**Transpressional deformation, strain partitioning and fold superimposition in the
southern Chinese Altai, Central Asian Orogenic Belt**

Pengfei Li ^{1*}, Min Sun ¹, Gideon Rosenbaum ², Keda Cai ³, Ming Chen ¹, Yulin He ¹

¹ Department of Earth Sciences, The University of Hong Kong, Pokfulam Road, Hong Kong,
China

² School of Earth Sciences, The University of Queensland, Brisbane 4072, Queensland,
Australia

³ Xinjiang Research Center for Mineral Resources, Xinjiang Institute of Ecology and
Geography, Chinese Academy of Sciences, Urumqi 830011, China

*Corresponding author: Department of Earth Sciences, The University of Hong Kong,
Pokfulam Road, Hong Kong, China. Email: pengfei@hku.hk; pengfeili2013@gmail.com

Abstract: Transpressional deformation has played an important role in the late Paleozoic evolution of the western Central Asian Orogenic Belt (CAOB), and understanding the structural evolution of such transpressional zones is crucial for tectonic reconstructions. Here we focus on the transpressional Irtysh Shear Zone with an aim at understanding amalgamation processes between the Chinese Altai and the West/East Junggar. We mapped macroscopic fold structures in the southern Chinese Altai and analyzed their relationships with the development of the adjacent Irtysh Shear Zone. Structural observations from these macroscopic folds show evidence for four generations of folding and associated fabrics. The earlier fabric (S_1), is locally recognized in low strain areas, and is commonly isoclinally folded by F_2 folds that have an axial plane orientation parallel to the dominant fabric (S_2). S_2 is associated with a shallowly plunging stretching lineation (L_2), and defines ~NW-SE tight-close upright macroscopic folds (F_3) with the doubly plunging geometry. F_3 folds are superimposed by ~NNW-SSE gentle F_4 folds. The F_3 and F_4 folds are kinematically compatible with sinistral transpressional deformation along the Irtysh Shear Zone and may represent strain partitioning during deformation. The sub-parallelism of F_3 fold axis with the Irtysh Shear Zone may have resulted from strain partitioning associated with simple shear deformation along narrow mylonite zones and pure shear-dominant deformation (F_3) in fold zones. The strain partitioning may have become less efficient in the later stage of transpressional deformation, so that a fraction of transcurrent components was partitioned into F_4 folds.

Key words: Central Asian Orogenic Belt; Chinese Altai; Irtysh Shear Zone; Transpression;
Strain partitioning

1. Introduction

Transpressional deformation in orogenic belts commonly results from oblique plate convergence (Harland, 1971; Carosi and Palmeri, 2002; Holdsworth et al., 2002; Sarkarinejad et al., 2008; Zhang et al., 2010; Díaz-Azpiroz et al., 2014; Xu et al., 2015), and is characterized by general shear strain that involves both simple and pure shear components (Sanderson and Marchini, 1984; Fossen and Tikoff, 1993; Lin et al., 1998). Theoretically, the general shear strain in transpressional zones can be either distributed homogeneously (Fig. 1a), or partitioned across the deformation zone due to the presence of mechanical anisotropies (Fig. 1b, c) (Fossen et al., 1994; Tikoff and Teyssier, 1994; Jones and Tanner, 1995; Dewey et al., 1998; Schulmann et al., 2003; Massey and Moecher, 2013). The simplest partitioning model involves a series of discrete slip planes that accommodate simple shear deformation, with pure shear deformation occurring in-between contractional domains (Fig. 1b; e.g. Dewey et al., 1998). Alternatively, discrete slip planes may only accommodate a fraction of the lateral displacement, with the remaining transcurrent component distributed in contraction-dominant domains between discrete slip planes (Fig. 1c) (Tikoff and Teyssier, 1994; Teyssier et al., 1995; Miller, 1998). The degree of strike-slip partitioning (i.e. the relative amount of lateral displacement partitioning into discrete slip planes) is dependent on the orientation of the convergence vector relative to the boundary (Fossen et al., 1994; Tikoff

and Teyssier, 1994; Teyssier et al., 1995), and the rheological weakening along the shear zone (Mount and Suppe, 1987; Zoback et al., 1987; Mount and Suppe, 1992).

The Central Asian Orogenic Belt (CAOB) is one of the largest accretionary orogens in the world. It underwent a prolonged history of accretion from the late Mesoproterozoic to Mesozoic (Zonenshain et al., 1990; Şengör et al., 1993; Zorin, 1999; Khain et al., 2002; Xiao et al., 2003; Buslov et al., 2004a; Buslov et al., 2004b; Yakubchuk, 2004; Windley et al., 2007; Wilhem et al., 2012; Han et al., 2015; Xiao et al., 2015; Zhang et al., 2015b), and was characterized by multiple phases of deformation (e.g. Qu and Zhang, 1994; Allen et al., 2001; Lin et al., 2009; Wang et al., 2010; Zhang and Cunningham, 2012). Recognizing the styles of deformation is crucial for reconstructing the orogenic history and its associated episodes of terrane accretion. Within the CAOB, multiple generations of folds have been recognized, and the origin of these structures was taken into account in large scale tectonic reconstructions (Lehmann et al., 2010; Tian et al., 2013; Guy et al., 2014). For example, macroscopic Type 2 fold interference patterns in the Beishan region of the southern CAOB were inferred as indicating two episodes of orthogonal convergence of the Tarim-North China Craton, following the closure of the Paleo-Asian Ocean (Tian et al., 2013). Transpressional deformation was particularly common during the late Paleozoic amalgamation of the western CAOB, and this deformation has apparently modified the earlier accretion-related architecture and generated superimposed fold structures (Qu and Zhang, 1994; Allen et al., 2001; Choulet et al., 2012; Li et al., 2015a). Understanding the structural evolution and strain

patterns across major transpressional zones is crucial for reconstructing the amalgamation processes of the western CAO in the late Paleozoic and the earlier accretion history.

This paper focuses on fold structures in the area of the Irtysh Shear Zone, southern Chinese Altai (Fig. 2). This sinistral shear zone marks the front of an oblique collision between the Chinese Altai and the intra-oceanic arc system of the West/East Junggar, extending >1000 km from NE Kazakhstan, through NW China, to western Mongolia (Qu and Zhang, 1991; Qu and Zhang, 1994; Laurent-Charvet et al., 2003; Liu et al., 2013; Li et al., 2015a). A series of macroscopic (~10-50 km) doubly plunging folds that commonly show a dome shape in map view, occur adjacent to the Irtysh Shear Zone (Figs. 2c and 3). It remains unclear whether these doubly plunging folds resulted from superimposed folding or they were produced by a single phase of deformation. Available chronological data (e.g. Li et al., 2015b) suggest that macroscopic folds in this area were developed in the Permian, which is the timing of transpressional deformation along the Irtysh Shear Zone (Laurent-Charvet et al., 2003; Briggs et al., 2007). It seems, therefore, that the doubly plunging folds are genetically linked to the development of the Irtysh Shear Zone. As such, understanding the origin of these macroscopic folds and their relationship with the regional shear zone could provide rigorous constraints on the convergence angle and amalgamation processes between the Chinese Altai and the West/East Junggar. To achieve this aim, we have conducted detailed structural mapping around the macroscopic doubly plunging folds in the Qiongkuer Domain of the southern Chinese Altai (Fig. 3).

102

103 **2. Geological setting**

104 The Chinese Altai is located in the junction between the peri-Siberian, West
105 Junggar-Kazakhstan-Tianshan, and Xing'an-Mongolian-East Junggar orogenic systems (Fig.
106 2a). It records Paleozoic accretionary processes along the southwestern margin of the
107 Siberian Craton, and the subsequent late Paleozoic convergence history with the intra-oceanic
108 arc systems of the West/East Junggar (Windley et al., 2002; Xiao et al., 2008).

109

110 The Chinese Altai mainly consists of Paleozoic meta-sedimentary/volcanic rocks, and is
111 divided into four ~NW-SE tectonic domains (Fig. 2c) (He et al., 1990; Windley et al., 2002;
112 Cai et al., 2011a). Based on the regional geological map (Li et al., 2008), these four tectonic
113 domains can be traced along the strike into Mongolia, Russia and Kazakhstan. The Northern
114 Altai Domain is the northernmost unit in the Chinese Altai, extending eastward into Mongolia.
115 It mainly includes Late Devonian to Early Carboniferous metasedimentary and metavolcanic
116 rocks with a metamorphic grade up to greenschist facies. The Central Altai Domain, which is
117 separated from the Northern Altai Domain by a normal fault, occupies the major part of the
118 Chinese Altai (Fig. 2c) as well as the Altai-Mongolian terrane in Russia and Mongolia. Rocks
119 in this domain are predominantly represented by Cambrian to Silurian marine-facies
120 turbidites and pyroclastic rocks of the Habahe Group and the Kulumuti Group, which were
121 interpreted as an accretionary complex developing along the adjacent Tuva-Mongolian
122 microcontinent (Fig. 2b) (Long et al., 2012). Metamorphism in this domain is characterized

by zonal metamorphic sequences centered by gneissic granitoids (Zhuang, 1994). Late Ordovician to Devonian felsic volcanic rocks of the Dongxileke Formation and marine-facies clastic rocks of the Baihaba Formation unconformably overlie the Habahe Group in the westernmost Central Altai Domain (Fig. 2c) (Long et al., 2010). Farther south, the Qiongkuer Domain is mainly occupied by Devonian volcanic and sedimentary rocks of the Kangbutiebao and Altai formations, which were subjected to high temperature metamorphism, locally up to granulite facies (Wang et al., 2009; Li et al., 2014; Tong et al., 2014a; Yang et al., 2015b). The Qiongkuer Domain is possibly correlated with the Rudny Altai in NE Kazakhstan and the Tseel terrane in Mongolia (Badarch et al., 2002). The Southern Altai Domain is the southernmost tectonic unit of the Chinese Altai, and is represented by schist, para-/ortho-gneiss, amphibolite, migmatite and metaschert of the Irtysh Complex (Qu and Zhang, 1991; Briggs et al., 2007; Li et al., 2015a). The origin of these rocks is considered to be an accretionary complex (O'Hara et al., 1997; Xiao et al., 2009), which can be traced into NE Kazakhstan, termed as the Irtysh-Zaisan Complex (Windley et al., 2007) or the Kalba-Naryn terrane (Buslov et al., 2004a; Safonova, 2013).

Rocks in the Qiongkuer and Southern Altai domains, as well as in the southern part of the Central Altai Domain, show evidence for late Paleozoic transpressional deformation along the sinistral Irtysh Shear Zone (Qu and Zhang, 1991; Qu and Zhang, 1994; Laurent-Charvet et al., 2002; Laurent-Charvet et al., 2003; Liu et al., 2013; Li et al., 2015a; Zhang et al., 2015a; Li et al., 2016). This deformation phase was associated with a major uplift event in

the Permian as indicated by $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronological data (Laurent-Charvet et al., 2003; Briggs et al., 2007; Li et al., 2015b). The transpressional deformation has been attributed to oblique convergence between the Chinese Altai and the West/East Junggar (Qu and Zhang, 1994; Li et al., 2015a), which was accompanied in the southern Chinese Altai by high temperature metamorphism during the Permian (Wang et al., 2009; Li et al., 2014; Wang et al., 2014b; Yang et al., 2015a; Yang et al., 2015b). In the Mesozoic and Cenozoic, the Irtysh Shear Zone and/or other ~NW-SE faults in the southern Chinese Altai (Figs. 2 and 3) may have been reactivated as the uplift of the Altai Mountain (Yuan et al., 2006; Glorie et al., 2012; Delvaux et al., 2013).

A large number of granitoids occur in the Chinese Altai (Fig. 2c) and their ages are mostly Devonian and Permian (Zou et al., 1988; Liu, 1990; Yuan et al., 2007; Sun et al., 2009; Cai et al., 2011b; Kröner et al., 2014; Tong et al., 2014b). The Devonian granitoids occur throughout the Chinese Altai and were supposedly emplaced along a convergent continental margin (Wang et al., 2006; Sun et al., 2008). In contrast, Permian granitoids are mainly distributed along the southern Chinese Altai (Fig. 2c), and their origin is normally attributed to the collision of the Chinese Altai with the West/East Junggar (Tong et al., 2014b). In addition, a few Mesozoic granitic plutons were emplaced with a possible non-orogenic origin (Li et al., 2013; Wang et al., 2014a).

3. Lithostratigraphy of the Qiongkuer Domain in the Fuyun area

The Qiongkuer Domain in the Fuyun area is occupied by two lithostratigraphic units: the Kangbutiebao Formation and the Altai Formation (Figs. 2c and 3). A conglomerate layer (Fig. 4a) is recognized between the two formations, which together with the presence of an underlying paleo-weathering layer, suggest an unconformable contact between the two formations (BGMRX, 1978a). The spatial distribution of lithostratigraphic units is controlled by macroscopic folds with the Kangbutiebao and Altai formations dominantly distributed in the core areas of antiforms and synforms, respectively (Fig. 3).

Rocks of the Kangbutiebao Formation in the Fuyun area are metamorphosed up to granulite facies (Li et al., 2014; Yang et al., 2015b), and are dominantly represented by migmatized orthogneisses interlayered with minor amphibolite and metasedimentary rocks. The orthogneiss was likely derived from a felsic volcanic protolith (BGMRX, 1978a), as indicated by observed volcanic textures in the lower grade part of the Kangbutiebao Formation in the Aletai region (Fig. 2c) (Chai et al., 2009). U-Pb zircon geochronology from orthogneiss and metavolcanic rocks indicates an Early Devonian age for the Kangbutiebao Formation (Chai et al., 2009; He et al., 2015).

The Altai Formation in the Fuyun area is lithologically distinct from the rest of the Altai Formation in the Chinese Altai (Fig. 2c). It is characterized by the widespread occurrence of amphibolite and amphibole schist, which rarely occur within the turbidite-dominant sequence in other parts of the Altai Formation. Based on geochemical considerations, Xu et al. (2003)

suggested that the protoliths of these amphibolites originated in a back arc environment. The internal stratigraphy of the Altai Formation is not well recognized due to the intense deformation. Based on field observations and the interpretation of satellite images (Appendix A), we here subdivide the Altai Formation in the Fuyun area into four lithostratigraphic units (units 1-4; Fig. 3), which are approximately similar to the units in the 1:200,000 geological map (BGMRX, 1978a). The lower part of the Altai Formation (unit 1) is characterized by diagnostic white and black layers of quartzofeldspathic gneiss, amphibolite and banded chert (particularly typical south of Kuerti, Figs. 3 and 4b, c). Rocks in unit 2 are relatively homogeneous, and are dominated by amphibole schist and amphibolite with minor porphyritic metavolcanic rocks and quartzite (Fig. 4d, e), in which deformed pillow basalt was recognized (Fig. 4f). Further up, unit 3 is characterized by a thin layer of gray-white quartz schist. Unit 4, at the top of the sequence, mainly comprises gray-green chlorite actinolite schist, meta-siltstone and porphyritic metavolcanic rocks.

4. Structural observations from the Qiongkuer Domain (Fuyun area)

A series of macroscopic doubly plunging fold structures with different sizes were recognized in the Qiongkuer Domain of the southern Chinese Altai (Fuyun area, Fig. 3). Here we focus on two sets of doubly plunging folds, termed the eastern and western doubly plunging folds, respectively (Fig. 3). The eastern doubly plunging folds are characterized by a map-view dome shape with >50 km long axis (Fig. 3). In order to constrain its geometry, we conducted two structural transects in the hinge areas of this macroscopic fold structure (Figs. 3 and 5),

complemented by satellite image interpretation (Appendix A) (Fig. 5). In addition, detailed structural observations were made throughout the western doubly plunging folds (Figs. 6 and 7).

4.1. Structures of the eastern doubly plunging folds

4.1.1. Mesoscopic structures (S_1 and S_2)

Two generations of outcrop-scale fabric were recognized around the eastern doubly plunging folds. S_1 is locally recognized, and isoclinally folded with the axial plane parallel to S_2 (Fig. 8a). S_2 is a dominant fabric, and represents an axial planar fabric of F_2 folds (Fig. 8a, b). It is predominantly associated with symmetric lens-shaped clasts and conjugate shear sets (Fig. 8c, d). A shallowly plunging stretching lineation (L_2), which is defined by preferred alignment of amphiboles or stretched quartz and feldspar aggregates, is recognized within S_2 (Fig. 8e). On a larger scale, S_2 shows variable orientations that define the ~NW-SE macroscopic folds of F_3 (Fig. 5). Metamorphism associated with S_2 reaches the amphibolite facies in the core of the macroscopic antiform as indicated by the mineral assemblage of amphibole, plagioclase and garnet. The metamorphic grade progressively decreases to greenschist facies in the core of macroscopic synform (i.e. unit 4 of the Altai Formation).

4.1.2. Macroscopic fold structures (F_3)

Macroscopic F_3 antiforms and synforms are characterized by tight to close inter-limb angles (Fig. 5c, d). In the western transect (Figs. 3 and 5a), S_2 defines F_3 folds with β axis of 14-306

(plunge angle-plunge) (Fig. 5e), which together with the map-view axial trace of F_3 (azimuth 126°), indicates that the F_3 axial plane must be vertical (Fig. 5g). Stretching lineations (L_2) in this area gently plunge to ~NW (~11-304). Similarly, in the eastern transect, S_2 defines F_3 folds with β axis of 19-115 and a map-view axial trace of 114° , which constrain the F_3 axial plane at 87-204 (Fig. 5h, j). The mean orientation of stretching lineations (L_2) across this transect is 20-120 (Fig. 5i). Overall, L_2 stretching lineations are generally subparallel to the axis of F_3 that shows similar axial plane orientations across the two transects. The opposite plunging orientations of both F_3 fold axes and stretching lineations (L_2) in the eastern and western transects (Fig. 5a, b), reveal the doubly plunging geometry of macroscopic F_3 folds.

At the outcrop scale, the dominant S_2 fabric is folded by F_3 and overprinted by S_3 (Fig. 8f, g). The F_3 folds are commonly tight to close and rounded with a steeply dipping axial plane. Folded L_2 stretching lineation around F_3 folds is also observed (Fig. 8h).

4.2. Structural observations from the western doubly plunging folds

The western doubly plunging folds are separated from the eastern doubly plunging folds by ~E-W faults, which show sinistral strike-slip movements as manifested by the offset of lithological layers (Fig. 3). The western doubly plunging folds are characterized by a map-view dome shape with a ~15 km long axis and a ~4 km short axis (Figs. 3 and 6). Rocks are gneissic granitoids in the core of the structure, surrounded by the Kangbutiebao and Altai formations in the rim (Fig. 6).

249

250 **4.2.1. Mesoscopic structures (S_1 and S_2)**

251 Two generations of fabric (S_1 and S_2) were recognized around the western doubly plunging
 252 folds (Fig. 6). S_1 is locally recognized in low strain areas (Fig. 9a), but is commonly
 253 transposed and overprinted by S_2 , which is the dominant fabric throughout the mapping area
 254 (Fig. 9b, c). S_2 is parallel to the axial plane of F_2 folds, and is associated with a stretching
 255 lineation (L_2 , Fig. 9d). The development of S_2 in this area involved high temperature
 256 metamorphism (amphibolite facies) and migmatization. In the area of the western doubly
 257 plunging folds (Fig. 6), S_2 shows variable orientations that define macroscopic F_3 and F_4 fold
 258 structures.

259

260 **4.2.2. Macroscopic folds (F_3 and F_4)**

261 Two generations of macroscopic folds were recognized with earlier ~NW-SE tight to close F_3
 262 folds overprinted by ~NNW-SSE F_4 folds (Figs. 6 and 7). The map-scale superimposition of
 263 F_4 and F_3 is manifested in the southwestern part of the western doubly plunging folds (bottom
 264 left corner in Fig. 6). In order to understand the macroscopic geometry of F_3 and F_4 , we
 265 divided the map area into four domains (Fig. 6), with domains 1 and 4 dominated by the
 266 geometry of F_3 and domains 2 and 3 dominated by the F_4 geometry.

267

268 The macroscopic geometry of F_3 is best recognized in domains 1 and 4. Domain 1 shows a
 269 series of map-scale ~NW-SE antiforms and synforms (Fig. 6). 58 measurements of S_2 define

a fold hinge at 31-311, which together with the map-view axial trace of F_3 (azimuth 121°), constrain the dip and dip direction of the axial plane (74-031) (Fig. 6). In domain 4, only one map-scale F_3 antiform was mapped, with a strike orientation of 104° . The calculated F_3 hinge in this domain is 53-099, which together with the fold trace, constrain an axial plane of F_3 at 86-014 (Fig. 6). The opposite hinge orientations of F_3 in domains 1 and 4 show the doubly plunging geometry of macroscopic F_3 folds, consistently with the opposite plunging orientations of L_2 in domains 1 and 4 (see stereonets in Fig. 6). A series of ~NE-SW granitic dykes occurs in domain 1, cutting macroscopic F_3 folds (Fig. 6). At the outcrop scale, F_3 in domain 1 is commonly tight and rounded to angular, whereas it is generally open and rounded in domain 4 (Fig. 9e, f). Spaced cleavage (S_3) was locally recognized in the core areas of macroscopic F_3 folds (Fig. 9g), with an orientation that is approximately parallel to the F_3 axial plane.

Macroscopic F_4 folds were mapped in domains 2 and 3. In domain 2, the calculated F_4 fold hinge (B_{42}) is 61-352 (Fig. 10a), which together with $\sim 150^\circ$ map-view axial trace of F_4 , defines an axial plane of 78-060 (Fig. 10a). F_4 in domain 3 yielded an axial plane of 81-240, based on a calculated fold hinge (B_{42}) of 73-298 and the map-view axial trace of $\sim 150^\circ$ of F_4 (Fig. 10b). Macroscopic F_4 folds in both domains 2 and 3 show an asymmetric geometry with sinistral vergence. Similarly, outcrop-scale F_4 folds are predominantly asymmetric, and 10 out of 12 observations show S-shaped folds, illustrating sinistral vergence (e.g. Fig. 9h). These folds are commonly open and rounded to angular, and locally develop as F_4

crenulations (Fig. 9h). Throughout the map area, the measured axial plane of outcrop-scale F_4 is consistently trending ~NNW-SEE (Fig. 10c). In the northern limb of the macroscopic F_3 fold, direct measurements of the F_4 axial plane show a steep dip to NE (77-055), consistent with the calculated F_4 axial plane in domain 2. Similarly, in the southern limb, the F_4 dips to the SW (77-235) (Fig. 10c), consistently with the calculated F_4 axial plane in domain 4. Field measurements for the hinge of F_4 (B_{42}) show variable orientations.

5. Geochronology

In order to constrain the protolith age of the amphibolite-rich units of the Altai Formation in the Fuyun area, and to provide a maximum age constraint for deformation, we collected two samples for zircon U-Pb geochronology. Sample L14FY33 is a quartzite from unit 2 of the Altai Formation (Fig. 5b; GPS: 47°2'37"N, 89°21'57"E). The sample is characterized by a weak S_2 fabric, and predominantly comprises quartz and some feldspar and amphiboles. Sample L15FY22 is a gneissic granitoid with a pervasive S_2 fabric from the core of the macroscopic doubly plunging fold (Fig. 6; GPS: 47°20'18"N, 88°50'56"E).

Zircon grains were separated using conventional crushing, heavy liquid and magnetic techniques, and then mounted in epoxy resin and polished to expose equatorial section. Cathodoluminescence (CL) images were taken at the Department of Earth Sciences, the University of Hong Kong (HKU). Zircon U-Pb isotopic analysis was conducted at the same lab of HKU via a Nu Instruments MC-ICP-MS attached to a Resonetics Resolution M-50-HR

Excimer Laser Ablation System. The analytical procedure and instrument parameters follow Xia et al. (2011) and Geng et al. (2014). We used the ICPMSDataCal software (Liu et al., 2010) to process data, and the ISOPLOT program (Ludwig, 2003) for the weighted mean age calculation and Concordia plots.

Zircons from both samples L14FY33 and L15FY22 are characterized by euhedral shape and oscillatory zoning, indicating an igneous origin. U-Pb analytical results are presented in Appendix B. The 17 analyses for gneissic granitoid sample (L15FY22) yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of $419.6 \pm 2.1\text{Ma}$ (MSWD=0.84, Fig. 11a). As for quartzite sample (L14FY33), the concordant $^{206}\text{Pb}/^{238}\text{U}$ ages (<10% discordance) are shown in histogram/probability density plots. The 49 analyses for sample L14FY33 yielded a single age peak of ~392 Ma (Fig. 11b).

6. Discussion

6.1. Protolith age and the timing of deformation

Zircons from the quartzite (sample L14FY33) yielded a single age peak of ~392 Ma, indicating that the protolith of this sample is either a meta-volcanic rock or a meta-sedimentary rock with a local source. This single zircon age population constrains the maximum protolith age of the Altai Formation in the Fuyun area, which is consistent with middle Devonian fossil assemblages within the limestone layers of the Altai Formation in the Aletai area (Fig. 2c) (BGMRX, 1978b).

333

334 The timing of deformation is unfortunately poorly constrained. Structures around both the
335 eastern and western doubly plunging folds can be correlated with each other based on the
336 observations that the dominant fabric (S_2) in both areas are associated with shallowly
337 plunging stretching lineations and F_3 in both areas are characterized by doubly plunging
338 geometry and steeply dipping axial planes trending ~NW-SE. Our new zircon ages from the
339 gneissic granitoid and quartzite indicate that this deformation must have occurred after 392
340 Ma. $^{40}\text{Ar}/^{39}\text{Ar}$ step heating on syn- S_2 amphibole from an amphibolite (Fig. 3) yielded a
341 plateau age of ~270 Ma (Li et al., 2015b). This age is consistent with other $^{40}\text{Ar}/^{39}\text{Ar}$
342 amphibole ages from the Qiongkuer Domain and the Southern Altai Domain (Fig. 2) (Briggs
343 et al., 2007; Shen et al., 2013), but is younger than the age of metamorphic zircons (~300-280
344 Ma), which mark the time of peak high temperature metamorphism in the Permian (Li et al.,
345 2014; Wang et al., 2014b). Therefore, the ~270 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ amphibole ages were interpreted
346 to indicate cooling of the southern Chinese Altai through the amphibole closure temperature
347 (Li et al., 2015b). This cooling has been attributed to a regional uplift event associated with
348 the development of ~NW-SE fold and fault system (Figs. 2c and 3), which is interpreted to
349 represent transpressional deformation associated with the Irtysh Shear Zone at ~290-252 Ma
350 (Qu and Zhang, 1994; Li et al., 2015a). We therefore think that F_3 folds in our study area
351 developed at ~270 Ma, which together with strike-slip and reverse movements of the Irtysh
352 Shear Zone (Qu and Zhang, 1994; Li et al., 2015a), were responsible for regional uplift. This
353 suggestion for the timing of deformation is also supported by the age of ~NE-SW granitic

dykes (~252 Ma, Zhang et al., 2012) that crosscut F₃ fold structures (Figs. 3 and 6). The timing of F₄ is not well constrained, but most likely is in the Permian given that the ~252 Ma dykes are not affected by this generation of folds (Fig. 6).

The Middle Permian ⁴⁰Ar/³⁹Ar amphibole age (~270 Ma), which was obtained by ⁴⁰Ar/³⁹Ar step heating (Shen et al., 2013; Li et al., 2015b), contrasts with published ⁴⁰Ar/³⁹Ar amphibole ages of ~249-244 Ma around the area of Fig. 5b (Laurent-Charvet et al., 2003).

The reason for this discrepancy, in our opinion, is associated with the fact that Laurent-Charvet et al. (2003) have dated minerals using an *in situ* laser probe total fusion technique. This analytic method could yield geologically younger ages, because the argon system of the analyzed sample is commonly partially thermally disturbed leading to extraction and measurement of mixed gases from different reservoirs (see discussion in Li et al., 2015b). Using a similar technique, Laurent-Charvet et al. (2003) have also obtained two ~245 Ma ⁴⁰Ar/³⁹Ar biotite ages from the macroscopic F₃ folds in the area of Fig. 5b, but the geological interpretation of these ages are equally enigmatic. On a larger scale, available ⁴⁰Ar/³⁹Ar biotite ages from the Qiongkuer Domain are variable and are predominantly Triassic (Laurent-Charvet et al., 2003; Briggs et al., 2009; Li et al., 2015b). This variability may either represent a thermal disturbance associated with the emplacement of Triassic granitoids, or along-strike variations of cooling and exhumation (Li et al., 2015b).

6.2. Structural interpretation and strain partitioning

Our structural observations indicate that the Qiongkuer Domain in the Fuyun area (Fig. 3) was subjected to four phases of folding. The available age constraints indicate that deformation occurred at ~392-252 Ma. During this period, the tectonic setting of the Chinese Altai changed from subduction to collision, following the consumption of the Ob-Zaisan Ocean (Laurent-Charvet et al., 2003; Xiao et al., 2004; Briggs et al., 2007; Xiao et al., 2009; Li et al., 2015a; Xiao et al., 2015). The exact age of this collision is not well constrained. A number of authors have suggested, based on the occurrence of collision-related magmatic rocks, that the collision initiated in the Late Carboniferous or the latest Early Carboniferous (Buslov et al., 2004a; Glorie et al., 2012; Safonova, 2013; Kuibida et al., 2016). However, some ophiolitic rocks along the collisional zone may be as young as the early Permian, which seems to support a Permian collision (e.g. Xiao et al., 2015). The strike-slip displacement along the Irtysh Shear Zone likely occurred after collision, based on the fact that the shear zone bounds the Chinese Altai with the West/East Junggar. Therefore, the major phase of shearing at ~290-252 Ma (Laurent-Charvet et al., 2003; Buslov et al., 2004a; Briggs et al., 2007; Li et al., 2015a and reference therein), could provide a minimum timing constraint for the collision, which is consistent with the fact that the youngest component of the Irtysh accretionary complex along the southern Chinese Altai is Late Carboniferous (Alexeiev and Gegtyarev, 2008; Li et al., 2015a). In the following section, we discuss the structural interpretation of four generations of structures (D_1 - D_4) in the Qiongkuer Domain, as well as their tectonic significance in the context of the subduction and collision history of the southern Chinese Altai.

396

397 The tectonic origin of S_1 and S_2 is enigmatic. The penetrative fabric (S_2) was likely flat-lying
398 prior to F_3 folding, given the doubly plunging geometry of F_3 as discussed below. Such
399 flat-lying structures are also recognized in the Southern Altai Domain (Figs. 2c and 3).
400 Together with evidence for orogen-parallel stretching lineations, this penetrative fabric (S_2) is
401 interpreted to represent an episode of orogenic collapse during the convergence between the
402 Chinese Altai and the East Junggar in the late Paleozoic (Li et al., 2015a). Accordingly, the
403 earlier transposed fabric (S_1) in the Southern Altai Domain is interpreted to represent an
404 episode of contraction, which facilitated the subsequent crustal thinning as represented by the
405 flat-lying fabric and related stretching lineations (Li et al., 2015a). A similar stretching
406 lineation (L_2) oriented parallel to the orogenic structural grains (~NW-SE) is also recognized
407 in the Qiongkuer Domain (Fuyun area) (Qu and Zhang, 1994; Laurent-Charvet et al., 2003;
408 Wang and Xia, 2005a; Li and Sun, 2014 and this study), which together with flat-lying S_2
409 foliations provides evidence for orogen-parallel extension. The available chronological data
410 constrain the timing of S_1 and S_2 at ~392-270 Ma, but it remains unclear whether these
411 phases of deformation represent processes associated with the collision between the Chinese
412 Altai and the East/West Junggar (e.g. Li et al., 2015a), or whether they are related to the
413 earlier period of subduction accretion. Laurent-Charvet et al. (2003) reported that the
414 penetrative S_2 fabric around the macroscopic F_3 folds in Fig. 5b is indicative of top-to-west
415 shearing. However, such observations were neither recognized in earlier work by Qu and
416 Zhang (1994) nor confirmed by our study. In contrast, we notice that S_2 in the map area of

Fig. 5b is predominantly associated with symmetric lens-shaped clasts and conjugate shear sets (Fig. 8c, d), which are indicative of symmetric deformation that contrasts with the interpretation of top-to-the-W shearing.

The regional structural style is controlled by F_3 folds, which show a doubly plunging geometry (Fig. 3). We attribute the doubly plunging geometry to the fold growth in response to buckling. According to Dubey and Cobbold (1977), folds nucleate with non-cylindrical geometry and propagate laterally. When the lateral propagation of a nucleated fold is blocked, the fold hinge would become highly curved (Holdsworth et al., 2002). Such an effect on lateral fold propagation could result from variations in mechanical properties along deformation zones (e.g. Ramsay and Huber, 1987). In the map area, the Altai Formation is rich in amphibolite, which is mechanically different than the turbidite-dominant sequence in the adjacent areas. The contrast in competency between the two lithologies may have blocked the lateral propagation of macroscopic F_3 folds, resulting in the doubly plunging geometry. Alternatively, the doubly plunging geometry of F_3 may have partly been inherited from a primary dome structure within the flat-lying S_2 . Such domes are common in gneiss and migmatite domains that were subjected to regional-scale sub-vertical flattening deformation in response to crustal thinning (e.g. Whitney et al., 2004). An additional interpretation is that the doubly plunging F_3 folds are associated with a Type 1 F_3/F_4 fold interference pattern. We think that this explanation is unlikely, because map-scale F_3 folds show close to tight geometry (Figs. 5, 6 and 7), which tends to form a Type 2 interference pattern (Grujić, 1993).

Furthermore, the F_4 folds appear to be kinematically linked to sinistral shearing subparallel to F_3 axial planes (see discussion below), which cannot generate a Type 1 fold interference pattern.

On a larger scale, the doubly plunging F_3 folds in the Qiongkuer Domain (Fuyun area) are part of a ~NW-SE fold and fault system associated with the development of the Irtysh Shear Zone (Fig. 3) (Qu and Zhang, 1994). This shear zone, which is characterized by a series of ~NW-SE fold zones bounded by sinistral mylonite zones with minor reverse components (Fig. 3), was interpreted to be a transpressional system in response to oblique convergence between the Chinese Altai and the intra-oceanic arc system of the West/East Junggar (Qu and Zhang, 1991; Qu and Zhang, 1994; Li et al., 2015a). Within ~NW-SE fold zones of the Irtysh Shear Zone, fold axial planes are steeply dipping to the northeast (Li et al., 2015a), and are consistent with the steeply dipping axial planes of macroscopic F_3 folds in the Qiongkuer Domain, suggesting that ~NW-SE folds in both domains resulted from the same transpressional event. On the other hand, chronological constraints on the formation of the doubly plunging F_3 folds at ~270 Ma in the Qiongkuer Domain (Section 4.1) are approximately similar to the timing of activity along the sinistral Irtysh Shear Zone (~290-252 Ma) (Laurent-Charvet et al., 2003; Buslov et al., 2004a; Briggs et al., 2007; Li et al., 2015a and reference therein), thus further supporting the interpretation that ~NW-SE macroscopic folds across the southern Chinese Altai are genetically linked with the Irtysh Shear Zone. A series of ~NW-SE reverse faults were also recognized in the southern Chinese

Altai (Fig. 3) (Qu and Zhang, 1994; Wang and Xia, 2005b; Briggs et al., 2007; Briggs et al., 2009). These faults are kinematically compatible with the transpressional deformation, but the timing of fault activity is relatively poorly constrained. They were possibly active in the Cenozoic given the topographic change across fault zones and the evidence for uplift of the Altai Mountain in the Cenozoic (Yuan et al., 2006; Glorie et al., 2012).

F₄ folds in the area of the western doubly plunging folds (Fig. 6) are superimposed on the F₃ structures and are characterized by ~NNW-SSE axial planes. The angular relationships between the F₄ axial plane and the Irtysh Shear Zone, together with predominant sinistral vergence of asymmetric F₄ folds (Section 4.2.2), suggest that F₄ folding corresponds to the strain associated with the sinistral movement along the Irtysh Shear Zone. This interpretation is consistent with constraints on the timing of F₄ folding (Permian and prior to the intrusion of unfolded ~NE-SW granitic dykes), which overlap with the timing of activity along the Irtysh Shear Zone (Laurent-Charvet et al., 2003; Buslov et al., 2004a; Briggs et al., 2007).

The spatial occurrence of folded zones bounded by sinistral mylonite zones across the Qiongkuer and Southern Altai domains (Fuyun area) suggests that transpressional strain was partitioned across the deformation zone, as commonly observed in oblique convergence boundaries (Miller, 1998; Teyssier and Tikoff, 1998; Norris and Cooper, 2001). The transcurrent component in a transpressional zone can either be accommodated in discrete slip planes (Fig. 1b), or partitioned into both internal deformation of folded zones and slip on

discrete faults (Fig. 1c). The former is characterized by fold axial planes subparallel to discrete faults, whereas the latter would result in oblique relationships between fold axial planes and faults (Fig. 1b, c). In the Qiongkuer Domain, the axial plane of F_3 is sub-parallel to the Irtysh Shear Zone. One possible interpretation is that F_3 folds were initially oblique relative to the Irtysh Shear Zone, but were then rotated towards the orientation of the flow apophyses. However, wrench-related folds commonly shows *en echelon* patterns, and progressive fold rotation would lead to hinge-parallel extensional structures (e.g. Jamison, 1991; Teyssier and Tikoff, 1998; Titus et al., 2007), which are not observed in our study area (apart of the L_2 stretching lineation which cannot be attributed to extension during F_3 folding).

An alternative interpretation is that sub-parallelism between the F_3 axial plane and the Irtysh Shear Zone may have resulted from the partitioning of transpressional strain into simple shear components along ~NW-SE mylonite zones and pure shear-dominant deformation in ~NW-SE fold zones (e.g. F_3 , Figs. 1b and 12a). F_3 folds are superimposed by F_4 folds that show kinematic features of sinistral shearing, suggesting that a fraction of transcurrent components may have been partitioned into ~NNW-SSE F_4 folds in the later stage of transpressional deformation. According to Tikoff and Teyssier (1994), one factor controlling the efficiency of strain partitioning along discrete faults is the convergence angle, with the transpressional strain being sufficiently partitioned into discrete faults when such an angle is less than 20° . In the San Andreas fault system (central California), the plate convergent angle

is oriented $\sim 5^\circ$ relative to the fault zone, leading to a significant amount of simple shear (>95%) that is accommodated by discrete strike-slip faults, and a pure shear-dominated deformation in folded zones that are sub-parallel to the fault (Tikoff and Teyssier, 1994; Teyssier et al., 1995; Teyssier and Tikoff, 1998; Argus and Gordon, 2001). A less efficient strain partitioning in Sumatra is associated with convergence angle of $\sim 50^\circ$ that resulted in only $\sim 33\%$ of the strike-slip components accommodated by discrete faults (Tikoff and Teyssier, 1994; McCaffrey et al., 2000). Therefore, we postulate that the convergence angle between the Chinese Altai and the East/West Junggar may have increased in a later stage of transpressional deformation along the Irtysh Shear Zone (Fig. 12), thus leading to a decrease in the partitioning of transcurrent components into the mylonite zones.

The degree of decoupling and weakening along the shear zones can also account for strain partitioning. A weak fault zone can potentially generate independent contractional and transcurrent zones that do not mechanically interact with each other (Mount and Suppe, 1987; Zoback et al., 1987; Mount and Suppe, 1992). It is thus possible that during the early stage of activity along the Irtysh Shear Zone, the shear zone was sufficiently weak and thus accommodated transcurrent components along narrow mylonite zones and distributed contractional deformation in $\sim$ NW-SE fold zones (e.g. F_3 in the Qiongkuer Domain). An enhanced coupling in the later stage of transpressional deformation may have resulted in a less efficient strain partitioning that led to a fraction of transcurrent components partitioned into $\sim$ NNW-SSE F_4 folds. Indeed, structural observations show that the Irtysh Shear Zone

evolved from a high temperature stage (commonly associated with migmatization) to a more brittle setting (Li et al., 2015a), thus leading to strain hardening and enhanced coupling.

Brittle strike-slip structures and their associated paleostress fields have been studied by Glorie et al. (2012) along the Kazakhstan segment of the Irtysh Shear Zone. However, the temporal relationships between the interpreted paleostress orientations and the superimposed F_3/F_4 folds studied by us, remains an open question. Glorie et al. (2012) have suggested that an earlier ~E-W orientation of the maximum horizontal principal stress axis was subsequently rotated into a ~ENE-WSW orientation. The latter could theoretically account for the development of the ~NNW-SSE F_4 folds, but the earlier ~NW-SE F_3 folds are not compatible with the earlier ~E-W orientation of the maximum horizontal principal stress axis. The timing of brittle deformation is relatively poorly constrained, and it is likely that the data documented by Glorie et al. (2012) represent paleostress orientations that are younger than the Permian (~270-252 Ma) ductile deformation associated with the superimposed folds.

7. Conclusions

Four generations of structures were recognized in the Qiongkuer Domain (Fuyun area) of the Chinese Altai. S_1 is locally recognized in low strain areas and was isoclinally folded with axial planes parallel to the dominant fabric (S_2). S_2 is the dominant fabric and is associated with an orogen-parallel stretching lineation. On a larger scale, S_2 defines ~NW-SE macroscopic upright folds with the doubly plunging geometry (F_3), which is overprinted by

~NNW-SSE F_4 folds. Both F_3 and F_4 folds are kinematically compatible and overlap in time with the adjacent Irtysh Shear Zone, and thus we interpret that these two phases of folds were genetically associated with the transpressional deformation along this shear zone. The sub-parallelism between F_3 and the Irtysh Shear Zone may have resulted from a high degree of strain partitioning, which led to pure shear-dominant deformation in fold zones (e.g. F_3) and simple shear deformation along narrow mylonite zones. Strain partitioning may have become less efficient in the later stages of transpressional deformation, thus forming F_4 folds oblique to the Irtysh Shear Zone.

Acknowledgements: This study was financially supported by the Major Basic Research Project of the Ministry of Science and Technology of China (Grant: 2014CB448000 and 2014CB440801), Hong Kong Research Grant Council (HKU705311P, HKU704712P and HKU17303415), National Science Foundation of China (41273048), HKU seed funding (201111159137) and a HKU small grant (201309176226). We thank Rod Holcombe for the discussion with structural interpretations, and Wenjiao Xiao, Damien Delvaux and Jianhua Li for the helpful comments. Jean Wong and Hongyan Geng are acknowledged for their help during the geochronological analysis. This work is a contribution of the Joint Laboratory of Chemical Geodynamics between HKU and CAS (Guangzhou Institute of Geochemistry), IGCP 592 and PROCORE France/Hong Kong Joint Research Scheme.

References:

- Alexeiev, D., Gegtyarev, K., 2008. Accretionary tectonics of Kazakhstan: Main features and principles of reconstruction, 33rd International Geological Congress, Oslo, Norway.
- Allen, M.B., Alsop, G.I., Zhemchuzhnikov, V.G., 2001. Dome and basin refolding and transpressive inversion along the Karatau Fault System, southern Kazakhstan. *J. Geol. Soc.* 158, 83-95.
- Amante, C., Eakins, B.W., 2009. ETOPO1 1 arc-minute global relief model: procedures, data sources and analysis. NOAA Technical Memorandum NESDIS NGDC-24, 19 pp.
- Argus, D.F., Gordon, R.G., 2001. Present tectonic motion across the Coast Ranges and San Andreas fault system in central California. *Geol. Soc. Am. Bull.* 113, 1580-1592.
- Badarch, G., Dickson Cunningham, W., Windley, B.F., 2002. A new terrane subdivision for Mongolia: implications for the Phanerozoic crustal growth of Central Asia. *J. Asian Earth Sci.* 21, 87-110.
- BGMRX, 1978a. Geological Map of the Fuyun Sheet: scale 1:200,000. Unpublished.
- BGMRX, 1978b. Geological Map of the Aletai Sheet: scale 1:200,000. Unpublished.
- Briggs, S.M., Yin, A., Manning, C.E., Chen, Z.-L., Wang, X.-F., Grove, M., 2007. Late Paleozoic tectonic history of the Ertix Fault in the Chinese Altai and its implications for the development of the Central Asian Orogenic System. *Geol. Soc. Am. Bull.* 119, 944-960.
- Briggs, S.M., Yin, A., Manning, C.E., Chen, Z.-L., Wang, X.-F., 2009. Tectonic development of the southern Chinese Altai Range as determined by structural geology,

- 583 thermobarometry, $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology, and Th/Pb ion-microprobe monazite
584 geochronology. *Geol. Soc. Am. Bull.* 121, 1381-1393.
- 585 Buslov, M.M., Fujiwara, Y., Iwata, K., Semakov, N.N., 2004a. Late Paleozoic-Early
586 Mesozoic Geodynamics of Central Asia. *Gondwana Res.* 7, 791-808.
- 587 Buslov, M.M., Watanabe, T., Fujiwara, Y., Iwata, K., Smirnova, L.V., Safonova, I.Y.,
588 Semakov, N.N., Kiryanova, A.P., 2004b. Late Paleozoic faults of the Altai region,
589 Central Asia: tectonic pattern and model of formation. *J. Asian Earth Sci.* 23, 655-671.
- 590 Cai, K., Sun, M., Yuan, C., Long, X., Xiao, W., 2011a. Geological framework and Paleozoic
591 tectonic history of the Chinese Altai, NW China: a review. *Russ. Geol. Geophys.* 52,
592 1619-1633.
- 593 Cai, K., Sun, M., Yuan, C., Zhao, G., Xiao, W., Long, X., Wu, F., 2011b. Prolonged
594 magmatism, juvenile nature and tectonic evolution of the Chinese Altai, NW China:
595 Evidence from zircon U–Pb and Hf isotopic study of Paleozoic granitoids. *J. Asian
596 Earth Sci.* 42, 949-968.
- 597 Carosi, R., Palmeri, R., 2002. Orogen-parallel tectonic transport in the Variscan belt of
598 northeastern Sardinia (Italy): implications for the exhumation of medium-pressure
599 metamorphic rocks. *Geol. Mag.* 139, 497-511.
- 600 Chai, F., Mao, J., Dong, L., Yang, F., Liu, F., Geng, X., Zhang, Z., 2009. Geochronology of
601 metarhyolites from the Kangbutiebao Formation in the Kelang basin, Altay Mountains,
602 Xinjiang: implications for the tectonic evolution and metallogeny. *Gondwana Res.* 16,
603 189-200.

- 604 Chen, J.-F., Han, B.-F., Ji, J.-Q., Zhang, L., Xu, Z., He, G.-Q., Wang, T., 2010. Zircon U–Pb
605 ages and tectonic implications of Paleozoic plutons in northern West Junggar, North
606 Xinjiang, China. *Lithos* 115, 137-152.
- 607 Choulet, F., Faure, M., Cluzel, D., Chen, Y., Lin, W., Wang, B., 2012. From oblique accretion
608 to transpression in the evolution of the Altaid collage: new insights from West Junggar,
609 northwestern China. *Gondwana Res.* 21, 530-547.
- 610 Díaz-Azpiroz, M., Barcos, L., Balanyá, J., Fernández, C., Expósito, I., Czeck, D., 2014.
611 Applying a general triclinic transpression model to highly partitioned brittle-ductile
612 shear zones: a case study from the Torcal de Antequera massif, external Betics,
613 southern Spain. *J. Struct. Geol.* 68, 316-336.
- 614 Delvaux, D., Cloetingh, S., Beekman, F., Sokoutis, D., Burov, E., Buslov, M., Abdrakhmatov,
615 K., 2013. Basin evolution in a folding lithosphere: Altai–Sayan and Tien Shan belts in
616 Central Asia. *Tectonophysics* 602, 194-222.
- 617 Dewey, J.F., Holdsworth, R.E., Strachan, R.A., 1998. Transpression and transtension zones.
618 *Geol. Soc. Spec. Publ.* 135, 1-14.
- 619 Dubey, A.K., Cobbold, P.R., 1977. Noncylindrical flexural slip folds in nature and experiment.
620 *Tectonophysics* 38, 223-239.
- 621 Fossen, H., Tikoff, B., 1993. The deformation matrix for simultaneous simple shearing, pure
622 shearing and volume change, and its application to transpression-transtension
623 tectonics. *J. Struct. Geol.* 15, 413-422.
- 624 Fossen, H., Tikoff, B., Teyssier, C., 1994. Strain modeling of transpressional and

- 625 transtensional deformation. *Norsk Geologisk Tidsskrift* 74, 134-145.
- 626 Geng, H., Brandl, G., Sun, M., Wong, J., Kröner, A., 2014. Zircon ages defining deposition of
627 the Palaeoproterozoic Soutpansberg Group and further evidence for Eoarchaeon crust
628 in South Africa. *Precambrian Res* 249, 247-262.
- 629 Glorie, S., De Grave, J., Delvaux, D., Buslov, M.M., Zhimulev, F.I., Vanhaecke, F., Elburg,
630 M.A., Van den haute, P., 2012. Tectonic history of the Irtysh shear zone (NE
631 Kazakhstan): New constraints from zircon U/Pb dating, apatite fission track dating
632 and palaeostress analysis. *J. Asian Earth Sci.* 45, 138-149.
- 633 Grujić, D., 1993. The influence of initial fold geometry on Type 1 and Type 2 interference
634 patterns: an experimental approach. *J. Struct. Geol.* 15, 293-307.
- 635 Guy, A., Schulmann, K., Clauer, N., Hasalová, P., Seltnann, R., Armstrong, R., Lexa, O.,
636 Benedicto, A., 2014. Late Paleozoic–Mesozoic tectonic evolution of the Trans-Altai
637 and South Gobi Zones in southern Mongolia based on structural and geochronological
638 data. *Gondwana Res.* 25, 309-337.
- 639 Han, Y., Zhao, G., Sun, M., Eizenhöfer, P.R., Hou, W., Zhang, X., Liu, D., Wang, B., Zhang,
640 G., 2015. Paleozoic accretionary orogenesis in the Paleo-Asian Ocean: Insights from
641 detrital zircons from Silurian to Carboniferous strata at the northwestern margin of the
642 Tarim Craton. *Tectonics* 34, doi:10.1002/2014TC003668.
- 643 Harland, W., 1971. Tectonic transpression in caledonian Spitsbergen. *Geol. Mag.* 108, 27-41.
- 644 He, G., Han, B., Yue, Y., Wang, J., 1990. Tectonic division and crustal evolution of Altay
645 orogenic belt in China. *Geoscience of Xinjiang* 2, 9-20 (In Chinese with the English

- 646 abstract).
- 647 He, Y., Sun, M., Cai, K., Xiao, W., Zhao, G., Long, X., Li, P., 2015. Petrogenesis of the
- 648 Devonian High-Mg rock association and its tectonic implication for the Chinese Altai
- 649 orogenic belt, NW China. *J. Asian Earth Sci.* 113, 61-74.
- 650 Holdsworth, R., Tavarnelli, E., Clegg, P., Pinheiro, R., Jones, R., McCaffrey, K., 2002.
- 651 Domainal deformation patterns and strain partitioning during transpression: an
- 652 example from the Southern Uplands terrane, Scotland. *J. Geol. Soc.* 159, 401-415.
- 653 Jamison, W.R., 1991. Kinematics of compressional fold development in convergent wrench
- 654 terranes. *Tectonophysics* 190, 209-232.
- 655 Jiang, Y., Štípská, P., Sun, M., Schulmann, K., Zhang, J., Wu, Q.H., Long, X., Yuan, C.,
- 656 Racek, M., Zhao, G., Xiao, W., 2015. Juxtaposition of Barrovian and migmatite
- 657 domains in the Chinese Altai: a result of crustal thickening followed by doming of
- 658 partially molten lower crust. *J. Metamorph. Geol.* 33, 45-70.
- 659 Jones, R., Tanner, P.W., 1995. Strain partitioning in transpression zones. *J. Struct. Geol.* 17,
- 660 793-802.
- 661 Khain, E., Bibikova, E., Kröner, A., Zhuravlev, D., Sklyarov, E., Fedotova, A.,
- 662 Kravchenko-Berezhnoy, I., 2002. The most ancient ophiolite of the Central Asian fold
- 663 belt: U–Pb and Pb–Pb zircon ages for the Dunzhugur Complex, Eastern Sayan,
- 664 Siberia, and geodynamic implications. *Earth Planet. Sci. Lett.* 199, 311-325.
- 665 Kröner, A., Kovach, V., Belousova, E., Hegner, E., Armstrong, R., Dolgoplova, A., Seltnann,
- 666 R., Alexeiev, D.V., Hoffmann, J.E., Wong, J., Sun, M., Cai, K., Wang, T., Tong, Y.,

- 667 Wilde, S.A., Degtyarev, K.E., Rytisk, E., 2014. Reassessment of continental growth
668 during the accretionary history of the Central Asian Orogenic Belt. *Gondwana Res.* 25,
669 103-125.
- 670 Kuibida, M., Safonova, I.Y., Yermolov, P., Vladimirov, A., Kruk, N., Yamamoto, S., 2016.
671 Tonalites and plagiogranites of the Char suture-shear zone in East Kazakhstan:
672 implications for the Kazakhstan-Siberia collision. *Geoscience Frontiers* 7, 141–150.
- 673 Laurent-Charvet, S., Charvet, J., Shu, L., Ma, R., Lu, H., 2002. Palaeozoic late collisional
674 strike-slip deformations in Tianshan and Altay, Eastern Xinjiang, NW China. *Terra*
675 *Nova* 14, 249-256.
- 676 Laurent-Charvet, S., Charvet, J., Monié, P., Shu, L., 2003. Late Paleozoic strike - slip shear
677 zones in eastern Central Asia (NW China): new structural and geochronological data.
678 *Tectonics* 22, doi:10.1029/2001TC901047.
- 679 Lehmann, J., Schulmann, K., Lexa, O., Corsini, M., Kröner, A., Štípská, P., Tomurhuu, D.,
680 Otgonbator, D., 2010. Structural constraints on the evolution of the Central Asian
681 Orogenic Belt in SW Mongolia. *Am. J. Sci.* 310, 575-628.
- 682 Li, P., Sun, M., 2014. Structural evolution of the southern Chinese Altai: implication for the
683 tectonic evolution, 7th National Symposium on Structure Geology & Geodynamics,
684 Qindao, p 37.
- 685 Li, P., Sun, M., Rosenbaum, G., Cai, K., Yu, Y., 2015a. Structural evolution of the Irtysh
686 Shear Zone (northwestern China) and implications for the amalgamation of arc
687 systems in the Central Asian Orogenic Belt. *J. Struct. Geol.* 80, 142-156.

- 688 Li, P., Yuan, C., Sun, M., Long, X., Cai, K., 2015b. Thermochronological constraints on the
689 late Paleozoic tectonic evolution of the southern Chinese Altai. *J. Asian Earth Sci.* 113,
690 51-60.
- 691 Li, P., Sun, M., Rosenbaum, G., Jiang, Y., Cai, K., 2016. Structural evolution of zonal
692 metamorphic sequences in the southern Chinese Altai and relationships to Permian
693 transpressional tectonics in the Central Asian Orogenic Belt. *Tectonophysics*, in press.
- 694 Li, S., Wang, T., Wilde, S.A., Tong, Y., 2013. Evolution, source and tectonic significance of
695 Early Mesozoic granitoid magmatism in the Central Asian Orogenic Belt (central
696 segment). *Earth Sci. Rev.* 126, 206-234.
- 697 Li, T., Daukeev, S., Kim, B., Tomurtogoo, O., Petrov, O., 2008. Atlas of geological maps of
698 Central Asia and adjacent areas (1:2500000). Geological Publishing House, Beijing,
699 China.
- 700 Li, Z., Yang, X., Li, Y., Santosh, M., Chen, H., Xiao, W., 2014. Late Paleozoic
701 tectono-metamorphic evolution of the Altai segment of the Central Asian Orogenic
702 Belt: constraints from metamorphic P-T pseudosection and zircon U-Pb dating of
703 ultra-high-temperature granulite. *Lithos* 204, 83-96.
- 704 Lin, S., Jiang, D., Williams, P.F., 1998. Transpression (or transtension) zones of triclinic
705 symmetry: natural example and theoretical modelling. *Geol. Soc. Spec. Publ.* 135,
706 41-57.
- 707 Lin, W., Faure, M., Shi, Y., Wang, Q., Li, Z., 2009. Palaeozoic tectonics of the south-western
708 Chinese Tianshan: new insights from a structural study of the

- 709 high-pressure/low-temperature metamorphic belt. *Int. J. Earth Sci.* 98, 1259-1274.
- 710 Liu, F., Wang, Z., Lin, W., Chen, K., Jiang, L., Wang, Q., 2013. Structure deformation and
711 tectonic significance of Erqis fault zone in the southern margin of Chinese Altay. *Acta*
712 *Petrologica Sinica* 29, 1811-1824 (In Chinese with the English abstract).
- 713 Liu, W., 1990. Petrogenetic epochs and peculiarities of genetic types of granitoids in the Altai
714 Mountains, Xinjiang Uygur Autonomous Region, China. *Geotectonica et*
715 *Metallogenia* 14, 44-56 (In Chinese with the English abstract).
- 716 Liu, Y., Gao, S., Hu, Z., Gao, C., Zong, K., Wang, D., 2010. Continental and Oceanic Crust
717 Recycling-induced Melt–Peridotite Interactions in the Trans-North China Orogen:
718 U–Pb Dating, Hf Isotopes and Trace Elements in Zircons from Mantle Xenoliths. *J.*
719 *Petrol.* 51, 537-571.
- 720 Long, X., Yuan, C., Sun, M., Xiao, W., Zhao, G., Wang, Y., Cai, K., Xia, X., Xie, L., 2010.
721 Detrital zircon ages and Hf isotopes of the early Paleozoic flysch sequence in the
722 Chinese Altai, NW China: new constrains on depositional age, provenance and
723 tectonic evolution. *Tectonophysics* 480, 213-231.
- 724 Long, X., Yuan, C., Sun, M., Xiao, W., Wang, Y., Cai, K., Jiang, Y., 2012. Geochemistry and
725 Nd isotopic composition of the Early Paleozoic flysch sequence in the Chinese Altai,
726 Central Asia: evidence for a northward-derived mafic source and insight into Nd
727 model ages in accretionary orogen. *Gondwana Res.* 22, 554-566.
- 728 Ludwig, K.R., 2003. *Isoplot/Ex Version 3.0: A geochronological toolkit for Microsoft Excel.*
729 Berkeley Geochronological Centre Special Publication, 70pp.

- 730 Massey, M.A., Moecher, D.P., 2013. Transpression, extrusion, partitioning, and lateral escape
731 in the middle crust: Significance of structures, fabrics, and kinematics in the Bronson
732 Hill zone, southern New England, USA. *J. Struct. Geol.* 55, 62-78.
- 733 McCaffrey, R., Zwick, P.C., Bock, Y., Prawirodirdjo, L., Genrich, J.F., Stevens, C.W.,
734 Puntodewo, S., Subarya, C., 2000. Strain partitioning during oblique plate
735 convergence in northern Sumatra: Geodetic and seismologic constraints and
736 numerical modeling. *J. Geophys. Res.* 105, 28363-28376.
- 737 Miller, D.D., 1998. Distributed shear, rotation, and partitioned strain along the San Andreas
738 fault, central California. *Geology* 26, 867-870.
- 739 Mount, V.S., Suppe, J., 1987. State of stress near the San Andreas fault: Implications for
740 wrench tectonics. *Geology* 15, 1143-1146.
- 741 Mount, V.S., Suppe, J., 1992. Present - day stress orientations adjacent to active strike - slip
742 faults: California and Sumatra. *J. Geophys. Res.* 97, 11995-12013.
- 743 Norris, R.J., Cooper, A.F., 2001. Late Quaternary slip rates and slip partitioning on the Alpine
744 Fault, New Zealand. *J. Struct. Geol.* 23, 507-520.
- 745 O'Hara, K.D., Yang, X.-Y., Guoyuan, X., Li, Z., 1997. Regional $\delta^{18}\text{O}$ gradients and
746 fluid-rock interaction in the Altay accretionary complex, northwest China. *Geology* 25,
747 443-446.
- 748 Qu, G., Zhang, J., 1991. Irtys structural zone. *Geoscience of Xinjiang* 3, 115-131 (in Chinese
749 with the English abstract).
- 750 Qu, G., Zhang, J., 1994. Oblique thrust systems in the Altay orogen, China. *Journal of*

- 751 Southeast Asian Earth Sciences 9, 277-287.
- 752 Ramsay, J.G., Huber, M.I., 1987. The techniques of modern structural geology. Volume 2:
753 Folds and fractures. Academic press, London, 391 pp.
- 754 Safonova, I., 2013. The Russian-Kazakh Altai orogen: an overview and main debatable issues.
755 Geoscience Frontiers, 537-552.
- 756 Sanderson, D.J., Marchini, W., 1984. Transpression. J. Struct. Geol. 6, 449-458.
- 757 Sarkarinejad, K., Faghih, A., Grasemann, B., 2008. Transpressional deformations within the
758 Sanandaj–Sirjan metamorphic belt (Zagros Mountains, Iran). J. Struct. Geol. 30,
759 818-826.
- 760 Schulmann, K., Thompson, A.B., Lexa, O., Ježek, J., 2003. Strain distribution and fabric
761 development modeled in active and ancient transpressive zones. J. Geophys. Res. 108,
762 2023, doi: 2010.1029/2001JB000632.
- 763 Şengör, A.M.C., Natal'in, B.A., Burtman, V.S., 1993. Evolution of the Altaid tectonic collage
764 and Palaeozoic crustal growth in Eurasia. Nature 364, 299-307.
- 765 Shen, X.-m., Zhang, H.-x., Ma, L., 2013. Zircon U-Pb and amphibole $^{40}\text{Ar}/^{39}\text{Ar}$
766 geochronology of Kuerti Ophiolite in Altay and geological implication. Journal of
767 Guilin University of Technology 33, 394-405 (in Chinese with the English abstract).
- 768 Sun, M., Yuan, C., Xiao, W., Long, X., Xia, X., Zhao, G., Lin, S., Wu, F., Kröner, A., 2008.
769 Zircon U–Pb and Hf isotopic study of gneissic rocks from the Chinese Altai:
770 progressive accretionary history in the early to middle Palaeozoic. Chem. Geol. 247,
771 352-383.

- 772 Sun, M., Long, X., Cai, K., Jiang, Y., Wang, B., Yuan, C., Zhao, G., Xiao, W., Wu, F., 2009.
773 Early Paleozoic ridge subduction in the Chinese Altai: insight from the abrupt change
774 in zircon Hf isotopic compositions. *Science in China Series D: Earth Sciences* 52,
775 1345-1358.
- 776 Teyssier, C., Tikoff, B., Markley, M., 1995. Oblique plate motion and continental tectonics.
777 *Geology* 23, 447-450.
- 778 Teyssier, C., Tikoff, B., 1998. Strike-slip partitioned transpression of the San Andreas fault
779 system: a lithospheric-scale approach. *Geol. Soc. Spec. Publ.* 135, 143-158.
- 780 Tian, Z., Xiao, W., Shan, Y., Windley, B., Han, C., Zhang, J.e., Song, D., 2013. Mega-fold
781 interference patterns in the Beishan orogen (NW China) created by change in plate
782 configuration during Permo-Triassic termination of the Altaids. *J. Struct. Geol.* 52,
783 119-135.
- 784 Tikoff, B., Teyssier, C., 1994. Strain modeling of displacement-field partitioning in
785 transpressional orogens. *J. Struct. Geol.* 16, 1575-1588.
- 786 Titus, S.J., Housen, B., Tikoff, B., 2007. A kinematic model for the Rinconada fault system in
787 central California based on structural analysis of en echelon folds and
788 paleomagnetism. *J. Struct. Geol.* 29, 961-982.
- 789 Tong, L., Xu, Y.G., Cawood, P.A., Zhou, X., Chen, Y., Liu, Z., 2014a. Anticlockwise P-T
790 evolution at ~280 Ma recorded from ultrahigh-temperature metapelitic granulite in the
791 Chinese Altai orogenic belt, a possible link with the Tarim mantle plume? *J. Asian*
792 *Earth Sci.* 94, 1-11.

- 793 Tong, Y., Wang, T., Jahn, B.M., Sun, M., Hong, D.W., Gao, J.F., 2014b. Post-accretionary
794 permian granitoids in the Chinese Altai orogen: geochronology, petrogenesis and
795 tectonic implications. *Am. J. Sci.* 314, 80-109.
- 796 Wang, B., Faure, M., Shu, L., de Jong, K., Charvet, J., Cluzel, D., Jahn, B.m., Chen, Y.,
797 Ruffet, G., 2010. Structural and geochronological study of high - pressure
798 metamorphic rocks in the Kekesu section (northwestern China): Implications for the
799 late Paleozoic tectonics of the Southern Tianshan. *J. Geol.* 118, 59-77.
- 800 Wang, J.-c., Xia, B., 2005a. Geological evidences for post-orogenic extensional collapse in
801 Altaides, NW China. *Journal of Guilin University of Technology* 25, 267-273 (In
802 Chinese with the English abstract).
- 803 Wang, J.-c., Xia, B., 2005b. Large-scale overthrust structure in Altay Orogen, Northwest
804 China. *Journal of Guilin University of Technology* 25, 135-140 (In Chinese with the
805 English abstract).
- 806 Wang, T., Hong, D.W., Jahn, B.M., Tong, Y., Wang, Y.B., Han, B.F., Wang, X.X., 2006.
807 Timing, petrogenesis, and setting of Paleozoic synorogenic intrusions from the Altai
808 Mountains, Northwest China: implications for the tectonic evolution of an
809 accretionary orogen. *J. Geol.* 114, 735-751.
- 810 Wang, T., Jahn, B.-m., Kovach, V.P., Tong, Y., Wilde, S.A., Hong, D.-w., Li, S., Salnikova,
811 E.B., 2014a. Mesozoic intraplate granitic magmatism in the Altai accretionary orogen,
812 NW China: implications for the orogenic architecture and crustal growth. *Am. J. Sci.*
813 314, 1-42.

- 814 Wang, W., Wei, C., Wang, T., Lou, Y., Chu, H., 2009. Confirmation of pelitic granulite in the
815 Altai orogen and its geological significance. *Chin Sci Bull.* 54, 2543-2548.
- 816 Wang, W., Wei, C., Zhang, Y., Chu, H., Zhao, Y., Liu, X., 2014b. Age and origin of sillimanite
817 schist from the Chinese Altai metamorphic belt: implications for late Palaeozoic
818 tectonic evolution of the Central Asian Orogenic Belt. *Int. Geol. Rev.* 56, 224-236.
- 819 Whitney, D.L., Teyssier, C., Vanderhaeghe, O., 2004. Gneiss domes and crustal flow, in:
820 Whitney, D.L., Teyssier, C., Siddoway, C.S. (Eds.), *Gneiss domes in orogeny*.
821 Geological Society of America Special Paper 380, pp. 15-33.
- 822 Wilhem, C., Windley, B.F., Stampfli, G.M., 2012. The Altaids of Central Asia: A tectonic and
823 evolutionary innovative review. *Earth Sci. Rev.* 113, 303-341.
- 824 Windley, B.F., Kröner, A., Guo, J., Qu, G., Li, Y., Zhang, C., 2002. Neoproterozoic to
825 Paleozoic Geology of the Altai Orogen, NW China: New Zircon Age Data and
826 Tectonic Evolution. *J. Geol.* 110, 719-737.
- 827 Windley, B.F., Alexeiev, D., Xiao, W., Kroner, A., Badarch, G., 2007. Tectonic models for
828 accretion of the Central Asian Orogenic Belt. *J. Geol. Soc.* 164, 31-47.
- 829 Xia, X., Sun, M., Geng, H., Sun, Y., Wang, Y., Zhao, G., 2011. Quasi-simultaneous
830 determination of U-Pb and Hf isotope compositions of zircon by excimer
831 laser-ablation multiple-collector ICPMS. *J. Anal. Atom. Spectrom* 26, 1868-1871.
- 832 Xiao, W., Windley, B.F., Hao, J., Zhai, M., 2003. Accretion leading to collision and the
833 Permian Solonker suture, Inner Mongolia, China: termination of the central Asian
834 orogenic belt. *Tectonics* 22, 1069.

- 835 Xiao, W., Windley, B.F., Badarch, G., Sun, S., Li, J., Qin, K., Wang, Z., 2004. Palaeozoic
836 accretionary and convergent tectonics of the southern Altaids: implications for the
837 growth of Central Asia. *J. Geol. Soc.* 161, 339-342.
- 838 Xiao, W., Han, C., Yuan, C., Sun, M., Lin, S., Chen, H., Li, Z., Li, J., Sun, S., 2008. Middle
839 Cambrian to Permian subduction-related accretionary orogenesis of Northern
840 Xinjiang, NW China: implications for the tectonic evolution of central Asia. *J. Asian
841 Earth Sci.* 32, 102-117.
- 842 Xiao, W., Windley, B., Yuan, C., Sun, M., Han, C., Lin, S., Chen, H., Yan, Q., Liu, D., Qin, K.,
843 Li, J., Sun, S., 2009. Paleozoic multiple subduction-accretion processes of the
844 southern Altaids. *Am. J. Sci.* 309, 221-270.
- 845 Xiao, W., Windley, B., Sun, S., Li, J., Huang, B., Han, C., Yuan, C., Sun, M., Chen, H., 2015.
846 A Tale of Amalgamation of Three Collage Systems in the Permian–Middle Triassic in
847 Central-East Asia: Oroclines, Sutures, and Terminal Accretion. *Annual Review of
848 Earth and Planetary Sciences* 43, doi: 10.1146/annurev-earth-060614-105254.
- 849 Xu, J.F., Castillo, P.R., Chen, F.R., Niu, H.C., Yu, X.Y., Zhen, Z.P., 2003. Geochemistry of
850 late Paleozoic mafic igneous rocks from the Kuerti area, Xinjiang, northwest China:
851 implications for backarc mantle evolution. *Chem. Geol.* 193, 137-154.
- 852 Xu, Z., Wang, Q., Cai, Z., Dong, H., Li, H., Chen, X., Duan, X., Cao, H., Li, J., Burg, J.-P.,
853 2015. Kinematics of the Tengchong Terrane in SE Tibet from the late Eocene to early
854 Miocene: Insights from coeval mid-crustal detachments and strike-slip shear zones.
855 *Tectonophysics* 665, 127–148.

- 856 Yakubchuk, A., 2004. Architecture and mineral deposit settings of the Altaid orogenic collage:
857 a revised model. *J. Asian Earth Sci.* 23, 761-779.
- 858 Yang, T., Li, J., Liang, M., Wang, Y., 2015a. Early Permian mantle–crust interaction in the
859 south-central Altaids: High-temperature metamorphism, crustal partial melting, and
860 mantle-derived magmatism. *Gondwana Res.* 28, 371-390.
- 861 Yang, X., Li, Z., Wang, H., Chen, H., Li, Y., Xiao, W., 2015b. Petrology and geochemistry of
862 ultrahigh-temperature granulites from the South Altay orogenic belt, northwestern
863 China: Implications for metamorphic evolution and protolith composition. *Island Arc*
864 24, 169-187.
- 865 Yuan, C., Sun, M., Xiao, W., Li, X., Chen, H., Lin, S., Xia, X., Long, X., 2007. Accretionary
866 orogenesis of the Chinese Altai: insights from Paleozoic granitoids. *Chem. Geol.* 242,
867 22-39.
- 868 Yuan, W., Carter, A., Dong, J., Bao, Z., An, Y., Guo, Z., 2006. Mesozoic–Tertiary exhumation
869 history of the Altai Mountains, northern Xinjiang, China: New constraints from
870 apatite fission track data. *Tectonophysics* 412, 183-193.
- 871 Zhang, B., Zhang, J., Zhong, D., 2010. Structure, kinematics and ages of transpression during
872 strain-partitioning in the Chongshan shear zone, western Yunnan, China. *J. Struct.*
873 *Geol.* 32, 445-463.
- 874 Zhang, C.L., Santosh, M., Zou, H.B., Xu, Y.G., Zhou, G., Dong, Y.G., Ding, R.F., Wang, H.Y.,
875 2012. Revisiting the “Irish tectonic belt”: Implications for the Paleozoic tectonic
876 evolution of the Altai orogen. *J. Asian Earth Sci.* 52, 117-133.

- 877 Zhang, J., Cunningham, D., 2012. Kilometer-scale refolded folds caused by strike-slip
878 reversal and intraplate shortening in the Beishan region, China. *Tectonics* 31, TC3009,
879 doi:3010.1029/2011TC003050.
- 880 Zhang, J., Sun, M., Schulmann, K., Zhao, G., Wu, Q., Jiang, Y., Guy, A., Wang, Y., 2015a.
881 Distinct deformational history of two contrasting tectonic domains in the Chinese
882 Altai: their significance in understanding accretionary orogenic process. *J. Struct.*
883 *Geol.* 73, 64-82.
- 884 Zhang, X., Zhao, G., Sun, M., Eizenhöfer, P.R., Han, Y., Hou, W., Liu, D., Wang, B., Liu, Q.,
885 Xu, B., 2015b. Tectonic evolution from subduction to arc-continent collision of the
886 Junggar ocean: Constraints from U-Pb dating and Hf isotopes of detrital zircons from
887 the North Tianshan belt, NW China. *Geol. Soc. Am. Bull.*, doi:
888 10.1130/B31230.31231.
- 889 Zhuang, Y.X., 1994. The pressure-temperature-space-time (PTSt) evolution of metamorphism
890 and development mechanism of the thermal-structural-gneiss domes in the Chinese
891 Altai. *Acta Geologica Sinica* 68, 35-47 (In Chinese with the English abstract).
- 892 Zoback, M.D., Zoback, M.L., Mount, V.S., Suppe, J., Eaton, J.P., Healy, J.H., Oppenheimer,
893 D., Reasenber, P., Jones, L., Raleigh, C.B., 1987. New evidence on the state of stress
894 of the San Andreas fault system. *Science* 238, 1105-1111.
- 895 Zonenshain, L.P., Kuzmin, M.I., Natapov, L.M., Page, B.M., 1990. *Geology of the USSR: a*
896 *plate - tectonic synthesis*. American Geophysical Union 21, 242 pp.
- 897 Zorin, Y.A., 1999. Geodynamics of the western part of the Mongolia–Okhotsk collisional belt,

- 898 Trans-Baikal region (Russia) and Mongolia. *Tectonophysics* 306, 33-56.
- 899 Zou, T., Cao, H., Wu, B., 1988. Orogenic and anorogenic granitoids of the Altay Mountains,
900 Xinjiang and their discrimination criteria. *Acta Geologica Sinica* 62, 228-245 (In
901 Chinese with the English abstract).
- 902
- 903

Figure caption

Fig. 1. Conceptual models showing strain styles in monoclinic tranpressional zones (Sanderson and Marchini, 1984; Tikoff and Teyssier, 1994; Dewey et al., 1998). (a) Non-partitioning general shear strain; (b) High degree of strain partitioning with simple shear components along discrete slip planes and pure shear deformation in contraction-dominant domains, which results in the parallelism between fold axis and the deformation zone; (c) Low degree of strain partitioning distributing simple shear components in both discrete slip planes and contraction-dominant domain, which leads to obliquity of initial fold axis relative to the deformation zone.

Fig. 2. Geological maps of the research area. (a) A simplified tectonic map showing major tectonic segments of the CAOBS. In this paper, we follow Li et al. (2015a) to refer the Xing'an-Mongolian-East Junggar orogenic system (XMEJOS) to represent arc systems in the East Junggar and southern Mongolia as well as NE China, whereas the West Junggar-Kazakhstan-Tianshan orogenic system (WJKTOS) and the peri-Siberian orogenic system (PSOS) represent arc systems to the south and north of the Irtysh/Chara shear zones, respectively. The topographic image is from Amante and Eakins (2009). (b) A simplified tectonic map showing the major tectonic units around the Chinese Altai (based on Li et al., 2008). ISZ: Irtysh Shear Zone; NEF: North-East Fault; CSZ: Chara Shear Zone. (c) Geological map in the area of the Chinese Altai after Li et al. (2015b). The time range of granitoids is based on the summary by Tong et al. (2014b) and Chen et al. (2010), whereas

major faults are based on Qu et al. (1994), Laurent-Charvet et al. (2003), Briggs et al. (2007), Jiang et al. (2015) and Zhang et al. (2015a).

Fig. 3. Geological map in the southern Chinese Altai (Fuyun area) based on 1:20 000 geological maps. Note that the map highlights major structural elements and published geochronological data (Laurent-Charvet et al., 2003; Briggs et al., 2007; Li et al., 2014; Li et al., 2015b). The geometry of the Irtysh Shear Zone is after Qu and Zhang (1991) and Li et al. (2015a).

Fig. 4. Photographs of representative rocks in the Qiongkuer Domain (Fuyun area). (a) A pebble conglomerate layer marking the boundary of the Kangbutiebao and Altai formations; (b) White and black layers of quartzofeldspathic gneiss and amphibolite in unit 1 of the Altai Formation; (c) Banded chert in unit 1 of the Altai Formation; (d) Amphibolite/amphibole schist in unit 2 of the Altai Formation; (e) A quartzite layer in unit 2 of the Altai Formation, which was folded with the axial plane parallel to S_2 ; (f) Deformed pillow basalt within unit 2 of the Altai Formation. Note that the location information of all photos in this paper (Figs. 4, 8 and 9) is presented in Appendices A and C.

Fig. 5. (a-d) Structural maps and transects around two hinge areas of eastern doubly plunging folds (See the location in Fig. 3). (e-j) Equal-area lower hemisphere stereonet plots for dominant fabric (S_2) and associated stretching lineation (L_2). Note that the axial plane of

macroscopic F_3 is calculated based on map-view fold trace of F_3 and calculated fold hinge.

Fig. 6. Structural map in the area of western doubly plunging folds (see the location in Fig. 3).

Four domains were divided for geometric analysis and stereonet projection (equal area, lower hemisphere).

Fig. 7. Two structural transects across doubly plunging F_3 folds in Fig. 6. See the legend in Fig. 6.

Fig. 8. Photographs of representative structures around the eastern doubly plunging F_3 folds.

(a) A local S_1 fabric that was folded with the axial plane parallel to regional penetrative S_2 (unit 1 of the Altai Formation). (b) A F_2 defined by folded veins with the occurrence of the axial planar fabric (S_2 , unit 1 of the Altai Formation). (c) Conjugate shear bands indicating pure shear-dominant deformation during the development of S_2 (unit 1 of the Altai Formation). (d) A symmetric lithic clast suggesting pure shear-dominant origin of S_2 (unit 1 of the Altai Formation). (e) Stretching lineations (L_2) defined by extended feldspar grains (unit 1 of the Altai Formation). (f) Outcrop-scale F_3 folds in unit 3 of the Altai Formation. (g) Outcrop-scale F_3 folds close to the contact of the Kangbutiebao Formation with the Altai Formation. (h) Deflection of stretching lineations (L_2) by F_3 within unit 3 of the Altai Formation.

Fig. 9. Photographs of representative structures around the western doubly plunging F_3 folds.

(a) S_1 fabric overprinted and transposed by S_2 . (b-c) Penetrative S_2 in amphibolite of the Altai Formation and gneissic granitoids, respectively. (d) A stretching lineation (L_2) defined by extended quartz/feldspar porphyroblasts in an orthogneiss of the Kangbutiebao Formation. (e) Tight F_3 folds in domain 1 of Fig. 6. (f) Gentle F_3 folds in domain 4 of Fig. 6. (g) Spaced S_3 fabric locally occurs in the core area of macroscopic F_3 . (h) S_4 crenulations overprinting S_3 (Altai Formation). Note sinistral vergence of asymmetric F_4 folds.

Fig. 10. Structural maps of F_4 folds in the area of the western doubly plunging folds (see the figure location of Fig. 10a, b in Fig. 6). (a) F_4 folds in domain 2 of Fig. 6. F_4 fold hinge (B_{42}) can be constrained to be 61° - 352° , which together with $\sim 150^\circ$ map-view trace of F_4 defines an axial plane at 78° - 060° . (b) F_4 in domain 3 of Fig. 6 yielded an axial plane of 81° - 240° in the basis of the fold hinge (B_{42}) of 73° - 298° and the map view trace of $\sim 150^\circ$. (c) Stereonet projection of outcrop-scale S_4 (equal area, lower hemisphere) in the area of the western doubly plunging folds.

Fig. 11. (a) Concordia diagrams for zircon U-Pb analyses of sample L15FY22 (gneissic granitoid); (b) Age probability diagrams of zircons U-Pb analyses for quartzite (sample L14FY33) in unit 2 of the Altai Formation.

Fig. 12. Schematic diagrams showing the development of $\sim$ NW-SE doubly plunging F_3 folds

988 and ~NNW-SSE F_4 folds, as well as the relationship with the evolution of the Irtysh Shear
989 Zone. Note that ~NW-SE reverse faults were possibly active during the development of the
990 Irtysh Shear Zone. (a) The sub-parallelism of F_3 fold axis with the Irtysh Shear Zone may
991 have resulted from strain partitioning associated with simple shear deformation along narrow
992 mylonite zones and pure shear-dominant deformation (F_3) in fold zones; (b) The obliquity of
993 F_4 with the shear zone likely results from the less efficient strain partitioning in the later stage
994 of transpressional deformation, so that a fraction of transcurrent components were partitioned
995 into F_4 folds.

996

997

*Figure 1

Fig. 1

*Figure 2

Fig. 2

*Figure 3

Fig. 3

*Figure 4

Fig. 4

*Figure 5

Fig. 5

*Figure 6

Fig. 6

*Figure 7

Fig. 7

*Figure 8

Fig. 8

*Figure 9

Fig.9

*Figure 10

Fig. 10

*Figure 11

Fig. 11

*Figure 12

Fig. 12