



Nature and development of the South Tianshan-Solonker suture zone

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ABSTRACT

The Central Asian Orogenic Belt (CAOB) was generated through multiple collisional and accretionary events in the Paleo-Asian Ocean, a major global ocean that existed from the late Neoproterozoic into the late Paleozoic. Nevertheless, the question of when the Paleo-Asian Ocean finally closed has notoriously been enigmatic, especially due to the absence of large-scale investigations. The South Tianshan-Solonker Suture, is the largest and southernmost suture within the CAOB, and records the ultimate collision between the Tarim-North China cratons with the Siberia craton, and is commonly interpreted as marking the eventual closure of the Paleo-Asian Ocean. In this paper, we synthesize and evaluate relevant Chinese papers on the area, which are not readily available to an international audience, and extend this across the full length of the suture zone. Based on this review, we can divide the suture zone into four distinct segments, which are, from west to east: the South Tianshan Belt, the Beishan Belt, the Solonker Belt and the Yanji Belt. This enables us to provide a more systematic understanding of the nature and development of the South Tianshan-Solonker suture.

Geochronological data from Paleozoic cover rocks and high-pressure metamorphic rocks show that during the late Carboniferous, the western section of the Paleo-Asian Ocean closed when the Tarim Craton moved northward to collide with the Kazakhstan-Yili Block, thus marking the initial development of the South Tianshan Belt. The Beishan Belt to the east formed when the Dunhuang-Alxa blocks were transported northward to collide with the Tuva-Mongolia Block, with the youngest zircon U–Pb data from ophiolites indicating that closure of the local Beishan Ocean was in the early-middle Permian, although its final closure may not have been until the early late Permian along the northern margin of the Alxa Block, thus making it slightly younger than the South Tianshan Belt. Further to the east, the available petrographic, geochronological and paleontological data from the Solonker Belt indicate that this belt was formed in the middle-late Permian, during which time bi-directional subduction occurred and the North China and Siberia cratons were amalgamated during an ‘Appalachian-type’ orogeny. Furthermore, we observe that a transition in polarity from northward to bi-directional subduction occurred along the boundary between the Beishan and Solonker belts, which may be related to the East Gobi Transform Fault. In contrast, under the influence of the westward subduction of the Paleo-Pacific Ocean, the Jiamusi-Khanka Block moved southwestward along the Yilan-Yitong Fault in the middle-late Triassic (230–220 Ma) to amalgamate with the North China Craton.

The formation of the South Tianshan-Solonker Suture can be characterized by four distinct west-to-east accretion/collision events that lasted from the late Carboniferous to the late Permian. The South Tianshan Belt, the Beishan Belt and the Solonker Belt therefore demarcate, respectively, the amalgamation between the Tarim Craton and the Kazakhstan-Yili Block, the Dunhuang-Alxa Block and the Tuva-Mongolia Block, and the North China Craton with the Siberia Craton, representing the final closure of the Paleo-Asian Ocean. During the middle-late Triassic, the Yanji Belt in the far east was influenced by the Paleo-Pacific tectonic domain as a result of amalgamation between the Jiamusi-Khanka Block and the North China Craton.

1. Introduction

As the world's largest accretionary orogenic belt, the nature and

evolution of the Central Asian Orogenic Belt (CAOB) remains an important subject for geological research. The CAOB was generated by accretion of several microcontinents, arc-back arc systems, oceanic

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islands/plateaus, ophiolites, and subduction-accretion complexes, marking the transition from the Pangea supercontinent to the evolving Eurasian continent (Şengör et al., 1993; Jahn et al., 2000; Jahn, 2004; Kröner et al., 2009, 2011, 2013; Windley et al., 2007). It stretches from the Caspian Sea in the west to the Pacific Ocean in the east, passing through Russia, Kazakhstan, Mongolia and northern China, and covering an extensive region between the Siberia, North China and Tarim cratons (Jian et al., 2013; Kröner et al., 2007; Safonova et al., 2011; Şengör and Natal'in, 1996; Şengör et al., 1993; Şengör and Okurogullari, 1991; Windley et al., 1990; Yin et al., 1998). The evolution of the CAO was a long and intricate process, which was initiated at ca. 1 Ga (Khain et al., 2002) and persisted until ca. 250 Ma (Windley et al., 2007). Since the Devonian, the CAO witnessed multiple stages of microcontinent convergence, compression and accretionary orogeny, driven by the development and ultimate closure of the Paleo-Asian Ocean, which is recorded in numerous suture zones characterized by ophiolites, high-pressure metamorphic belts, and large-scale faults (Zhou et al., 2018).

In recent years, many workers have carried out detailed studies of the CAO, and made significant advances in terms of tectonic subdivision, magmatism, geochronology, and geochemistry (e.g., Burianek et al., 2017; Chen et al., 2021a; Degtyarev et al., 2015; Degtyarev and Ryzantsev, 2007; Demoux et al., 2009a, 2009b; Khain et al., 2003; Kröner et al., 2009, 2011, 2013; Wilde et al., 2000, 2003; Wu et al., 2000, 2005,

2011a; Xiao et al., 2009b, 2010b, 2013; Zhou et al., 2009a, 2010a, 2010c). However, most of those studies were focused on the investigation of particular aspects in localized areas and therefore did not provide a comprehensive analysis at the regional scale. This has resulted in multiple hypotheses as to the nature and timing of closure of the Paleo-Asian Ocean, leaving it open to serious debate.

The South Tianshan-Solonker Suture, is the largest and southernmost suture within the CAO (Fig. 1), and records the final collision between the Tarim-North China cratons to the south with the Siberia Craton to the north, and therefore its timing is critical to defining the ultimate fate of the Paleo-Asian Ocean. Here, by integrating the latest geochronological, geochemical, paleontological and paleomagnetic data, we re-investigate the nature of this suture zone and divide it into four major sections, and thereby provide a synthesis of the subduction characteristics and closure times. The results provide an in-depth understanding of the nature and development of the South Tianshan-Solonker suture zone, and also the accretionary history of the CAO, helping to construct a more viable tectonic model for the Paleo-Asian Ocean.

2. Subdivision of the South Tianshan-Solonker Suture Zone

The South Tianshan-Solonker Suture, also called the South Tianshan-Beishan-Solonker-Yanji suture, extends over 5500 km from South Tianshan in Xinjiang Province, via Beishan, to Solonker at the border



Fig. 1. Schematic tectonic map showing the main subdivisions of the Central Asian Orogenic Belt (basemap from website <http://www.ngdc.noaa.gov/mgg/global>). **Micro-continental blocks:** ①Erguna; ②Xing'an; ③Songliao; ④Jiamusi-Khanka; ⑤Kazakhstan-Yili; ⑥Junggar; ⑦Tuva-Mongolia; ⑧Dunhuang; and ⑨Alxa. **Suture zones:** I. South Tianshan-Solonker Suture; II. North Tianshan-Hegenshan-Heihe Suture; III. Erqis-Toudaoqiao Suture. **Strike-slip Faults:** F1 = Altyn Tagh Fault; F2 = Xingxing Xia Fault; F3 = East Gobi Fault; F4 = Tanlu Fault; F5 = Yilan-Yitong Fault; F6 = Dunhua-Mishan Fault.

between China and Mongolia, where it then extends further east to the Yanji area in NE China, with a width of over 1100 km at its widest point (Fig. 1; Han and Zhao, 2018). As noted above, it is the southernmost suture within the CAO and is characterized by the distribution of ophiolites, high-pressure metamorphic rocks, collision-related intrusive rocks, and accretionary complexes. The evolution of the South Tianshan-Solonker Suture has undergone a lengthy geological history, from incorporating the closure of the Paleo-Asian Ocean to the initiation of the Paleo-Pacific tectonic domain, and is the result of the differential accretion/collision events between various continental blocks represented by the Tarim Craton, the Dunhuang-Alxa Block and the North China Craton in the south and the Kazakhstan-Yili Block, the Tuva-Mongolia Block, and the Siberia Craton in the north. Based on a review of the literature, it is possible to identify that particular sections of the suture zone were influenced by different events, based on differences in rock composition, closure timing and tectonic characteristics. As outlined above, we therefore propose that the South Tianshan-Solonker Suture can be subdivided into four segments, which are, from west to east: the South Tianshan Belt, the Beishan Belt, the Solonker Belt and the Yanji Belt. These belts record discrete interactions between various crustal blocks, and mark the progressive west-to-east closure of the Paleo-Asian Ocean during the late Paleozoic. Below, we outline the unique features that support this conclusion.

3. The South Tianshan Belt

3.1. Regional geology

The South Tianshan Belt extends in an E-W direction from Uzbekistan to the border between Xinjiang and Gansu provinces in China, stretching for about 2500 km and reaching 50–80 km in width. It consists of a complex mixture of Precambrian micro-continents, Paleozoic island arcs and accretion complexes (Zhou et al., 2018) that separate three major geological units: the Kazakhstan-Yili Block, the Central Tianshan Block and the Tarim Craton (Fig. 2; Gao et al., 2008, 2011, 2015).

The Kazakhstan-Yili Block is separated from the Central Tianshan block by the Nikolaev Line-North Nalati Suture to the south (Fig. 2; He et al., 2001; Li and Li, 1997; Lomize et al., 1997). The former is a composite terrane with Precambrian metamorphic basement. Granitic

gneisses with zircon U–Pb age of 798 Ma and Sm–Nd model ages of 2.01–1.93 Ga are reported in the Yili Block (Chen et al., 1999), and they are covered by Meso–Neoproterozoic carbonates, clastic rocks, Cambrian–Ordovician siltstones, mudstones, volcanics and early Carboniferous intermediate-basic rock (Gao et al., 1998; Liu et al., 1994; Zhai et al., 2006). Furthermore, extensive late Carboniferous–early Permian alkali-rich volcanic rocks outcrop along its southern margin (Zhao et al., 2000, 2003). In the Kazakhstan block, 2791 ± 24 Ma and 2187–1789 Ma granitic gneisses within the Precambrian basement have been reported (Kröner et al., 2007), which are covered by Cambrian–early Ordovician oceanic and island arc complexes (Mikolaichuk et al., 1997), all of which were intruded by middle Ordovician–early Silurian granites (Kiselev, 1999; Konopelko et al., 2008). Along the southern boundary of the Kazakhstan-Yili Block, high-grade metamorphic rocks formed during plate subduction, including granulites and blueschists (Han et al., 2011; Zhang et al., 2013). Ophiolites are mainly distributed along the Nikolaev Line–North Nalati Suture, and are considered to be remnants of the Neoproterozoic–early Ordovician Kyrgyz-Terskey Ocean (Lomize et al., 1997; Gao et al., 2008). They are thrust over the early Paleozoic island–arc complexes in the Northern Tianshan of Kyrgyzstan (Bazhenov et al., 2003; Qian et al., 2009).

The Central Tianshan block consists of two major units, the Kyrgyz Middle Tianshan and the Chinese Central Tianshan, which are connected by a narrow Precambrian metamorphic belt at the southern margin of the Yili Block (Gao et al., 2008). The Central Tianshan block is bounded by the Nikolaev Line–North Nalati Suture to the north and the South Central Tianshan Suture to the south (Fig. 2). However, some authors suggest that the Nikolaev Fault and North Nalati Fault are not a continuous suture zone, and that the Nikolaev Fault may disappear near the border between China and Kyrgyzstan, and that the two Central Tianshan units may be two separate tectonic units (Biske and Seltmann, 2010). In this paper, we consider the two units as a unified tectonic unit. The Central Tianshan block has a Meso- to Neoproterozoic metamorphic basement (Li and Li, 1997; Xiao et al., 1992) with zircon U–Pb ages of 1910 Ma for granulites (LNIGS (Liao Ning Institute of Geological Survey), 2005) and Sm–Nd model ages of 2.65 Ga for gneisses (Chen et al., 2000b). The Cambrian–Ordovician rocks in the Kyrgyz Middle Tianshan consist of carbonates and flysch (Milanovsky, 1996), but strata of this period are missing in the Chinese segment, where late Silurian island-arc type volcanics and volcanoclastic rocks were thrust over the Precambrian

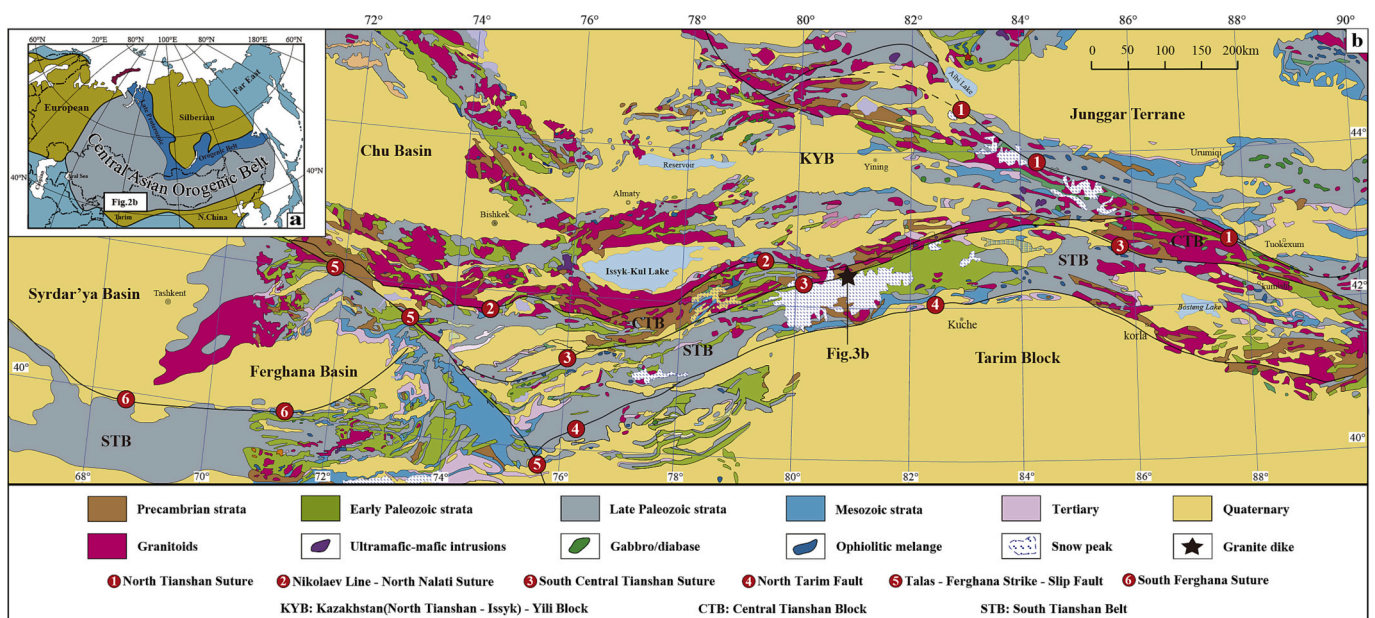


Fig. 2. (a) Tectonic sketch map of the Central Asian Orogenic Belt showing the location of the South Tianshan Belt (modified from Gao et al., 2011; Yakubchuk, 2004); (b) Geological map of the South Tianshan Belt (modified from Alexeiev et al., 2015; Gao et al., 2011).

metamorphic basement (Gao et al., 1998), and Carboniferous island-arc type volcanics unconformably overlying the late Silurian strata or Precambrian basement (Zhu et al., 2005). In addition, Paleozoic granites and diorites crop out extensively in the Central Tianshan block (Gao et al., 2006; Han et al., 2004; Yang et al., 2006; Zhu et al., 2006). The formation of the Central Tianshan block is considered to be related to subduction of the Terskey Ocean during the Neoproterozoic-early Ordovician (Gao et al., 2011), from which MORB-type basalts of 516 ± 7 Ma and adakitic diorites of 470 ± 12 Ma are reported (Qian et al., 2009). It is also associated with the subduction of the South Tianshan Ocean during the Silurian-late Carboniferous, with abundant tholeiitic to calc-alkaline I-type granitoids with an age of 430–330 Ma being reported (Gao et al., 2008; Han et al., 2004; Shi et al., 2007).

At the northern margin of the Tarim Craton, the bedrock is less well exposed and consists mainly of gneisses, leptynite, quartz schists and sandstones (Gao et al., 2009). The Sm–Nd model age of granitic gneisses is 1.9 Ga, and the zircon U–Pb age is 707 ± 13 Ma (Chen et al., 2000c). Granulite xenoliths from the mantle and lower crust exhibit zircon U–Pb ages of 770–690 Ma and Hf model age of 1700–1300 Ma (Zheng et al., 2006a, 2006b). The lower Paleozoic strata consist of upper Ordovician marble, clastic rocks and Silurian clastic rocks and limestones (Allen et al., 1993; Carroll et al., 1995). The upper Paleozoic strata consist of lower-middle Devonian limestones, siliceous rocks, intermediate-basic volcanic rocks, middle-upper Carboniferous carbonate rocks and clastic rocks (Gao et al., 2009). Silurian-early Carboniferous ophiolitic klippe were thrust over the Paleozoic strata at the northern margin of the Tarim Craton (Gao et al., 1998). Extensive granite intrusions occur in the Paleozoic strata, some of which are early Paleozoic in age, including 418–387 Ma granodiorite (Zhu et al., 2008), 475 Ma granite, 457 Ma granitic gneiss (Han et al., 2004) and 475 Ma granodiorite (Hopson et al., 1989), whereas some intrusions are Permian in age (Konopelko et al., 2008). Silurian-Devonian granitoids from Korla city show geochemical evidence that they formed in a continental arc setting (Gao et al., 2012; Ge et al., 2012; Jiang et al., 2001; Lin et al., 2008b).

The South Tianshan Belt, representing the suture zone between the Central Tianshan Block and the Tarim Craton, consists of Paleozoic ophiolites and HP-LT metamorphic rocks, indicating that an ancient ocean once existed in this region: the South Tianshan Ocean (STO) (Ai et al., 2005; Gao et al., 1999; John et al., 2008), which is a part of the broader Paleo-Asian Ocean. Gabbro zircon ages of 600–590 Ma (Yang et al., 2005) and granulite zircon ages of 640–452 Ma (Wang et al., 1998, 1999; Zhou et al., 2004) suggest that the STO already existed in the Precambrian-early Paleozoic. A huge number of fossils, including Devonian-Carboniferous radiolarians and conodonts (Liu, 2001; Tang et al., 1995a), together with late Permian terrestrial-facies molasses (Li and Wang, 2000; Lu et al., 2001), and paleomagnetic evidence (Bazhenov et al., 2003; Wang et al., 2007) indicate that the evolution of STO may have lasted into the late Paleozoic. The HP-LT metamorphic rocks, e.g., greenschist, blueschist and eclogite, indicate that the area experienced intense plate subduction (Ai et al., 2005; Gao and Klemd, 2003; Li et al., 2002, 2005). The eclogites were formed by the dehydration of blueschists at depth (Gao and Klemd, 2001; John et al., 2008), with prograde metamorphic conditions of 480–580 °C at 1.9–2.1 GPa (Klemd et al., 2002; Wei et al., 2003). Multiple geochronological evidence from the South Tianshan Belt indicates that final collision occurred in the late Paleozoic, with Sm–Nd isochron ages of 319 ± 4 Ma and 343 ± 44 Ma for gabbro (Gao and Klemd, 2003; Hegner et al., 2010), ^{40}Ar – ^{39}Ar ages of 316 ± 3 Ma and 331–310 Ma for phengite, 344 ± 3 Ma for crosstie, and 327–324 Ma for phengite and glaucophane (Gao and Klemd, 2003; Gao et al., 2011; Klemd et al., 2005; Simonov et al., 2008; Wang et al., 2010). There is also a U–Pb age of 320 Ma from metamorphic zircon rims (Su et al., 2010). Exotic blocks of granulite and remnants of blueschists are preserved in the ophiolite matrix, and their peak metamorphic ages are 364 ± 5 Ma (Wang et al., 1998) and 360 ± 2 Ma (Liu and Qian, 2003), respectively. In addition, the stratigraphic section in

the central part of the South Tianshan Belt preserves abundant Silurian-Carboniferous foraminifera, corals, conodonts, brachiopods, and amphipora fossils (Alexeiev et al., 2015, 2019).

3.2. Formation age of the South Tianshan Belt

Previous studies considered that the STO eventually closed in the late Devonian-early Carboniferous (Allen et al., 1993; Bai et al., 2015; Charvet et al., 2007; Wang et al., 2010, 2011a; Xia et al., 2008, 2012; Xu et al., 2013b). However, ophiolites with ages of 430–330 Ma present in the South Tianshan Belt suggest that it was not completely closed at this time (Jiang et al., 2014; Li et al., 2015; Xu et al., 2006). Some researchers consider that the STO may have lasted until the late Permian-early Triassic, because eclogites from a high-pressure metamorphic belt in South Tianshan recorded U–Pb zircon rim ages of 230–220 Ma (Zhang et al., 2005a, 2007a), and late Permian radiolarian fossils were also reported from this region (Li et al., 2002, 2005). However, other researchers argued that the occurrence of younger zircons may be due to fluid-mediated recrystallization of zircon grains (Jong et al., 2009; Rasmussen, 2005; Su et al., 2010) or rejuvenation caused by intra-continental deformation (Gilotti and McClelland, 2007). This is supported by the absence of eclogite-facies minerals in the rim domain of these zircon grains (Gao et al., 2009). The occasional presence of young zircons may record the superposition of Triassic tectonic events (Alexeiev et al., 2019), and Zhang et al. (2013) subsequently revised the metamorphic age of the eclogite to 320 ± 4 Ma, consistent with the widely reported eclogite ages of 320–310 Ma elsewhere in the region (Klemd et al., 2011; Li et al., 2011a; Su et al., 2010). Shu et al. (2007) observed that the late Permian radiolarian fossils were poorly preserved, and cannot be accurately identified to the species level, so they cannot be used as the basis for determining the closure time of the STO. Alexeiev et al. (2019) consider that it is unreliable to constrain the age by single radiolarians, because the same rock contains numerous well-preserved radiolarian fossils of the same genus, which were re-deposited in the early Carboniferous, and there are no reliable data on foraminifera, conodonts or other representative groups of fossils younger than the Serpukhovian-Bashkirian. Han et al. (2011) suggested that tectonic activity was relatively quiet during the Mesozoic, and that the STO is unlikely to have existed into the Triassic.

The balance of evidence shows that most workers now consider that the South Tianshan Belt formed in the late Carboniferous (Biske and Seltmann, 2010; Coleman, 1989; Gao and Klemd, 2003; Gao et al., 1998, 2008; Su et al., 2010; Zhou et al., 2001). Furthermore, reports of terrestrial-facies molasse in Permian strata (Fig. 3a; Gao et al., 1998; Li and Wang, 2000; Lu et al., 2001; Wang et al., 1990; Wang et al., 1994) and paleomagnetic data (Bazhenov et al., 2003; Wang et al., 2007) indicate that the Tarim Craton had merged with the Kazakhstan-Yili Block by the early-middle Permian. Li et al. (2006b) considered that the Tianshan Mountains region was under an extensional setting during the late Carboniferous, so that during the Permian it was undergoing regional extension and strike-slip faulting. The presence of intrusive and volcanic rocks supports this view, for example, the volcanic rocks show bimodal features, with basic dykes widely developed in the area and the intrusive rocks are mostly small stocks. What is more, calc-alkaline volcanics and granitoids with ages of 315–300 Ma are reported from the Central Tianshan block, which have distinct affinities with typical subduction-related rocks. During the late Carboniferous-early Permian, a large number of post-collisional intrusions have been reported in the South Tianshan Belt, including high-potassium I-type, A-type granites and rare peraluminous S-type granites, with magmatic activities culminated between 298 and 280 Ma, reflecting the upwelling of large-scale magmas under the control of trans-crustal strike-slip motions after the closure of the STO (Seltmann et al., 2011). Biske and Seltmann (2010) reported post-collisional granitic magmatism occurring during the early Permian, represented by many magma types from A-type granites to nepheline syenites, which were concentrated in the narrow

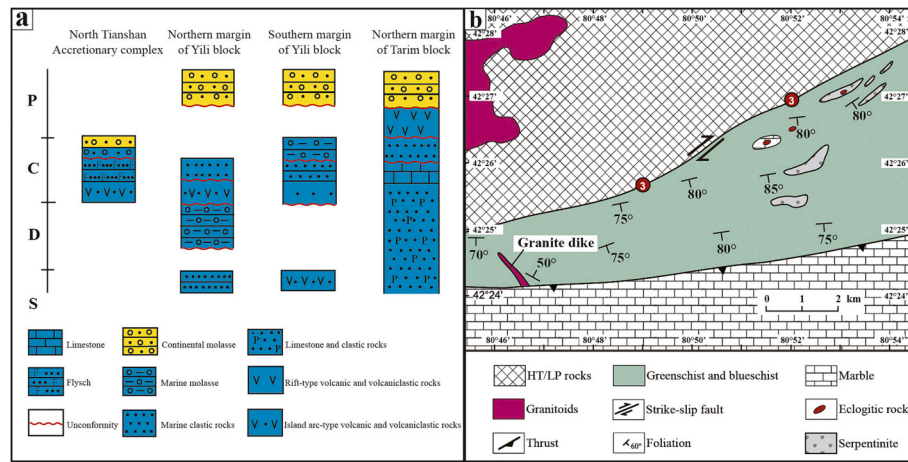


Fig. 3. (a) Schematic stratigraphic diagram showing the late Paleozoic strata in the North Tianshan accretionary complexes, the northern margin of the Yili Block, the southern margin of the Yili Block and the northern margin of the Tarim Craton (modified from Gao et al., 1998; Wang et al., 1990); (b) Geological map showing the occurrence of a granite dike cross-cutting the South Tianshan Belt (modified from Gao et al., 2011).

time span from 300 to 270 Ma and were influenced by post-collisional transtension. Han et al. (2016) also noted that Carboniferous shallow-marine carbonate successions are overlain by Permian siliciclastic-volcanic sequences in the Tarim and South Tianshan areas, and they appear to record the transition from a remnant ocean to lacustrine or fluvial conditions during the late Carboniferous-Permian. Zircon $\epsilon_{\text{Hf}}(t)$ data from the Central Tianshan block show a trend towards negative values at ca. 310 Ma, which most likely reflects the tectonic transformation from oceanic subduction to continental collision. In addition, the distribution of turbidites within the South Tianshan Belt can also provide an age constraint. The age range of turbidites within the foreland basin located at the northern margin of the Tarim Craton is from the Moscovian to Asselian. In addition, the overlying siliciclastic rocks are no younger than Asselian (Biske and Seltnann, 2010). Therefore, we conclude that the complete closure of the STO occurred in the late Carboniferous, when the regional tectonic setting changed from convergent collisional tectonics to regional extension.

Gao et al. (2011) found a granite dike that crosscuts the HP-LT metamorphic assemblage in the South Tianshan Belt recorded a zircon U—Pb age of 285 Ma (Fig. 3b). Geochemical data show that this granite dike has a strongly peraluminous bulk rock chemistry and likely formed in a post-collisional setting. In the central area, alkaline gabbro with an age of 284 Ma and A-type granite with an age of 280 Ma intruded into the Paleozoic strata of the South Tianshan (Seltnann et al., 2011). At the northern margin of the Tarim Craton, granites and alkaline gabbro with ages of 275–262 Ma cut through deformed marine turbidites and other sediments (Han et al., 2011; Li and Xu, 2007). In addition, abundant shallow intrusive rocks were developed in this region with ages ranging from 306 Ma to 250 Ma, which also indicate an extensional environment during the late Carboniferous-early Permian (Zhao et al., 2003). From the Sm—Nd isochron age of 343 ± 44 Ma for eclogite and the $^{40}\text{Ar}/^{39}\text{Ar}$ age of 345–340 Ma for glaucophane, we can deduce that the peak of metamorphism was in the early Carboniferous, therefore the time of collision should be later than this (Gao and Klemd, 2003; Gao et al., 2006). The $^{40}\text{Ar}/^{39}\text{Ar}$ ages for white mica of 335–310 Ma (Gao et al., 2006) and Rb—Sr whole rocks ages of 313–302 Ma (Klemd et al., 2005) represent late collision or early post-collision. Zircon core ages of 291 Ma for rodingite in the Changawuzi ophiolite (Li et al., 2003, 2010) may represent the time of retrogression from former eclogite, meaning that the peak eclogite-facies metamorphism occurred earlier than the Permian (Gao et al., 2006).

3.3. Subduction polarity of the South Tianshan Belt

There are different views concerning the subduction polarity of the South Tianshan Belt. Some researchers claim that subduction occurred in a southward direction based on regional structural features, such as an unconformity with the overlying late Paleozoic strata (Charvet et al., 2007, 2011; Lin et al., 2013; Wang et al., 2011a). Other workers advocated that northward subduction occurred in this region (Biske and Seltnann, 2010; Gao et al., 2008; Han et al., 2011; Han et al., 2016; Windley et al., 1990; Xiao et al., 2013), whereas Alexeiev et al. (2015) suggested that a diachronous and bilateral subduction may have occurred.

Based on the following evidence, we support the view of northward subduction. The South Tianshan Belt separates the Kazakhstan-Yili Block to the north from the Tarim Craton to the south. The Tarim Craton underwent south-to-north directional compression during the middle Paleozoic, resulting in strong stress modification on both its southern and northern margins, where the tectonic deformation and the migration of the various facies in the north are considered to be a response to the tectonic events of the South Tianshan Belt (He et al., 2016). During the Carboniferous, the northern margin of the Tarim Craton is considered to be a passive continental margin (Gao et al., 2009, 2011; Han et al., 2011), with the southern part of the South Tianshan Belt overlain by a succession of Paleozoic sedimentary strata (Allen et al., 1993; Carroll et al., 1995). At the same time, the southern margin of the Kazakhstan-Yili Block was an active continental margin, with accretionary complexes (Gao et al., 2009) and arc magmatic rocks, especially calc-alkaline magmatic arcs rocks (Han et al., 2016; Xiao et al., 2013), widely distributed, and followed by LP-HT granulite facies metamorphism (Lü et al., 2012; Zhang et al., 2013). Hence, the subduction direction of the Southern Tianshan Ocean was northward. There is also a magmatic belt formed by the northward subduction of the STO along the southern margin of the Yili-Central Tianshan Block, which is reported to contain arc plutons formed during the middle Silurian-early Carboniferous (Gao et al., 2008; Long et al., 2007; Qian et al., 2009; Xu et al., 2010; Yang and Zhou, 2009). During the collision, the accretionary complex on the southern margin of the Kazakhstan-Yili Block and ophiolitic mélanges were thrust over the passive margin sediments on the northern margin of the Tarim Craton (Han et al., 2011).

4. The Beishan Belt

4.1. Regional geology

The Beishan Belt forms the middle segment of the South Tianshan-Solonker Suture. It separates the Tuva-Mongolia Block to the north from the Dunhuang Block to the south. To the west, it is separated from the Central Tianshan Block by the Xingxingxia Fault, whereas its eastern margin with the Alxa Block is obscured by the Badain Jaran Desert (Fig. 4; Liu and Wang, 1995; Song et al., 2014; Xin et al., 2020). This belt was previously considered to be the eastern extension of Central Tianshan Block (Li, 1980; Liu and Wang, 1995), however, many workers now consider that Beishan should be a separate geological unit. It is composed of accretionary complexes (Fang et al., 2015; Liu et al., 2015; Song et al., 2013a, 2013b; Xiao et al., 2010b) that record a long process of subduction and accretion since the Cambrian (Ao et al., 2011; Song et al., 2015; Xiao et al., 2010c), and is a typical multicycle orogenic belt (Gong et al., 2002; Nie, 2002). Because of its position, it is a key suture zone for studying the closure of the Paleo-Asian Ocean.

A large number of ophiolites are distributed within the Beishan Belt (Wang et al., 2021a; Xiao et al., 2010c; Xin et al., 2020; Yang et al., 2010; Zuo et al., 1990b), and can be divided into four major ophiolitic mélange units, which are from north to south: the Hongshishan mélange, the Xingxingxia-Shibanjing mélange, the Hongliuhe-Xichangjing mélange, and the Liuyuan mélange (Fig. 4). They are

surrounded by magmatic arcs formed during the progressive closure of the Paleo-Asian Ocean, and named from north to south (Fig. 5; Xiao et al., 2010c): the Queershan arc, the Heiyingshan-Hanshan arc, the Mazongshan arc, the Shuangyingshan-Huaniushan arc, and the Shibanshan arc.

The Hongshishan mélange is mainly composed of marine carbonate rocks, siliciclastic rocks and basal ophiolite. The most typical ophiolites are located in the Hongshishan area (Fig. 4), and consist of cumulate gabbro, gabbro, and basalt, with some calc-alkaline volcanic rocks (Zuo et al., 1990a, 1990b, 2003). Geochemical characteristics indicate the basalts formed at a mid-ocean ridge, whereas the calc-alkaline volcanic rocks formed at an active continental margin (Huang and Jin, 2006a, 2006b; Wei et al., 2004). Based on zircon U–Pb ages (280–270 Ma; Ao et al., 2010; Jiang et al., 2006; Li et al., 2009a), the Hongshishan mélange is considered to have formed in the late Carboniferous-Permian (Gong et al., 2002; Huang and Jin, 2006a, 2006b; Wei et al., 2004).

The Xingxingxia-Shibanjing mélange is mainly composed of meta-ultramafic rocks, gabbro, basalt, with some clastic rocks. Ophiolitic rocks are mainly distributed in the Shibanshan and Xiaohuangshan areas (Fig. 4), and most of them have generally undergone amphibolite facies metamorphism (Zuo et al., 1990a, 1990b, 2003; Nie, 2002). Recent studies of the gabbros suggest that they may be part of an early Carboniferous ophiolitic mélange belt (Li et al., 2012a; Wu et al., 2011b).

The Hongliuhe-Xichangjing mélange mainly consists of Cambrian-

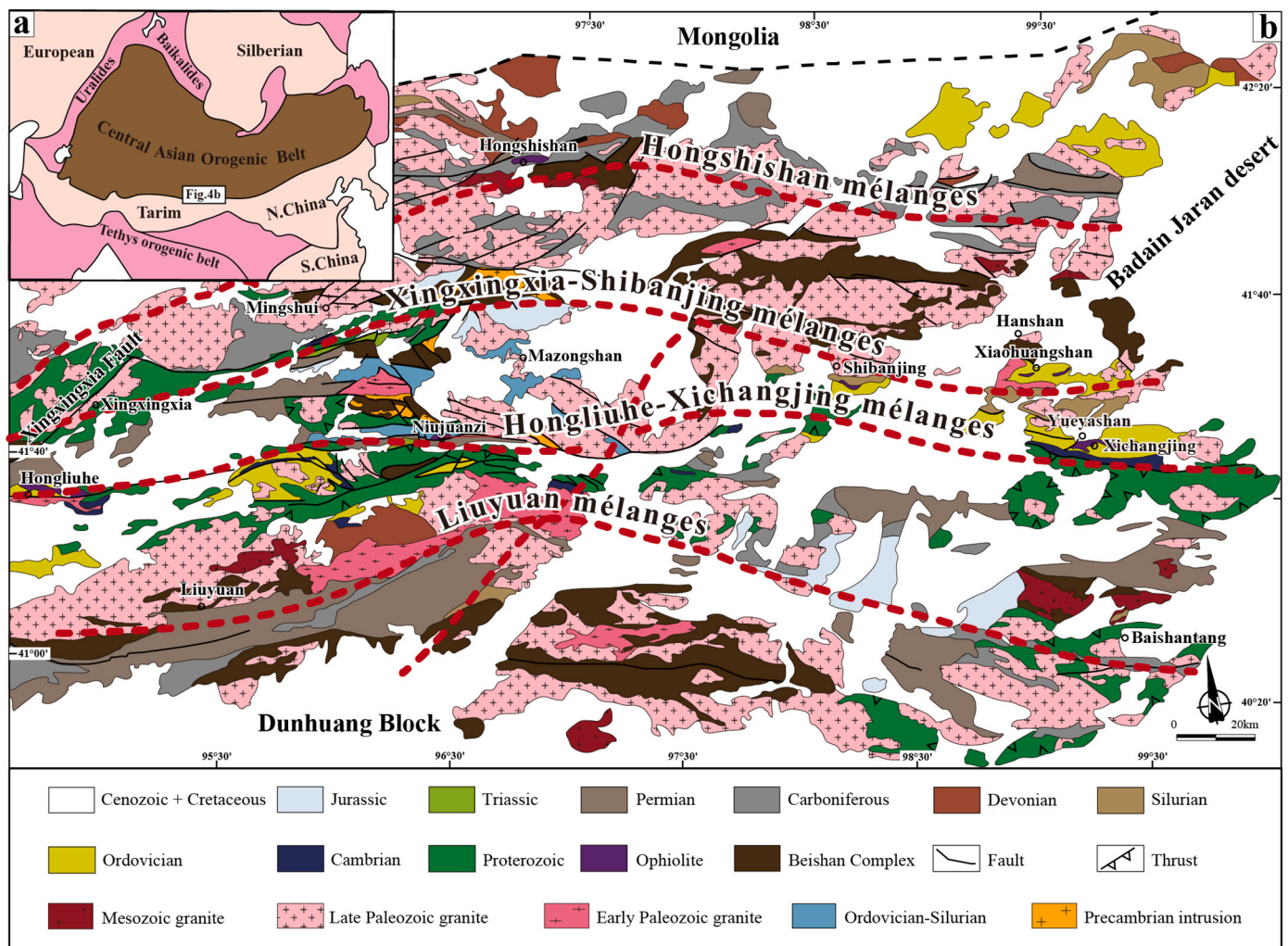


Fig. 4. (a) Tectonic sketch map of the Central Asian Orogenic belt showing the location of the Beishan belt (modified from Wang et al., 2021b); (b) Geological map of the Beishan belt, showing the four major ophiolitic mélanges units (modified from Song et al., 2013b, 2014).

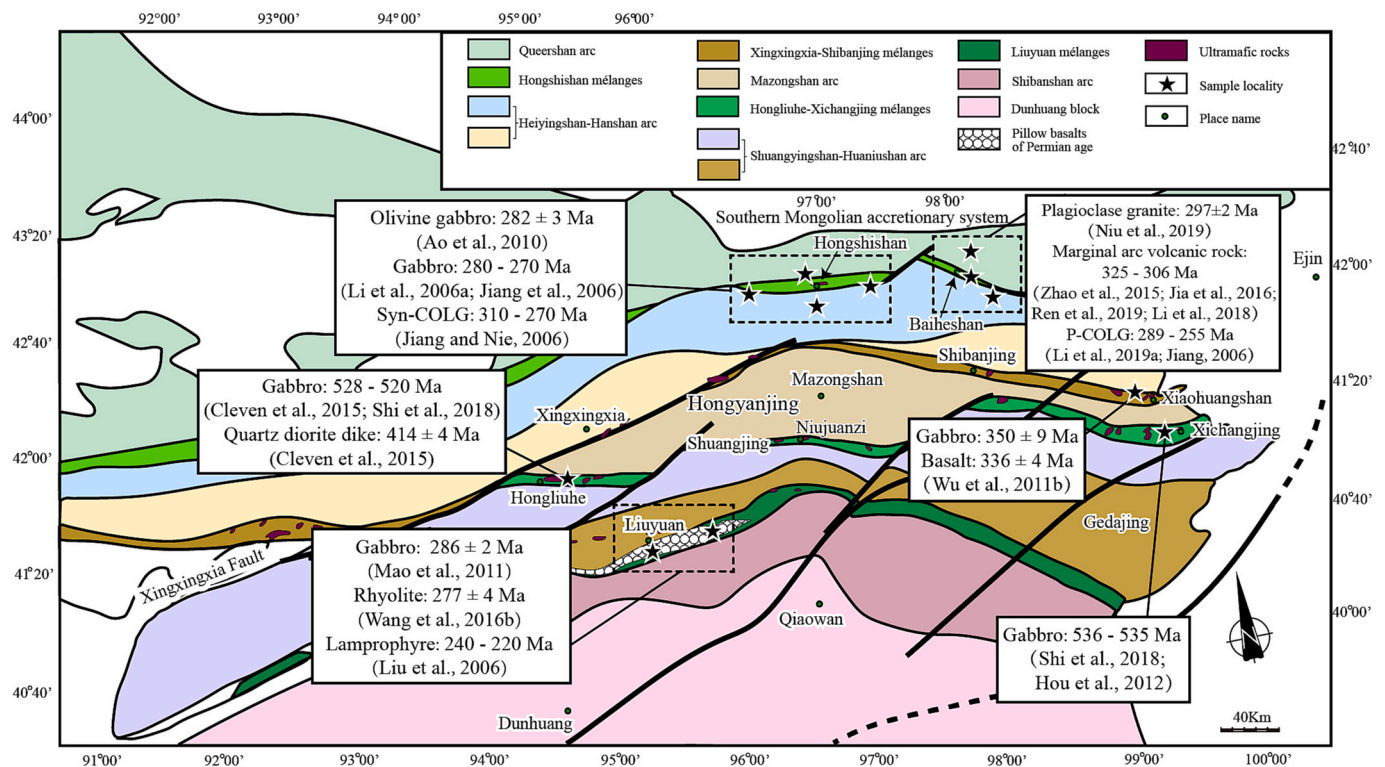


Fig. 5. Simplified tectonic map of the Beishan belt, showing location of representative geochronological data (modified from Xiao et al., 2010c, 2018).

Silurian volcanic clastic rocks, early Paleozoic ophiolites and late Paleozoic conglomerates and sandstones. The ophiolites are mainly exposed around Hongliuhe, Niujuanzi and Xichangjing (Fig. 4), and the lithologies consist of ultramafic intrusive rocks, basalt and siliciclastic rocks (Xiao et al., 2010c). The gabbro age of 536–520 Ma indicates that the Hongliuhe-Xichangjing mélangé formed in a paleo-ocean that opened in the early Cambrian (Cleven et al., 2015; Shi et al., 2018).

The Liuyuan mélangé mainly consists of peridotite and pillow basalt, with some mafic dikes. Geochemical characteristics show that the basalts and gabbros formed in a fore-arc setting (Mao et al., 2012; Xiao et al., 2010c). The Liuyuan mélangé is the southernmost mélangé unit in the Beishan area, and zircon ages of 286 ± 2 Ma obtained from the gabbro indicate that it formed in the Permian (Mao et al., 2012; Xiao et al., 2010c, 2015b).

4.2. Closure of the middle section of the Paleo-Asian Ocean

The four mélangé units in the Beishan area vary greatly in terms of their formation ages, and different workers have different views about the closure time and location of the Beishan Ocean (a branch of the Paleo-Asian Ocean). Zuo et al. (1990b) suggested that the Xingxingxia-Shibanjing mélangé was formed in the Ordovician-Silurian based on evidence from the Sinian-Ordovician surrounding rocks, fossil evidence from volcano sedimentary rocks, and the fact that the Beishan Belt was already formed in the early Paleozoic. However, the latest isotopic ages for gabbro and basalt indicate that they formed in the early Carboniferous (Fig. 5; Wu et al., 2011b; Li et al., 2012a), Yang et al. (2010) suggested that the mélangé may have formed in a back-arc basin. The ages of gabbro in the Hongliuhe-Xichangjing mélangé are 536–520 Ma (Fig. 5; Cleven et al., 2015; Shi et al., 2018), indicating that a paleo-ocean was already open in the early Cambrian. However, the radiolarian fossils obtained from the siliceous rocks are no younger than the Devonian (He et al., 2004), suggesting that oceanic subduction ended before the middle Devonian (Wang et al., 2021b). The age of an undeformed quartz diorite dike intruding the overlying strata of the

Hongliuhe-Xichangjing mélangé is 414 ± 4 Ma, suggesting that the closure of the paleo-ocean was prior to the early Devonian (Cleven et al., 2015). Therefore, the Hongliuhe-Xichangjing mélangé is generally considered to be an early Paleozoic ophiolite belt (Sun et al., 2017; Yang et al., 2008, 2010). However, it is evident that the complete closure of the middle section of the Paleo-Asian Ocean could not have ended in the early Paleozoic. From the regional tectonic perspective of the Beishan area, with the South Tianshan Belt on its western side and the Solonker Belt on its eastern side, all the ophiolites are considered to be the result of progressive oceanic subduction during the late Paleozoic-early Mesozoic (Gao et al., 1999; Gao and Klemm, 2001, 2003; Xiao et al., 2003, 2004, 2008, 2009a, 2010a, 2010b; Zhang et al., 2005b, 2007a), so it would be difficult to explain this if the Beishan Ocean closed in the early Paleozoic. In addition, from the paleomagnetic data, it is evident that a remnant ocean basin still existed in the Beishan region during the early Permian (Xu et al., 2019a). Therefore, more and more researchers now consider that the Beishan Belt experienced multiple accretionary and collisional processes during the Paleozoic (Guo et al., 2014; Song et al., 2013a, 2013b; Tian et al., 2014; Xiao et al., 2010c), and that the Hongshishan and Liuyuan mélanges may therefore represent the youngest suture related to closure of the Beishan Ocean (Xiao et al., 2010b).

4.3. Formation age of the Beishan Belt

The Hongshishan mélangé lies on the north side of the Beishan Mountains (Fig. 5), and represents a suture zone formed during subduction of the Beishan Ocean. The area has a huge gravity lineament and magnetic anomaly belts, consistent with the geophysical characteristics of a suture zone (He et al., 2002). It also marks the boundary between Cathaysian flora to the south and Angara flora to the north; the paleontological characteristics of a suture zone (Wang and Cheng, 2001). The gabbros show negative anomalies in high field strength elements (Ao et al., 2010), typical characteristics of arc magmas (Pearce and Peate, 1995), and most of the gabbros are calc-alkaline arc (Ao et al., 2010), suggesting that the Hongshishan mélangé formed in an island arc

environment.

The Hongshishan mélange is generally considered to have formed during the late Carboniferous-Permian (Gong et al., 2002; Huang and Jin, 2006a, 2006b; Wei et al., 2004). The zircon weighted mean age of an olivine gabbro is 282 ± 3 Ma, and the geochemical characteristics indicate that it formed in an island arc setting (Ao et al., 2010), with ophiolites in the area recording ages of 280–270 Ma (Jiang et al., 2006; Li et al., 2009a). The age of plagioclase granite from the ophiolite mélange is 297 ± 2 Ma (Niu et al., 2019). Based on the view that the age of the youngest ophiolite mélange in an orogenic belt represents the time of final closure of an ancient ocean (Li, 2004; Xiao et al., 2019), the formation of the suture zone was most likely in early-middle Permian. When subduction of the Beishan Ocean entered its final phase in the Carboniferous-Permian, it led to the generation of massive granites and intermediate-felsic volcanic rocks in this region. The zircon U–Pb weighted mean ages are 306 ± 1 Ma for quartz diorite (Zhao et al., 2015), 319 ± 1 Ma for rhyolite (Jia et al., 2016), 325–313 Ma for intermediate-felsic volcanic rocks (Ren et al., 2019), and 310 ± 1 Ma and 306 ± 1 Ma for granodiorite and monzogranites (Li et al., 2018), which, together with the geochemical characteristics of these rocks, indicate that they all formed in a continental arc setting. Hence, oceanic subduction was still occurring in the Beishan area during the late Carboniferous (Jia et al., 2016; Li et al., 2018; Ren et al., 2019; Zhao et al., 2015). The ages of syn-collisional granite in the Hongshishan area range from 310 Ma to 270 Ma (Jiang and Nie, 2006). The age of alkali feldspar granites here is 289 ± 1 Ma (Li et al., 2019), and for granodiorite and syenogranite it is 262 ± 4 Ma and 255 ± 4 Ma, respectively (Jiang and Nie, 2006). These rocks all formed in a post-collisional setting. Therefore, the final closure of the Beishan Ocean occurred before the middle Permian.

The Liuyuan mélange lies on the south side of the Beishan Mountains (Fig. 5), with the surrounding area covered by early Permian turbidites that formed in an oceanic island arc and at an Andean continental margin setting (Guo et al., 2012, 2013). Pillow basalts are cut by lamprophyre dikes whose ages range from 240 Ma to 220 Ma, representing the transition of the magma source from depleted mantle to enriched mantle, related to crustal thinning (Liu et al., 2006). In this region, the zircon U–Pb ages of gabbros are 286 ± 2 Ma (Mao et al., 2012), whose chemistry is consistent with rocks formed in a supra-subduction zone setting. An overlying rhyolite records an age of 277 ± 4 Ma (Wang et al., 2016b), with these volcanic rocks unconformably overlain by late Permian-early Triassic sandstone (Wang et al., 2016b). The Permian magmatic and clastic rocks show features characteristic of a subduction zone (Song et al., 2018). Therefore, the Liuyuan mélange was also formed during the early-middle Permian (Mao et al., 2012; Xiao et al., 2010c, 2015b). The geochemical features of mafic rocks in this belt match those of MORB, and likely formed in a fore-arc setting as part of the ophiolite suite (Mao et al., 2012; Xiao et al., 2010c, 2015a).

However, some controversial views have emerged recently. Based on studies of Permian ultramafic-mafic volcanic rocks, some researchers advocate for a Permian post-collisional setting of the Beishan Belt (Chen et al., 2016; Zhang et al., 2011). Liu et al. (2019) reported that the Permian clastic deposits on both sides of the Liuyuan mélange record a thick sequence of conglomerate at the base and thin-bedded turbidite sequences in the upper section, they may have resulted from a progressively deepening of a basin within an extensional post-collision regime. These controversies arise mainly from controversial, multi-explanatory geochemical and provenance analyses of magmatic and sedimentary rocks. Song et al. (2016b) reported peak ages of 276 Ma and 286 Ma for detrital zircons in metamorphic complexes of the Beishan Belt, and they both show positive $\varepsilon_{\text{Hf}(t)}$ values, indicating significant crustal growth in the Permian. Furthermore, all basalts in the Liuyuan mélange are thrust imbricated against fossiliferous Permian sediments (Mao et al., 2012). Rocks in the Beishan Belt were therefore deformed in a compressional tectonic regime. If the Permian tectonic setting of the Beishan Belt is simply considered extensional, some of the late

Carboniferous-Permian geological components cannot be explained (Song et al., 2016b). Therefore, some scholars tend to interpret it as a local extension under an overall subduction setting (Sun et al., 2020a, 2020b; He et al., 2018; Zheng et al., 2020, 2021).

Paleostratigraphic, paleontological, and paleomagnetic evidence can also provide a constraint on the closure time of the Beishan Ocean. For example, the lower-middle Permian strata in the northern region are littoral-neritic facies belonging to the Shuangputang Formation, whereas the middle Permian strata are terrestrial volcanic rocks of the Jinta Formation (Wei et al., 2020). Similarly, the lower-middle Permian strata in the southern area are marine strata of the Shuangputang Formation, and the upper Permian strata are terrestrial volcanic rocks of the Fangshankou Formation (Xu et al., 2019b). The northern margin of the Dunhuang Block, dominated by Cathaysian flora, began to mix with northern Angara flora from the middle Permian (Fang, 1997; Guo, 2001; Li et al., 1999; Ou et al., 1993, 2004; Wu, 1993; Zhu, 1997, 2001). The landmass movement trajectory inferred from paleomagnetic data shows that there was still a latitudinal difference between the Dunhuang Block and the Tuva-Mongolia Block in the early Permian, suggesting that the closure occurred after the early Permian (Li et al., 1999; Xu et al., 2019a).

The microcontinental Alxa Block is situated at the border between the Beishan Belt and the North China Craton. Two ophiolitic mélange belts are distributed along its northern margin, and they are referred to as the Engger Us mélange and the Quagan Qulu mélange, extending in a north-east direction along the Engger Us Fault and the Badain Jaran Fault, respectively. They are the suture zones that connect the Beishan Belt to the west with the Solonker Belt to the east, and represent the suture zone and the post-arc basin formed during the closure process of the Paleo-Asian Ocean (Zheng et al., 2014). The Engger Us mélange is mainly composed of a carbonate matrix that hosts mafic-ultramafic tectonic blocks, including gabbro and basalt. The zircon U–Pb age obtained from the basalt is 302 ± 14 Ma (Zheng et al., 2014). Xie et al. (2014) reported the youngest Permian radiolarian fossils from this unit, whose ages range from the middle Permian to the early late Permian. The Quagan Qulu mélange is mainly composed of a micro-clastic matrix enclosing tectonic blocks, with the latter consisting of ultramafic rocks, gabbros and cherts. The zircon U–Pb age obtained from boninitic rocks in this mélange is 275 ± 3 Ma (Zheng et al., 2014). In addition, Permian magmatic activity, which peaked at ca. 276 Ma, was identified in the northern margin of the Alxa Block (Chen et al., 2021b). Adakitic granitic rocks with ages of 268–255 Ma indicate that the continent/arc collision occurred during the middle-late Permian (Hui et al., 2021). Hence, we consider that the progressive closure of the Paleo-Asian Ocean at the northern margin of the Alxa Block continued into the early late Permian.

4.4. Subduction polarity of the Beishan Belt

We consider that the subduction polarity of the Beishan Belt was to the north. According to previous research reports, oceanic crust, a fore-arc accretionary complex, and magmatic arcs in the Hongliuhe-Xichangjing area are sequentially distributed from south to north (Song et al., 2018). All of them were formed between 446 and 442 Ma, and are the remnants of northward subduction (Song et al., 2014, 2015). Ductile deformation in the middle part of the Beishan Belt also become progressively younger from north to south, which is considered to be caused by the southward accretion of the orogenic belt during northward subduction (Song et al., 2018). Furthermore, geophysical profiles show that most of the thrust faults within the Beishan region have southward movement, which also supports the conclusion of northward subduction (Xiao et al., 2010c).

5. The Solonker Belt

5.1. Regional geology

The Solonker Belt is located east of the Beishan Belt and extends from the Solonker region eastward to near Changchun. It separates the Siberia Craton to the north from the North China Craton to the south (Fig. 6; Han et al., 2015, 2017a, 2017b; Jahn et al., 2000; Li, 2006; Li et al., 2012b; Natal'in, 1991; Ren et al., 1999b; Xiao et al., 2003). A series of ophiolitic mélanges, radiolarian-bearing cherts, *syn*-collisional granites and some dynamically metamorphosed rocks are distributed along the Solonker Belt, and led most researchers to consider that this belt represents a suture zone associated with the Paleo-Asian Ocean (Li et al., 2007b; Sun et al., 2004; Tang et al., 1983; Wang and Liu, 1986; Wang and Fan, 1997; Wu et al., 2007a). Because the Solonker Belt lies in the eastern part of the CAOB, it is an extremely critical area to explore when the Paleo-Asian Oceanic tectonic domain terminated, and when the Paleo-Pacific domain started to influence East Asia. Based on previous studies (Xiao et al., 2003), the Solonker Belt is located between what has been referred to as the North and South Orogenic belts (Fig. 6). These latter two areas are significant in establishing the evolution of the suture zone.

The North Orogenic belt is bounded by the Hegenshan-Heihe Suture to the north and the Xilinhot Fault to the south (Fig. 6), and is considered to have formed by northward subduction of the Paleo-Asian Ocean (Xiao et al., 2003). In the western Hutag Uul terrane (Fig. 6), garnet-bearing gneisses are in fault contact with blueschist and ophiolitic mélange, with an age of 383 ± 13 Ma for the blueschist (Xu et al., 2001) and 770 Ma for the gneisses (Kozakov et al., 1999), which both formed by accretion related to Paleo-Asian subduction. Along the belt to the east, large areas of amphibolite facies gneisses occur in the Sunid Zouqi and

Xilinhot areas (Fig. 6; Hsu et al., 1991), with ophiolites and blueschists exposed as lenses in late Paleozoic clastic sediments (Wang and Liu, 1986). Large accretionary magmatic arcs are exposed in the Baolidao area, consisting mainly of deformed metaluminous to weakly peraluminous gabbroic diorite, quartz diorite and granodiorite (Chen et al., 2000a). There are many outcrops of ophiolites in the Hegenshan area in the east (Fig. 6), some of which reach 20 km in width, with the thickest part of the units consisting of harzburgite, dunite and basalt (Xiao et al., 2003). The ophiolitic rocks are mainly found with Carboniferous-early Permian strata and were thrust southward over the Mesozoic sediments (Hsu et al., 1991).

The South Orogenic belt is bounded by the Linxi Fault to the north and the Chifeng-Bayan Obo Fault to the south (Fig. 6). The Bainaimiao arc on the southern side formed at an active continental margin before the Mesozoic, and consists of calc-alkaline tholeiitic basalt, alkali basalt, intermediate-felsic lavas and some granite intrusions. To the north, andesite, rhyolite and clastic rocks formed a northward-vergent imbricate zone (Xiao et al., 2003). To the north of the Bainaimiao arc is the Ondor Sum accretionary complex (Fig. 6), which shows a trend of increasing deformation from south to north (Xiao et al., 2003). The southern part consists of undeformed pillow basalt, facing towards the north. In the central part is a weakly deformed remnant island arc consisting of calc-alkaline basalt, andesite, and rhyolite. The northern part consists of mylonites, which have strong mineral lineations that plunge to the north. This indicates that southward accretion occurred in the Ondor Sum region.

The Solonker Belt is about 700 km long and 60 km wide (Fig. 6; Xiao et al., 2003), and consists of accretionary complexes, ophiolites, blueschists, and marbles; it is a typical accretionary wedge. The northern part consists of more highly tectonized rocks, including ophiolites, marbles and arc volcanic rocks (Wang and Liu, 1986), while the southern part is

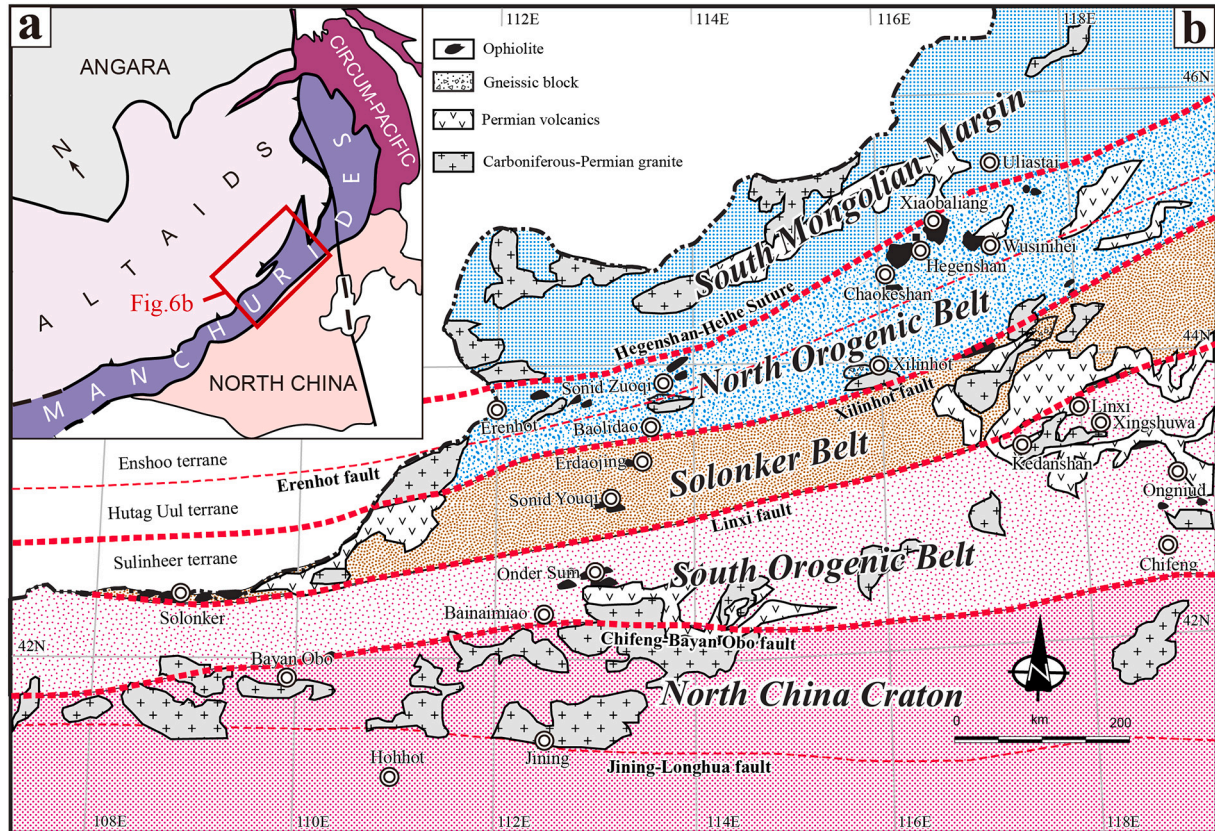


Fig. 6. (a) Tectonic sketch map of the Central Asian Orogenic Belt showing the location of the Solonker Belt (modified from Şengör, 1996); (b) Regional geological map of the Solonker Belt (modified from Xiao et al., 2003).

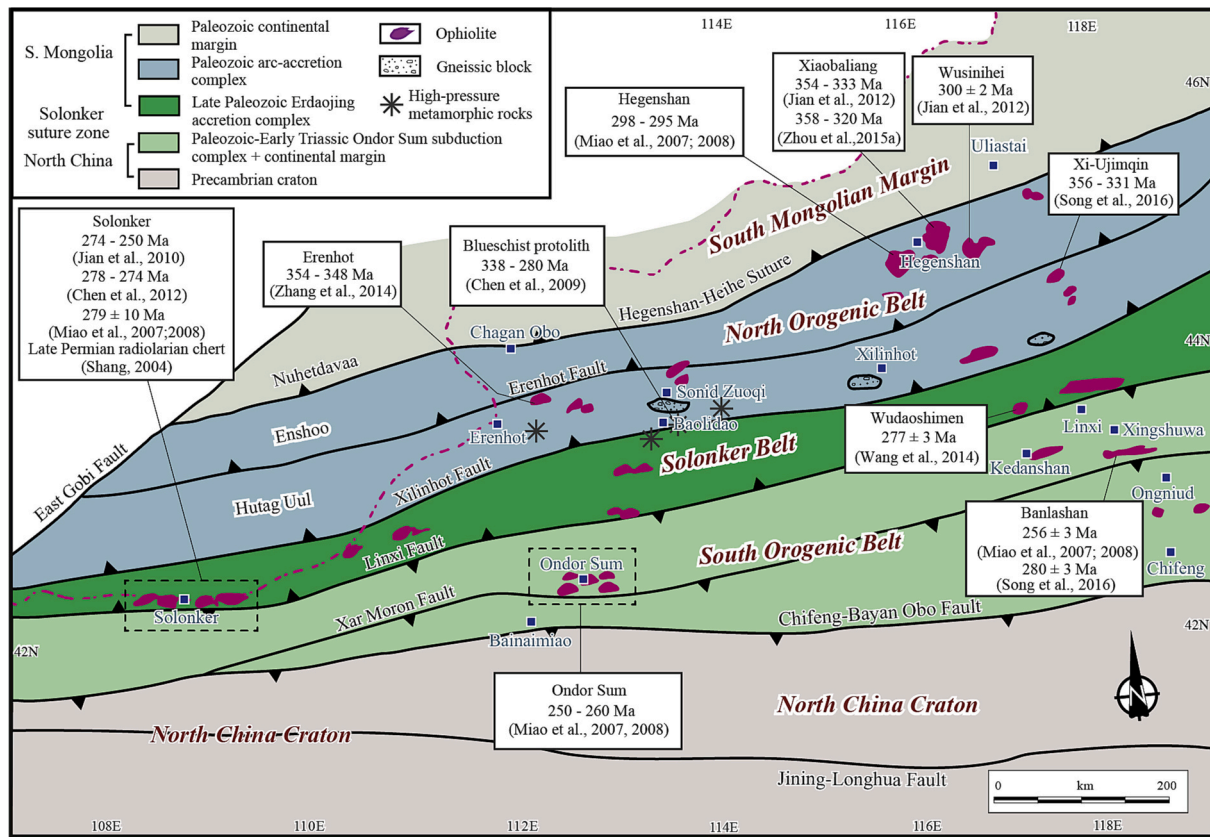


Fig. 7. The tectonic subdivisions of the Solonker Belt, including the location and age of significant lithologies (modified from Song et al., 2016b; Xiao et al., 2003, 2018).

less tectonized and is mainly composed of lenticular mafic-ultramafic rocks encased by an argillite matrix (Xiao et al., 2003). There is a discontinuous ophiolitic belt extending longitudinally from Sunid Youqi to Linxi (Fig. 6; Wang and Liu, 1986). The western part of the ophiolite is mainly composed of harzburgite, dunite, and gabbroic cumulates, with pillow basalt, and overlain by lower Permian clastic sediments. In the eastern part, lenticular ophiolites with thicknesses up to 1.1 km have been reported, which are composed of pyroxenite, layered gabbro, and basalt, and occur within early Permian clastic sediments (Xiao et al., 2003). Mélanges in the Erdaojiang area are tectonic mélanges associated with coherent turbidites that are imbricated with ophiolitic rocks, cherts, marbles, and arc volcanic rocks (Xiao et al., 2018). The mélanges distributed within the Solonker and Sulinheer terrane consist mainly of pillow basalt, chert, tuff and limestone, and the ophiolites are dismembered into serpentinite, dunite and gabbro fragments at the China-Mongolia border (Fig. 6; Xiao et al., 2003).

5.2. Closure of the eastern section of the Paleo-Asian Ocean

There has been considerable controversy about the location of the actual suture zone in the eastern section of the CAO. Some workers suggest that the Paleo-Asian Ocean closed along the Hegenshan-Heihe suture (Fig. 6; Kravchinsky et al., 2002; Mueller et al., 1991; Zhang and Tang, 1989; Zorin, 1999), which was also known as the fault that separates the Songliao Block from the Xing'an Block. Zircon U—Pb ages of 354–333 Ma (Jian et al., 2012), 358–320 Ma (Zhou et al., 2015a), and 298–295 Ma (Miao et al., 2007, 2008) were obtained from ophiolitic rocks and detrital zircon ages of 279 ± 2 Ma (Zhou et al., 2015a) and 308 ± 5 Ma (Pei et al., 2016) were obtained from the overlying strata, indicating that this suture was formed in the Bashkirian-Asselien (Zhou and Wilde, 2013; Zhou et al., 2012b). However, other researchers considered that the final closure of the Paleo-Asian ocean was located

further south in the Solonker Belt (Chen et al., 2009; Jian et al., 2008, 2010; Jong et al., 2006; Li et al., 2006b; Lin et al., 2008a; Miao et al., 2008; Shen et al., 2006; Wang and Liu, 1986; Wang and Fan, 1997; Wu et al., 2002, 2007b; Xiao et al., 2003; Zhang et al., 2007b, 2009c).

It is evident from the literature that there are still numerous reports of Permian ophiolites within the Solonker Belt (Fig. 7; Chen et al., 2012; Jian et al., 2010; Miao et al., 2007, 2008; Song et al., 2016a; Wang et al., 2014a), which are linearly distributed and extend from Solonker through Erdaojiang to Changchun. Radiolarian fossils found in the Xilinhot (Shang, 2004) and Xingshuwa regions (Wang and Fan, 1997) suggested that the Paleo-Asian Ocean was still not fully closed in the Permian. Furthermore, the Solonker Belt also marks the dividing line between Angara flora to the north and Cathaysian flora to the south, and this occurred before the Permian (Sun et al., 2018; Xiao et al., 2015b). Therefore, we consider that the older Hegenshan-Heihe Suture is a local suture zone between the Songliao and Xing'an blocks, and that the younger Solonker Belt is the location marking the final closure of the Paleo-Asian Ocean, separating the two fundamental crustal units, the North China and Siberia cratons.

5.3. Formation age of the Solonker Belt

According to previous research, early Permian marine to shallow marine volcano-clastic strata were well developed in this area (Eizenhöfer and Zhao, 2018; Zhu, 2015). This means that the Paleo-Asian Ocean was not completely closed at this time. In the late Permian, the regional tectonic setting began to change as the northern and southern blocks gradually approached and collided with each other, and northern sediments began to mix with material from the North China Craton (Zhao et al., 2016). During the early-middle Triassic, terrestrial-facies deposits, including lacustrine, fluvial and alluvial fans, were deposited in this region (Li et al., 2014; Zhu, 2015). In the middle-

late Triassic, all sediments in this region were terrestrial (Li et al., 2014), and regional extensional basins were widely developed (Wang et al., 2011b), implying that the Paleo-Asian Ocean was closed at this time. Paleontological evidence shows that in the western region, early-middle Permian radiolarian fossils are present in the Solonker and Bayan Obo regions (Li et al., 2017b; Wang et al., 2005), and in the eastern region, middle Permian radiolaria were found within the Xingshuwa ophiolite (Wang and Fan, 1997), and are also preserved in Permian marine argillaceous sediments in the central region (Shang, 2004). There are no Triassic radiolarian fossils reported in the region. Palaeontological and lithostratigraphical data from the northern margin of the North China Craton and several areas around the Solonker Belt indicate that early-middle Permian strata are characterized by marine fossils such as brachiopods, corals, fusulinoideans, locally associated with ammonoids, conodonts and marine bivalves, while the late Permian Lopingian strata is dominated by fossil plants and some non-marine bivalves and fusulinoids (Shen et al., 2006; and references therein). There is a clear boundary between Angara flora in the north and Cathaysia flora in the south during the middle Permian (Wang et al., 2008), whereas the two biota mixed together before the Triassic (Sun et al., 2018; Xiao et al., 2015b). Paleomagnetic data indicate that during the late Permian, a relative rotation occurred between the Songliao Block and the North China Craton as they approached each other (Ren et al., 2020), and that the collision between the North China and Siberia cratons commenced in the late Permian (Qin et al., 2013).

Eizenhöfer et al. (2014) limited the closure time along the Solonker Belt to 270–240 Ma by determining the age of undeformed intrusions cutting sedimentary rocks and andesitic pyroclastic rocks. Major and trace element data from the sandstones of the Linxi Formation indicate that they have a greywacke protolith and were deposited during a period of intense tectonic activity. The sandstone samples show a youngest peak age of ca. 252 Ma, indicating that they were deposited mainly in the late Permian, hence the final collisional time of the Solonker Belt should be the late Permian (Han et al., 2015). Xiao et al. (2003) further indicated that traces of late Permian Andean-type magmatism and middle-late Triassic S-type granite intrusions (Chen et al., 2000a) are preserved in this region, while the Triassic sedimentary record is lacking (BGMIRM (Bureau of Geology and mineral resources of Inner Mongolia), 1991), which may represent the shift in tectonic setting from subduction to collision, with later uplift preventing deposition of sediments in the area.

A series of collision-related intrusive rocks developed in this region. For example, the zircon U–Pb ages of S-type granites are 272 ± 2 Ma and 265 ± 2 Ma (Li et al., 2009b) in the Linxi area, which are considered to represent the peak of magma intrusion, after which the rocks were transformed into granitic gneiss by deformation and metamorphism under tectonic stresses, and large-scale collisional activities would have occurred later than this. The 229 ± 4 Ma and 238 ± 3 Ma S-type granites from the same area are considered to be produced by re-melting of thickened crust during the post-collisional phase (Li et al., 2007b). Ages of A-type granites are 233–220 Ma, indicating an extensional tectonic setting at this time (Chen et al., 2009; Hu et al., 2015; Shi et al., 2004). High Sr/Y granites are reported to have ages of 264–256 Ma in the central region (Liu et al., 2011) and 251–245 Ma in the eastern region (Li et al., 2017a), and they both represent a post-collisional setting. On the north bank of the Xar Moron Fault, a biotite–plagioclase schist with an age of 298 ± 2 Ma was intruded by granite of 272 ± 2 Ma, indicating that the closure of Paleo-Asian Ocean was later than this (Li et al., 2011b). Post-collisional granites of 249–241 Ma are also developed along this fault, which may be derived from the melting of mixed mafic oceanic crust and subducted sediments, representing the latest time of collision (Zhao et al., 2019). At the northern margin of the North China Craton, the age of the undeformed granitic pluton is 252–248 Ma, and the geochemical characteristics indicate that it was derived from metamorphic mafic rocks in the thickened lower continental crust, with the youngest detrital zircon within the surrounding rocks recording an

age of 274 ± 4 Ma, so 274–252 Ma may represent the timing of collision between the North China Craton and the Mongolian arcs (Wang et al., 2021a). Similar ages of 254–237 Ma are reported for post-collisional granites at the northern margin of the North China Craton (Zhang et al., 2009b), together indicating that the region had changed from a collisional orogenic regime to an extensional one by the early-middle Triassic.

A series of ophiolites are also distributed along the Solonker suture zone (Fig. 7). In the Solonker area, a fore-arc diabase records ages of 274–250 Ma (Jian et al., 2010), a cumulate gabbro has an age of 279 ± 10 Ma (Miao et al., 2007, 2008), and ages of gabbro and basalt are 279 ± 3 Ma and 274 ± 1 Ma, respectively, with geochemical characteristics indicating that their magma source was influenced by plate subduction (Chen et al., 2012). At Ondor Sum, the age of pillow basalt is 260 Ma (Miao et al., 2007). The pillow basalt at Wudaoshimen yields a concordant age of 277 ± 3 Ma (Wang et al., 2014a), and cumulate gabbro and rodingite at Banlashan yield zircon U–Pb ages of 256 ± 3 Ma (Miao et al., 2007, 2008) and 280 ± 3 Ma (Song et al., 2016a), respectively. These data show that the ages of ophiolitic mélanges in the Solonker Belt are mainly distributed between 280 and 250 Ma, indicating the Solonker Belt was accreted between the middle-late Permian.

5.4. Subduction polarity of the Solonker Belt

The subduction polarity of the Solonker Belt is still the focus of debate (Chen et al., 2009; Jian et al., 2008, 2010; Li, 2006; Miao et al., 2008; Song et al., 2016a; Windley et al., 2007; Xiao et al., 2003, 2009a; Xu et al., 2013a, 2014). Some believe it underwent unidirectional subduction to the south, with bimodal volcanic rocks developed on its northern side (Jian et al., 2008, 2010; Li, 2006). Others consider that bidirectional subduction occurred along the Solonker Belt (Eizenhöfer and Zhao, 2018; Eizenhöfer et al., 2014; Xiao et al., 2003, 2018). Based on the following features, we support the second view. First of all, there is no obvious overthrusting and crustal thickening in the area, and there are no reports of high-pressure granulite facies metamorphic rocks (Song et al., 2016a). Hence, the features are more consistent with the ‘Appalachian-type’ collision model in which there was no passive continental margin, and subduction zones are developed on both sides of the oceanic lithosphere (Draut and Clift, 2001, 2013). Secondly, there are many outcrops of mélanges exposed on both sides of the belt, indicating bi-directional subduction (Fig. 7). It is noteworthy that the age of mélanges in the north of the belt is mainly Carboniferous (Chen et al., 2009; Miao et al., 2007, 2008; Pei et al., 2016; Song et al., 2016a; Zhang et al., 2014; Zhou et al., 2015a), while the age of mélanges in the south of the belt is mainly middle-late Permian (Chen et al., 2012; Jian et al., 2010; Miao et al., 2007, 2008; Song et al., 2016a; Wang et al., 2014a). The age difference between the two differs by about 50 Ma, and this may be due to the fact that northward subduction ceased earlier than in the south.

5.5. Possible subduction polarity transition

We identified that a transition in subduction polarity occurred near the Solonker region, with northward subduction in both the South Tianshan Belt and the Beishan Belt, and a change to bi-directional subduction in the Solonker Belt. At the western edge of the Solonker Belt, a giant structural corridor called the East Gobi Fault crosses the China-Mongolia border, and it is considered to be a further northeastward extension of the Altyn Tagh Fault, with strike-slip characteristics (Fig. 1; Darby et al., 2005; Heumann et al., 2014; Johnson and Ritts, 2011; Webb and Johnson, 2006; Yue and Liou, 1999; Yue et al., 2001). This fault is considered to have developed during the amalgamation of the North China and Siberia cratons and controlled a series of intra-continental deformation events that subsequently occurred during the Mesozoic-Cenozoic (Heumann et al., 2014, 2018; Johnson and Ritts, 2011; Webb et al., 2010).

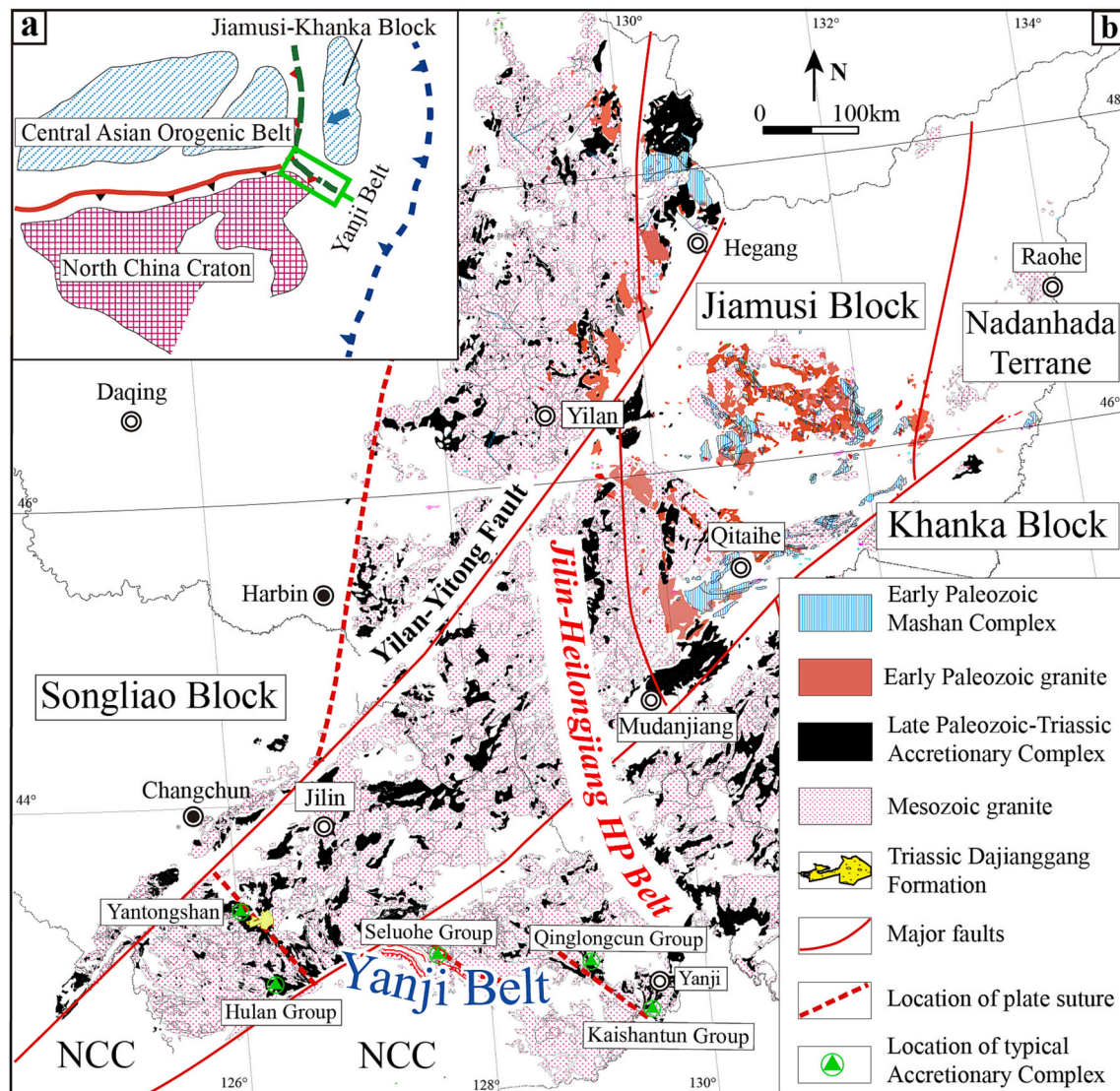


Fig. 8. (a) Tectonic sketch map of the Central Asian Orogenic Belt showing the location of Yanji Belt; (b) Regional geological map of the Yanji Belt and the location of major accretionary complexes (modified from Zhou et al., 2013).

Around the East Gobi Fault, the late Permian fluvial-lacustrine sediments are overlain by strongly deformed and metamorphosed early Triassic turbidite successions, marking sea level decline and ocean closure (Johnson and Ritts, 2011), which is consistent with the view that the middle section of the South Tianshan-Solonker Suture closed in the early-middle Permian. The Paleozoic intermediate-felsic volcanic intrusions were also cut by the East Gobi Fault (Badarch et al., 2002; Heumann et al., 2018), which may be related to the ocean-continent convergence and collision. Hence, we suggest that there is a geotectonic correlation between the East Gobi Fault and the closure of the Paleo-Asian Ocean.

Despite the lack of sufficient geophysical data, the evolutionary history of the East Gobi Fault and its geotectonic attributes suggest that the transition in subduction polarity of the Paleo-Asian Ocean is related to the development of the East Gobi Fault, although additional evidence is required to substantiate this view.

6. The Yanji Belt

6.1. Regional geology

The Yanji Belt is a typical accretionary complex that lies between the

Jiamusi-Khanka Block and the North China Craton, extending from Changchun eastward to the Yanji area (Fig. 8; Zhou and Wilde, 2013). It lies at the eastern edge of Eurasian continent, and was influenced by both the north-south closure of the Paleo-Asian Ocean (Kröner et al., 2014; Şengör, 1996; Şengör et al., 1993; Windley et al., 2007; Wu et al., 2002, 2011a; Zhou and Wilde, 2013; Zhou et al., 2011a, 2011b, 2012a) and the westward subduction of Paleo-Pacific Ocean (Faure and Natal'in, 1992; Maruyama, 1997; Ren et al., 1999a; Tang, 1990; Wilde et al., 2000, 2003; Zhou et al., 2009a, 2010a, 2010c), hence it is a critical region for studying the switch in tectonics between these two oceanic domains.

The Yanji Belt is characterized by a series of accretionary complexes (Fig. 8; Zhou et al., 2013; Zhou and Li, 2017). In the Yantongshan area in the west, piemontite schists are coeval with strongly-deformed chlorite schists, and the typical metamorphic minerals include phengite, piemontite, and garnet, representing HP metamorphism (Cao et al., 2019; Liu, 2000; Zhang et al., 2009a). Further east in the Hulan area, there are strongly deformed felsic gneisses, mica schist, metamorphosed gabbro and marble, all of which have undergone low greenschist-amphibolite facies metamorphism and multi-phase deformation, and have the characteristics of tectonic mélanges (Xi et al., 2003; Zhou and Wilde, 2013). Still further east in the Seluohe area, there is a set of late

Table 1
Isotopic geochronology data of the Changchun-Yanji suture.

Unit	Location	Lithology	Method	Age (Ma)	2σ	Interpretations	References
Hulan Complex	Hongqiling	Pelitic-schist	Zircon LA-ICPMS	287	6	Protolith age	(Wu et al., 2007b)
	Hongqiling	Hornblende-gneiss	Zircon SHRIMP	272	4	Protolith age	(Lü et al., 2011)
	Panshi	Piedmontite-schist	Zircon LA-ICPMS	239	11	Protolith age	(Zhang et al., 2009a)
	Panshi	Piedmontite-schist	Zircon LA-ICPMS	237	5	Protolith age	(Cao et al., 2019)
	Hongqiling	Quartz schist	Zircon LA-ICPMS	245	5	Protolith age	(Zhou et al., 2013)
Seluohe Complex	Seluohu	High-Mg dacite	Zircon SHRIMP	252	5	Protolith age	(Li et al., 2007a)
	Seluohu	Rhyolite	Zircon LA-ICPMS	264	10	Protolith age	(Zhang et al., 2008b)
	Banmiaozhi	Slate	Zircon LA-ICPMS	264	3	Protolith age	(Zhou et al., 2013)
	Banmiaozhi	Sandstone	Zircon LA-ICPMS	255	3	Protolith age	(Zhou et al., 2013)
	Qinglongcun	Amphibolite	Zircon LA-ICPMS	274	6	Protolith age	(Zhang et al., 2009a)
Qinglongcun Complex	Qinglongcun	Biotite-gneiss	Zircon LA-ICPMS	250	4	Protolith age	(Zhou et al., 2013)
	Qinglongcun	Gabbro	Zircon LA-ICPMS	248	1	Protolith age	(Zhou et al., 2013)
	Shimen	Sandstone	Zircon LA-ICPMS	246	2	Protolith age	(Zhou et al., 2013)
Kaishantun Complex	Kaishantun	Slate	Zircon LA-ICPMS	234	4	Protolith age	(Zhou et al., 2013)
	Zidong	Chlorite-schist	Zircon LA-ICPMS	292	2	Protolith age	(Zhou et al., 2013)
	Hongqiling	Amphibolite	Hornblende Ar-Ar	208	2	Metamorphic age	(Lin et al., 2008a)
Hulan Complex (metamorphic age)	Hongqiling	Amphibolite	Hornblende Ar-Ar	214	3	Metamorphic age	(Lin et al., 2008a)
	Piaohechuan	Diorite	Hornblende Ar-Ar	189	1	Metamorphic age	(Lin et al., 2008a)
	Piaohechuan	Diorite	Biotite Ar-Ar	187	1	Metamorphic age	(Lin et al., 2008a)
	Piaohechuan	Diorite	Hornblende Ar-Ar	192	2	Metamorphic age	(Lin et al., 2008a)
	Hongqiling	Gabbro	Muscovite Ar-Ar	187	1	Metamorphic age	(Lin et al., 2008a)
	Hongqiling	Garnet-biotite gneiss	Biotite Ar-Ar	224	1	Metamorphic age	(Xi et al., 2003)
	Hongqiling	Two-mica schist	Muscovite Ar-Ar	229	5	Metamorphic age	(Xi et al., 2003)
	Hongqiling	Amphibolite	Hornblende Ar-Ar	228	3	Metamorphic age	(Liu et al., 2010)
	Hongqiling	Quartz-schist	Muscovite Ar-Ar	220	2	Metamorphic age	(Liu et al., 2016)
	Hongqiling	Muscovite-schist	Muscovite Ar-Ar	221	3	Metamorphic age	(Liu et al., 2016)
	Hongqiling	Muscovite-schist	Muscovite Ar-Ar	221	3	Metamorphic age	(Liu et al., 2016)

Paleozoic-early Mesozoic mélanges, consisting of amphibolite, diorite, gabbro, garnet-bearing biotite schist and garnet-bearing muscovite schist (Table 1). Most of the rocks have undergone strong mylonitization, and some were tectonically transported over TTG complexes of the North China Craton in the form of nappes or klippe (Zhou et al., 2020). To the east in the Qinglongcun area, there are many different types of metamorphic rocks, including phyllonite, amphibolite, gneiss and marble, which have undergone multiple phases of metamorphism and deformation (Zhou et al., 2020). In the far eastern segment of the Yanji Belt at Kaishantun, the rocks of a tectonic mélange are in the form of olistostromes in the south, with imbrication to the north (Tang and Zhao, 2007). The mélange was affected by different degrees of shear deformation and metamorphism, and some of rocks contain high-pressure metamorphic minerals, including phengite and barroisite (Tang and Zhao, 2007; Wu et al., 2003).

6.2. Formation age of the Yanji Belt

Along the Yanji Belt, the series of accretionary complexes (Fig. 8; Zhou et al., 2013, 2020) are mixtures of rocks of different ages and origins (Wu et al., 2007b, 2011a). Numerous geochronological data from these complexes have been reported by previous researchers (Table 1). In the Hulan area, ages of clastic sediments are 287–245 Ma (Lü et al., 2011; Wu et al., 2007b; Zhou et al., 2013), protoliths of piemontite schist are 237 ± 5 Ma (Cao et al., 2019) and 239 ± 11 Ma (Zhang et al., 2009a), and metamorphic ages obtained from the complexes are in the range of 229–188 Ma (Liu et al., 2010; Liu et al., 2016; Xi et al., 2003). These data imply that the Hulan Complex was formed in the middle-late Triassic, rather than early Paleozoic as previously thought. Li et al. (2007a) reported a zircon U–Pb age of 252 ± 5 Ma for high-Mg andesites from the Seluohe area, and these are considered to have formed by partial melting of the mantle wedge with fractional crystallization of the magma caused by fluids derived from the subducted plate. In addition, ages of clastic sediments from the Seluohe area are 255 ± 3 Ma and 264 ± 3 Ma (Zhou and Wilde, 2013), establishing the Seluohe Complex is of middle Triassic age. Ages of magmatic protoliths of metamorphic rocks in the Qinglongcun area are 248 ± 1 Ma (Zhou et al., 2013) and 274 ± 6 Ma (Zhang et al., 2009a), whereas ages of clastic sediments from the Kaishantun area range from 292 ± 2 Ma to 234 ± 4 Ma (Zhou

et al., 2013). Hence, the accretionary complexes distributed along the Yanji Belt mainly formed in the middle-late Triassic.

Additional evidence can also help us to constrain the age of the Yanji Belt. For example, the weighted mean age of an intrusive felsic dike from the Yantongshan area is 217 ± 5 Ma (Fig. 9a, c, d; Cao et al., 2019), whereas ultramafic-mafic intrusions in the Hongqiling area, which indicate an extensional setting, have an age of 216 ± 2 Ma (Wu et al., 2007b). In the overlying Triassic strata, late Triassic terrestrial plant fossils (Li et al., 1985; Sun et al., 1983; Xin et al., 2011) and molasses (Wang et al., 2016a) are present, and the youngest zircon age obtained from molasse is 225 ± 1 Ma (Fig. 9b, e, f). This evidence, supports the contention that the Yanji Belt formed in the middle-late Triassic (230–220 Ma).

6.3. Tectonic setting of the Yanji Belt

As noted above, the Yanji Belt is located between the Jiamusi-Khanka Block and the North China Craton. It was previously considered to be the eastern extension of the Solonker Belt, and therefore part of the Paleo-Asian Oceanic tectonic domain (Li, 2006; BGMRJP (Bureau of Geology and mineral resources of Jilin Province), 1988; Tang, 1990; Tang et al., 1995b; Wang and Fan, 1997; Wu et al., 2007b; Xu et al., 2018). However, later studies indicated that the Jiamusi-Khanka Block originally formed the eastern part of Songliao block that was separated from it during the Permian, only to be later influenced by westward subduction of the Paleo-Pacific, to finally re-amalgamate with the Songliao Block and North China Craton during the Triassic (Li et al., 2021; Zhou et al., 2009a, 2010b, 2014). Under the influence of the Yilan-Yitong Fault (Fig. 8), a collision between the southern margin of the Jiamusi-Khanka Block and the North China Craton occurred, forming the Yanji Belt (Zhou et al., 2020). The Yilan-Yitong Fault is considered to have the attributes of a strike-slip fault (Aouizerat et al., 2020).

After comprehensive analysis, we consider that the Yanji Belt is not part of the eastern extension of the Solonker Belt, but an independent tectonic unit. It was formed during the collision of the Jiamusi-Khanka Block with the North China Craton around the middle-late Triassic, and was influenced by the Paleo-Pacific tectonic domain. The reasons are: (1) Lithospheric structural research shows that the thickness differs between the east and west sides of the Yanji Belt, being generally >130

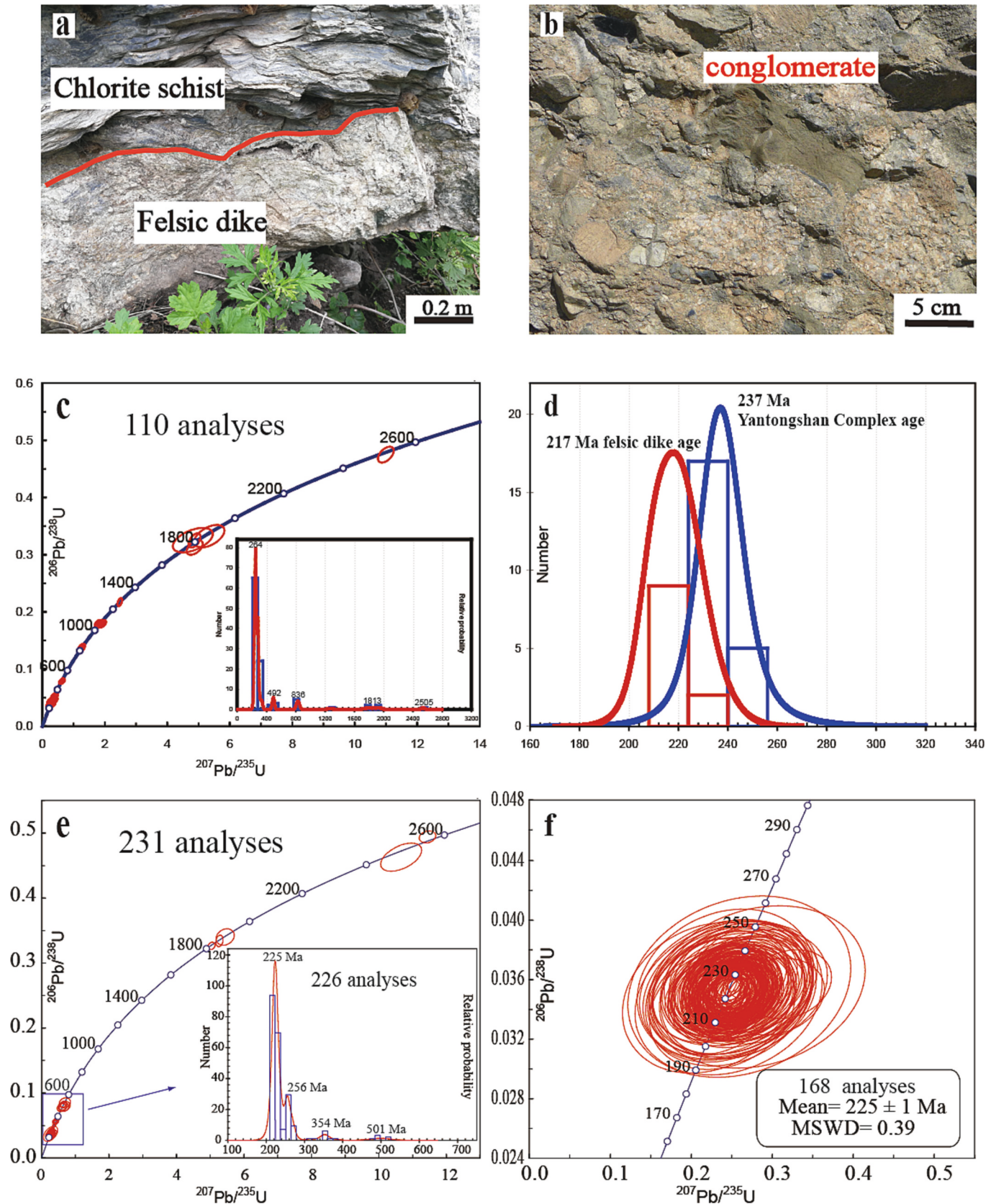


Fig. 9. (a) Felsic dike cutting chlorite schist at Yantongshan; (b) Conglomerate in the Triassic Dajianggang molasse; (c) U—Pb concordia diagram for zircon data from the Yantongshan Complex; (d) Relative age probability plots of zircon U—Pb ages from the felsic dike and the Yantongshan Complex; (e) Concordia diagram and relative probability plot of all concordant zircon U—Pb ages for the Dajianggang Formation; (f) Concordia diagram of the youngest age group from the Dajianggang Formation. (a,c,d plots modified from [Cao et al., 2019](#); b,e,f plots modified from [Wang et al., 2016a](#)).

km in the west and much less than this in the east (60–100 km; Zhang et al., 2008a); (2) The Yanji Belt was formed in the middle-late Triassic (230–220 Ma), which is 20–30 Ma later than the age of the Solonker Belt and closure of the Paleo-Asian Ocean; (3) In the Heilongjiang area (Fig. 8), the protolith ages of blueschist are 208 ± 6 Ma, 213 ± 2 Ma and 224 ± 7 Ma, and the metamorphic age of glaucophane is 185 ± 4 Ma (Zhou et al., 2009a, 2013), indicating a late Triassic-early Jurassic metamorphic event in the area. The similar age relationship between accretionary complexes of the Yanji Belt and blueschists of the Heilongjiang area suggests that they were formed in the same geological event (Zhou et al., 2013), with the latter known to be influenced by Paleo-Pacific subduction; and most importantly, (4) the collision between the Jiamusi-Khanka Block and the North China Craton resulted from the westward subduction of the Paleo-Pacific plate, and is unrelated to the Paleo-Asian Ocean (Zhou et al., 2009b, 2010a).

Hence, we consider that the Yanji Belt, which is a typical accretionary complex at the southern edge of the Jiamusi-Khanka Block, is related to the westward subduction of the Paleo-Pacific plate. The Yanji Belt, along with the Heilongjiang blueschist belt, are also both part of the Jilin-Heilongjiang high-pressure metamorphic belt (Ji-Hei HP belt, Zhou et al., 2018, 2020).

7. Discussion

7.1. Locating the final closure of the Paleo-Asian Ocean

There are three major suture zones in the CAOB, which from north to south are: the Erqis-Toudaoqiao Suture, the North Tianshan-Hegenshan-Heihe Suture and the South Tianshan-Solonker Suture (Fig. 1). They are broadly parallel to each other and are the result of multi-phase assemblage of microcontinents, arcs-back arcs, oceanic plateaus, subduction-accretion complexes and ophiolites. Together they record the complex history of the Paleo-Asian Ocean. To answer the question of where and when the final closure of the Paleo-Asian Ocean occurred, we need to discuss the tectonic attributes of all three sutures.

The Erqis-Toudaoqiao Suture, also called the Erqis-South Mongolian-Toudaoqiao-Xinlin Suture, is a huge fault zone which extends from western Mongolia eastward to the Xinlin area of Heilongjiang Province, China, and separates the Tuva-Mongolia and Erguna blocks to the north from the Junggar and Xing'an blocks to the south (Fig. 1). This suture is characterized by the occurrence of ophiolitic rocks and high-grade metamorphic rocks, including blueschists, gneisses, amphibolites and marbles. The ophiolitic rocks are the result of plate collision in the early Paleozoic (Li, 1991), and the blueschists are metamorphosed basalts with N-MORB and OIB affinities (Zhou et al., 2015b). In the Toudaoqiao-Xinlin region, zircon U–Pb ages of 517 ± 2 Ma and 517 ± 7 Ma were obtained from the accretionary complexes, with an age of 492 ± 1 Ma obtained from a cross-cutting granite dike in the sequence (Zhou et al., 2015b). Along the boundary between the Erguna and Xing'an blocks, sillimanite gneiss with an age of 496 ± 3 Ma, hornblende-plagioclase gneiss with an age of 496 ± 7 Ma (Zhou et al., 2011b), and post-collisional granite dated at 494–480 Ma (Ge et al., 2005, 2007; Wu et al., 2011b), place constraints on the age of the suture. However, there may be controversy here, because some early Carboniferous ophiolites have also been found in this area, and Zhang et al. (2018) recently reported zircon U–Pb ages of 344–328 Ma for E-MORB gabbro. In the Erqis region, several high-resolution geochronological data provide constraints on the tectonic evolutionary history of the belt. Wang et al. (2012) obtained zircon U–Pb ages of 364 ± 5 Ma from MORB-type gabbro and ca. 355 Ma from supra-subduction/syn-collisional granites. Zhang et al. (2003) obtained a SHRIMP zircon U–Pb age of 372 ± 19 Ma from plagiogranite in the ophiolitic fragment. Wu et al. (2006) reported tholeiite at 352 ± 4 Ma, which is geochemically similar to OIB/IAB. The widely reported Cambrian-early Carboniferous ophiolites in the Erqis-Toudaoqiao Suture therefore suggest its complex evolutionary history, which needs to be further investigated. However, it can be

clearly stated that, overall, the formation age of this suture zone is older than that of the South Tianshan-Solonker Suture.

The North Tianshan-Hegenshan-Heihe Suture extends from North Tianshan in Xinjiang Province, through Hegenshan, to Heihe in NE China, and separates the Xing'an Block to the north from the Songliao Block to the south, and also the Junggar Block to the north from the Central Tianshan Block to the south. It is also known by earlier researchers as the boundary between the Altai and the Manchurides (Şengör and Natal'in, 1996; Şengör et al., 1993). In its eastern section, it is characterized by ophiolitic rocks, the most famous of which is the Hegenshan ophiolite, which is one of the largest oceanic lithospheric fragments in the CAOB, composed of harzburgite, dunite and basalt (Xiao et al., 2003; Zhou et al., 2015a). Zircon U–Pb ages of 354–333 Ma (Jian et al., 2012), 358–320 Ma (Zhou et al., 2015a), and 298–295 Ma (Miao et al., 2007, 2008) were obtained from ophiolitic rocks at Hegenshan, and detrital zircon ages of 279 ± 2 Ma (Zhou et al., 2015a) and 308 ± 5 Ma (Pei et al., 2016) were obtained from the overlying strata. We previously discussed the tectonic significance of its eastern extension, i.e., the Hegenshan-Heihe Suture, in Section 5.2, and concluded that it is the local suture zone between the Xing'an Block and the Songliao Block. With respect to the western extension, in the Chinese North Tianshan, the ophiolites are dated at 347–325 Ma and may represent remnants of the Paleo-Tianshan Ocean (Jian et al., 2005; Wang et al., 2009, 2014b; Xu et al., 2006). The age of a post-collisional A-type granite intrusion within the ophiolites is ca. 316 Ma (Han et al., 2010), whereas arc-related felsic volcanics are 319–315 Ma (Xie et al., 2016). Granitic gneisses derived in a continental arc setting have crystallization ages of 325–320 Ma (Zhang et al., 2015). Regional migmatization and coeval granitic plutonism occurred at 322–297 Ma and may be related to crustal thinning due to post-collision extension (Ni et al., 2021). These lines of evidence suggest that the formation age of the Chinese North Tianshan approximately dates to the late Carboniferous, which is similar to that of the Beishan Belt (Ao et al., 2010; Mao et al., 2012; Xiao et al., 2010c). Furthermore, the two are geographically close to each other, with the northwestern side of the Beishan adjacent to Bogda Shan in the Chinese North Tianshan. Therefore, we suggest that the Beishan Belt not only constitutes the middle section of the South Tianshan-Solonker Suture, but also can be considered as the location of the middle section of the North Tianshan-Hegenshan-Heihe Suture. The middle section of the two suture zones was influenced by north-south plate subduction in the late Paleozoic.

The South Tianshan-Solonker Suture is the southernmost suture zone within the CAOB, and extends from northwest China to northeast China, spanning >5500 km (Han and Zhao, 2018). It is the most prominent suture zone in the CAOB, and separates the North China and Tarim cratons to the south from the Siberia craton to the north. Abundant high-pressure metamorphic rocks and ophiolites formed during subduction of the intervening ocean (Gao et al., 2008; Mao et al., 2012; Xiao et al., 2018; Zhou et al., 2018). Compared to the Erqis-Toudaoqiao and North Tianshan-Hegenshan-Heihe sutures, it represents the main plate boundary within the CAOB and is younger than both of them. It corresponds to the boundary between the Cathaysia flora and the Angara flora, and the two biotas began to mix together during the progressive closure of the Paleo-Asian Ocean from the late Carboniferous (Johnson et al., 2008; Sun et al., 2018; Wang and Cheng, 2001; Xiao et al., 2015b; Zhang et al., 2020). We consider that during the formation of the CAOB, multiple phases of accretion occurred, and that the Erqis-Toudaoqiao and North Tianshan-Hegenshan-Heihe sutures may have been re-activated during the early process of plate collision, whereas the youngest South Tianshan-Solonker Suture is interpreted as representing the final closure of Paleo-Asian Ocean (Eizenhöfer et al., 2014; Eizenhöfer and Zhao, 2018; Xiao et al., 2003).

7.2. Development of the South Tianshan-Solonker Suture

As the position of final closure of the Paleo-Asian Ocean, the

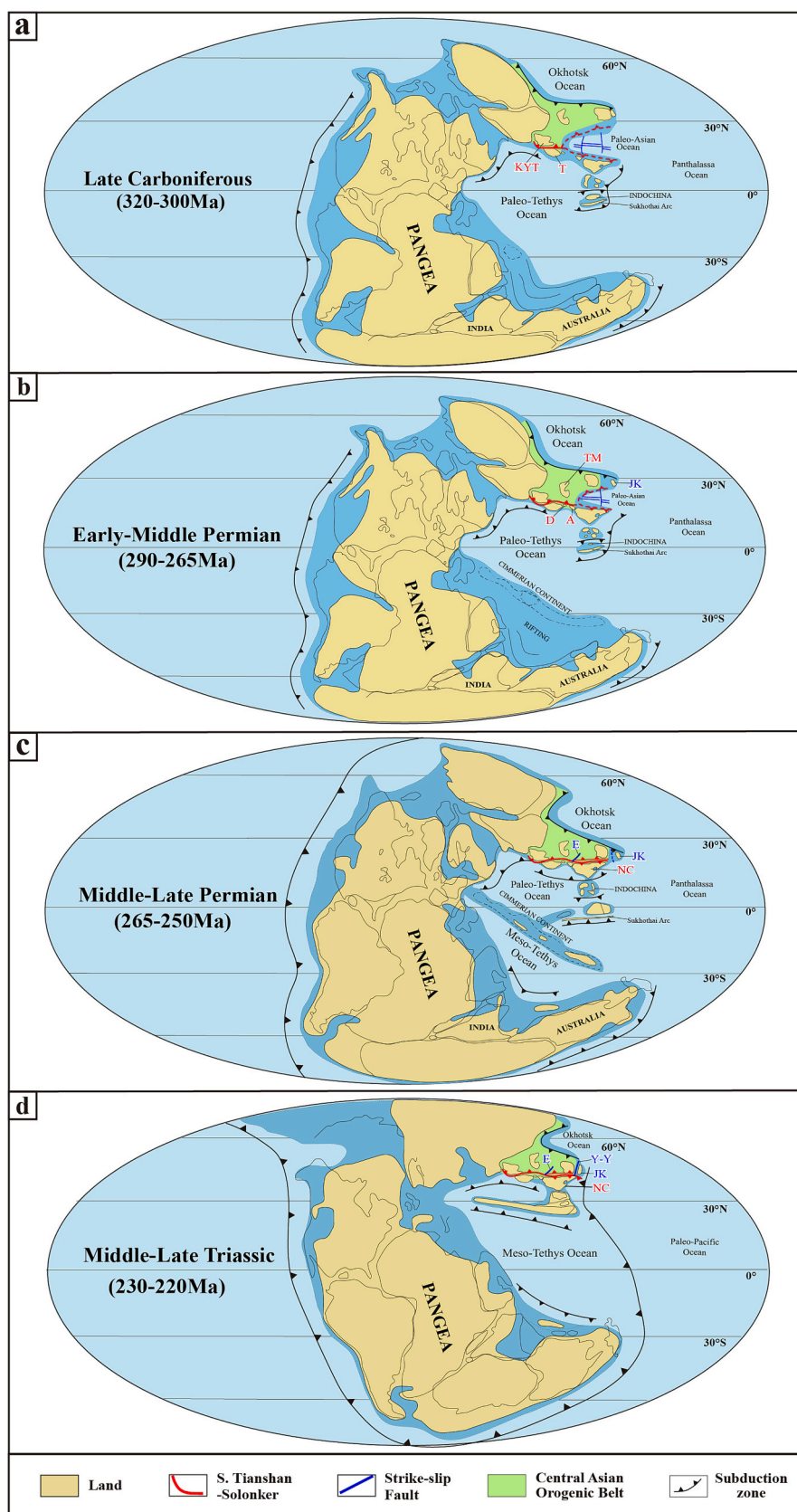


Fig. 10. Pangea from amalgamation to break-up (modified from Metcalfe, 2021), showing the evolution of South Tianshan-Solonker Suture. A = Alxa block; D=Dunhuang block; E = East Gobi Fault; JK = Jiamusi-Khanka block; KYT = Kazakhstan-Yili block; T = Tarim block; TM = Tuva-Mongolia block; Y-Y=Yilan-Yitong Fault; NC=North China Craton; SC=South China Block.

evolution of the South Tianshan-Solonker Suture has long been widely discussed. Available petrographic, geochronological, geochemical, paleontological, and paleomagnetic data enable us to systematically review the nature of the final closure of the Paleo-Asian Ocean at a broader scale. Based on the differences in tectonic setting, rock composition, closure timing and characteristics, we have for the first time divided the South Tianshan-Solonker Suture zone into four segments, which, from west to east are: the South Tianshan, Beishan, Solonker and Yanji belts.

The western segment of the South Tianshan-Solonker Suture, here named the South Tianshan Belt, records strong plate collision before the late Carboniferous, forming a series of high-pressure metamorphic rocks, characterized by the presence of blueschist, eclogite and mica schist (Gao and Klemd, 2003; Gao et al., 2006; Klemd et al., 2005). In the Permian, the whole area was under a relatively stable post-orogenic setting, with the formation of bimodal volcanic rocks, post-collisional granites, and terrestrial molasses (Gao et al., 1998; Li and Wang, 2000; Li et al., 2006b; Wang et al., 1990; Wang et al., 1994). Sedimentary facies gradually changed from marine to either lacustrine or fluvial during this period (Han et al., 2016). An early Permian granite dike crosscuts the HP metamorphic belt (Gao et al., 2011), and the HP metamorphic rocks also underwent retrograde metamorphism at this time (Gao et al., 2006), indicating the formation of the South Tianshan Belt was earlier than the Permian. Combined with related studies of subduction polarity (Gao et al., 2009; Han et al., 2011; Zhang et al., 2013), we interpret that during the late Carboniferous, the closure of the western section of the Paleo-Asian Ocean was completed, with the Tarim Craton transported northward to collide with the Kazakhstan-Yili Block, leading to the formation of the South Tianshan Belt (Fig. 10a).

The Beishan Belt lies east of the South Tianshan Belt. Before the closure of the Paleo-Asian Ocean, two remnant ocean basins existed in this region: the Hongshishan Ocean and the Liuyuan Ocean (Guo et al., 2014; Song et al., 2013a, 2013b; Tian et al., 2014; Xiao et al., 2010c). Age data obtained from ophiolites (Ao et al., 2010; Jiang et al., 2006; Li et al., 2009a) and collision-related granites (Jia et al., 2016; Jiang and Nie, 2006; Li et al., 2018, 2019; Ren et al., 2019; Zhao et al., 2015) indicate that the Hongshishan ocean closed in the early-middle Permian. The early Permian ophiolites (Mao et al., 2012; Wang et al., 2016b) in the Liuyuan area are unconformably overlain by late Permian sediments (Wang et al., 2016b), and the final closure of the Liuyuan Ocean is also considered to be early-middle Permian. Combining geochronological evidence with stratigraphic, paleontological, and paleomagnetic data (Fang, 1997; Guo, 2001; Li et al., 1999; Ou et al., 1993, 2004; Wei et al., 2020; Wu, 1993; Xu et al., 2019b; Zhu, 1997, 2001), we suggest that the evolution of the middle section of the South Tianshan-Solonker Suture ended in the early-middle Permian, although it may have continued into the early late Permian at the northern margin of the Alxa Block (Chen et al., 2021b; Hui et al., 2021; Xie et al., 2014; Zheng et al., 2014). This makes it slightly younger than the South Tianshan Belt. Following this, the Dunhuang-Alxa blocks were transported northward to collide with the Tuva-Mongolia Block (Fig. 10b).

After the progressive closure of the western and middle sections of the Paleo-Asian Ocean, there were still remnant ocean basins within the Solonker region (Eizenhöfer and Zhao, 2018; Zhu, 2015). Stratigraphic studies show that since the late Permian, the sedimentary facies began to change from marine to terrestrial (Li et al., 2014; Zhao et al., 2016; Zhu, 2015). At this time, the Angara and Cathaysia floras, which were distributed on either side of the suture zone, began to mix together (Sun et al., 2018; Xiao et al., 2015b), but it was not until the early-middle Triassic that terrestrial sediments began to be widely developed (Li et al., 2014; Zhu, 2015). The occurrence of ophiolites (Chen et al., 2012; Jian et al., 2010; Miao et al., 2007, 2008; Wang et al., 2014a; Song et al., 2016a) and collision-related intrusive rocks (Chen et al., 2009; Hu et al., 2015; Li et al., 2007b; Li et al., 2009b; Shi et al., 2004; Zhang et al., 2009b) show that the closure of the eastern section of Paleo-Asian Ocean was completed in the middle-late Permian, at which time bi-directional

subduction occurred (Eizenhöfer and Zhao, 2018; Eizenhöfer et al., 2014; Xiao et al., 2003, 2018) and the North China and Siberia cratons were amalgamated in an 'Appalachian-type' collision (Fig. 10c). A transition in subduction polarity occurred between the Beishan Belt and the Solonker Belt, with this phenomenon likely related to the development of the East Gobi Fault.

The Yanji Belt has traditionally been considered to be the eastern extension of Solonker Belt (Li, 2006; BGMRJP (Bureau of Geology and mineral resources of Jilin Province), 1988; Tang, 1990; Tang et al., 1995b; Wang and Fan, 1997; Wu et al., 2007b; Xu et al., 2018). However, the ages of accretionary complexes within this belt are all middle-late Triassic (Table 1), and the molasse in the overlying strata (Wang et al., 2016a) and the intrusive dikes (Cao et al., 2019; Wu et al., 2007b) were both formed in the late Triassic, so we deduce that the formation of the Yanji Belt was most likely in the middle-late Triassic (230–220 Ma), 20–30 Ma later than amalgamation of the Solonker Belt. Based on our comprehensive analysis, we consider that the Yanji Belt belongs to an independent tectonic unit, rather than the eastward extension of the Solonker Belt, and that it was influenced by the Paleo-Pacific tectonic domain. During the middle-late Triassic, the Jiamusi-Khanka Block moved southwestward along the Yilan-Yitong Fault and amalgamated with the North China Craton (Fig. 10d).

Thus, the formation of the South Tianshan-Solonker Suture can be characterized by four distinct west-to-east accretion/collision events that lasted from the late Carboniferous to the late Permian, but with various reactivation of the earlier components. The South Tianshan Belt, the Beishan Belt, and the Solonker Belt demarcate the amalgamation between the Tarim Craton and the Kazakhstan-Yili Block, the Dunhuang-Alxa Block and the Tuva-Mongolia Block, and the North China Craton with the the Siberia Craton, respectively, representing the progressive closure of the Paleo-Asian Ocean. During the middle-late Triassic, the transition from the Paleo-Asian Oceanic tectonic domain to the Paleo-Pacific tectonic domain occurred, and NE China began to be influenced more strongly by the latter, with the Yanji Belt recording part of this process. Although traditionally considered part of the South Tianshan-Solonker Suture zone, we propose that the Yanji Belt, although it appears to link with the eastern extremity of this zone, should be considered a separate entity as it post-dates closure of the Paleo-Asian Ocean and instead is directly related to subduction of the Paleo-Pacific Ocean.

8. Conclusions

We have identified four distinct accretion/collisional belts within the South Tianshan-Solonker Suture, which are, from west to east: the South Tianshan Belt, the Beishan Belt, the Solonker Belt and the Yanji Belt.

(1) In the South Tianshan Belt, geochronological data from Paleozoic cover rocks and high-pressure metamorphic rocks show that during the late Carboniferous, the western section of the Paleo-Asian Ocean was closed, resulting in the northward transport of the Tarim Craton to collide with the Kazakhstan-Yili Block, thus forming the South Tianshan Belt.

(2) The Beishan Belt to the east was formed by the northward transport of the Dunhuang-Alxa Block, resulting in its collision with the Tuva-Mongolia Block. The youngest zircon U–Pb age from ophiolites indicate that closure of the Paleo-Asian Ocean in this middle section of the South Tianshan-Solonker Suture was in the early-middle Permian, but may have continued into the early late Permian at the northern edge of the Alxa Block.

(3) The available petrographic, geochronological, and paleontological data from the Solonker Belt indicate that it formed in the middle-late Permian, during which time bi-directional subduction of the Paleo-Asian Ocean occurred and the North China and Siberia cratons were amalgamated in an 'Appalachian-type' collisional orogeny. Furthermore, we found evidence that the transition in subduction polarity from northward to bi-directional occurred along the boundary between the Beishan

and Solonker belts, which may be related to the East Gobi Transform Fault.

(4) The Yanji Belt post-dates the other segments of the South Tianshan-Solonker Suture and was not related to the closure of the Paleo-Asian Ocean. Instead, it was influenced by Paleo-Pacific subduction and resulted from amalgamation between the Jiamusi-Khanka Block and the North China Craton during the middle-late Triassic (230–220 Ma). It is, therefore, not technically part of the South Tianshan-Solonker Suture, although it does link with the eastern extremity of this suture.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

No data was used for the research described in the article.

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