ho
9
Ha
On
NORTH CHINA

Fig.16

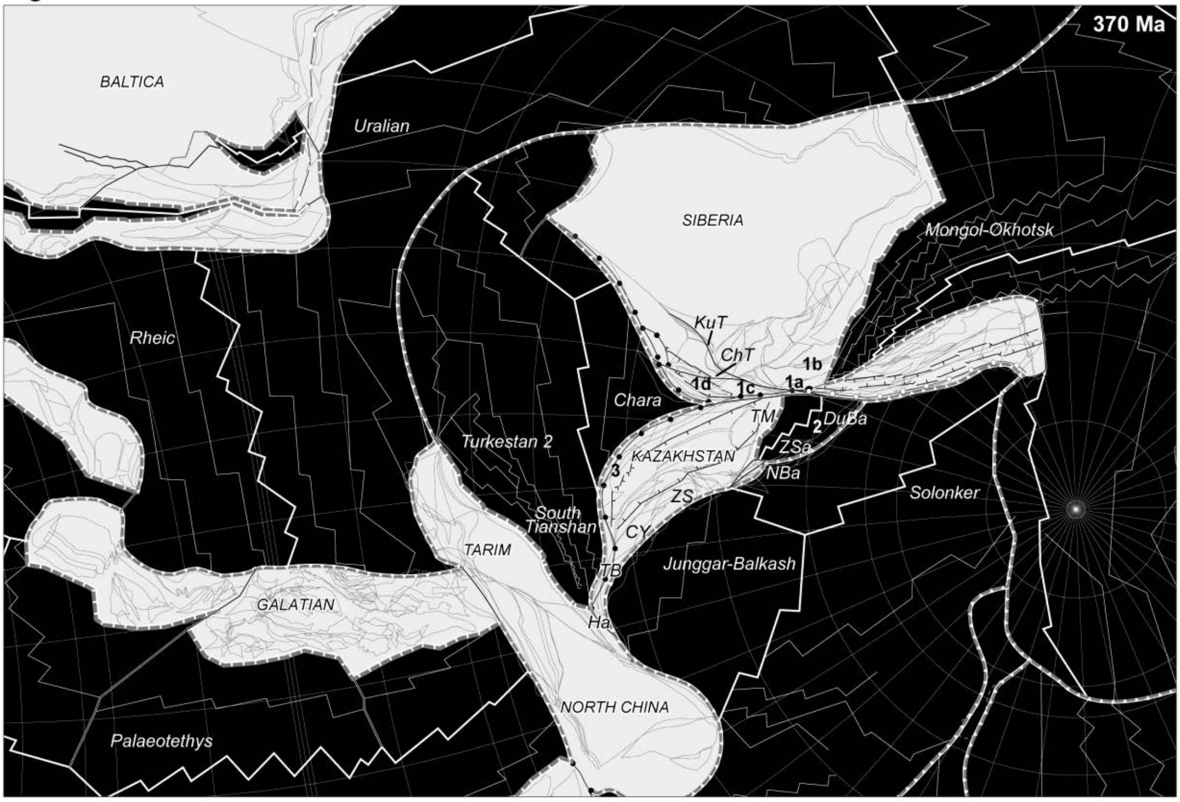
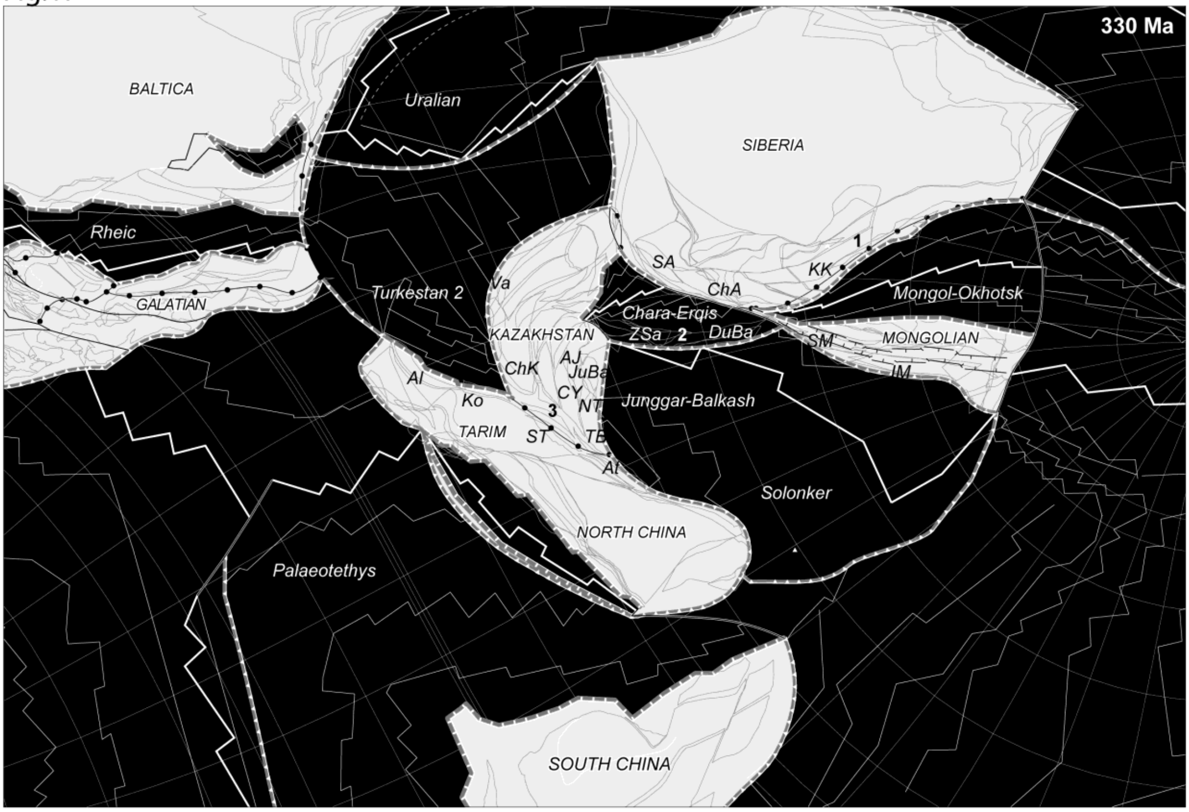
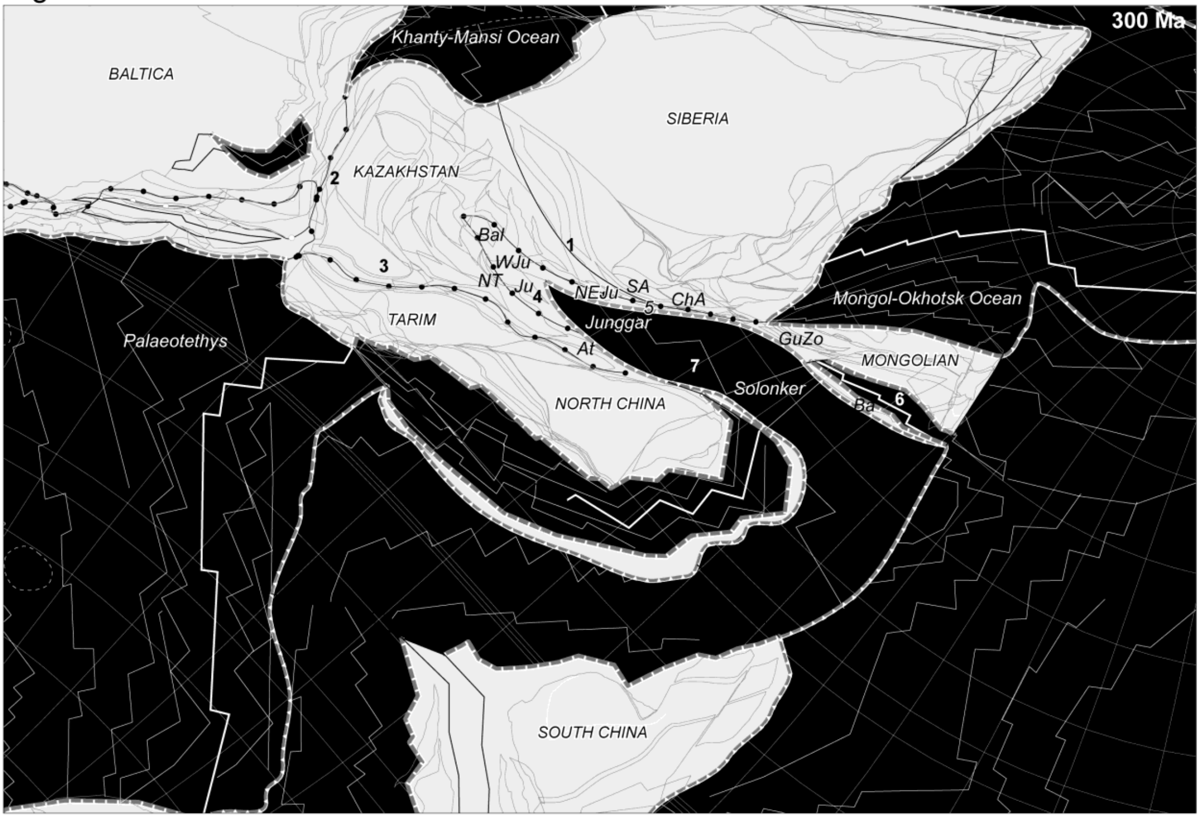

370 Ma
BALTICA
Uralian
SIBERIA
Mongol-Okhotsk
Rheic
KuT
ChT
1b
1d
1c
1a
Chara
TM
DuBa
2
ZSa
Turkestan 2
KAZAKHSTAN
3
NBa
Solonker
ZS
South
Tianshan
CY
TARIM
Junggar-Balkash
TB
GALATIAN
Ha
NORTH CHINA
Palaeotethys

Fig.17



330 Ma
BALTICA
Uralian
SIBERIA
Rheic
GALATIAN
Turkestan 2
Va
SA
ChA
KK
1
Mongol-Okhotsk
Chara-Erqis
KAZAKHSTAN
ZSa
2
DuBa
SM
MONGOLIAN
IM
AJ
JuBa
ChK
Al
Ko
CY
NT
3
Junggar-Balkash
TARIM
ST
TB
Al
Solonker
NORTH CHINA
Palaeotethys
SOUTH CHINA

Fig.18


300 Ma
Khanty-Mansi Ocean
BALTICA
SIBERIA
KAZAKHSTAN
2
Bal
1
WJu
3
NT
Ju
4
NEJu
SA
5
ChA
Mongol-Okhotsk Ocean
TARIM
Junggar
Palaeotethys
At
GuZo
MONGOLIAN
7
Solonker
6
NORTH CHINA
SOUTH CHINA