Late Carboniferous collision between the Tarim and Kazakhstan–Yili terranes in the western segment of the South Tian Shan Orogen…

Article *in* Earth-Science Reviews · December 2011

DOI: 10.1016/j.earscirev.2011.09.001

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Earth-Science Reviews 109 (2011) 74–93

Contents lists available at SciVerse ScienceDirect

Earth-Science Reviews

journal homepage: www.elsevier.com/locate/earscirev

Review

Late Carboniferous collision between the Tarim and Kazakhstan–Yili terranes in the western segment of the South Tian Shan Orogen, Central Asia, and implications for the Northern Xinjiang, western China

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ARTICLE INFO

Article history:
Received 18 June 2009
Accepted 14 September 2011
Available online 21 September 2011

Keywords:
South Tian Shan
Late Carboniferous
Collisional orogen
Central Asia

ABSTRACT

The Tian Shan of Central Asia is located in the southwestern part of the Central Asian Orogenic Belt (CAOB, also known as the Central Asian Orogenic System or CAOS). Formation of the South Tian Shan Orogen is a diachronous, scissors-like process during the Paleozoic and its western segment in China–Kyrgyzstan contiguous regions is accepted as the site of the final collision zone between the Tarim craton to the south and the Kazakhstan–Yili terrane to the north in the Late Paleozoic. However, when the final collision occurred is still in hot debate. Particularly, an end-Permian to Triassic collisional model is recently proposed for the western segment of the South Tian Shan Orogen. This even leads to the speculation that the complicated accretion–collision processes in the Northern Xinjiang of western China, which involved the terrane amalgamation in the East and West Junggar and the collision between the Altai and Kazakhstan terranes and between the Yili–Central Tian Shan and Junggar terranes, were finally terminated during the end-Permian to mid-Triassic, rather than the Late Paleozoic as usually accepted. Obviously, the western segment of the South Tian Shan Orogen also presents the key issue associated with the termination time of accretion–collision processes in the Northern Xinjiang. A collisional model that is derived from the knowledge of the Himalayan Orogen is helpful for establishing a sequence of major tectonothermal events in the western segment of the South Tian Shan Orogen and constraining the time of collision between the Tarim craton and the Kazakhstan–Yili terrane.

For the western segment of the South Tian Shan Orogen, the end-Permian to Triassic collisional model is mainly based on Triassic zircon U–Pb ages of 234 to 226 Ma from the West Tian Shan eclogite and two suspected Late Permian radiolarian specimens *Albaillella excelsa* Ishiga, Kito and Imoto (?) from the Baleigong ophiolitic mélange. Actually, the poor preservation of the two radiolarian specimens and the lack of a ventral wing make their identifications difficult. Furthermore, the Baleigong ophiolitic mélange was intruded by one granite pluton with a zircon age of 273 Ma, and this provides geological evidence against the reliability of the Late Permian radiolarian specimens. Because the Triassic zircons contain no index mineral inclusions such as omphacite and coesite grown under high to ultrahigh pressure conditions, it is difficult to link their ages to high to ultrahigh pressure peak metamorphism. In addition, this model is not compatible with extensive Permian plutonism and molasse sedimentation and Triassic to Jurassic tectonomagmatic quiescence and continental deposits in the collisional zone and adjacent tectonic units.

In contrast, new U–Pb ages of the zircon domains containing omphacite and phengite inclusions and Sm–Nd and rutile U–Pb ages of eclogite samples from the western segment of the South Tian Shan Orogen consistently indicate that high pressure peak metamorphism of subducted oceanic material occurred at ~319 Ma (the end of the Early Carboniferous). This and the youngest Early Carboniferous radiolarian and conodonts fossils from ophiolitic mélanges show that the collision must have taken place after the Early Carboniferous, whereas the oldest stitching granitic plutons in the collisional zone place an upper-age bound of ~300 Ma (the end of the Late Carboniferous) for the collision. These specify that the final collision in the western segment of the South Tian Shan took place in the Late Carboniferous rather than the end-Permian to Triassic. Noticeably, syn-collisional granitoids are rare, but Permian post-collisional plutonism and molasse sedimentation are widespread in the western segment of the South Tian Shan and adjacent tectonic units, and the oldest post-collisional plutons were nearly concurrent with low pressure, high temperature metamorphism in the south edge of the Kazakhstan–Yili terrane. All these suggest a significant geodynamic change at ~300 Ma, which may be caused by delamination of the thickened lithospheric root and asthenospheric upwelling. Such a process might have provided heat for low pressure, high temperature metamorphism and triggered partial melting of the

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0012-8252/$ – see front matter
doi:10.1016/j.earscirev.2011.09.001

lower crust and underlying lithosphere in the western segment of the South Tian Shan Orogen and adjacent tectonic units. The Late Carboniferous collisional model is also compatible with the Triassic to Jurassic tectonomagmatic quiescence and continental deposits in the western segment of the South Tian Shan Orogen and adjacent tectonic units.
For the South Tian Shan Orogen, the final collision in the western segment occurred in the Late Carboniferous, significantly younger than that in the eastern segment. In the Northern Xinjiang, the Late Carboniferous collision in the western segment of the South Tian Shan Orogen was nearly simultaneous with the final collision in the North Tian Shan collisional zone between the Yili–Central Tian Shan and Junggar terranes and in the Irtysh–Zaysan collisional zone between the Altai and Kazakhstan terranes, and these collisional events postdated the terrane amalgamation in the East and West Junggar. Therefore, the accretion–collision processes in the Northern Xinjiang were finally terminated during the Late Carboniferous rather than the end-Permian to mid-Triassic.

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1. Introduction

The Central Asian Orogenic Belt (CAOB) or the Central Asian Orogenic System (CAOS) is situated between the European (also known as the Baltic) craton to the west, Siberian craton to the east, and the Tarim and North China cratons to the south (Fig. 1a). It is the largest Paleozoic accretionary orogen in the world and its history was not only dominated by accretion of arcs, accretionary complexes, seamounts, and oceanic plateaus (Coleman, 1989; Zonenshain et al., 1990; Mossakovsky et al., 1993; Fedorovskii et al., 1995; Buslov et al., 2001, 2004; Khain et al., 2002, 2003; Kheraskova et al., 2003) but also involved collision and suturing of island arcs and large continental blocks (Badarch et al., 2002; Buchan et al., 2002; Xiao et al., 2004, 2008, 2009; Yakubchuk, 2004; Briggs et al., 2007; Windley et al., 2007; Kelty et al., 2008), and locally Early Paleozoic rifting was followed by Late Paleozoic accretion (Kröner et al., 2010). The complicated accretion–collision processes are generally thought to have been related to closure of the Paleo-Asia Ocean and have finally resulted in the largest tectonic assembly of continental and oceanic terranes and the largest region of Phanerozoic continental growth in the world (Coleman, 1989; Zonenshain et al., 1990; Mossakovsky et al., 1993; Şengör et al., 1993; Şengör and Natal'in, 1996; Jahn et al., 2000; Kovalenko et al., 2004; Windley et al., 2007). However, progressive duplication of a single and long-evolving island arc system (Şengör et al., 1993; Şengör and Natal'in, 1996) for the development of the CAOB has faced many difficulties in light of recent geologic observations (e.g., Bykadorov et al., 2003; Windley et al., 2007; Kröner et al., 2010).

The Northern Xinjiang is located in the southwestern part of the CAOB, including the Chinese Altai, Junggar and Tian Shan tectonic domains from north to south. It has long been held that all of these tectonic domains were amalgamated together as a result of accretion–collision processes during the Late Paleozoic (e.g., Xiao et al., 1990, 1992; He et al., 1994; Shu et al., 2001; Li et al., 2006; also see Han et al., 2010a, 2010b, and references therein). However, Xiao et al. (2008, 2009) recently propose that the accretion–collision processes were finally terminated during the end-Permian to mid-Triassic, for which two pieces of key evidence such as the Late Permian radiolarian specimens and Triassic zircon ages for high to ultrahigh pressure metamorphism come from the Chinese South Tian Shan close to the China–Kyrgyzstan border. It is apparent that the western segment of the South Tian Shan Orogen in China–Kyrgyzstan contiguous regions is the key to when the accretion–collision processes in the Northern Xinjiang were finally terminated.

The Tian Shan (also spelled Tianshan or Tien Shan) in Central Asia extends in the east–west direction for more than 2500 km from Uzbekistan in the west across Tajikistan, Kyrgyzstan, and southern Kazakhstan in the center to Xinjiang in western China in the east. The Tian Shan is a Late Paleozoic orogen, but it was strongly modified by large-scale strike-slipping faulting in an intracontinental setting during the Late Permian and Early Triassic (Laurent-Charvet et al., 2002, 2003; Buslov et al., 2004; Natal'in and Şengör, 2005; Van der Voo et al., 2006), significantly later than the disappearance of any marine basins from the Tian Shan, northern Tarim, eastern Kazakhstan, and Altai (e.g., BGMRXUAR, 1993; Bakirov and Kakitaev, 2000). The present-day Tian Shan has resulted from a relatively late-stage response to the Cenozoic India–Asia collision (e.g., Avouac et al., 1993; Hendrix et al., 1994; Abdrakhmatov et al., 1996; Yin et al., 1998; Bullen et al., 2001, 2003; Charreau et al., 2005, 2006; De Grave et al., 2007; Sun et al., 2009).

The Kyrgyzstan Tian Shan, east of the Cenozoic NW-striking Talas–Fergana dextral strike-slip fault, is usually divided into three tectonic

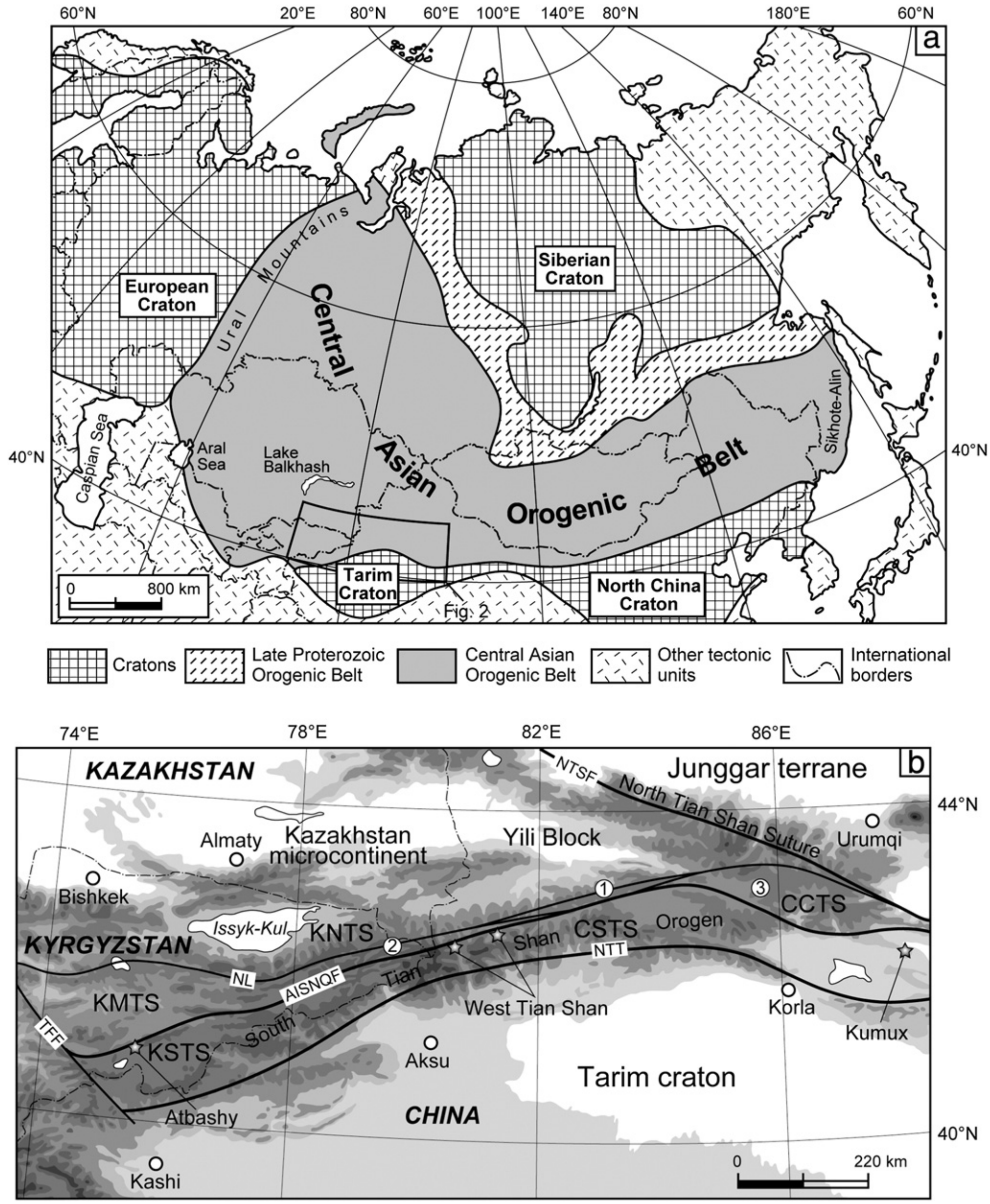

Fig. 1. (a) The Central Asian Orogenic Belt is the tectonic assembly of continental and oceanic terranes between the European craton in the west, the Siberian craton in the east, and the North China and Tarim cratons in the south due to closure of the Paleo-Asian Ocean in the Phanerozoic (modified from Şengör et al., 1993; Jahn et al., 2000). (b) Topographic and sketch tectonic map of the western segment of the Tian Shan in China–Kyrgyzstan contiguous regions. KNTS — Kyrgyzstan North Tian Shan, KMTS — Kyrgyzstan Middle Tian Shan, KSTS — Kyrgyzstan South Tian Shan, and CSTS — Chinese South Tian Shan, AISNQF — Atbashy–Inylchek–South Nalati–Qawablak Fault, TFF — Talas–Fergana Fault, NL — Nikolaev Line, NTT — North Tarim Thrust, and NTSF — North Tian Shan Fault. The Atbashy and West Tian Shan high to ultrahigh pressure metamorphic rocks are indicated by stars. See text for discussion on tectonic correlation between the Kyrgyzstan and Chinese Tian Shan. Especially, the Chinese Central Tian Shan is a continuous strip along the southern margin of the Yili Block, which is truncated eastward by the NTSF and extends westward to the Kyrgyzstan Middle Tian Shan (1) (He et al., 2001; Qian et al., 2009; Gao et al., 2009). The Yili Block and the Chinese Central Tian Shan were amalgamated together as the 'Yili–Central Tian Shan' terrane in the Early Silurian (Zhang et al., 2007b, 2007c; Burtman, 2008; Yang and Zhou, 2009; Hegner et al., 2010). Alternatively, the Kyrgyzstan Middle Tian Shan may extend eastward across the border into China (e.g., Bakirov and Kakitaev, 2000) and then thin out at the eastern end of the Atbashy–West Tian Shan high to ultrahigh pressure metamorphic belt along the Atbashy–Inylchek–South Nalati–Qawablak Fault (2) (e.g., Burtman, 2006; Hegner et al., 2010). In any case, the Kazakhstan–Yili terrane is made up of the Kazakhstan microcontinent, Yili Block, Kyrgyzstan North and Middle Tian Shan, and Chinese Central Tian Shan. By contrast, the Chinese Central Tian Shan may be a separate block east of the Yili Block (3), but it is considered as part of the South Tian Shan Orogen (Charvet et al., 2007, 2011; Wang et al., 2007a, 2010a, 2011; Lin et al., 2009).

units: the North, Middle and South Tian Shan separated by major suture zones. The Kyrgyzstan North and Middle Tian Shan, southern Kazakhstan, and the Yili Block and Central Tian Shan of China constitute the Kazakhstan–Yili terrane with a continental basement (Fig. 1b). The Kazakhstan–Yili terrane was formed by closure of the Terskey Ocean during the Ordovician to Silurian times (Zonenshain et al., 1990; Gao et al., 2009; Glorie et al., 2010). The Ordovician collision occurred between the Kyrgyzstan North Tian Shan arc and the Kazakhstan microcontinent to the north (Lomize et al., 1997; Mikolaichuk et al., 1997; Konopelko et al., 2008; Biske and Seltmann, 2010), followed by the Early Silurian collision between the Kyrgyzstan Middle and North Tian Shan (Mikolaichuk et al., 1997; Konopelko et al., 2008; Biske and Seltmann, 2010; Glorie et al., 2010).

The Chinese Tian Shan is generally divided into eastern and western segments roughly along longitude 88° E. The West Tian Shan is usually divided into four main tectonic units: the North Tian Shan, Yili Block, Central and South Tian Shan from north to south. The Chinese North Tian Shan is mainly occupied by a west-northwest-striking ophiolitic mélange zone between the Yili–Central Tian Shan terrane to the south and the Junggar terrane to the north, which is usually accepted as the youngest suture zone in the Northern Xinjiang and formed at 325 to 316 Ma (Han et al., 2010b, and references therein). The Yili Block is the eastern part of the Kazakhstan microcontinent, but the Chinese Central Tian Shan may be the eastern extension of the Kyrgyzstan Middle Tian Shan (e.g., He et al., 2001; Qian et al., 2009; Gao et al., 2009) or of the Kyrgyzstan North Tian Shan (e.g., Pickering et al., 2008; Biske and Seltmann, 2010; Hegner et al., 2010) or of both (Burtman, 2006). The Kyrgyzstan Middle Tian Shan may thin out eastward near the China–Kyrgyzstan border (e.g., Biske and Seltmann, 2010) or extend eastward across the border into China (e.g., Bakirov and Kakitaev, 2000) and then thin out at the eastern end of the Atbashy–West Tian Shan high to ultrahigh pressure metamorphic belt along the Atbashy–Inylchek–South Nalati–Qawablak Fault (e.g., Burtman, 2006; Hegner et al., 2010; Fig. 1b). The Chinese Central Tian Shan was originally part of the northern margin of the Tarim craton and at present it is a separate block east of the Yili Block (Charvet et al., 2007, 2011; Wang et al., 2007a, 2010a, 2011; Lin et al., 2009) or a continuous strip along the southern margin of the Yili Block as eastern extension of the Kyrgyzstan Middle Tian Shan (He et al., 2001; Qian et al., 2009; Gao et al., 2009).

It seems clear that the Chinese and Kyrgyzstan South Tian Shan are eastern and western parts of one single Late Paleozoic orogen between the Tarim craton to the south and the Kazakhstan–Yili terrane to the north (Fig. 2). It is generally accepted that the South Tian Shan Orogen resulted from closure of the South Tian Shan (also called the Turkestan) Ocean followed by collision between the Tarim craton and the Kazakhstan–Yili terrane during the Late Paleozoic. The closure of the South Tian Shan Ocean may be a diachronous, scissors-like process, which occurred first in the east and ended in the west (Chen et al., 1999a; Dong et al., 2011). The earliest closure took place in the Hongliuhe area of the East Tian Shan in the Early Devonian, where the Hongliuhe ophiolite is crosscut by the undeformed granite dyke with a zircon U–Pb age of 405 ± 5 Ma (SHRIMP, Zhang and Guo, 2008), while the collision in the Kumux area occurred during the Late Devonian to Early Carboniferous (e.g., Charvet et al., 2007; Dong et al., 2011), and the final closure of the South Tian Shan Ocean in the west should not be later than Early Permian (Dong et al., 2011) or in the end-Permian to Triassic (Li et al., 2002, 2005a, 2005b, 2009; Zhang et al., 2007b, 2007c). This suggests that the western segment of the South Tian Shan Orogen in China–Kyrgyzstan contiguous regions is the key site for constraining the time of the final collision and suturing of the Tarim craton and the Kazakhstan–Yili terrane.

The collision between the Tarim craton and the Kazakhstan–Yili terrane might occur during the Late Paleozoic (e.g., Burtman, 1975, 2008; Khain, 1985; Windley et al., 1990; Allen et al., 1992; Xiao et al., 1992; Biske, 1995; Carroll et al., 1995; Gao et al., 1998, 2009; Chen et al., 1999a; Bakirov and Kakitaev, 2000; Yakubchuk, 2004; Pickering et al., 2008; Konopelko et al., 2009; Biske and Seltmann, 2010; Hegner et al., 2010). Most researchers place the collisional event in the Late Devonian (Xiao et al., 1992; Che et al., 1994; Wang et al., 1994; Xia et al., 2004), Late Devonian to Early Carboniferous (Allen et al., 1992; Carroll et al., 1995; Chen et al., 1999a, 1999b; Charvet et al., 2007, 2011; Wang et al., 2007a, 2010a, 2011; Lin et al., 2009), Early to Middle Carboniferous (Coleman, 1989; Windley et al., 1990; Gao et al., 1998; Zhou et al., 2001; Bykadorov et al., 2003; Gao and Klemd, 2003a, 2003b; Yang and Zhou, 2009), Late Carboniferous (Bakirov and Kakitaev, 2000; Burtman, 2006, 2008; de Jong et al., 2009; Gao et al., 2009; Hegner et al., 2010; Su et al., 2010), Middle Carboniferous to Early Permian (Biske, 1995; Bykadorov et al., 2003; Biske and Seltmann, 2010), and the latest Carboniferous to Early Permian (Zonenshain et al., 1990; Shi et al., 1994; Chen et al., 1999a, 1999b; Heubeck, 2001; Bazhenov et al., 2003; Xiao et al., 2004; Klemd et al., 2005; Gao et al., 2006; Li et al., 2006; Konopelko et al., 2007; Solomovich, 2007; Wang et al., 2007b; Li et al., 2008; Pickering et al., 2008), but some researchers place the major collisional event in the latest Permian to earliest Triassic (Li et al., 2002, 2005a, 2005b, 2009), Triassic (Zhang et al., 2005, 2007a, 2007b) or the end-Permian to mid Triassic (Xiao et al., 2008, 2009), for which the arguments are two Late Permian radiolarian specimens from the Baleigong ophiolitic mélange (Li et al., 2002, 2005a, 2005b) and Triassic zircon U–Pb ages from the West Tian Shan high to ultrahigh pressure metamorphic

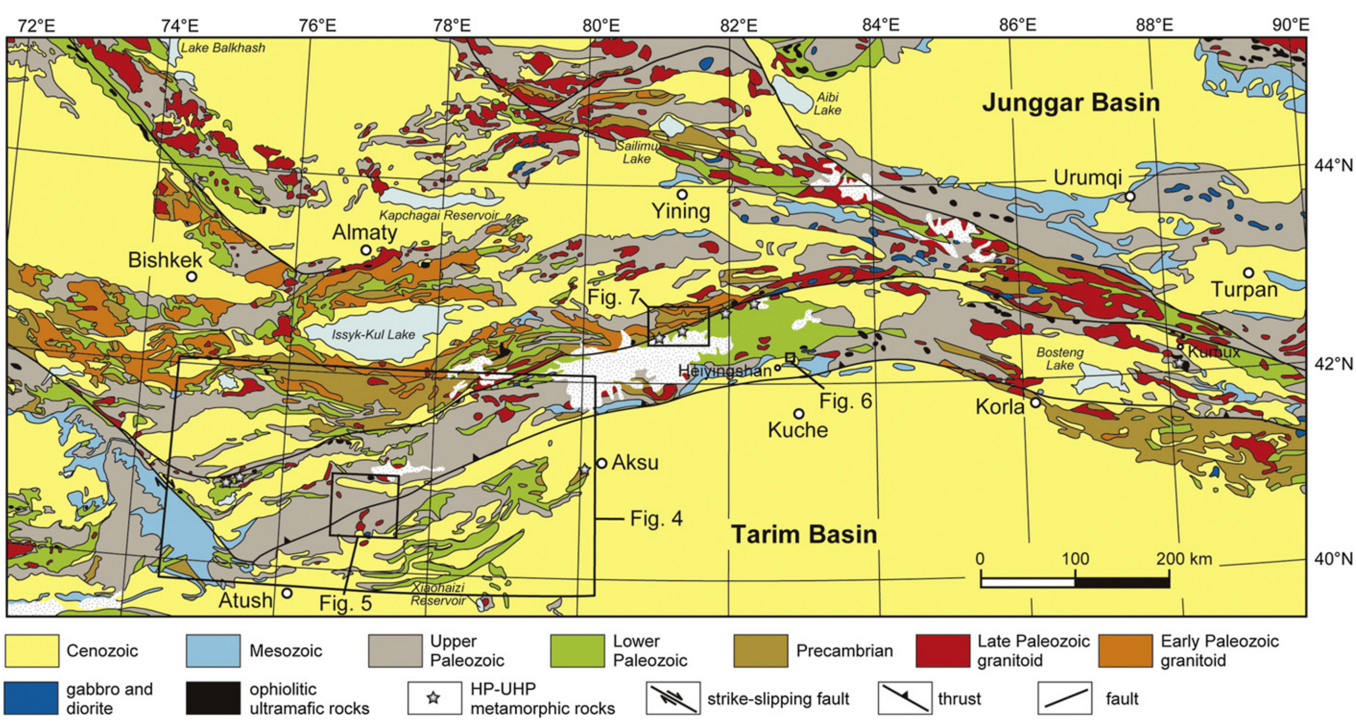

Fig. 2. Geological map of the western segment of the South Tian Shan Orogen and adjacent tectonic units (modified from IGCAGS, 2006). AISNQF — Atbashy–Inylchek–South Nalati–Qawablak Fault, TFF — Talas–Fergana Fault, NL — Nikolaev Line, and NTT — North Tarim Thrust.

rocks (Zhang et al., 2005, 2007a, 2007b). This has led to a hot debate on the timing of collision between the Tarim craton and the Kazakhstan–Yili terrane (Gao et al., 2006, 2009; de Jong et al., 2009; Han et al., 2010a; Hegner et al., 2010; Shu et al., 2011a; Wang et al., 2010a, 2011).

In addition to the controversy on the timing of collision, the subduction direction of the South Tian Shan Ocean is also in discussion (for details see Wang et al., 2010b). Based upon the structural, sedimentological, and magmatic data, most researchers (e.g., Windley et al., 1990; Allen et al., 1992; Xiao et al., 1992; Gao et al., 1993, 1998; Biske, 1995; Carroll et al., 1995; Chen et al., 1999a; Liu, 2001; Zhou et al., 2001; Bykadorov et al., 2003; Xiao et al., 2004; Yang and Wang, 2006; Zhang et al., 2007a, 2007b; Burtman, 2008; Pickering et al., 2008; Gao et al., 2009; Yang and Zhou, 2009; Zhu et al., 2009; Biske and Seltmann, 2010; Hegner et al., 2010; Su et al., 2010; Dong et al., 2011) favor the northward subduction of the South Tian Shan Ocean beneath the Kazakhstan–Yili terrane, which was followed by collision of the Tarim craton with the Kazakhstan–Yili terrane and finally resulted in the South Tian Shan Orogen. Alternatively, a southward subduction model is also proposed for the South Tian Shan Ocean. This model is mainly based on kinematic analyses of deformation structures, in which N-vergent (D1) thrust sheets are documented in the Chinese Central and South Tian Shan (Charvet et al., 2007, 2011; Wang et al., 2007a, 2010a, 2011; Lin et al., 2009), but those contrast with S-vergent (D1) thrust sheets in the Kyrgyzstan South Tian Shan (Biske and Seltmann, 2010), so further structural studies are needed to confirm the two competing tectonic models (Hegner et al., 2010). Additionally, the Chinese Central Tian Shan is considered as either part (Charvet et al., 2007, 2011; Wang et al., 2007a, 2010a, 2011; Lin et al., 2009) or whole (see Fig. 1A in Wang et al., 2010b) of the South Tian Shan Orogen, it was separated by the South Tian Shan back-arc basin from the Tarim carton during the Middle Silurian to Late Devonian. The final collision between the Yili Block and the Central Tian Shan and concurrent closure of the South Tian Shan back-arc basin took place during the Late Devonian to Early Carboniferous (Charvet et al., 2007, 2011; Wang et al., 2007a, 2010a, 2010b, 2011; Lin et al., 2009). Whether the Chinese Central Tian Shan is part of the South Tian Shan Orogen strongly depends upon tectonic division. If the Atbashy–Inylchek–South Nalati–Qawablak Fault is accepted as the northern boundary of the South Tian Shan Orogen (e.g., Burtman, 2006, 2008; Gao et al., 2009; Biske and Seltmann, 2010; Dong et al., 2011), the Chinese Central Tian Shan, regardless of a separate block (Charvet et al., 2007, 2011; Wang et al., 2007a, 2010a, 2011; Lin et al., 2009) or a continuing strip (He et al., 2001; Qian et al., 2009; Gao et al., 2009), had been amalgamated together with the Yili Block before the Middle Silurian (Zhang et al., 2007a, 2007b; Burtman, 2008; Gao et al., 2009; Yang and Zhou, 2009; Biske and Seltmann, 2010; Hegner et al., 2010; Dong et al., 2011). They are together referred to as the 'Yili–Central Tian Shan' terrane (Zhang et al., 2007b; Burtman, 2008; Yang and Zhou, 2009; Hegner et al., 2010). In other words, the Chinese Central Tian Shan was part of the Kazakhstan–Yili terrane in the Late Paleozoic. In this case, two Silurian to Early Carboniferous magmatic belts along the southern and northern margins of the Yili–Central Tian Shan terrane may be related to southward subduction of the North Tian Shan Ocean (Han et al., 2010b, and references therein; Dong et al., 2011) and northward subduction of the South Tian Shan Ocean (Yang and Wang, 2006; Zhang et al., 2007a, 2007b; Gao et al., 2009; Zhu et al., 2009; Xu et al., 2010; Dong et al., 2011), respectively.

On the basis of integrating various available data into a collisional model and critically evaluating the Triassic zircon U–Pb ages for high to ultrahigh pressure metamorphic rocks (Zhang et al., 2005, 2007b, 2007c) and the Late Permian radiolarian specimens (Li et al., 2002, 2005a, 2005b), both of which are often cited as the key evidence for the end-Permian to Triassic collision of the western segment of the South Tian Shan Orogen (Li et al., 2002, 2005a, 2005b; Zhang et al., 2005, 2007b, 2007c) and even for the Northern Xinjiang (Xiao et al., 2008, 2009), this work focuses on the timing of the collision between the Tarim craton and the Kazakhstan–Yili terrane in the western segment of the South Tian Shan Orogen and then simply discusses the termination time of the accretion–collision processes in the Northern Xinjiang.

2. Methodology

Orogens around the world may be divided into two fundamental types: collisional and accretionary (Murphy and Nance, 1991; Windley, 1992). Accretionary orogens develop during long-term subduction and plate convergence without the collision of continental blocks and usually contain little reworked older crust (Windley, 1992), but collision and suturing of largely juvenile terranes with oceanic affinities (ophiolites, island arcs, oceanic plateaus, etc.) may cause successive juxtaposition and amalgamation of these terranes into the margin of the continental block (Dickinson, 2008). The timing of orogenic events in accretionary orogens can be broadly defined by the ages of post-accretion stitching plutons and regional unconformities (Jones et al., 1983; Schermer et al., 1984; Gardner et al., 1988; Gray and Foster, 1997; Dickinson, 2008).

In contrast, collisional orogens involve oceanic subduction followed by collision of two continental blocks at the completion of a Wilson Cycle. During subduction of the oceanic crust beneath one continental block, part of the overlying sediments can be decoupled from the oceanic crust and accreted onto the margin of the overriding plate to form an accretionary wedge; subduction-related magmatism occurs only along the active margin of the overriding plate. Meanwhile, passive margin sedimentation develops on a passive margin of the shelf of the subducting plate. Collision of two continental blocks may result in involvement of remnants of oceanic lithosphere and materials from opposing margins in the collisional zone. Finally, the collisional zone may be intruded by plutons induced by either slab breakoff of the subducting plate or delamination of thickened mantle lithosphere beneath the overriding plate that was thickened by shortening and arc magmatism (e.g., Mikolaichuk et al., 1997; Whalen et al., 2006), and the oldest post-collisional stitching plutons can provide an upper-age bound for the collisional event (e.g., Han et al., 2010b, and references therein).

Continental collision commonly does not produce abundant hydrothermal solutions or H_2O-bearing melts rising into and forming a coeval volcanic–plutonic arc (Ernst, 2010), and high to ultrahigh pressure metamorphism within collisional orogens is seldom accompanied by coeval island-arc volcanic and plutonic rocks, whereas post-collisional or late-stage A-type granitic plutons are common (Liou et al., 2009). In the Himalayan collisional orogen, the high to ultrahigh pressure metamorphism at 52 to 46 Ma (see Liou et al., 2004, and references therein) were concurrent with the India–Asia collision at ~65–70 to ~45–40 Ma (e.g., Yin and Harrison, 2000; Mo et al., 2007). In contrast to the scarcity of syn-collisional magmatic rocks, post-collisional plutons are widespread in the Tethys Himalaya and the Higher Himalaya on the passive margin of the Indian plate (e.g., Simpson et al., 2000; Zhang et al., 2004a, 2004b; Searle et al., 2005; Chung et al., 2005; Mo et al., 2007, and references therein) and in the earlier Gandese arc (Hou et al., 2004; Chung et al., 2005; Mo et al., 2007, and references therein), almost concurrent with the Miocene low and medium pressure, high temperature metamorphism in the Himalaya (e.g., Simpson et al., 2000; Kohn and Parkinson, 2002; Liou et al., 2004; Searle et al., 2005) and adjacent South Karakorum (Rolland et al., 2001, 2009a, 2009b; Mahéo et al., 2002). The sequence of major tectonothermal events in the Himalayan Orogen has important implications for other collisional orogens.

The sequence of major tectonothermal events in collisional orogens may be the major manifestations of an abrupt change in deep dynamics as a natural consequence of ocean closure, such as slab breakoff or delamination of thickened lithospheric mantle within collisional zone and in neighboring tectonic domains (Bird, 1979; Sacks and Secor, 1990; Nelson, 1991; Davies and von Blanckenburg, 1995; Schott and Schmeling, 1998; Yin and Harrison, 2000; Rolland et al., 2001, 2009a, 2009b; Coulon et al., 2002; Kohn and Parkinson, 2002; Mahéo et al., 2002; Brouwer et al., 2004; Liou et al., 2004; Leech et al., 2005; Lustrino, 2005; Massonne, 2005; Whalen et al., 2006; Massonne et al., 2007). Possibly, the delamination process could have profound influences on a larger area beyond collisional zones and even resulted in occurrences of delamination magmatism not only in collisional zones but also in previous arc and passive margin, whereas the effects of slab breakoff could be limited to collisional zones.

The western segment of the South Tian Shan Orogen is the collisional zone between the Tarim craton and Kazakhstan–Yili terrane and it may record the sequence of tectonothermal events similar to the Himalayan Orogen. Particularly, the collision can be bracketed by the youngest ophiolitic mélanges and the oldest post-collisional stitching plutons within the collisional zone. This should be compatible with chronological data for arc magmatism, high to ultrahigh pressure and low pressure, high temperature metamorphism and continental sedimentation.

3. Geological background

Previous studies show that the Kazakhstan–Yili terrane was amalgamated prior to the Early Silurian (Biske, 1995; Lomize et al., 1997; Filippova et al., 2001; Bazhenov et al., 2003; Burtman, 2006; Long et al., 2007; Konopelko et al., 2008; Qian et al., 2009; Gao et al., 2009). The Kazakhstan–Yili terrane is mainly composed of Late Precambrian to Paleozoic rocks (Fig. 2). In the Kyrgyzstan North Tian Shan, the Late Precambrian to Early Paleozoic igneous rocks are widespread and have $^{207}Pb/^{206}Pb$ zircon ages of 1090 to 404 Ma (Mikolaichuk et al., 1997) and U–Pb zircon ages of 844 to 435 Ma (SHRIMP and LA-ICP-MS, Konopelko et al., 2008; Glorie et al., 2010; Rojas-Agramonte et al., 2010), with the oldest zircon U–Pb age of 2200 ± 50 Ma (Djenchuraeva et al., 2008). The age of the main deformation and metamorphism is constrained between 508 and 470 Ma (the latest Cambrian to earliest Ordovician) during which a large accretion–collision belt was formed (Rojas-Agramonte et al., 2010). In southern Kazakhstan between Bishkek and the Lake Balkhash (Fig. 2), Archean and Proterozoic granitic gneisses have zircon ages of 2791 ± 24 to 742 ± 1 Ma, and Early Paleozoic arc magmatism generated the dacites and granodiorites with zircon ages of 534 to 480 Ma (SHRIMP and single-zircon evaporation, Kröner et al., 2007; Alexeiev et al., 2011). High pressure metamorphism occurred at 490 Ma as dated by metamorphic zircons from garnet pyroxenite (SHRIMP, Alexeiev et al., 2011).

The Late Precambrian granitic gneisses in the Yili–Central Tian Shan terrane (Fig. 2) have zircon age of 798 and 882 Ma (TIMS, Chen et al., 1999b, 2000b), 969 ± 11 Ma (SHRIMP, Yang et al., 2008), and 926 ± 8 and 948 ± 8 Ma (SHRIMP, Chen et al., 2009). In addition, Early Paleozoic granitoids, generally showing a gneissic texture, are widespread across the Yili–Central Tian Shan terrane and they intrude into Meso- to Neoproterozoic metamorphic rocks and sometimes into Ordovician to Silurian volcano-sedimentary rocks. For example, gneissic and mylonitized granitoids have zircon U–Pb ages of 442 ± 4 to 396 ± 4 Ma (SHRIMP, Zhu and Song, 2006; Yang and Wang, 2006; Yang et al., 2006b; Shi et al., 2007), and amphibolites have zircon ages of 455 ± 3 and 451 ± 5 Ma (SHRIMP, Hu et al., 2008). Other zircon ages for the Early Paleozoic granitoids in the Yili–Central Tian Shan terrane are in a range of 447 to 395 Ma (Hopson et al., 1989; Han et al., 2004; Xu et al., 2006).

There are two magmatic belts along the northern and southern margins of the Yili–Central Tian Shan terrane. In the southern belt, the Middle Silurian to Early Carboniferous arc plutons produced by the northern subduction of the South Tian Shan Ocean have zircon ages of 485 to 352 Ma (LA-ICP-MS and SHRIMP, Long et al., 2007; Qian et al., 2009; Gao et al., 2009; Yang and Zhou, 2009; Wang et al., 2007b; Xu et al., 2010). One Late Carboniferous granite pluton with slight deformation in the belt yielded a zircon age of 313 ± 4 Ma (LA-ICP-MS, Wang et al., 2007b). This arc magmatic belt is intruded by Permian undeformed granitic plutons with zircon ages of 296 to 276 Ma (LA-ICP-MS, Gao et al., 2009; this study). The northern belt is composed of the Silurian to Early Carboniferous (440 to 325 Ma) arc plutons generated by the southern subduction of the North Tian Shan Ocean, but it is intruded by the Late Carboniferous to Early Permian (316 to 270 Ma) post-collisional granitoid plutons (see Han et al., 2010b, and references therein; Dong et al., 2011). In addition, the Visean unconformity between the Early Carboniferous and underlying older sequences is locally preserved in the Chinese Central Tian Shan (Charvet et al., 2007, 2011).

The Tarim craton is one of the largest Precambrian continental blocks in China, underlain by Neoarchean to Paleoproterozoic metamorphic basement with zircon ages of 2830 to 1900 Ma (TIMS and SHRIMP, Lu et al., 2008, and references therein; Guo et al., 2003; Deng et al., 2008; Shu et al., 2011b). Those granitic gneisses, supercrustal rocks, and plutons were covered by Mesoproterozoic sedimentary and volcanic deposits, and they were all affected by the latest Mesoproterozoic to Middle Neoproterozoic tectonothermal events (Lu et al., 2008, and references therein). These events are commonly thought to be related to the Rodinian supercontinent cycle (Guo et al., 2005; Deng et al., 2008; Lu et al., 2008; Zhu et al., 2008a; Xu et al., 2009; Zhang et al., 2009c). In the northwestern Tarim craton, Late Precambrain through Devonian strata include shallow marine to nonmarine limestone and sandstone, which are unconformably overlain by the Carboniferous and Permian deposits (Carroll et al., 1995). Particularly, the Permian continental deposits (Liao et al., 1990; BGMRXUAR, 1993), bimodal dykes (Yang et al., 2007) and a few plutons occur in the northern Tarim craton (Fig. 3a).

The western segment of the South Tian Shan Orogen is located between the Kazakhstan–Yili terrane to the north and the Tarim craton to the south and bounded by the Atbashy–Inylchek–South Nalati–Qawablak Fault in the north and by the North Tarim Thrust in the south (Fig. 1b). The Atbashy–West Tian Shan high to ultrahigh pressure metamorphic belt occurs along the Atbashy–Inylchek–South Nalati–Qawablak Fault, and many ophiolitic mélanges disperse in the orogen (Fig. 3b). As a whole, the western segment of the South Tian Shan Orogen may be considered as an accretion–collision belt (e.g., Windley et al., 2007) without participation of island arcs in the orogen (e.g., Zhang et al., 2007a, 2007b; Burtman, 2008; Gao et al., 2009; Yang and Zhou, 2009; Biske and Seltmann, 2010), only involving passive margin sediments and fragments of the Tarim craton, remnants of the South Tian Shan oceanic lithosphere, and active margin materials of the Kazakhstan–Yili terrane. The passive margin sediments of the Tarim craton are dominated by clastic rocks and limestone and they generally occur in the southern part of the orogen, where the Middle Devonian limestone unconformably overlies the Silurian and contain abundant coral and brachiopod fossils (Zhou and Chen, 1990; Zhang et al., 2004a, 2004b), the Lower Carboniferous clastic rocks and limestone contain fragmentary plant remains and abundant brachiopod, coral, bivalve, and gasteropod fossils, but the Upper Carboniferous bioclastic limestone appears as faulted blocks (Zhou and Chen, 1990). The active margin materials of the Kazakhstan–Yili terrane occur in the northern part of the orogen (Biske, 1995; Biske and Shilov, 1998; Gao et al., 1998; Burtman, 2006, 2008; Konopelko et al., 2008; Yang and Zhou, 2009; Biske and Seltmann, 2010), commonly as thrust sheets over the passive margin sediments (Gao et al., 1998; Bakirov and Kakitaev, 2000; Burtman, 2008; Biske and Seltmann, 2010). Probably, the Proterozoic metamorphic rocks with zircon ages of 1950 ± 30 (Biske and Shilov, 1998) and 707 ± 13 Ma (TIMS, Chen et al., 2000a) were fragments of the Tarim craton, which

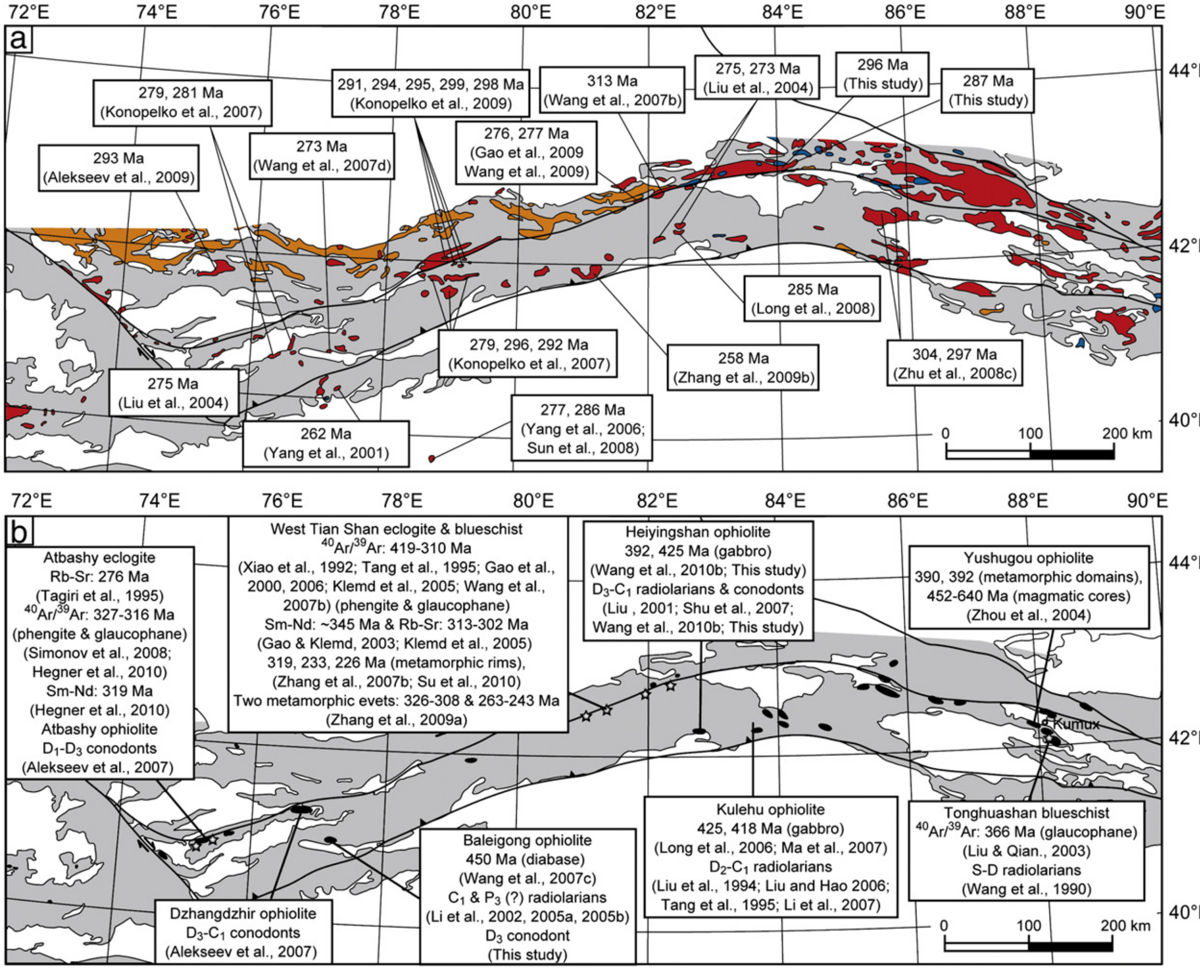

Fig. 3. Distributions of (a) ophiolitic rocks and (b) Paleozoic plutons in the western segment of the South Tian Shan Orogen and adjacent tectonic units, only the Late Carboniferous and Permian age data from the literature are shown for clarity (see text for details). Symbols as Fig. 2.

were involved in the western segment of the South Tian Shan Orogen during collision.

In the western segment of the South Tian Shan Orogen, the Permian is almost dominated by continental deposits (Zhou and Chen, 1990; BGMRXUAR, 1993; Bakirov and Kakitaev, 2000), followed by successive Triassic–Jurassic continental sequences (BGMRXUAR, 1993; Bakirov and Kakitaev, 2000; Li et al., 2004). The Permian granitic plutons are widespread (Fig. 3a) and continental volcanic rocks occur in places, but there have been no records of the Triassic–Jurassic magmatism (see below).

4. Ages of ophiolitic mélanges

Ophiolitic mélanges are widely dispersed in the western segment of the South Tian Shan Orogen (Fig. 3b). In the Kyrgyzstan South Tian Shan, the ophiolitic mélanges mainly occur along the east-northeast-striking Atbashy–Inylchek Fault (Fig. 4). The Aigyrbulak ophiolitic mélange occurs east of the Talas–Ferganan Fault and consists of mylonitized apoharzburgite serpentinites, flaser gabbro-amphibolites, basic volcanic rocks and cherts. The lavas contain Early Devonian coral-bearing limestone beds (Burtman, 2008). The Akbeit ophiolitic mélange includes serpentinized pyroxenites and peridotites, gabbro and gabbro-amphibolites and Devonian conodont-bearing cherts, tholeiitic basalts, hyalobasalts and their tuffs metamorphosed to greenschists and blueschists (Burtman, 2008). These rocks were locally thrust over the Permian molasse (Khristov et al., 1978). The Sarybulak ophiolitic mélange is composed of serpentinized peridotites, cumulate gabbroids, tholeiitic pillow basalts, and cherts (Burtman, 2008). The cherts are interbedded with basalts and yielded a plenty of Pragian to Famennian conodonts (Alekseev et al., 2007). The Dzhanydzher ophiolitic mélange is made up of serpentinized peridotites, gabbros, basalts, cherts, greenschists, and ophiolitoclastic breccia-conglomerates (Burtman, 2008). The cherts yielded Middle Devonian to Tournaisian conodonts and radiolarians (Alekseev et al., 2007; Burtman, 2008). The Karaarcha ophiolitic mélange comprises serpentinized ultrabasites, wehrlite–pyroxenite–gabbro cumulates, parallel dikes, basalts, hyaloclastites and cherts (Burtman, 2008).

In the Chinese South Tian Shan, the ophiolitic mélanges mainly occur in the Baleigong, Heiyingshan, Kulehu, and Kumux from west to east (Fig. 3a), and the former three may be the thrust sheets over the passive margin sediments of the northern Tarim craton. The Baleigong ophiolitic mélange (Fig. 5) consists of serpentinized peridotites, diabase-gabbros, basalts, and cherts, with metagreywacks and marls. Several serpentinized peridotite blocks occur along fault and the tholeiitic volcanic rocks have an OIB affinity (Wang et al., 2007c). The diabase-gabbro yielded a zircon age of 450 ± 2 Ma (LA-ICP-MS, Wang et al., 2007c), the cherts interbedded with basalts contain the Late Devonian index conodont *Palmatolepis* sp. (this study), abundant Early Carboniferous and one single Late Permian (*Albaillella* sp. cf. *excelsa* Ishiga, Kito and Imoto) radiolarians (Li et al., 2005a, 2005b). A similar assemblage of Early Carboniferous radiolarians was reported from another section about 15 km east of the first location, but no Permian radiolarians were found (Li et al., 2005a, 2005b), and the limestone only contains Early Carboniferous conodonts *Spathognathadus*, *Polygnathus communis communis*, and *Siphonodella duplicate* (this study).

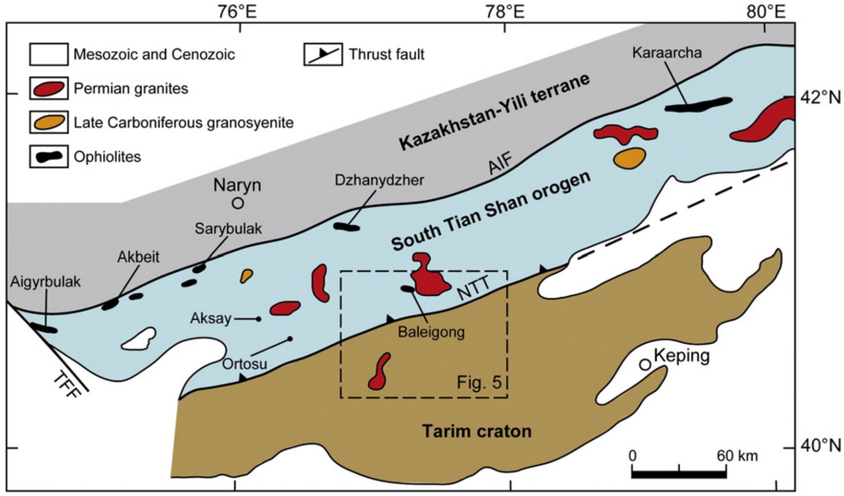

Fig. 4. Sketch map showing distributions of main ophiolites and the latest Carboniferous and Permian granitoid plutons in China–Kyrgyzstan contiguous regions (modified from Burtman, 2008). AIF — Atbashy–Inylchek Fault, TFF — Talas–Fergana Fault, and NTT — North Tarim Thrust.

From another location, about 25 km southeast of the first location, Li et al. (2002) also reported a similar radiolarian assemblage dominated by abundant Early Carboniferous radiolarians, with only one single species of Late Permian *Albaillella* sp. cf. *excelsa* Ishiga, Kito and Imoto (?).

The Heiyingshan ophiolitic mélange occurs in the Misibulake in the west (see Fig. 2 of Wang et al., 2011) and the Mandaleke in the east (Fig. 6). According to Shu et al. (2007) and Wang et al. (2011), the Misibulake ophiolitic mélange is composed of serpentinized peridotite, cumulate gabbro, sheeted diabase dikes, and massive or pillowed basalt, together with radiolarian chert, limestone and sandstone blocks. A cumulate gabbro yielded a zircon U–Pb age of 392 ± 5 Ma (LA-ICP-MS, Wang et al., 2011), two mylonitic metapelite samples yielded muscovite $^{40}Ar/^{39}Ar$ plateau ages of 359 and 356 Ma (Wang et al., 2011), and representative Famennian to Visean radiolarian species were obtained from the siliceous muddy matrix of the Misibulake ophiolitic mélange (Shu et al., 2007; Wang et al., 2011). The volcanic rocks show the chemical affinity to OIB and N-MORB and the mafic rocks display supra-subduction signatures (Wang et al., 2011). Similarly, the Mandaleke ophiolitic mélange includes serpentinized peridotite, foliated gabbro, and pillowed or massive basalt, together with thin-bedded chert, limestone, and amphibolite blocks (Fig. 6). A foliated gabbro sample yielded a TIMS zircon U–Pb age

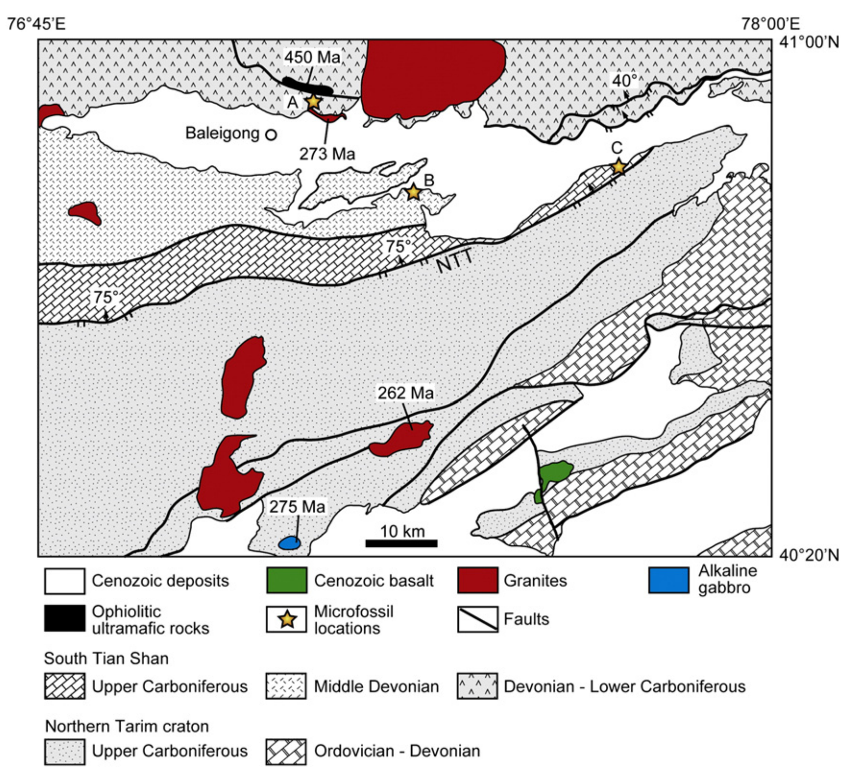

Fig. 5. Sketch geological map of the Baleigong and adjacent areas across the western segment of the South Tian Shan Orogen and the northern Tarim craton in the south, showing that the Baleigong ophiolitic mélange, in which the diabase yielded a zircon U–Pb age of 450 Ma (Wang et al., 2007c), is intruded by the K-feldspar granite pluton with a zircon U–Pb age of 273 Ma (Wang et al., 2007d). This age is similar to those of the K-feldspar granite and alkaline gabbro plutons (Liu et al., 2004) in the northern Tarim craton. The Early Carboniferous radiolarian fossils were found in locations A, B, and C, whereas the two suspected Late Permian radiolarian specimens were separately reported from locations A and C (Li et al., 2002, 2005a, 2005b). Also, the Late Devonian conodont *Palmatolepis* sp. was obtained from the chert interbedded with basalt in location A and the Early Carboniferous conodonts *Spathognathadus*, *Polygnathus communis communis*, and *Siphonodella duplicate* were separated from limestone in location B (this study). NTT — North Tarim Thrust.

of 425 ± 5 Ma and an amphibolite sample yielded an amphibole $^{40}Ar/^{39}Ar$ plateau age of 371 Ma (this study). In addition to a plenty of Tournaisian to Visean radiolarian assemblages (Liu, 2001), the chert also yielded representative Late Devonian to Early Carboniferous conodonts such as *Ancyrodella* sp., *Palmatolepis* sp., and *Polygnathus* sp. (this study). Some bioclastic limestone blocks are enclosed in the chert and they yielded the Middle Devonian brachipoda (*Acrospirifer*? sp.) and fragments of coral, atrypoid, rhynchonelloid, and abundant crinoid stems, and bryozoan (this study).

The Kulehu ophiolitic mélange is composed of diabase-gabbros, tholeiitic massive and pillowed basalts, and thin- to thick-bedded cherts, together with limestone blocks. The basaltic lavas have an N-MORB affinity (Long et al., 2006). A great number of Middle Devonian to Early Carboniferous radiolarians have been recognized from the cherts (Liu et al., 1994; Liu and Hao, 2006; Li et al., 2007a). Two gabbro samples yielded zircon ages of 425 ± 8 Ma (SHRIMP, Long et al., 2006) and 418 ± 3 Ma (LA-ICP-MS, Ma et al., 2007), respectively. Some limestone blocks contain Middle to Late Ordovician and Early Silurian brachipoda and conodonts such as *Paderodus* cf. gracilis Branson & Mehl, 1933 (this study). A gneissic granodiorite pluton to the east of the Kulehu ophiolitic mélange, which shows arc affinity and has a zircon age of 382 ± 6 Ma, is interpreted as part of thrust sheets from the southern arc magmatic belt of the Yili–Central Tian Shan terrane (SHRIMP, Zhu et al., 2008c).

The Kumux ophiolitic mélange occurs in the Yushugou, Tonghuashan, and Liuhuangshan from northwest to southeast, including peridotites, cumulate gabbros, and intermediate-basic volcanic rocks. Generally, these rocks are faulted blocks enclosed in Silurian to Devonian metasedimentary rocks and they are crosscut by several granitic plutons. One granodiorite pluton with arc affinity intrudes the Silurian strata and its zircon age is 424 ± 1 Ma (TIMS, Zhang et al., 2007a). In the Yushugou, the ophiolitic mélange is a ~10 km long and 1–3 km wide tectonic slice extending in NW–SE direction and mainly

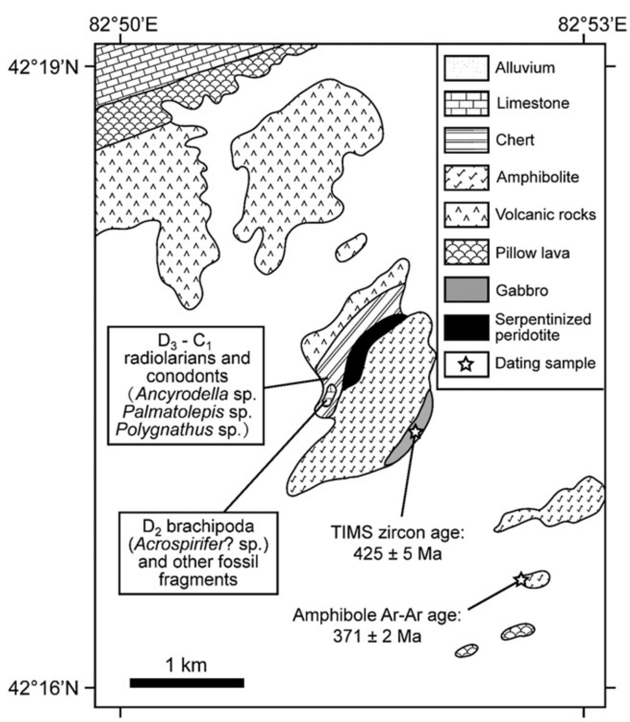

Fig. 6. Sketch geological map of the Madaleke ophiolitic mélange as the eastern part of the Heiyingshan ophiolitic mélange. The Madaleke ophiolitic mélange yielded abundant Late Devonian to Early Carboniferous conodonts and radiolarians (Liu, 2001; this study), the same as the Misibulake ophiolitic mélange (Shu et al., 2007; Wang et al., 2011) in the western part of the Heiyingshan ophiolitic mélange. These fossils consistently suggest that the Heiyingshan ophiolitic mélange was emplaced after the Visean.

consists of peridotites, two-pyroxene granulites, and intermediate-basic granulites and amphibolites, with minor metasedimentary interlayers, and protoliths of the metamorphic rocks are mafic cumulates, tholeiitic basalts, and clastic rocks (Wang et al., 1999). The peridotites represent the residual mantle after basaltic melts were extracted, and the mafic rocks show the chemical affinity to N-MORB (Dong et al., 2001). Metamorphic domains of zircons from two granulite sample yielded ages of 392 ± 7 and 390 ± 11 Ma, respectively, and ages from residual magmatic cores vary from 640 to 452 Ma (Zhou et al., 2004). In the Tonghuashan–Liuhuangshan area, the Lower Carboniferous is in fault contact with the Upper Devonian, the ophiolitic rocks and small blueschist blocks (Gao et al., 1993) mainly occur in the Devonian sequences. The cherts yielded Silurian to Devonian radiolarians (Wang et al., 1990) and crossite from the blueschist gave a Famennian $^{40}Ar/^{39}Ar$ age of 361 ± 2 Ma (Liu and Qian, 2003).

All the ophiolitic mélanges in the west segment of the South Tian Shan Orogen show a youngering trend from east to west. This is consistent with a diachronous, scissors-like closing process of the South Tian Shan Ocean (Chen et al., 1999a; Dong et al., 2011).

5. Ages of two contrasting types of metamorphism

There are two contrasting types of metamorphic belts in the western segments of the South Tian Shan Orogen and neighboring Kazakhstan–Yili terrane. One is a blueschist–eclogite–facies metamorphic belt and the other is a granulite–facies metamorphic belt. The granulite–facies metamorphic belt occurs locally in the southernmost edge of the Kazakhstan–Yili terrane, on the northern side of the Atbashy–Inylchek–South Nalati–Qawablak Fault (Fig. 7), in which the two-pyroxene granulite and cordierite–garnet–sillimanite gneiss underwent low pressure, high temperature metamorphism at P = 0.5 to 0.6 GPa and T = 681 to 705 °C (Li and Zhang, 2004) at 299 ± 5 Ma (SHRIMP zircon U–Pb age, Li and Zhang, 2004).

The blueschist–eclogite–facies metamorphic belt occurs on the southern side of the Atbashy–Inylchek–South Nalati–Qawablak Fault (Figs. 2 and 7). In addition to the small blueschist blocks in the Akbeit ophiolitic mélange of Kyrgyzstan (Khristov et al., 1978) and in the Tonghuashan ophiolitic mélange of China (Fig. 3b; Gao et al., 1993; Liu and Qian, 2003), the high to ultrahigh pressure metamorphic rocks mainly occur in the West Tian Shan of China and the Atbashy of Kyrgyzstan (Fig. 1b). The Atbashy high pressure metamorphic rocks include eclogites (Tagiri et al., 1995), glaucophane-schists and associated greenschists (Khristov et al., 1978; Dobretsov et al., 1987; Sobolev et al., 1989). The eclogite occurs as tectonic lenses and small discrete blocks in schists and quartzites and they have the geochemistry similar to mid-oceanic ridge basalts and oceanic plateau basalts (Sobolev et al., 1989; Simonov et al., 2008; Hegner et al., 2010). In addition to a Rb–Sr mineral-whole rock isochron age of 267 ± 5 Ma (Tagiri et al., 1995) and phengite $^{40}Ar/^{39}Ar$ plateau ages of 324 to 327 Ma (Simonov et al., 2008), a statistically robust Sm–Nd mineral-whole rock isochron age of 319 ± 4 Ma and a phengite $^{40}Ar/^{39}Ar$ cooling age of 316 ± 3 Ma are recently obtained from one single eclogite sample, suggesting the docking of the Tarim Craton with the Kazakhstan–Yili terrane during the Late Carboniferous (Hegner et al., 2010).

The West Tian Shan high to ultrahigh pressure metamorphic rocks (Fig. 7) include coesite-bearing eclogites (Lü et al., 2008), glaucophane-schists and associated greenschists (Gao et al., 1999; Gao and Klemd, 2003a, 2003b; Ai et al., 2006; Li et al., 2007b). Eclogite occurs as pods, boudins, thin layers or massive blocks within the blueschist, and a transition from eclogite to blueschist and pillow structures are well preserved (Gao et al., 1999, 2000; Gao and Klemd, 2003a, 2003b). Blueschist also occurs as small discrete blocks, lenses, bands and thick layers within greenschist-facies metasediments (e.g., Gao and Klemd, 2001; Klemd et al., 2002; Wei et al., 2003; Lin and

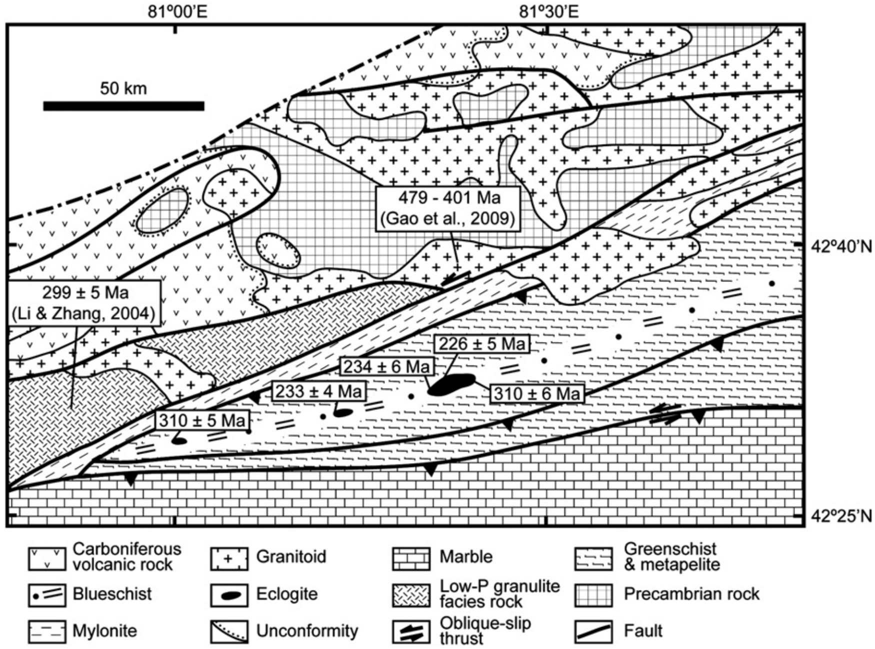

Fig. 7. Sketch geological map showing the distributions of blueschist–eclogite–facies granulite–facies metamorphic belts and the relationship between a stitching granitic pluton and high to ultrahigh pressure metamorphic rocks in the West Tian Shan, western China (modified from Zhang et al., 2007b). Note that the greenschist and metapelite also underwent high to ultrahigh-pressure metamorphism (Wei et al., 2009). Zircon U–Pb ages for eclogite are from Zhang et al. (2007b).

Enami, 2006; Gao et al., 2007; de Jong et al., 2009). Geochemically, the high to ultrahigh pressure metamorphic rocks have the OIB, E-MORB, and N-MORB affinity (Ai et al., 2006; Zhang et al., 2007b). A recent study reveals that the greenschist-facies rocks are subducted metasediments and they also underwent high to ultrahigh pressure prograde and post-peak retrograde metamorphism controlled by dehydration reactions followed by fluid-absent conditions (Wei et al., 2009).

For these eclogites and blueschists, many Sm–Nd and Rb–Sr isochron and $^{40}Ar/^{39}Ar$ ages have been obtained (Xiao et al., 1992; Gao et al., 1995, 2000, 2006; Tang et al., 1995; Gao and Klemd, 2003a, 2003b; Klemd et al., 2005; Wang et al., 2007b; Zhang et al., 2009a; Hegner et al., 2010; Li et al., 2010) but they show a large variation (Fig. 3b). Zircons from the eclogite have core-ages of 413 to 310 Ma and rim-ages of 234 to 226 Ma (SHRIMP, Zhang et al., 2007b, 2007c; also see Fig. 3b). The rodingitized eclogites have zircon U–Pb ages of 422 to 291 Ma (SHRIMP, Li et al., 2010). The youngest ages are assigned to mark a Triassic collision in the western segment of the South Tian Shan Orogen (Zhang et al., 2007b, 2007c). New SIMS zircon rim U–Pb ages as well as Lu–Hf and Sm–Nd mineral-whole rock isochron ages gave two separate metamorphic stages for the eclogitic rocks: one in the Carboniferous (327 to 309 Ma) and the other in the Permian to Triassic (263 to 243 Ma), but which date is due to peak high to ultrahigh pressure metamorphism is still uncertain (Zhang et al., 2009a). Other SIMS zircon U–Pb dating on two eclogite samples yielded three distinct age populations: the youngest concordia ages of 319.9 ± 2.7 and 319.8 ± 2.7 Ma from the metamorphic rims are suggested to record the Late Carboniferous peak metamorphism under high to ultrahigh pressure conditions, in addition to 2450 ± 31 to 1956 ± 26 Ma from the inherited cores and 486 ± 8 to 428 ± 6 Ma from the oscillatory inner zones (Su et al., 2010). Similarly, a SIMS rutile U–Pb age of 318 ± 7 Ma from the eclogite may record the time when the rocks cooled down to the closure temperature of Pb in rutile (ca. 500 °C), which may be very close to the time of eclogite–facies metamorphism due to fast exhumation of eclogite (Li et al., 2011). The new age data from the eclogite suggest a Late Carboniferous collision in the South Tain Shan Orogen (Hegner et al., 2010; Su et al., 2010; Li et al., 2011).

The two contrasting types of metamorphic belts are considered together as the paired metamorphic belts: the granulite–facies metamorphic belt was formed during the northern subduction of the South Tian Shan Ocean in the Early Permian, whereas the blueschist–eclogite–facies metamorphic belt was generated during the collision between the Tarim craton and the Kazakhstan–Yili terrane in the Triassic (Li and Zhang, 2004; Zhang et al., 2007b, 2007c).

6. Ages and tectonic implications of Permian plutons

Permian plutons are extensively developed in the western segment of the South Tian Shan Orogen and adjacent tectonic units (Fig. 3a). These plutons are intruded into Paleozoic sedimentary strata (Gao et al., 1998) and Late Paleozoic thrust sheets (Solomovich and Trifonov, 2002). In the Kyrgyzstan South Tian Shan, the Permian A-type rapakivi granites and leucogranites have zircon ages of 299 to 279 Ma (SHRIMP, Mao et al., 2004; Konopelko et al., 2007, 2009) and Rb–Sr isochron ages of 279 to 269 Ma (Solomovich and Trifonov, 2002). In the Chinese South Tian Shan, the Permian granitoids are generally massive, including nepheline syenites, aegirine syenites, diorites, granites, two-mica granites and alkali-feldspar granites. The nepheline and aegirine syenites have U–Pb zircon ages of 279 to 273 Ma (TIMS, Liu et al., 2004), and granites have zircon ages of 304 to 258 Ma (TIMS, SHRIMP and LA-ICP-MS, Jiang et al., 1999; Wang et al., 2007b; Long et al., 2008; Zhu et al., 2008b; Zhang et al., 2009b). Furthermore, the Permian granitoids also occur in the northern Tarim craton and the Kazakhstan–Yili terrane (Figs. 2 and 3a). In the southern part of the Kazakhstan–Yili terrane, a slightly deformed granite and a diorite enclave separately gave zircon ages of 291 ± 5 Ma and 294 ± 5.3 Ma (SHRIMP, Konopelko et al., 2009), one undeformed quartz syenite pluton, which intrudes into the Precambrian rocks, has a zircon age of 275 ± 3 Ma (LA-ICP-MS, Gao et al., 2009), another undeformed alkali-feldspar granite, which crosscuts the Late Carboniferous volcanic rocks with a zircon age of 313 ± 4 Ma (SHRIMP, Zhu et al., 2005), has a zircon age of 287 ± 3 Ma (LA-ICP-MS, this study). In the Kyrgyzstan North Tian Shan, two undeformed granodiorite plutons yielded zircon U–Pb ages of 280 ± 9 and 268 ± 1 Ma (Abeleira et al., 2000), and another granodiorite produced a zircon U–Pb age of 292 ± 4 Ma (SHRIMP, Glorie et al., 2010). A granite pluton in the Kyrgyzstan Middle Tian Shan has a zircon U–Pb age of 291 ± 1 Ma (TIMS, Alekseev et al., 2009). In the northern Tarim craton, the Xiaohaizi (also called Mazhashan) syenite pluton, which intrudes into the Silurian to Lower Carboniferous strata, has zircon ages of 277 ± 4 (SHRIMP, Yang et al., 2006a), 286 ± 3 (SHRIMP, Sun et al., 2008), 281 ± 4, and 282 ± 3 Ma (LA-ICP-MS, Li et al., 2007c), the Huoshibulak alkali-feldspar granite and an alkaline gabbro plutons, which crosscut the Upper Carboniferous to Permian strata, have zircon ages of 262 and 275 Ma (TIMS, Yang et al., 2001; Liu et al., 2004), respectively.

In addition, the Lower Permian rhyolite in the Chinese South Tian Shan has a $^{40}Ar/^{39}Ar$ age of 282 ± 2 Ma (Hendrix et al., 1992), which is roughly coeval with the basalts with a $^{40}Ar/^{39}Ar$ plateau age of 279 ± 2 Ma (Chen et al., 1997) and a zircon U–Pb age of 275 ± 13 Ma (LA-ICP-MS, Li et al., 2007c), diabase with a zircon age of 272 ± 6 Ma (Li et al., 2007c), and dacite and rhyolite from drill cores with a zircon U–Pb age range of 291 to 272 Ma (LA-ICP-MS, Li et al., 2007c; Tian et al., 2010) in the northern Tarim craton.

Permian plutons are spatially dispersed throughout the western segment of the South Tian Shan Orogen (Fig. 3a). They are generally undeformed and locally have intrusive relationships with ophiolitic mélanges (Wang et al., 2007b) and the Late Paleozoic thrust sheets (Solomovich and Trifonov, 2002), implying that they temporally post-dated the collision between the Tarim craton and the Kazakhstan–Yili terrane. Therefore, these plutons are usually interpreted as products of post-collisional magmatism (Solomovich and Trifonov, 2002; Mao et al., 2004; Konopelko et al., 2007, 2009; Solomovich, 2007; Wang et al., 2007d; Zhu et al., 2008b; Han et al., 2010a). More importantly, the Permian plutons also occur in other tectonic units in the neighborhood of the South Tian Shan Orogen (Fig. 3a). If the South Tian Shan is a Triassic orogen, the widely accepted northward subduction model (e.g., Windley et al., 1990, 2007; Gao et al., 1998, 2009; Yang and Wang, 2006; Zhang et al., 2007b, 2007c; Burtman, 2008; Konopelko et al., 2008; Yang and Zhou, 2009; Biske and Seltmann, 2010; Dong et al., 2011) may predict that Permian plutons might have been generated by subduction-related magmatism and therefore they must have been confined only to the southern margin of the Kazakhstan–Yili terrane. However, the spatial distribution of Permian plutons is not compatible with such a tectonic model of Triassic orogen. Importantly, the Permian plutons intrude into the collision zone and passive margin of the Tarim craton (Fig. 3a). Within-collision-zone magmatism may be induced by slab breakoff after collision (e.g., Coulon et al., 2002; Whalen et al., 2006) and the earliest magmatic age must have postdated slab breakoff (Davies and von Blanckenburg, 1995). Therefore, the Permian plutons in the western segment of the South Tian Shan Orogen must have been emplaced after collision, and the oldest plutons with zircon ages of ~300 Ma (Fig. 3a) can place an upper-age bound for the time of collision.

7. Permian molasse

Occurrence of molasse is one of the crucial evidence for constraining the timing of collisional orogeny. In the Atbashy Range of Kyrgyzstan, the Permian molasse are widespread (Khristov et al., 1978; Bakirov and Kakitaev, 2000) and locally they unconformably overlie the eclogite-bearing Choloktor complex (Tagiri et al., 1995; Tursungaziev and Petrov, 2008; Hegner et al., 2010). More importantly, the molasses contain not only the latest Carboniferous *Fusulina*-bearing limestone conglomerates (Hegner et al., 2010) but also abundant eclogite gravels and clasts (Baslakunov et al., 2007). This is compatible with the Devonian to Early Carboniferous conodonts from the Atbashy ophiolitic mélange (Alekseev et al., 2007).

In the Chinese South Tian Shan, the oldest molasse were deposited in the Early Permian and overlain by the Upper Permian to Triassic continental deposits (Shu et al., 2007). Besides, the Early Permian molasse also occurs in the Kazakhstan–Yili terrane (BGMRXUAR, 1993). These molasse deposits imply the pre-Permian collision in the South Tian Shan Orogen. This is in good agreement with the following facts: a foreland basin along the northern Tarim craton started to develop at the end of the Late Carboniferous (Allen et al., 1999), the nonmarine Upper Permian through Triassic sandstone and conglomerate unconformably overlie the older rocks (Carroll et al., 1995), and their paleocurrent analysis and sediment-dispersal patterns consistently indicate a trend away from the Tian Shan (Hendrix et al., 1992). Furthermore, Mesozoic sandstone from the northern Tarim craton is diverse in composition and lithic-rich, mainly has a recycled-orogen provenance from the Tian Shan (Hendrix, 2000), and the Triassic in the southern slope of the South Tian Shan is characterized by polymictic conglomerates and sandstones (Li et al., 2004).

8. Discussion

Most of pre-existing data from the ophiolitic mélanges, magmatic, metamorphic and sedimentary rocks in the western segment of the South Tian Shan Orogen and adjacent tectonic units indicate the possible closure of the South Tian Shan Ocean during the Late Carboniferous, except for the Late Permian radiolarian *Albaillella* sp. cf. *excelsa* Ishiga, Kito and Imoto (?) from the Baleigong ophiolitic mélange (Li et al., 2002, 2005a, 2005b) and zircon ages of 234 to 226 Ma from the West Tian Shan eclogite (Zhang et al., 2007a, 2007b), both are the key lines of evidence for the end-Permian to Triassic collision (Zhang et al., 2007b, 2007c; Xiao et al., 2008, 2009; Li et al., 2009).

8.1. Evaluation of Late Permian radiolarian specimens

If the collision between the Tarim craton and the Kazakhstan–Yili terrane occurred in the Late Carboniferous, it seems to severely conflict with the two Late Permian radiolarian specimens from the Baleigong ophiolitic mélange (Li et al., 2002, 2005a, 2005b).

Out of twenty-six representative radiolarians from the Baleigong area (Li et al., 2002, 2005a, 2005b), two Silurian–Devonian through Permian and five Carboniferous through Permian long-lived species are meaningless for the age of the ophiolitic mélange, other seventeen radiolarians are definitely of Early Carboniferous age, and only two were identified as the Late Permian specimens: one is suspected *Albaillella excelsa* Ishiga, Kito and Imoto (?) (Li et al., 2002) and the other is supposed to be suspected *Albaillella excelsa* Ishiga, Kito and Imoto (?) (Li et al., 2005a) or *Albaillella* sp. cf. *excelsa* Ishiga, Kito and Imoto (Li et al., 2005b). However, the two Late Permian radiolarian specimens were critically suspected. Shu et al. (2007) noted that both specimens are hard to be recognized correctly, because they are poorly preserved, the first (Li et al., 2002) has no ventral wing and the second (Li et al., 2005a, 2005b) has no ventral wing too, with a suspected dorsal wing. Therefore, the two specimens are unable to convincingly demonstrate the presence of a Late Permian ocean in the western segment of the South Tian Shan Orogen (Shu et al., 2007).

The Baleigong mélange is crosscut by several alkali-feldspar granite plutons (Fig. 5), and one of the plutons directly intrudes into the section (location A of Fig. 5) from which one of the Late Permian radiolarian specimens was obtained (Li et al., 2005a, 2005b). This undeformed pluton has a zircon age of 273 ± 2 Ma (Wang et al., 2007d), corresponding to the Kungurian Stage of Early Permian (Gradstein et al., 2008) and showing post-collisional granite affinity (Wang et al., 2007d). Such a crosscutting relationship between the granite pluton and the ophiolitic mélange strongly supports the conclusion of Shu et al. (2007) and convincingly indicates that the closure of the South Tian Shan Ocean in the Baleigong area occurred before the Middle Permian (Wang et al., 2007d). This is also in good agreement with the youngest radiolarians of Early Carboniferous

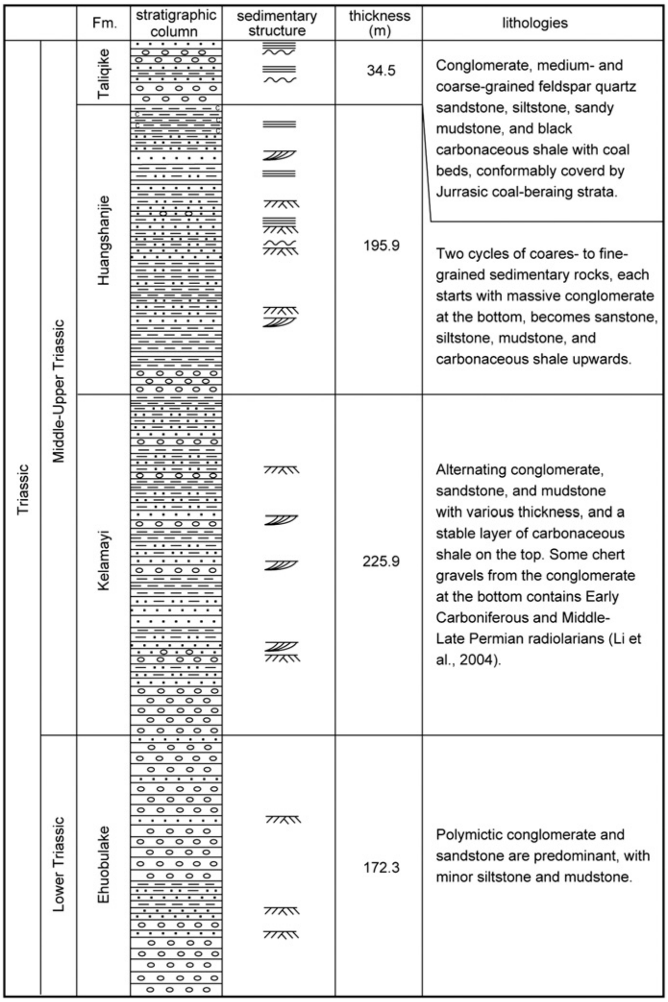

Fig. 8. Representative stratigraphic column for the Triassic continental deposits in the western segment of the South Tian Shan of China (simplified from Li et al., 2004). Note that there are no volcanic records during the Triassic.

age from other ophiolitic mélanges within the western segment of the South Tian Shan Orogen.

It is noteworthy that other two Middle–Late Permian radiolarian species, *Follicucullus* sp. and *Follicucullus* sp. aff. *Follicucullus bipartitus* Caridroit et DeWever (Li et al., 2004), were also cited in support of Triassic collision (Zhang et al., 2007b, 2007c). These radiolarians were separated from chert gravels of Middle Triassic molasses, while the gravels are thought to come from the ophiolitic mélanges of the South Tian Shan Orogen (Li et al., 2004). Unfortunately, except for the two suspected Late Permian radiolarian specimens from the Baleigong ophiolitic mélange as discussed above, no other Permian radiolarian species have been found in all of the other ophiolitic mélanges in the western segment of the South Tian Shan Orogen. In the Aksay, west of the Baleigong ophiolitic mélange (Fig. 4), the Silurian graptolite shales are overlain by a sequence of Early Devonian to Early Moscovian condensed sediments, among which the prevailing cherts contain Late Devonian to Serpukhovian radiolarians and conodonts, and gigantic limestone blocks in flysch contain the Ludlow to Early Moscovian brachiopods, corals, and foraminifers (Burtman, 2008). In the Ortosu, the Devonian siliceous and clayey shales are overlain by thick volcano-sedimentary strata in which the carbonate siliceous sediments contain Late Visean to Early Moscovian foraminifers (Burtman, 2008). Moreover, radiolarian-chert grains are very common in the Permian and Triassic sandstone of the northern Tarim craton (Hendrix, 2000; Li et al., 2004). In any case, microfossils from the ophiolitic mélanges as well as the Permian to Triassic molasse and continental sediments are not in support for the presence of a Permian ocean in the South Tian Shan. Furthermore, Triassic sequences in the western segment of the South Tian Shan Orogen are polymictic conglomerates, sandstones, siltstone, shale, and minor coal-beds with abundant plant fossils, without marine or volcanic rocks (BGMRXUAR, 1993; Fig. 8). This just suggests an intracontinental setting during the Triassic.

All these data consistently indicate that the western segment of the South Tian Shan Ocean had been already closed before the Permian.

8.2. Constraints on the time of high to ultrahigh pressure metamorphism

Another line of the key evidence for Triassic collision in the western segment of the South Tian Shan Orogen comes from two U–Pb zircon ages: 234 ± 7 Ma and 226 ± 4.6 Ma are interpreted as the times of peak high to ultrahigh pressure metamorphism and blueschist–facies retrograde metamorphism (Zhang et al., 2007b, 2007c), respectively. Regarding the earlier $^{40}Ar/^{39}Ar$, Rb–Sr and Sm–Nd ages for the high to ultrahigh pressure metamorphic rocks, Zhang et al. (2007a) argue that the phengite or glaucophane $^{40}Ar/^{39}Ar$ ages may be problematic due to the unknown excess argon in these metamorphic minerals, and the Sm–Nd and Rb–Sr isochron ages may be invalid because most of porphyroblastic garnets have zonation of major element compositions, which implies that the chemical equilibrium had not been reached homogeneously. If so, the Rb–Sr isochron age of 267 ± 5 Ma for the Atbashy eclogite (Tagiri et al., 1995) is invalid too, because the porphyroblastic garnet in the dated sample also shows strong zonation of major element compositions (Tagiri et al., 1995). Unfortunately, this age is often cited as support for Triassic collision in the western segment of the South Tian Shan Orogen (Zhang et al., 2007b; Li et al., 2009).

Because of no petrographic evidence, such as index metamorphic mineral inclusions of coesite, omphacite, and garnet in the dated zircon domains, Zhang et al. (2007b) also noted that the Triassic zircon ages do not definitely indicate the exact ages of either peak high to ultrahigh pressure metamorphism or blueschist–facies retrograde metamorphism.

In fact, the Permian to Triassic metamorphic ages of 263 to 226 Ma (Zhang et al., 2007b, 2009a, 2009b, 2009c) are much younger than the new ages recently obtained for the Atbashy and West Tian Shan eclogites (Hegner et al., 2010; Su et al., 2010; Li et al., 2011). Noticeably, the metamorphic zircon rims containing omphacite, phengite and rutile inclusions yielded a consistent concordia age of 319 Ma for the West Tain Shan eclogite (Su et al., 2010). It is well known that mineral inclusions in metamorphic zircons can give useful information on zircon growth conditions and provide an excellent link between zircon formation and P–T conditions, and thus between age and metamorphism (Tichomirowa et al., 2005; Rubatto and Hermann, 2007). Omphacite, phengite and rutile inclusions demonstrate that the metamorphic zircon rims grew during the peak high to ultrahigh pressure metamorphism and thus their U–Pb ages consistently indicate that the peak high to ultrahigh pressure metamorphism occurred at the end of the Early Carboniferous (Su et al., 2010). This is consistent with the rutile U–Pb age of 318 Ma (Li et al., 2011) and the Sm–Nd isochron age of 319 Ma (Hegner et al., 2010). These ages are much older than the Triassic U–Pb ages of the zircons containing no omphacite and phengite inclusions (Zhang et al., 2007b).

On the other hand, it is difficult to determine if zircon growth occurred during either peak high to ultrahigh pressure or retrograde metamorphism without index mineral inclusions in the dated zircon domains, because metamorphic zircon may occur at different times during petrogenesis (Hoskin and Black, 2000) and thus those zircon ages may provide information on the history of cooling from high temperatures (Roberts and Finger, 1997; Tichomirowa et al., 2005; Harley et al., 2007). Furthermore, most metamorphic zircon rims may form nearly instantaneously during fluid influx and hence zircon rim U–Pb ages may date such events (Ayers et al., 2003). In fact, eclogites from the western segment of the South Tian Shan Orogen did undergo fluid metasomatism at medium pressure (Simonov et al., 2008) or under greenschist-facies conditions (Li et al., 2007b) during exhumation. For example, zircons from partially and completely rodingitized eclogites due to fluid metasomatism show clear fluid channels from rim to core and yielded complex age patterns: a well-defined rim age of 422 ± 10 Ma and varied core ages of 422 to 291 Ma, and the youngest age of 291 Ma is from the hydrothermal zircons formed during pervasive rodingitization (Li et al., 2007b, 2010). If the hydrothermal zircons formed under greenschist-facies conditions during exhumation (Li et al., 2007b, 2010), the peak high to ultrahigh pressure metamorphism must have occurred prior to the hydrothermal event. This implies that the Permian to Triassic zircon ages (Zhang et al., 2007b, 2009a, 2009b, 2009c) need further investigations for their origin and tectonic significance (Hegner et al., 2010; Li et al., 2011). Alternatively, they may be re-interpreted as a result of younger fluid-induced recrystallization (Gao et al., 2006, 2009; de Jong et al., 2009).

In collisional orogens, high to ultrahigh pressure metamorphism was usually followed by low- and medium-pressure, high-temperature metamorphism and associated magmatism, such as the western and central Alps (e.g., Brouwer et al., 2004). In the Himalaya, Eocene high to ultrahigh pressure metamorphism (Parrish et al., 2006; Leech et al., 2005, 2007; Guillot et al., 2008; and references therein) was coeval with the Indian–Asian collision (Yin and Harrison, 2000; Liou et al., 2004; Yin, 2006; Leech et al., 2007), but Miocene medium-pressure, high-temperature metamorphism and associated leucogranites (Kohn and Parkinson, 2002, and references therein) in the Himalaya and adjacent South Karakorum (Rolland et al., 2001, 2009a, 2009b; Mahéo et al., 2002) are generally attributed to an abrupt change in deep dynamics, i.e. re-heating caused by slab breakoff during exhumation (Yin and Harrison, 2000; Rolland et al., 2001, 2009a, 2009b; Kohn and Parkinson, 2002; Mahéo et al., 2002; Brouwer et al., 2004; Leech et al., 2005). If similar processes occurred in the western segment of the South Tian Shan Orogen, the ~300 Ma low pressure, high temperature metamorphic rocks (Fig. 7; Li and Zhang, 2004) and coeval A-type granites and leucogranites with the oldest ages of ~300 Ma (Konopelko et al., 2007, 2009; Wang et al., 2007d; Zhu et al., 2008b) may have postdated the collision between the Tarim craton and the Kazakhstan–Yili terrane, the same as summarized by Liou et al. (2004, 2009) for collisional orogens. Therefore, an upper bound of ~300 Ma can be placed for the time of high to ultrahigh pressure metamorphism in the western segment of the South Tian Shan Orogen.

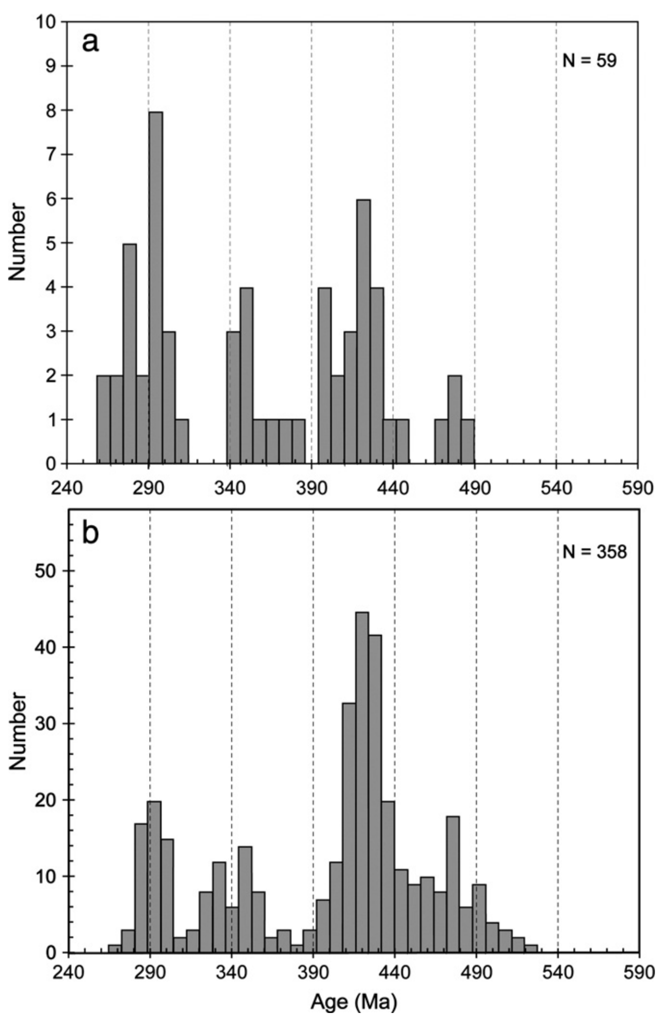

Fig. 9. Zircon age similarities between (a) Paleozoic plutons (data from the references cited in text) and (b) modern river sands (from Ren et al., 2010), showing rare Late Carboniferous and no Mesozoic magmatic records in the western segment of the South Tian Shan Orogen and adjacent southern arc magmatic belt of the Yili–Central Tian Shan terrane.

Regionally, the greenschist and metapelite around the high to ultrahigh pressure eclogite blocks in the West Tian Shan are crosscut by a granitic stitching pluton (Fig. 7). The relationship between the pluton and metamorphic rocks suggests that the metamorphic events must have occurred before the pluton emplacement. Although there has been no age constraint on the pluton until now, the fact that there are no records of Triassic to Jurassic magmatism in the western segment of the South Tian Shan Orogen and adjacent tectonic units (Fig. 9, Ren et al., 2010) suggests that the stitching pluton could be Permian or older in age. Along the N–S-trending Kekesu River profile, about 10 km east of Fig. 7, granitoid plutons in the southern magmatic belt of the Yili–Central Tian Shan terrane (Fig. 3a) have zircon ages of 433 to 338 Ma (Wang et al., 2007b; Gao et al., 2009; Xu et al., 2010), and one granitic pluton partially crosscutting the greenschist and metapelite in the high to ultrahigh pressure metamorphic belt yielded zircon ages of 276 ± 1 Ma (LA-ICP-MS, Gao et al., 2009) and 277 ± 3 Ma (LA-ICP-MS, Wang et al., 2009). About 25 km further east, another granitic pluton with a zircon age of 313 ± 4 Ma within the same magmatic belt is locally intruded into the greenschist and metapelite (see Fig. 2 in Wang et al., 2007b), but this Late Carboniferous pluton is slightly deformed (Wang et al., 2007b). Wei et al. (2009) confirm that the greenschist and metapelite around the eclogite blocks also underwent high to ultrahigh pressure and subsequent retrograde metamorphism. Recently, Gao et al. (2011) obtain a SIMS zircon U–Pb age of 285 ± 2 Ma for a NW–SE trending granite dike crosscutting the high to ultrahigh pressure metamorphic belt. All these data imply that the high to ultrahigh pressure metamorphism must have occurred prior to the Permian, probably during the Late Carboniferous. The slightly deformed granite (Wang et al., 2007b) may be one syn-collisional pluton.

It is known that high to ultrahigh pressure metamorphism in collisional orogens is seldom accompanied by coeval arc plutonism (Liou et al., 2004, 2009). In the western segment of the South Tian Shan Orogen and adjacent southern margin of the Kazakhstan–Yili terrane, the Late Carboniferous syn-collisional magmatism is rather rare (Fig. 9a) or even absent (Hegner et al., 2010). The Ordovician to Early Carboniferous plutons are predominant in the southern magmatic belt of the Yili–Central Tian Shan terrane. In contrast, the Permian plutons occur across the northern Tarim craton, South Tian Shan Orogen, and the Kazakhstan–Yili terrane from south to north (Fig. 3a). This provides further support for pre-Permian high to ultrahigh pressure metamorphism in the western segment of the South Tian Shan Orogen.

8.3. Mesozoic tectonomagmatic quiescence

According to the paradigm of plate tectonics, a collisional orogen generated by continent–continent collision is usually characterized by a series of magmatic pulses in different geodynamic settings. If the South Tian Shan is an end-Permian to Triassic collisional orogen (Li et al., 2002, 2005a, 2005b, 2009; Zhang et al., 2007a, 2007b; Xiao et al., 2009), it is expected at least to find a Late Carboniferous to Middle Permian arc magmatic belt along the southern margin of the Kazakhstan–Yili terrane, Late Permian to Late Triassic syn-collisional and Jurassic post-collisional magmatism in the western segment of the South Tian Shan Orogen and adjacent tectonic units, without Permian magmatism in the northern Tarim craton. In fact, up-to-date chronological data have not revealed any records of Triassic and Jurassic magmatism in the western segment of the South Tian Shan Orogen and adjacent tectonic units, and the Permian plutons are dispersed throughout the western segment of the South Tian Shan Orogen and even occur in the northern Tarim craton (Fig. 3a). Furthermore, a total of 500 detrital zircon U–Pb ages of modern river sands from the Tekes River contain no trace of Mesozoic magmatic events in China–Kyrgyzstan contiguous regions, among them the age spectra for the four branches running across the West Tian Shan high to ultrahigh pressure metamorphic belt and the southern magmatic belt of the Yili–Central Tian Shan terrane dominantly define Early Permian, Early Carboniferous, Late Silurian, and Early Ordovician peaks (Fig. 9b). All of these together suggest a major period of tectonomagmatic quiescence in the western segment of the South Tian Shan Orogen during the Mesozoic. This is not compatible with the Triassic collision model for the western segment of the South Tian Shan Orogen.

9. Tectonic evolution of the South Tian Shan Orogen

The Atbashy–West Tian Shan eclogites were derived from the subducted oceanic crust (Gao et al., 1999, 2000; Gao and Klemd, 2003a, 2003b; Zhang et al., 2007b, 2007c; Hegner et al., 2010) and the peak eclogite-facies metamorphism occurred at ~319 Ma (Hegner et al., 2010; Su et al., 2010; Li et al., 2011), whereas the Permian to Triassic

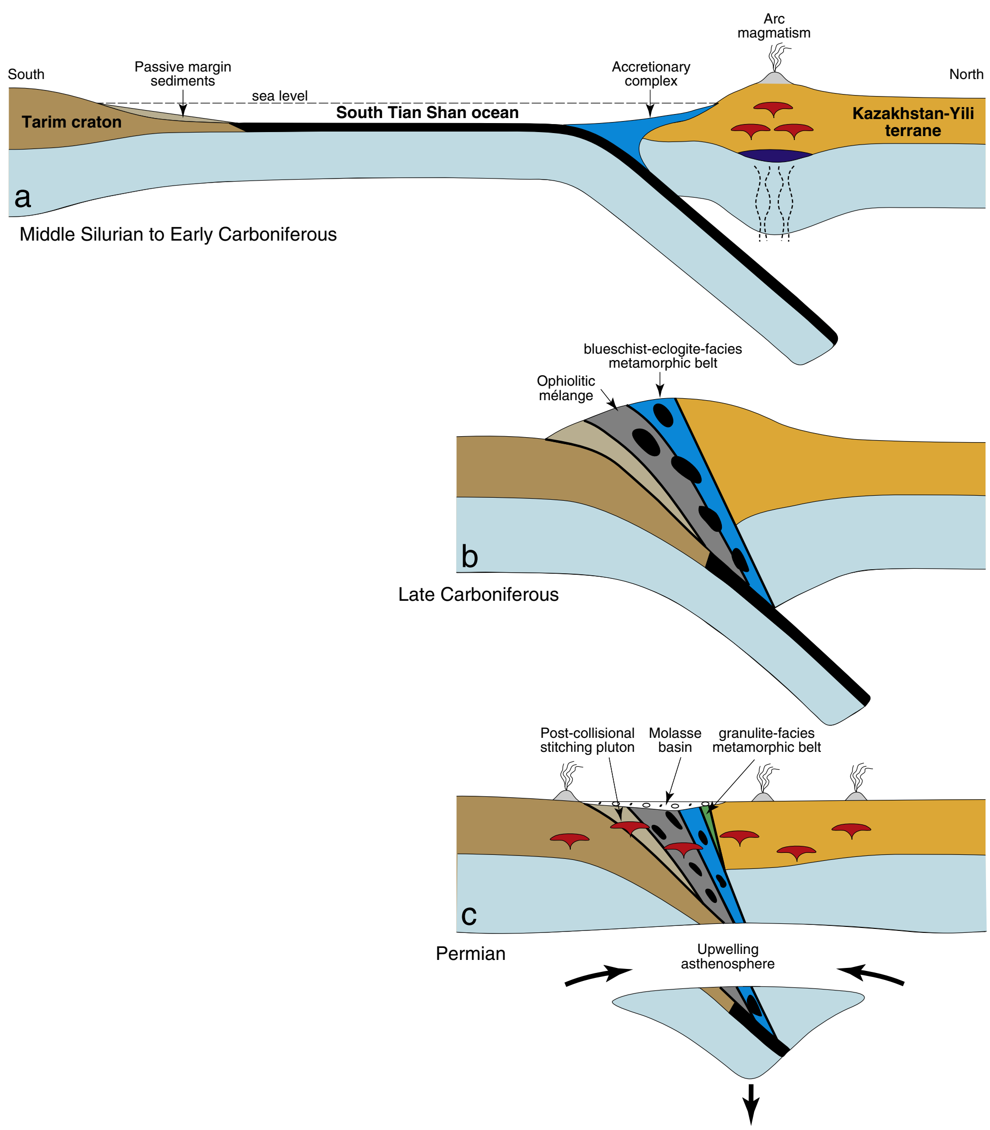

Fig. 10. Schematic model illustrating tectonic evolution of the western segment of the South Tian Shan Orogen in China–Kyrgyzstan contiguous regions. (a) Northward subduction of the South Tian Shan Ocean beneath the Kazakhstan–Yili terrane during the Middle Silurian to Early Carboniferous generated the southern arc magmatic belt of the Kazakhstan–Yili terrane; (b) exhumation of high to ultrahigh pressure metamorphic rocks during the Late Carboniferous collision between the Tarim craton and the Kazakhstan–Yili terrane, which were seldom accompanied by syn-collisional magmatism; and (c) Permian post-collisional plutonism, molasse sedimentation, and local low pressure, high temperature metamorphism. The post-collisional magmatism might be caused by delamination of the thickened lithospheric root, accompanied by upwelling of the asthenosphere, and the resultant plutons crosscut the high to ultrahigh pressure metamorphic belt, ophiolitic mélange and previous passive margin of the Tarim craton.

zircon ages for the eclogites (Zhang et al., 2007b; Li et al., 2010) may have resulted from younger fluid-induced recrystallization (Gao et al., 2006, 2009; de Jong et al., 2009; Li et al., 2010). This suggests that subduction of oceanic lithosphere was still ongoing at the end of the Early Carboniferous, and hence the collision between the Tarim craton and the Kazakhstan–Yili terrane must have occurred after that time. On the other hand, extensive Permian molasse, stitching plutons with the oldest ages of ~300 Ma and low pressure, high temperature metamorphism at 299 Ma in the western segment of the South Tian Shan Orogen and neighboring tectonic units definitely constrain the pre-Permian collision between the Tarim carton and the Kazakhstan–Yili terrane. Therefore, the final collision in the western segment of the South Tian Shan Orogen took place in the Late Carboniferous rather than the end-Permian to Triassic. This is in agreement with not only the well-preserved microfossils of Early Carboniferous age from the ophiolitic mélanges but also the scarcity of the Late Carboniferous plutons in the western segment of the South Tian Shan Orogen. The Late Carboniferous zircon U–Pb, mineral-whole rock Sm–Nd isochron, phengite Rb–Sr isochron and $^{40}Ar/^{39}Ar$ ages (Klemd et al., 2005; Zhang et al., 2007b, 2009a; Hegner et al., 2010; Wang et al., 2010a) may have provided the information on the cooling history and exhumation of the high to ultrahigh pressure metamorphic rocks.

The oldest ophiolites in the western segment of the South Tian Shan Orogen were generated at least in the Late Ordovician, possibly concurrent with the final amalgamation of the Kazakhstan–Yili terrane before the Silurian (Gao et al., 2009). This implies that the South Tian Shan Ocean was generated earlier. The ocean began to subduct northward beneath the Kazakhstan–Yili terrane prior to the Late Silurian (Gao et al., 2009) or earlier (Burtman, 2008), leading to a Middle Silurian to Early Carboniferous magmatic arc along the southern margin of the Kazakhstan–Yili Terrane (Fig. 10a; Gao et al., 2009; Zhu et al., 2009; Xu et al., 2010). The subducted oceanic materials underwent high to ultrahigh pressure metamorphism at the end of the Early Carboniferous, followed by the final closure of the South Tian Shan Ocean and the collision between the Tarim craton and the Kazakhstan–Yili terrane in the Late Carboniferous (Gao et al., 2009; Hegner et al., 2010; Su et al., 2010; Li et al., 2011). During the collision, the accretionary complex on the southern margin of the Kazakhstan–Yili terrane and ophiolitic mélanges were thrust over the passive margin sediments on the northern margin of the Tarim craton, with rather rare Late Carboniferous granitic plutonism and rapid exhumation of high to ultrahigh pressure metamorphic rocks (Fig. 10b). A significant geodynamic change from convergence to extension in the Tian Shan was initiated at ~300 Ma (de Jong et al., 2009; Su et al., 2010; Dong et al., 2011), which may be related to breakoff of the subducted oceanic slab or delamination of the thickened lithospheric root, accompanied by upwelling of the asthenosphere. Such a process may have provided significant heat and thus triggered partial melting of the lower crust and underlying lithosphere, leading to occurrences of the oldest stitching plutons within the collisional zone and coeval low pressure, high temperature metamorphism along the southern edge of the Yili–Central Tian Shan terrane, followed by widespread Permian molasse sedimentation, post-collisional magmatism and continental volcanic rocks across the South Tian Shan Orogen and adjacent tectonic units (Fig. 10c). If spatial distribution of the Permian plutonism in the northern Tarim craton is taken into account, a delamination model seems more feasible. During the Mesozoic, the amalgamated South Tian Shan Orogen and adjacent tectonic units are in a major period of tectonomagmatic quiescence, the Triassic polymictic and the Jurassic coal-bearing strata were successively deposited in an intracontinental environment.

10. Implications for the Northern Xinjiang

As discussed above, the suspected, poorly-preserved two Permian radiolarian specimens (Li et al., 2002, 2005a, 2005b) are incompatible with other geological data, and the Triassic zircons without index mineral inclusions from the high to ultrahigh pressure metamorphic rocks (Zhang et al., 2007b) are significantly younger than those containing omphacite and other mineral inclusions (Su et al., 2010) so they may have resulted from the subsequent hydrothermal effects and must have postdated the high to ultrahigh pressure metamorphism. Obviously, any such evidence cannot be cited as support for the end-Permian to Triassic final collision of the South Tian Shan Orogen (Li et al., 2009) and even for the end-Permian to mid-Triassic termination of the accretion–collision processes of the Northern Xinjiang (Xiao et al., 2008, 2009).

In the Northern Xinjiang, the Late Carboniferous collision between the Tarim craton and the Kazakhstan–Yili terrane was nearly concurrent with that between the Kazakhstan–Yili and Junggar terranes along the North Tian Shan suture zone at 325 to 316 Ma (Han et al., 2010a) and between the Altai and Kazakhstan terranes along the Irtysh–Zaysan suture zone at 321 to 307 Ma (Chen et al., 2010; Han et al., 2010a; and references therein). These events are younger than the amalgamations of the accretionary complexes and terranes in East and West Junggar before the Late Carboniferous (Han et al., 2010a, and references therein) and the collision in the eastern segment of the South Tian Shan Orogen (Charvet et al., 2007, 2011; Zhang and Guo, 2008; Dong et al., 2011). As a whole, the Northern Xinjiang was in a post-collisional setting during the Permian, characterized by widespread occurrences of A- and I-type granitoids (Han et al., 1997, 2006, 2010a; Su et al., 2006a, 2006b, 2008; Dong et al., 2011), continental deposits (Liao et al., 1990; BGMRXUAR, 1993) in all tectonic domains, and extension in the Tian Shan (de Jong et al., 2009; Shu et al., 2011b; Dong et al., 2011). Therefore, there were no the end-Permian to mid-Triassic accretion–collision processes in the Northern Xinjiang.

11. Conclusions

Formation of the South Tian Shan Orogen in Central Asia is a diachronous, scissors-like process, youngering from east to west, and the western segment of the South Tian Shan Orogen in China–Kyrgyzstan contiguous regions is the site of the final collision between the Tarim craton and the Kazakhstan–Yili terrane. Currently, every lines of evidence demonstrate that the western segment of the South Tian Shan Orogen finally formed after the high to ultrahigh pressure metamorphism of subducted oceanic material at the end of the Early Carboniferous. This is compatible with the youngest Early Carboniferous microfossils from the ophiolitic mélanges, providing the lower-age bound for the collision between the Tarim craton and the Kazakhstan–Yili terrane, whereas the oldest stitching plutons crosscutting the collisional zone place the upper-age bound of ~300 Ma for the collision. All these lead to the conclusion that the western segment of the South Tian Shan Orogen was formed in the Late Carboniferous rather than the end-Permian to Triassic. The Permian to Triassic zircon ages from the high to ultrahigh pressure metamorphic rocks may have resulted from later hydrothermal events at crustal levels after exhumation, and the suspected, poorly-preserved Late Permian radiolarian specimens previously reported are severely incompatible with other pieces of geological evidence. Furthermore, the western segment of the South Tian Shan Orogen and adjacent tectonic units all were in a major period of tectonomagmatic quiescence during the Triassic and Jurassic times, and this suggests no end-Permian to Triassic subduction and collision. In this case, the Triassic zircon ages for the high to ultrahigh pressure metamorphic rocks and Permian radiolarian specimens cannot be cited further as support for the end-Permian to Triassic collision in the western segment of the South Tian Shan Orogen in China–Kyrgyzstan contiguous regions. All these, together with available data from other tectonic domains of the Northern Xinjiang, also imply that the accretion–collision processes in the Northern Xinjiang were finally terminated during the Late Carboniferous.

Acknowledgments

We are grateful to Xiping Dong, Yuanlin Sun, and Jianbo Liu for their assistance in identifications of fossils. This paper benefited from valuable discussion with Chunjing Wei, Lifei Zhang and Jun Gao. Rob Van der Voo, An Yin, Wen-jiao Xiao and another anonymous reviewer are thanked for their detailed comments and constructive suggestions which have greatly helped us to improve our presentation and discussion. This work was supported by the Ministry of Science and Technology of China (Grant Nos. 2007CB411305 and 2001CB409802), the National Natural Science Foundation of China (Grant Nos. 41072168, 49872078, and 40772141) and China Geological Survey (Grant Nos. 1212010611803, 1212010611817, and 1212010050502).

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