Dunhuang Tectonic Belt in northwestern China as a part of the Central Asian Orogenic Belt: Structural and U-Pb geochronological evidence

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ABSTRACT

The Dunhuang Tectonic Belt (DTB) is located about 100 km south of the Beishan–Tianshan orogen in the Central Asian Orogenic Belt in NW China. It was previously considered as a part of the Tarim or North China craton.

Detailed structural analyses reveal two episodes of deformation in the central DTB, D1 and D2. D1 is a north-side-up reverse shear, and D2 a dextral strike slip. Mineral assemblages, microstructures and quartz C-axis patterns indicate that D1 deformation took place under amphibolite facies conditions (500 to 600ºC) and D2 mostly under greenschist-facies conditions (300–450ºC). U–Pb zircon dating of eight granitoid/intermediate intrusions (mostly dikes, with well constrained cross-cutting

relationships with the D1 and D2 structures) and an amphibolite gneiss indicates that D1 deformation took place before ca. 349 Ma and most likely at ca. 406 Ma, and D2 between ca. 249 Ma and ca. 241 Ma.

The DTB has a structural, metamorphic and magmatic signature in the Paleozoic–Mesozoic that is typical of an orogenic belt. It shares a similar geological history with the Beishan–Tianshan orogen and is likely a part of the Central Asian orogenic belt. The DTB and the Beishan-Tianshan orogen might represent two separate Paleozoic mountain belts that developed more or less synchronously on the south and north sides, respectively, of the last vestige of the Paleo-Asian Ocean before its terminal closure in the Permian. The D1 reverse shearing in the DTB is interpreted to be related to a Silurian–Devonian terrane accretion/collision and the D2 dextral strike slip to post-accretional/collisional movement among terranes in Late Permian–Middle Triassic time.

Keywords: Dunhuang Tectonic Belt; reverse shear; dextral strike slip; U–Pb zircon geochronology; Central Asian Orogenic Belt

1. Introduction

The Central Asian Orogenic Belt (CAOB, also referred to as the Altaids; Şengör et al., 1993; Wilhem et al., 2012) is one of the largest accretionary orogens on Earth (Fig. 1, inset). It is characterized by multiple Neoproterozoic–Late Paleozoic accretionary events which assembled island arcs, ophiolites, subduction–accretion complexes, seamounts and microcontinents along the southern margins of the Siberian and East European Cratons and the northern margins of the Tarim and North China Cratons (Windley et al., 2007; Xiao et al., 2010). The opening of the Paleo-Asian Ocean was initiated by at least 1.0 Ga (Khain et al., 2002) and the closure of the ocean terminated the accretionary history of the CAOB. Some researchers propose that the ocean was closed in the Early Paleozoic, and the CAOB subsequently underwent intraplate deformation, followed by continental rifting in the Permian (Zuo et al., 1990, 2003; He et al., 2005). However, evidence in Beishan orogen (Fig.

1B), including the formation of the Permian Liuyuan ophiolite complex, indicates that the Paleo-Asian Ocean continued to subduct in the Permian and orogenesis lasted until Triassic (Tian et al., 2013, 2015; Cleven et al., 2015; Xiao et al., 2010, 2015).

The Beishan–Tianshan orogen (Fig. 1) is traditionally considered to be the southernmost component of the CAOB (Xiao et al., 2010). More recently, the Dunhuang Tectonic Belt (DTB), located about 100 km south of the Beishan (Fig. 1), has been proposed as a part of the CAOB (Zhao et al., 2017; Shi et al., 2017). This interpretation is based on metamorphic, geochronological and geochemical data from the DTB (Zong et al., 2012; He et al., 2014; Peng et al., 2014; Zhao et al., 2016; Wang et al., 2016a, 2017a, 2017b), but a systematic structural study, a key to understanding any orogen, was lacking.

In this paper, we present results of a detailed field-based structural study in the DTB. We focus our attention to the Dashuixia valley–Qingshan area in the central part of the DTB (Figs. 2 and 3) where the rocks are well exposed but few structural data are available (Mei et al., 1997; Yu et al., 1998; Lu et al., 2008; Shi et al., 2017). We elucidate the kinematic history of the area using field and microstructural (including quartz c-axis) data and constrain the timing of deformation by U–Pb dating of pre-, syn- and post-tectonic intrusions and an amphibolite gneiss. We compare our results with the geological evolution of the Beishan–Tianshan orogen, and discuss their implications for the evolution of the DTB and the southern CAOB. We conclude that the DTB is a Paleozoic orogen that forms a part of the southern CAOB and that it developed on the south side of the last vestige of the Paleo-Asian Ocean before its final closure.

2. Geological setting

The DTB is situated immediately east of the Tarim Craton, between the Shulehe and the Altyn Tagh faults and the Beishan orogenic belt (Fig. 2). It covers an area of about 40,000 km^2 (Lu et al., 2006) and was previously referred to as the Dunhuang block. The DTB is dominated by extensive tonalite– trondhjemite–granodiorite intrusions (TTG) or TTG gneisses and metamorphosed supracrustal rocks (BGMRG, 1989; Mei et al., 1998; Lu et al., 2008). The metamorphosed supracrustal rocks have traditionally been referred to as the Dunhaung Group (BGMRG, 1989), dominated by sillimanite/kyanite-bearing metapelite, mafic granulite, amphibolite, quartzite and marble (Lu et al., 2008; Zong et al., 2012). The TTG gneisses have yielded Archean to Paleoproterozoic (ca. 3.1–1.85 Ga) ages (Zong et al., 2013; Long et al., 2014; Zhao et al., 2015a, 2015c). The presence of Paleoproterozoic (ca. 1.86–1.82 Ga) amphibolite to granulite-facies meta-mafic rocks with clockwise P–T paths suggests that the DTB was involved in Paleoproterozoic tectonothermal events, possibly related to the assembly of the Columbia supercontinent (Zhang et al., 2012, 2013; He et al., 2013; Wang et al., 2014).

Paleozoic metamorphic and magmatic rocks are widely distributed in the DTB (Zong et al., 2013; He et al., 2014; Wang et al., 2017b; Zhang et al., 2009; Meng et al., 2011; Zong et al., 2012; Peng et al., 2014; He et al., 2014; Zhao et al., 2016). The metamorphic rocks are mainly amphibolite- to high-pressure granulite-facies meta-mafic rocks, with metamorphic ages of ca. 440–400 Ma (Zong et al., 2012; He et al., 2014). Eclogite from a tectonic mélange in the southern DTB records a metamorphic event at ca. 428–391Ma (Wang et al., 2017a). Available data, including P–T paths of these metamorphic rocks, indicate that the DTB experienced subduction and subsequent rapid tectonic exhumation in Silurian–Devonian time (Zong et al., 2012; He et al., 2014; Peng et al., 2014). Paleozoic magmatism can be broadly divided into two distinct episodes, in the early Paleozoic and late Paleozoic, respectively. Early Paleozoic magmatism includes ca. 440 Ma TTG (Zhang et al., 2009) and ca. 430–410 Ma I-type granites (Zhao et al., 2017). Late Paleozoic magmatism includes Late Devonian I-type granites (Zhao et al., 2017) and Carboniferous adakitic intrusive rocks, mainly exposed in the southern part of the DTB (Zhu et al., 2014; Zhao et al., 2017; Bao et al., 2017). These adakitic intrusive rocks are suggested to have been generated by partial melting of a thickened lower crust (Zhu et al., 2014; Bao et al., 2017). In summary, the DTB experienced a major orogenic event(s) in the Paleozoic that may have continued to Late Devonian (Wang et al., 2016a, 2017a, 2017b) or middle Carboniferous (Zhao et al., 2017; Bao et al., 2017).

Rocks in the study area (Fig. 3) include Archean–Proterozoic basement (amphibolite gneiss and TTG gneiss) overlain by marble of unknown age, Silurian–Devonian high grade metamorphic rocks (amphibolite gneiss and TTG gneiss) and Permian–Triassic granitoid intrusions (280–240 Ma; our unpublished data).

3. Structural history

Our detailed structural analysis and overprinting relationships observed in the field reveal two major generations of deformation in the study area, D1 and D2 (see Fig. 4A-B for two examples of overprinting relationships), with contrasting kinematics, P–T conditions and timing of deformation. The main structural features of D1 and D2 are described in this section. Foliations and lineations associated with D1 are denoted as S1 and L1, and those with D2 as S2 and L2. To help constrain the deformation conditions and shear sense, we also did microstructural analysis and quartz-C axis measurements on selected samples.

Deformation is heterogeneous. For convenience of description, we divide the study area into three domains (A, B and C) based on lithology and structure (Fig. 3A). Rocks exposed in Domain A are mostly Archean–Proterozoic orthogneisses overlain by the marble. They are highly deformed and are characterized by a strong gneissosity or compositional layering that formed during D1 (Fig.5A) under amphibolite facies conditions (see below). Rocks exposed in Domain B mostly include Paleozoic TTG-like gneiss, amphibolite gneiss and some early Triassic granitoid plutons (Fig. 3A). Gneissosity similar to that in Domain A is also developed in Domain B. Rocks exposed in Domain C are mostly Permian granite that was strongly deformed during D2 under greenschist facies conditions (see below). The three domains are separated by two brittle faults (Fig. 3A) which coincide with prominent lineaments on satellite images. Foliation and lineations from the three domains are plotted in Fig. 3B.

3.1. D1: Top-to-SE reverse shear under lower amphibolite-facies conditions

D1 structures are best developed in Domain A, but also well developed in Domain B. S1 foliation is a well-developed gneissosity (Fig. 5A). It is defined by amphibolite-facies mineral assemblages (Fig. 4C-D), indicating D1 deformation took place under amphibolite facies conditions. S1 strikes ENE and dips steeply NW (Fig. 3B-a). A lineation, L1, is moderately to strongly developed on the foliation (Fig. 5B) and plunges steeply NW-W (Fig. 3B-a). It is defined by a preferred orientation of quartz, mica and hornblende crystals in the foliation, and by elongate feldspar augen. Shear sense indicators, such as σ- and δ-type feldspar porphyroclasts and S–C structure, indicate that D1 was associated with a top-to-SE (north-side-up) reverse shearing (Fig. 5C, D).

Microstructures of the D1 tectonites show that quartz is fully recrystallized, embedded in a polymineral matrix. Recrystallization was mainly accommodated by grain boundary migration mechanisms, with grains grown to medium size (0.5 mm-1 mm), consistent with deformation under amphibolite facies metamorphic conditions of 500–600 °C (Fig. 6A, C; Trouw et al., 2009). The polygonal granoblastic fabric in quartz is indicative of a static grain growth. The strain-free statically recrystallized quartz fabric also suggests that the temperature at the end of deformation remained at medium metamorphic conditions, probably close to the transition between greenschist and amphibolite facies (Trouw et al., 2009). Feldspar porphyroclasts mostly show undulose extinction and core-mantle structures due to partial recrystallization. Microstructures also indicate a top-to-SE (north-side-up) shear sense (Fig. 6A, B).

3.2.D2: Dextral strike-slip shearing under greenschist-facies conditions

D2 deformation is localized in the E-W-trending Qingshan shear zone (QSSZ; Domain C in Fig. 3A). The shear zone is over 3 km wide as exposed and is potentially much wider. It can be traced for over 24 km along strike before being covered beneath the desert. Mylonitic foliation in the shear zone is subvertical and stretching lineations are subhorizontal (Fig. 3B-d). They are defined by preferred orientation of quartz–feldspar aggregates, muscovite and biotite (Fig. 7A). Shear sense indicators are well developed and abundant, including S–C fabrics, σ- and δ-type feldspar porphyroclasts, shear bands (C’) and asymmetrically folded veins (Fig. 7B). They consistently indicate a dextral sense of shear.

D2 deformation is pervasively strong in Domain C and is much weaker in domains B and A. Near the northern boundary of Domain C, deformation is heterogeneous. Here, granitic intrusions vary from strongly to weakly deformed. In Domain B, granitoid intrusions show weak mylonitisation. Horizontal stretching lineations were developed in the quartz-feldspathic rocks (Fig. 8A). Mafic rocks deformed during D2 exhibit compositional layering with layers rich in biotite and/or hornblende alternating with those rich in quartz and feldspar (Fig. 8B). Some mafic rocks are boudinaged in the foliated granite (Fig. 8C). Strongly asymmetric folds indicate a dextral shear (Fig. 8D). Narrow D2 dextral shear zones occur in both Domains B and A (Fig. 3A).

In the granitic mylonite in the D2 Qingshan shear zone, quartz porphyroclasts display strong undulose extinction. Smaller new grains formed by bulging (BLG) recrystallization (Fig. 7C). Due to subgrain rotation recrystallization (SGR), recrystallized quartz grains in the quartz-rich band show a well-developed oblique foliation (Fig. 7D). The feldspar is mostly deformed by fracturing and separated into domino-type fragmented porphyroclasts (Fig. 7E). The narrow transition between protomylonite and ultramylonite is preserved, which is characteristic of low-grade mylonite (Trouw et al., 2009). All these features indicate that D2 deformation in the Qingshan shear zone took place under low-medium grade conditions (300–450°C) (Stipp et al., 2002a, 2002b; Passchier and Trouw, 2005). This conclusion is supported by the greenschist-facies mineral assemblages associated with D2 structures (Fig. 4E-F). Microscopic shear sense indicators such as K-feldspar porphyroclasts, shear bands and S–C fabric all indicate a dextral shear sense (Fig. 7D, E), consistent with the field observations described above.

Microstructures of the D2 tectonites in Domain B show that the polycrystalline quartz grains are large and very irregular in shape due to grain boundary migration recrystallization (Fig. 8E). K-feldspar porphyroclasts are characterized by bulging recrystallization indicating a deformation temperature of ~450°C (Fig. 8F, Passchier and Trouw, 2005). These features constrain the deformation temperature to 450–550°C.

3.3. Quartz C-axis fabric analysis

Quartz C-axes were measured by a universal stage on 11 oriented samples of tectonites, to help determine the shear sense and constrain the deformation temperature. The results are presented in Fig. 9. Pole figures show that the c-axis patterns are slightly asymmetric, which is consistent with a non-coaxial progressive deformation. A detailed review on deformation thermometry based on quartz c-axis fabrics and recrystallization microstructures is given in Law (2014).

Five samples of D1 tectonites from Domain A were measured (DHF 40-2, DHF39-1, DHF 37-2, DHF 41-6, DHF41-3). The pole figures (Fig. 9a–d) are characterized by single center girdles with dominant central maximum and secondary near-periphery maxima. The central maximum is indicative of dominant intracrystalline slip along prism planes, whereas the near-periphery maxima are indicative of dominant slip along rhomb planes (Passchier and Trouw, 2005). The quartz C-axis patterns of the four samples of mylonite (Fig. 9a-d) suggest deformation temperatures of ~500°C and that of sample DHF41-3 suggest deformation temperatures above ~500°C. These suggest quartz deformation under medium-grade metamorphic condition (~500–600°C).

Three samples of D2 mylonites from Domain C were analyzed (DHF21-5, DHF68-1, DHF54). Quartz c-axis fabrics (Fig. 9i–k) are characterized by double periphery maxima, indicating a dominant slip along basal planes and middle-low deformation temperatures of 300–450°C (Passchier and Trouw, 2005).

Three samples of D2 tectonites from Domain B were analyzed (DHF 26-7, DHF81-1, DHF 77-2). The quartz c-axis fabrics for samples DHF26-7 and DHF77-2 are dominated by central maxima (Fig. 9f & 9h), indicating deformation temperatures above ~500°C. The third sample (DHF81-1) shows a single center girdle with near-periphery maxima (Fig. 9g), indicating a dominant slip along the rhomb planes and deformation temperatures of 400–500°C. Compared with the samples in Domain A and C, samples in Domain B exhibit similar deformation temperature to those in Domain A and slightly higher deformation temperature than those in Domain C.

The asymmetric patterns of quartz LPOs from the D1 tectonites indicate a top-to-SE movement and those from D2 tectonites a dextral shear sense (Fig. 9). These are consistent with the kinematics observed in the field and in thin sections as described above.

4. U–Pb geochronology

Magmatic dikes and plutons having well-defined cross-cutting relationships with D1 and D2 structures were sampled for dating to constrain the timing of deformation. A total of nine samples were dated from the study area, including four deformed dikes/plutons, four undeformed dikes and one amphibolite gneiss. The locations of the samples are shown in Fig. 3A.

4.1. Analytical procedures

The data presented herein were collected by LA–ICP–MS (Laser Ablation Inductively Coupled Plasma Mass Spectrometry) at two laboratories. Samples DHF11 and DHF10 were dated at the Jack Satterly Geochronology Laboratory at the University of Toronto, Canada, using a New Wave UP-193 laser ablation system coupled to an Agilent 7900 ICP-MS. The primary standard used was TEMORA and secondary standard Ples. The laser beam spot diameter for zircon analysis was 20μm. Data reduction was carried out using an in-house program written by D. Davis. Detailed analytical procedure and data processing are described in Yin et al. (2013). The remaining seven samples were dated at the Hefei University of Technology Geochronology Laboratory in China. The LA-ICP-MS system consists of an Agilent 7500a ICP-MS equipped with a COMPex pro 102 ArF-Excimer laser source (λ =193 nm). The laser beam spot diameter for zircon analysis was 32μm. The 91500 zircon standard was used for standardization and NIST 610 glass standard was used for instrument optimization. Uncertainties on individual LA-ICP-MS analyses are reported at the 1σ level. Correction of common Pb follows the method of Andersen (2002). Isotopic ratios and element concentrations were analyzed using ICPMSDataCal 9.6 (Liu et al., 2010). All concordia diagrams were generated using Isoplot 4.15 (Ludwig, 2003). Field relationships of the dated samples are shown in Fig. 10, CL images of representative zircons in Fig. 11, and concordia diagrams in Fig. 12. U-Pb data table is given as a Supplementary File.

4.2. Sample descriptions and results

4.2.1. Sample DHF36-2: Post-D1 dike from Domain A

The sample is from a granite dike that cross-cuts the pervasive S1 gneissosity in Domain A (Fig. 10A). The age of dike would give a minimum age of the D1 deformation.

Zircon grains from the sample are irregular in shape and sizes range from 70 to 150μm in length and 70 to 100μm in width. CL images consistently exhibit dark overgrowth rims and inherited cores (Fig. 11A).

Eleven spots were analyzed on the cores. Their Th/U ratios range from 0.1 to 0.78, with moderate uranium concentrations (38.2–986 ppm). Their $^{207}Pb/^{206}Pb$ ages range from 1366–1928 Ma, which are interpreted as the crystallization ages of inherited zircon. The remaining eighteen spots were analyzed on the overgrowth rims. Their Th/U ratios range from 0.13 to 0.46, with very high uranium concentrations (715–7449 ppm). Sixteen of the analyses define a weighted mean $^{206}Pb/^{238}U$ age of 326 ± 7 Ma (MSWD=2.0; Fig. 12A), which is interpreted as the crystallization age of the dike.

4.2.2. Sample DHF38-2: Post-D1, pre-D2 aplite dike from Domain A

The sample is from an aplite dike in Domain A. It cuts the S1 gneissosity at a low angle (anticlockwise) and is weakly foliated and boudinaged (Fig. 10B). The boudinage is consistent with dextral shearing and is interpreted to be related to D2. The dike is thus interpreted to have been emplaced after D1 and before D2.

Zircon grains from the sample show two distinct types based on their morphology and internal structures: (1) subhedral prismatic grains with clear oscillatory zoning sizing from 180 to 200μm (Fig. 11B), and (2) subhedral or equant grains with clear core-rim structures sizing from 130 to 220μm. In the latter, weakly luminescent cores reveal patchy-zoning or sector-zoning, highly luminescent cores reveal oscillatory zoning and the overgrowth rims all are narrow, with weakly luminescent oscillatory zoning.

Ten analyses on the oscillatory zones of both types of zircon yield $^{207}Pb/^{206}Pb$ apparent ages from 1865–2036 Ma, interpreted to be inherited zircons. Two analyses on the type 2 weakly luminescent cores yield $^{207}Pb/^{206}Pb$ apparent ages 1846±19 Ma and 1831±11 Ma,. Eleven analyses on the rims of type 2 zircon have Th/U ratios from 0.03–0.10 and define a weighted mean $^{206}Pb/^{238}U$ age of 349±9 Ma (MSWD=1.4) (Fig. 12B), which is interpreted as the crystallization age of the dike.

4.2.3. Samples DHF38 and DHF11: Post-D1 granite intrusions from Domains A and B, respectively

Samples DHF38 and DHF11 are from a granitic dike and a small granite intrusion from domains A and B, respectively. They both cut the S1 gneissosity at a high angle, indicating that they were emplaced after D1 deformation (Fig. 10C & D). They are described together as they have similar field relationships, zircon morphology and ages.

Zircon grains are euhedral and prismatic. CL images consistently exhibit clear oscillatory zoning indicating an igneous origin (Fig. 11C & D). Thirty analyses from sample DHF38 yield a weighted mean $^{206}Pb/^{238}U$ age of 238±3 Ma (MSWD=0.93) (Fig. 12C), with Th/U ratios ranging from 0.7 to 2.6. Fourteen analyses from sample DHF11 define a weighted mean $^{206}Pb/^{238}U$ ages of 241±2 Ma (MSWD=9.4) (Fig. 12D), with Th/U ratios ranging from 1.31 to 1.72. They are interpreted as the crystallization ages of the two granite intrusions.

4.2.4. Sample DHF10: Amphibolite gneiss from Domain B

This sample of amphibolite gneiss is from the main gneiss body in Domain B, with well-developed S1 gneissosity (Fig. 10E). It is relatively homogeneous and consists of hornblende, plagioclase, biotite and minor quartz.

Zircon grains are mostly subhedral to euhedral, ranging from 80 to 350μm in length. CL images show well-developed core-rim structures, with highly luminescent cores surrounded by weakly luminescent overgrowth rims. The cores are rounded with fir-tree sectors, and the rims generally are structureless or patchy-zoned. The characteristics of the cores and rims indicate that the zircons are typical for high-grade metamorphic rocks (Corfu et al., 2003, Fig.11E). Twenty-eight spots were analyzed. Sixteen analyses on the cores have uranium concentrations ranging from 12 ppm to 429 ppm and Th/U ratios from 0.01 to 0.04. They define a weighted mean $^{206}Pb/^{238}U$ age of 406 ± 5 Ma, which is interpreted as a metamorphic age. Twelve analyses on the rims have very high uranium concentrations (565–9556 ppm) and Th/U ratios ranging from 0.01–0.07. They define a weighted mean $^{206}Pb/^{238}U$ age of 249 ± 4 Ma (Fig. 12E), which is interpreted as the metamorphic age of the amphibolite gneiss.

4.2.5. Sample DHF 22-2: Late syn-D2 dike from Domain B

Sample DHF22-2 was collected from a quartz-feldspathic/pegmatite dike from Domain B. It cuts the S1 gneissosity in the host rock and contains a weak S2 foliation (Fig. 10F). It is interpreted to have been emplaced late during D2 deformation.

Zircon grains from the sample are euhedral and prismatic, ranging in length from 80 to 180μm.They can be further divided into two groups based on the internal morphology revealed by CL imaging (Fig. 11F): (1) those with concentric cores with weakly luminescent rims of oscillatory zoning, and (2) those with strongly luminescent cores without/with narrow dark overgrowth rims.

Thirty-eight spots were analyzed. Twenty-two analyses on oscillatory zones of type-1 have concordant $^{206}Pb/^{238}U$ ages from 492±17 to 320±12 Ma, interpreted as the crystallization ages of xenocrystic zircon. Eleven spots on strongly luminescent cores of type-2 define a weighted mean $^{206}Pb/^{238}U$ age of 247±8 Ma (MSWD=2.0, Fig. 12F), with variable Th/U ratios from 0.099 to 0.963. It is interpreted as the crystallization age of the dike.

4.2.6. Sample DHF62: Post-D2 dioritic porphyry dike from Domain B

Sample DHF62 was collected from one of a series of undeformed dioritic porphyry dikes near the North Guanyinjing Fault in Domain B (Fig. 10G). Zircon grains can be grouped into two populations based on morphology and CL images (Fig. 11G): (1) rounded grains with oscillatory or sector zoning, and (2) irregular, euhedral or columnar grains with/without weak oscillatory zoning. A total of 40 spots were analyzed. Fourteen analyses on type-1 zircons yield older ages in two loosely defined clusters, eight analyses with $^{207}Pb/^{206}Pb$ apparent ages between 1831 ± 31 and 1402 ± 31 Ma, and six between 462 ± 14 and 391 ± 12 Ma; all interpreted as the ages of xenocrystic zircons. Twenty-four most concordant analyses on type-2 zircons define a weighted mean $^{206}Pb/^{238}U$ ages of 217 ± 4Ma (MSWD=1.6) (Fig. 13A), interpreted as the emplacement age of the dike.

4.2.7. Sample DHF68-1: Mylonitized granite from Domain C

Sample DHF68-1 is from a mylonitized granite from the Qingshan shear zone (Fig. 3A, 8D), with well-developed foliation and stretching lineations. The granite is interpreted to have been emplaced before D2 dextral shearing.

Zircons from this sample can be grouped into three populations based on morphology and CL images (Fig. 11H): (1) almost black, euhedral and prismatic grains with weak oscillatory zoning, (2) moderately luminescent, mainly euhedral and prismatic grains with well-developed oscillatory zoning, and (3) strongly luminescent, long elliptical grains with weak oscillatory zoning. A total of 40 spots were analyzed. Thirty-eight analyses on type-1 and type-2 zircons are mostly concordant and define a weighted mean $^{206}Pb/^{238}U$ age of 258 ± 5 Ma (MSWD=4.4; Fig. 13B), interpreted as the emplacement age of the granite. Two analyses on type-3 zircon yield $^{207}Pb/^{206}Pb$ apparent ages of 2380±17 Ma and 1941±21 Ma, respectively, which are interpreted as the ages of xenocrystic zircon.

4.2.8. Sample DHF21: Post- D2 undeformed dioritic porphyry dike from Domain C

This sample is from an undeformed dike that cuts the S2 foliation in a tonalitic mylonite in Domain C (Fig. 10H). The dike was thus emplaced after D2 deformation.

Zircons from this sample can be grouped into two types (Fig. 11I): (1) subhedral prismatic grains with rounded terminations sizing from 150 to 300μm in length, commonly weakly luminescent with clear oscillatory zoning, and (2) mainly needle-shaped, highly luminescent grains, sizing from 80 to 280μm in length and 70 to 120μm in width. Twenty-two spots made on the oscillatory zones of type-1 zircon yield $^{206}Pb/^{238}U$ ages ranging from 451 to 261 Ma, interpreted as the crystallization ages of xenocrystic zircon. Fourteen most concordant analyses on the type-2 zircon define a weighted mean $^{206}Pb/^{238}U$ ages of 242 ± 5Ma (MSWD=0.39) (Fig. 13C), interpreted as the crystallization age of the dioritic porphyry dike.

5. Discussion

5.1. Kinematics of deformation

Our structural analysis reveals two major generations of deformation, D1 and D2 (Fig. 14). D1 comprises penetrative ductile deformation associated with north-over-south reverse movements and D1 structures are well developed in both domains A and B. D2 comprises deformation associated with dextral strike slip. The Qingshan shear zone along the southern edge of the study area (Domain C) is a major D2 structure. Narrow D2 shear zones also occur in Domain B, and less commonly in Domain A.

Syn-D1 mineral assemblages, microstructures and quartz C-axis patterns indicate that D1 deformation took place under amphibolite facies conditions in both Domains A and B. In contrast, D2 structures in Domains B and C formed under different metamorphic conditions; those in Domain B exhibit higher-temperature fabrics, coeval with the ca. 249 +/- 4 Ma amphibolite facies metamorphism, whereas those in Domain C show lower temperature fabric under greenschist facies. This indicates a significant uplift of Domain B relative to Domain C. This is likely a result of late syn- to post-D2 movement along the South Guanyinjing fault separating the two domains (Fig. 3).

5.2. Timing of deformation

U–Pb zircon ages of intrusions dated during this study and their timing relationships with respect to D1 and/or D2 structures are summarized in Fig. 15. The four post-D1 dikes (samples DHF38-2, DHF36-2, DHF11 and DHF38) yield emplacement ages of 349 +/- 9, 326 +/- 7, 241+/-2 and 238 +/- 3 Ma, respectively, indicating that D1 deformation took place before ca. 350 Ma. Metamorphic zircon from an amphibolite gneiss in Domain B (sample DHF10) yields an age of 406 +/- 5 Ma (Early Devonian), which is considered as the most likely age of the amphibolite facies metamorphism. This is also considered as the most likely age of the D1 deformation as D1 took place under amphibolite facies conditions.

A granite deformed during D2 (sample DHF68-1) yields a protolith age of 258 +/- 5 Ma, a syn-D2 dike (sample DHF22-2) 247 +/- 8 Ma, and two post-D2 dikes (samples DHF21 and DHF 62) give 242 +/- 5 Ma and 217 +/- 4 Ma, respectively. This constrains the age of D2 deformation to between ca. 258 and 242 Ma, or Late Permian to Middle Triassic. This age is consistent with that of the younger metamorphic zircon from sample DHF10, 249 +/- 4 Ma.

5.3. *Deformation phases and tectonic significance*

5.3.1. *D1: Paleozoic reverse shearing (ca. 406–349 Ma)*

The D1 north-side-up reverse shear described herein formed under amphibolite facies condition, which took place at ca. 406 Ma. This is approximately coeval with the regional retrograde metamorphism, which occurred at ca. 403 Ma and overprinted the ca. 431 Ma HP granulite-facies metamorphism (He et al., 2014). Ductile deformation that occurs during plate convergence and crustal shortening commonly has fabrics related to folding and associated reverse shearing/thrusting. Consequently, we tentatively relate the D1 reverse shearing to an accretional/collisional event (see below).

We agree with the previous interpretation that Precambrian basement rocks in the DTB, represented by Domains A & B in the study area, could form part(s) of a microcontinent or microcontinents (Zhao et al., 2016). During the middle Silurian to middle Devonian (ca. 430–390 Ma), these basement rocks were affected by a prolonged period of convergence, collision and accretion, resulting in the extensive steeply-dipping S1 gneissosity and steeply-plunging

L1 stretching lineations preserved in Domains A and B. Our kinematic interpretation is supported by the presence of north-dipping shear zones, duplexes and imbricate thrust faults in the Hongliuxia area to the south (Fig. 2) (Shi et al., 2017).

5.3.2. D2: Late Permian–Early Triassic dextral strike slip (ca. 258–241 Ma)

The Permian–Triassic is an important period in the tectonic evolution of the CAOB. However, the presence of a significant Permian–Triassic tectono-thermal event(s) in the DTB was not recognized until this study. Although Wang et al. (2016a) reported ca. 249 Ma zircons from a medium-pressure mafic granulite near Qingshigou (Fig. 2), they attributed the zircon growth to a local thermal perturbation and did not provide any additional interpretation.

In this study, we documented a major dextral strike-slip ductile shearing event (D2) in the study area and constrained its age to between ca. 258 and 241 Ma, or Late Permian to Middle Triassic, an age consistent with that of the younger metamorphic zircon from sample DHF10 (249 +/- 4 Ma). It should be noted that our unpublished data show that the abundant granitoid plutons that comprise the main body of Domain C and the granitoid plutons that intrude the TTG felsic and mafic amphibolite gneisses in Domain B are all Early Permian to Early Triassic in age, supporting the presence of a significant tectono-thermal event in the DTB in this time period.

5.4. Comparison between the DTB and the Beishan–Tianshan orogen

The Beishan orogen is generally regarded as the eastern extension of the Chinese Tianshan (Fig. 1; Xiao et al., 2010). It records accretionary events during the early Paleozoic, terminating in collisional tectonics during the late Paleozoic to early Mesozoic (Cleven et al., 2016). The Tianshan orogen records two main accretionary events, in the middle and late Paleozoic, respectively (Windley et al., 1990; Charvet et al., 2007, 2011). The DTB shows many similarities to the Beishan–Tianshan orogen, as summarized below and in Fig. 15.

5.4.1. Similarities in the Paleozoic

Both the DTB and the Beishan–Tianshan orogen experienced strong deformation due to a prolonged period of convergence, accretion and continental collision in the Paleozoic.

In the Ordovician to Early Carboniferous, the amalgamation of Yili–North Tianshan and Tarim blocks gave birth to the Eo-Tianshan Mountains, which are characterized by north-verging thrust sheets of ophilitic mélange, HP and UHP metamorphic nappes and molasse (Charvet et al., 2011). Similarly, Paleozoic deformation in the DTB is characterized by extensive north-side-up reverse shear, duplexes and imbricate thrust faults in gneisses (this study; Shi et al., 2017). Dextral transpression also affected the Mazongshan terrane in Beishan (Cleven et al., 2016).

Similar metamorphic events have also been documented in both areas. For example, ~465 Ma HP eclogites have been reported west of Liuyuan in Beishan (Fig. 2; Liu et al., 2011; Qu et al., 2011), and ~440–430 Ma HP granulite (Zong et al., 2012; He et al., 2014) and ca. 428–391 Ma eclogite (Wang et al., 2017a) have been identified in the DTB. The metamorphism associated with D1 in the DTB is slightly younger than that in Beishan.

The Beishan–Tianshan orogen and the DTB also share a similar Paleozoic magmatic history: ca. 438–397 Ma granitoids are present in Beishan (Zhao et al., 2007; Zhang and Guo, 2008; Liu et al., 2011) and Silurian (ca. 428 Ma) and Late Devonion (368–361 Ma) granitoids are found in the central Tianshan (Shi et al., 2007); ca. 440–410 Ma granitoids and ca. 370–360 Ma intrusive rocks occur within the DTB (Zhang et al., 2009; Wang et al., 2016b, 2016c; Zhao et al., 2017).

5.4.2. Similarities in the Permian–Triassic

During the Permian–Triassic, both the DTB and Beishan–Tianshan experienced strike-slip deformation.

In the Beishan, collision with the Tarim craton forced inboard convergence, initiating syn-orogenic sedimentation and fold-and-thrust belt deformation in the Mazongshan terrane (Fig. 2; Zhang and Cunningham, 2012; Tian et al., 2013; Cleven et al., 2015). This was followed by sinistral strike-slip-related tectonics, and there is evidence that sinistral deformation overprints an early dextral deformation (Cleven et al., 2016). Wang et al. (2010) conclude that the NE-striking Xingxingxia sinistral shear zone initiated at ~240-235 Ma. Final amalgamation of the Shibanshan and Huaniushan arcs in late Permian led to the formation of the Liuyuan ophiolitic complex (Mao et al., 2012a).

The Tianshan orogen records several ductile deformation events between 290 and 245 Ma (Laurent-Charvet et al., 2003). The formation of the E-W trending ductile strike-slip faults (Laurent-Charvet et al., 2002, 2003; Wang et al., 2008) is significant as they also constitute the main boundaries between different terranes in the Tianshan. During the Early Permian (ca. 290–280 Ma), the collision between Junggar and Tianshan terranes and the accompanying relative rotations between stable blocks induced a sinistral transcurrent event along the Erqishi–Irtysh Shear zone (Laurent-Charvet et al., 2002). The last large-scale transcurrent deformation, dextral strike-slip in the Tianshan, occurred at ca. 250–245 Ma ($^{40}Ar/^{39}Ar$ ages; Laurent-Charvet et al., 2003; Wang et al., 2002, 2010). Charvet et al. (2007) suggested that the major strike-slip movement, dextral in Tianshan and sinistral in the Mongolian fold belt, was due to opposite motions of the Siberia and Tarim cratons.

The Qingshan shear zone in the DTB is a major dextral strike slip structure. It formed at ca. 249 Ma, coeval with the regional strike-slip deformation in Tianshan.

Permian–Triassic magmatism is present in both the Beishan–Tianshan orogen and the DTB. For example, ca. 279–275 Ma high-K alkaline granitoids in Beishan were possibly emplaced in a post-collisional extensional setting (Li et al., 2013b), and Early–Middle Triassic (ca, 250–230 Ma) granitoids have been reported in the Beishan and eastern Tianshan (Li et al., 2012, 2013a). Post-collisional ca. 295–293 Ma bimodal volcanic rocks (Chen et al., 2011) and ca. 286 Ma gabbros and rhyolites (Mao et al., 2014) are present in Tianshan. Similarity, Permian–Early Triassic (280–240 Ma) granitoids are widespread in the study area of the DTB.

5.4.3. DTB as part of the CAOB

The DTB has structural, metamorphic and magmatic characters in the Paleozoic–Mesozoic that are typical of an orogenic belt. The similarities in geological histories between the DTB and the Beishan–Tianshan orogen summarized above support the interpretation that the DTB is a part of the CAOB (Zhao et al., 2017; Wang et al., 2017a).

The DTB was previously considered to constitute a small Precambrian block with Archean basement and was considered as either the easternmost part of the Tarim craton (BGMRX, 1993; Lu et al., 2008) or the westernmost part of the North China Craton (Zhang et al., 2013). In light of evidence for Paleozoic orogenic events documented in the DTB in this and other studies summarized above, we propose that the Precambrian rocks in the DTB were parts of a microcontinent or microcontinents derived from the Tarim or the North China craton and the Paleozoic orogenic events in the DTB resulted from accretion or collision of these microcontinent(s) with other terranes (including other microcontinents and/or arcs and the Tarim and/or North China cratons). It is therefore likely that the DTB itself formed by accretion-collision of multiple terranes. The recent discovery of ~365 Ma plagiogranites in the Sanweishan area (Fig. 2), interpreted to have developed in a back-arc basin (Zhao et al., 2015a), is consistent with this interpretation.

It has been proposed the Liuyuan complex in the Beishan (Fig. 2) is a Permian ophiolitic fore-arc sliver (286+/-2 Ma; Mao et al., 2012) and was part of the Liuyuan mélange that formed as a result of the final closure of the Paleo-Asian Ocean (Xiao et al., 2010). If this interpretation is correct, it implies that the Paleozoic orogenic events in the DTB took place to the south of the final suture (marked by the Liyuan complex) before the final closure of the Paleo-Asian Ocean and were unrelated to the Paleozoic tectonic events in the Beishan-Tianshan orogen that took place to the north of the suture, in spite of the similarities summarized above. In other words, the DTB and the Beishan orogen might represent two separate Paleozoic mountain belts that developed more or less synchronously on the south and north sides of the Paleo-Asian Ocean, respectively, before closure of the last vestiges of the ocean in the Permian.

The Permian-Triassic strike-slip deformation in the DTB (D2) and Beishan was probably kinematically related to post-collisional adjustment among the various terranes.

6. Conclusions

(1) Detailed structural analyses reveal two episodes of deformation in the central DTB, D1 and D2. D1 is a north-side-up reverse shear, and D2 is a dextral strike slip, concentrated in the Qingshan shear zone.

(2) Mineral assemblages, microstructures and quartz C-axis patterns indicate that D1 took place under amphibolite facies conditions (500 to 600ºC) and D2 mostly under greenschist-facies conditions (300–450ºC).

(3) U–Pb zircon geochronology indicates that D1 deformation took place before ca. 349 Ma and most likely at ca. 406 Ma, and D2 between ca. 249 Ma and ca. 241 Ma.

(4) The DTB is likely a part of the CAOB. The D1 reverse shearing may have been induced by Silurian–Devonian terrane accretion/collision. The D2 dextral strike slip is interpreted as a product of adjustment of terranes, resulting from N-S compression and rotation of the terranes in Late Permian–Middle Triassic, after terrane amalgamation.

(5) The DTB and the Beishan orogen might represent two separate Paleozoic orogens that developed more or less synchronously on the south and north sides of the Paleo-Asian Ocean before closure of the last vestiges of the ocean in Permian.

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