Late Cenozoic Clockwise Rotations in the Westernmost Part of the Arcuate Qiulitage Fold-and-Thrust Belt of Southern Tian Shan Foreland and Its Tectonic Implications

Zhiliang Zhang1, Jemin Sun1,2,3, Lixing Lü1,3, Shengchen Tian1,3, Mengmeng Cao1,3, Bai Su1,3, Jiawei Li1,3, and Yousheng Li1,3

1Key Laboratory of Cenozoic Geology and Environment, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing, China, 2CAS Center for Excellence in Tibetan Plateau Earth Sciences, Chinese Academy of Sciences, Beijing, China, 3University of Chinese Academy of Sciences, Beijing, China

Abstract Curvature of fold-and-thrust belts (FTBs) is a rather common feature in foreland basins. The problem of how curves in FTBs originate is essential for understanding the propagation of deformation and tectonic history of orogens. In this study, we carried out systematic paleomagnetic studies in the westernmost part of the Qiulitage FTB, southern Tian Shan foreland, where thrusting, strike-slip fault, and tectonic boundary coexist. Our new results suggest that the study area has been subjected to ~20° clockwise rotations after ~5 Ma. Oroclinal test of the paleomagnetic data across the FTB suggests that oroclinal bending caused by the formation of the curved Qiulitage FTB is the dominate reason for these tectonic rotations. The Kalayuergun dextral strike-slip transfer fault delimiting the western boundary of the Qiulitage FTB is a thin-skinned structure accommodating the discrepancy in horizontal displacement on both sides of it. These results also suggest that the formation of the curved Qiulitage FTB should not be older than ~5 Ma, indicating that the southern Tian Shan foreland has experienced significant tectonic shortening since the latest Miocene to early Pliocene. Our new results, together with previous studies on deformation history, estimates of crustal shortening, GPS observations and earthquake records, suggest that the southern Tian Shan foreland has been subjected to significant deformation during the past ~5 million years, and the crustal shortening is still ongoing.

1. Introduction

Foreland fold-and-thrust belts (FTBs) record preorogenic and synorogenic sedimentation and deformation between an orogen and its undeformed foreland (Reiter et al., 2011). Thin-skinned arcuate FTB in map view that extends farther toward the foreland is a rather common feature in forelands and is strongly linked to the magnitude, direction, and mechanism of crustal shortening (Eichelberger & McQuarrie, 2015). The problem of how curves in FTBs originate is essential for understanding the propagation of deformation and tectonic history of orogens (Marshak, 2004).

In recent years, numerous kinematic models have been proposed to explain the formation of such curved orogens, which can be generally subdivided into two main categories depending on whether the curvature results from vertical-axis rotations or not (Carey, 1955; Johnston et al., 2013; Macedo & Marshak, 1999; Marshak, 2004; Weil & Sussman, 2004). The rotational models suggest that the curvature is acquired during a subsequent rotation of an originally linear FTB (orocline; Musgrave, 2015; van der Voo, 2004; Weil & Sussman, 2004) or coeval with the main orogenic deformation (progressive arc; Meijers et al., 2015; Sussman et al., 2004), whereas the nonrotational model (primary arc) postulates that the curvature is inherited from preexistent geometries during initial deformation and experience no appreciable tightening during subsequent deformation (Hindle et al., 2000; Marshak, 2004; Weil & Sussman, 2004).

It has been proven that paleomagnetism is the primary method for determining the distribution and magnitudes of vertical-axis rotations in curved FTB. The technique has been widely used to evaluate vertical-axis rotations of different arc-shaped sectors, such as the Pennsylvania salient in the central Appalachians (Kent, 1988; Schwartz & van der Voo, 1983), the Umbrian-Marches orogenic belt of the Northern Apennine (Hirt & Lowrie, 1988; Speranza et al., 1997), the frontal thrust systems of SW Sicily
(Speranza et al., 1999), the Bolivian and Patagonian oroclines in the Andes (Maffione et al., 2009, 2010, 2015; Poblete et al., 2014, 2016), and the eastern Pontides-Lesser Caucasus FTB (Meijers et al., 2015). By comparing the timing, directions, and magnitudes of vertical-axis rotations in different parts of foreland FTBs, the best fit model can be identified for explaining the mechanisms of arcuate FTBs origin.

The Tian Shan Range, which extends over 2,500 km from east to west, was once an accretionary orogen that formed in the late Paleozoic (Xiao et al., 2013). After a long period of tectonic quiescence in the Mesozoic, it was eroded into a peneplain landscape until the reactivation since late Miocene caused by the far-field effects of the Indian-Eurasian collision (Allen et al., 1999; Burchfiel et al., 1999; Cunningham et al., 2003; Molnar & Tapponnier, 1975; Sun et al., 2009; Sun & Zhang, 2009; Yin et al., 1998; Zhang et al., 2016). Therefore, Tian Shan is a key tectonic region not only for the study of the Paleozoic multiple accretion and collisional events but also for the understanding of the Cenozoic intracontinental deformation within the Asian interior.

Due to the intracontinental deformation resulted by the Indian-Eurasian collision, thick Cenozoic detrital sediments from the denudation of the adjacent orogens accumulated in the southern and northern forelands of Tian Shan. The basinward migration of thrust loads resulted in the formation of several rows of FTBs (Figure 1) in the forelands (Molnar & Tapponnier, 1975). These foreland FTBs, which are roughly parallel to adjacent orogens, are important geological evidence for N-S crustal shortening and for understanding the reactivation of Tian Shan in response to the Indian-Eurasian collision.

The Baicheng Depression, characterized by arcuate Qiulitage FTB to the south (Figures 1 and 2), is one of the major Cenozoic depocenters on the southern periphery of Tian Shan (Huang et al., 2006; Yin et al., 1998). In the western end of the depression, it is characterized by en echelon folds linked to the Kalayuergun dextral strike-slip transfer fault, which delimits the tectonic boundary between the depression and areas to the west. The tectonic history of the depression has been reviewed by Liu et al. (2000), and some important geological features have been studied by other authors (e.g., Lu et al., 1999; Wang et al., 2011). In recent years, numerous studies have been carried out on magnetostratigraphy (Charreau et al., 2006; Huang et al., 2006, 2010; Jing et al., 2011; T. Zhang et al., 2014; Z. Zhang et al., 2015, 2016), estimates of crustal shortening (Burchfiel et al., 1999; Tian et al., 2016; Zhang et al., 2014), low-temperature thermochronology (Chang et al., 2017; Yu et al., 2014), and deformation timing (Hubert-Ferrari et al., 2007; Saint-Cardier et al., 2016; Sun et al., 2009; Zhang et al., 2016), most of which were concentrated on the eastern part of the depression. To date, although the overall structural pattern and deformation timing of the Qiulitage FTB have been constrained, its mechanism and tectonic effects of the arcuate FTB have not been studied.

In this study, we focus on the tectonic rotations of the arcuate FTB during its formation. We selected a key site located in the western end of the Qiulitage FTB, where strike-slip fault, thrusting, and the boundary of structural units merged. We carried out paleomagnetic investigations of the continental sediments ranging

Figure 1. Simplified tectonic map of the Tian Shan Range and its adjacent areas (including the Junggar and Tarim basins). INC, IND, SCB, NCB, SIB, and KAZ at the top right corner of the graph represent Indo-China, India, South China Block, North China Block, Siberia, and Kazakhstan, respectively. The black rectangle shows the location of the Baicheng-Kuqa Depression.
Figure 2. (a) DEM image of the Baicheng-Kuqa Depression showing the five tectonic units within it. The green lines (A-A’, B-B’, C-C’, and D-D’) represent the four seismic reflection profiles across the depression. The circles in different color are the epicenters of recorded earthquakes acquired from China Earthquake Networks Center (Tang et al., 2017). The squares in different colors represent locations of previous paleomagnetic studies, and the arrows are the tectonic rotations. The red star represents the sampled section in this study. KLYF represents the Kalayuergun dextral strike-slip transfer fault; (b) four representative seismic reflection profiles across the Baicheng-Kuqa Depression. WSP, BZDS, ST, QSB, KYSB, KLYSB, BD, and NMB represent the Wensu Paleouplift, Bozidun syncline, southern Tian Shan, Kelasu-Yiqikelike Structural Belt, Kalayuergun Structural Belt, Baicheng Depression, and Northern Monoclinic Belt, respectively.
from Miocene to Pliocene in order to provide new constraints on the vertical-axis rotations. Then we combine our results with a regional review of existing paleomagnetic results in the same depression to constrain the rotation direction, magnitude, and timing of deformation and ultimately evaluate the formation of the arcuate Qiulitage FTB and its tectonic implications for the deformation in the southern Tian Shan foreland.

2. Geological Setting and Sampling

The Baicheng Depression, stretching above 470 km from east to west (Korla to Aksu), is recognized as a typical rejuvenated foreland basin in the middle part of the southern Tian Shan (Lu et al., 2000). As a result of the migrating thrust loads caused by crustal shortening, five rows of structural belts formed within the depression (Figure 2): the Northern Monocline belt to the northernmost, Kelasu-Yiqikelike Structural Belt, Baicheng-Yangxia Sag, Qiulitage Structural Belt, and the Yaken Structural Belt to the south, respectively (Zhang et al., 2015). Among them, the Qiulitage Structural Belt is a curve-shaped FTB (Figures 1 and 2), which was maybe caused by the differences in crustal shortening rates from east to west (Tian et al., 2016). Different from other parts, the western end of the Qiulitage Structural Belt is characterized by en echelon folds and the Kalayuergun dextral strike-slip transfer fault (Figure 2). In addition, the trends of the strata are variable (from roughly E-W to NW-SE in the western and eastern parts, respectively) according to our field investigation and measurements, which are also visible in Google Earth images (Figure 3).

In the study area, the Cenozoic strata consist of the Oligocene-Miocene Jidike, the late Miocene Kangcun, and the Pliocene Kuqa formation, which were loosely constrained by fossil spore-pollen assemblages (Editing Committee of the Stratigraphy of China, 1999). The Jidike Formation is dominated by lacustrine reddish mudstones/siltstones interbedded with thin greenish siltstones and occasionally conglomerates. The Kangcun Formation is mainly lacustrine-fluvial brownish siltstones/sandstones occasionally interbedded with conglomerates, whereas the Kuqa Formation is dominated by fluvial-alluvial light-brownish siltstones/sandstones interbedded with upward coarsening conglomerates. According to our field investigation and previous studies, numerous salt-related structures developed in the depression, especially the western part of it (Wu et al., 2014; Zhao & Wang, 2016). In the core of the Awate anticline, it is characterized by
large-scale outcrops of Paleocene-Eocene evaporites (Figure 3), which have been thought to be the detachment layer in the whole depression (Wang et al., 2011; Wu et al., 2014; Zhao & Wang, 2016).

A well-exposed section, which is located on the eastern flank of the Awate River, was selected for magnetosтратigraphic sampling. Our sampling was from the lower part of the Jidike formation to the lowest part of the Kuqa formation. Generally, all the Miocene-Pliocene strata dip to NE with relatively constant dip angles between 46° and 52°. We sampled fine-grained reddish-brownish sandstones, siltstones, and mudstones along the section, whereas the coarse-grained gravels and the strata covered by weathered debris, which are not suitable for the analysis of paleomagnetism, were avoid. Generally, the sampling intervals were between ~3 and ~5 m, but they may be as large as ~10 m where the strata were dominated by conglomerates. Totally, 713 oriented core samples were collected from the ~4,800 m-long section using a portable petrol-powered drill and oriented with a magnetic compass corrected for the local magnetic declination for year 2017 (3°55′E, http://www.magnetic-declination.com). Additionally, 22 paleomagnetic sites within the section (see the detailed locations in Table 1) with variable sampling distances were sampled for studying the tectonic rotations in this area. At each site we drilled 10–17 samples along different stratigraphic levels as much as possible in order to average out the secular variation of the geomagnetic field.

3. Methods
3.1. Rock Magnetic Measurements
Standard paleomagnetic samples about ~2 cm in length and ~2.5 cm in diameter were cut in laboratory. All the samples were first subjected to anisotropy of magnetic susceptibility (AMS) measurements using a KLY-4s Kappabridge. Representative samples from different levels were chosen for other rock magnetic measurements, including hysteresis loops and thermal demagnetization of three-component isothermal remanent magnetization (IRM; Lowrie, 1990). Hysteresis loops were measured using a MicroMag 3900 Vibrating Sample Magnetometer. The applied magnetic field was cycled between ±1.5 T. The demagnetization of

<table>
<thead>
<tr>
<th>Age</th>
<th>Site ID</th>
<th>Lat/Long</th>
<th>Strike/Dip</th>
<th>n/m0</th>
<th>Dg/lg</th>
<th>Ds/Is</th>
<th>α95g/α95s</th>
<th>κg/κs</th>
<th>R ± ΔR (°)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Miocene</td>
<td>AT01 41.605/80.7810</td>
<td>318/49</td>
<td>8/0</td>
<td>316.2/75.1</td>
<td>28.5/39.8</td>
<td>12.7/10.7</td>
<td>16.2/27.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT02 41.610/80.808</td>
<td>319/51</td>
<td>9/6</td>
<td>312.7/74.5</td>
<td>29.0/39.0</td>
<td>14.1/9.0</td>
<td>14.2/34.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT03 41.611/80.786</td>
<td>319/51</td>
<td>10/8</td>
<td>311.1/77.3</td>
<td>32.6/39.6</td>
<td>14.8/10.8</td>
<td>11.6/20.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT04 41.611/80.808</td>
<td>317/50</td>
<td>10/3</td>
<td>288.1/79.7</td>
<td>34.4/44.3</td>
<td>9.0/8.4</td>
<td>17.0/17.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT05 41.613/80.808</td>
<td>320/48</td>
<td>9/7</td>
<td>329.1/77.9</td>
<td>34.5/39.0</td>
<td>10.7/11.9</td>
<td>16.5/19.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT06 41.614/80.808</td>
<td>322/52</td>
<td>10/6</td>
<td>256.7/68.6</td>
<td>35.9/56.6</td>
<td>9.1/10.8</td>
<td>16.9/15.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT07 41.615/80.789</td>
<td>320/48</td>
<td>8/8</td>
<td>260.8/74.9</td>
<td>36.8/54.3</td>
<td>7.5/9.6</td>
<td>21.7/18.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT08 41.616/80.808</td>
<td>319/50</td>
<td>7/0</td>
<td>328.5/73.5</td>
<td>28.9/35.5</td>
<td>17.2/15.5</td>
<td>13.3/15.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT09 41.617/80.808</td>
<td>321/49</td>
<td>10/7</td>
<td>297.6/71.7</td>
<td>26.4/45.8</td>
<td>2.6/6.6</td>
<td>37.8/20.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT10 41.618/80.790</td>
<td>319/50</td>
<td>10/8</td>
<td>260.3/66.0</td>
<td>25.0/58.6</td>
<td>8.4/8.5</td>
<td>17.7/17.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT11 41.621/80.793</td>
<td>324/47</td>
<td>10/7</td>
<td>312.9/65.6</td>
<td>20.5/42.8</td>
<td>6.5/8.2</td>
<td>20.6/18.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT12 41.622/80.792</td>
<td>326/49</td>
<td>17/11</td>
<td>313.4/62.7</td>
<td>19.5/41.2</td>
<td>4.2/11.6</td>
<td>19.8/11</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT13 41.624/80.793</td>
<td>320/52</td>
<td>10/8</td>
<td>314.7/66.8</td>
<td>20.8/36.5</td>
<td>13.9/3</td>
<td>14.7/27.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT14 41.626/80.793</td>
<td>319/51</td>
<td>8/6</td>
<td>298.4/67.9</td>
<td>20.1/43.3</td>
<td>12/14.1</td>
<td>16.7/16.3</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT15 41.627/80.793</td>
<td>323/48</td>
<td>10/6</td>
<td>338.7/68.2</td>
<td>27.7/33.1</td>
<td>5.3/7.5</td>
<td>23.3/18.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT16 41.628/80.793</td>
<td>319/47</td>
<td>10/3</td>
<td>320.9/67.8</td>
<td>20.2/38.5</td>
<td>11.8/14.7</td>
<td>14.7/13.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT17 41.630/80.793</td>
<td>322/49</td>
<td>10/8</td>
<td>341.2/59.4</td>
<td>19.7/26.0</td>
<td>11.9/8.1</td>
<td>17.5/37.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td></td>
<td></td>
<td></td>
<td>307.6/72.3</td>
<td>26.8/42.1</td>
<td>4.6/4.2</td>
<td>62.0/73.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pliocene</td>
<td>AT18 41.631/80.793</td>
<td>326/47</td>
<td>10/6</td>
<td>311.0/71.3</td>
<td>31.8/38.7</td>
<td>19.7/14.5</td>
<td>8.8/12.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT19 41.632/80.791</td>
<td>319/50</td>
<td>9/7</td>
<td>347.1/79.3</td>
<td>37.6/34.4</td>
<td>18.4/15.9</td>
<td>8.8/11.5</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT20 41.633/80.791</td>
<td>324/48</td>
<td>10/3</td>
<td>4.2/72.1</td>
<td>38.4/29.3</td>
<td>17.1/9.8</td>
<td>9.0/25.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT21 41.633/80.791</td>
<td>320/46</td>
<td>16/10</td>
<td>357.8/66.7</td>
<td>29.3/27.6</td>
<td>12.0/9.3</td>
<td>10.5/16.8</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>AT22 41.634/80.791</td>
<td>321/47</td>
<td>9/3</td>
<td>339.1/59.7</td>
<td>18.0/28.3</td>
<td>7.5/8.8</td>
<td>20.2/18.4</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean</td>
<td></td>
<td></td>
<td></td>
<td>343.7/70.7</td>
<td>30.9/31.9</td>
<td>9.5/8.1</td>
<td>66.4/90.6</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Note. Site ID, site identification; Lat/Long, the latitude/longitude of the sampling site; Strike/Dip, strike azimuth and dip of bed; n/m0, number of samples used to calculate the site-mean direction/samples showing normal polarity; Dg, Ig, α95g and κg (Ds, Is, α95g and κs) represent declination and inclination of paleomagnetic direction, 95% confidence limit, precision parameter of Fisher statistics in situ (after tilt correction). R ± ΔR, calculated rotation and associated confidence limit according to Butler (1992); positive values mean clockwise rotations.
three-component IRM was measured using a 2G Enterprises Pulse Magnetizer and 2G-755 cryogenic magnetometer. The hard, medium, and soft components were treated in DC fields of 2.7, 0.5, and 0.15 T along the z, y, and x axes, respectively. The samples were then subjected to progressive thermal demagnetization with 18 steps.

### 3.2. Demagnetization

All the standard paleomagnetic samples were then subjected to progressive thermal demagnetization in a TD-48 thermal demagnetizer with 17–18 steps. Generally, demagnetization was performed with temperature intervals 50–100 °C below 585 °C and 10–30 °C above 585 °C. The remanent magnetization was measured by both a 2G-760 U-channel system and a 2G-755R cryogenic magnetometer in the paleomagnetism and Geochronology Laboratory, Institute of Geology and Geophysics, Chinese Academy of Sciences. Demagnetization results were evaluated by stereographic projections and orthogonal vector diagrams (Zijderveld, 1967), and the paleomagnetic directions were determined by principal component analysis suggested by Kirschvink (1980) using four consecutive points at least. The mean directions were calculated using Fisher statistics (Fisher, 1953).

### 4. Results

#### 4.1. Rock Magnetic Results

Hysteresis loops (Figures 4a–4f) of the selected samples are typically wasp waisted in shape illustrating the coexistence of at least two magnetic components with strongly contrasting coercivities or a ferrimagnetism dominated by a single ferromagnet with a broad grain size distribution (Roberts et al., 1995). The results of thermal demagnetization of the three-component IRM (Figures 4g–4l) suggested that three components show distinct unblocking temperatures, with the low-coercivity component unblocked at ~585 °C indicating the presence of magnetite, whereas the high- and medium-coercivity components exhibit an unblocking temperature around ~675–680 °C corresponding to the unblocking temperature of hematite. The above evidence suggests that both magnetite and hematite are present as remanence carriers.

#### 4.2. Demagnetization of the Natural Remanent Magnetization

Progressive thermal demagnetization diagrams of the representative samples are shown in Figure 5. Two or three directions can be isolated for most of the samples: a low-temperature component was removed below ~300 °C, whereas the middle component can be acquired between 400 and 585 °C. After removing the one or two components in low to medium temperature, a high-temperature component can be isolated (usually between 610 and 680 °C), which decays straightforward to the origin. The high unblocking temperature (~680 °C) indicates that hematite is the predominate magnetic mineral in the sediments. The middle- and high-temperature components yield similar paleomagnetic directions, which is mainly due to the overlapping unblocking temperatures of the two main components. Although chemical remanent magnetization (CRM) usually overprints the primary remanent magnetization of red beds due to chemical weathering (Dekkers, 2012; Deng et al., 2007; Huang & Opdyke, 1996), we suggest that the high-temperature components in this study are primary (detrital remanent magnetization, DRM), rather than overprinted by CRM based on the following reasons: (1) they have different unblocking temperature spectrum, DRM in red beds mostly decays in the 600–680 °C range, whereas CRM decays gradually between 200 and 650 °C (Jiang et al., 2015; Kodama, 2012), and (2) thermal demagnetization curves of DRM have a convex shape, while those of CRM are concave in shape (Jiang et al., 2015). The stepwise thermal demagnetization curves (Figure 5) suggest that the remanent magnetizations of all the samples are mostly unblocked between 650 and 680 °C, and the curves are all convex in shape. Therefore, the high-temperature component identified as the characteristic remanent magnetization (ChRM) is interpreted as of primary origin. The ChRM directions are determined by at least four successive demagnetization steps decaying toward the origin. Samples with the maximum angular deviation larger than 15° were rejected for further analyses. In addition, outliers and transitional directions lying over 45° from the mean were systematically discarded (Deenen et al., 2011). Among the 713 demagnetized samples, ChRM directions could be isolated from 432 samples (supporting information Table S1).
Figure 4. Rock magnetic measurements of representative samples. Hysteresis loops before (left corner, a–f) and after slope correction. Progressive thermal demagnetization of a composite IRM (Lowrie, 1990) produced by magnetizing samples from the study area in 0.15, 0.5, and 2.7 T along the x, y, and z axes, respectively (g–l).
4.3. Reliability Tests

The normal and reversed ChRM directions after tilt correction were used to calculate the Fisher mean directions for the reversal test based on the method suggested by McFadden and McElhinny (1990). The samples have a mean normal direction of D/I = 20.9°/36.9° (α95 = 2.9°, κ = 11.7, n = 225) and reversed direction of D/I = 197.7°/−40.2° (α95 = 2.9°, κ = 12.1, n = 207; Figure 6). The angle between the mean normal and reversed directions is 3.4°, indicating that the data set defines a positive reversal test. In addition, the magnetic directions both before and after tilt correction are relatively far from the geocentric axial dipole and present geomagnetic field directions expected in the study area (Figure 6), then excluding a possible recent magnetic overprint. The eigen analysis approach proposed by Tauxe and Watson (1994) was used for the fold

Figure 5. Orthogonal (Zijderveld) vector plots and the attenuation curves of magnetic remanence of the representative specimens from different levels. Directions are plotted in situ; solid and open circles represent vector endpoints projected onto horizontal and vertical planes, respectively. NRM is the natural remanent magnetization. GAD, geocentric axial dipole field direction; PGF, present geomagnetic field direction.
However, the result showed that the remanence achieved its tightest grouping at 65% unfolding. Despite that the negative fold test maybe due to the relatively constant dip angles across the section, the positive reversal test, together with the abundant number of reversals, is taken as an indication of the primary nature of the magnetic remanence.

4.4. Age Determinations of the Awate Anticline

The tilt-corrected 432 ChRM directions representing 432 sampling levels were used to calculate the virtual geomagnetic pole latitudes, together with the declinations and inclinations, establishing a detailed magnetozone sequence (Figure 7). Excluding magnetic polarity zones defined by just one horizon, a total of 28 reversed and 29 normal polarity zones can be identified (Figure 7). Unfortunately, we have not found any mammal fossil helping to constrain the age range of the section. Therefore, we correlate the magnetozone sequence to the geomagnetic polarity timescale (Gradstein et al., 2012) using the following criteria: (1) the ostracoda found in the Jidike formation (Cadona (candona) leei, Paracandona eulectella, Potamocypris reflexa, and Cypria acuta) suggest an Oligocene-Miocene age range (Editing Committee of the Stratigraphy of China, 1999); (2) the presence of vertebrate fossil Hipparion chiai in the upper part of the Jidike formation (Kuchetawu section) suggests an average age of 11–9 Ma (Sun et al., 2009); (3) the assemblages of fossils of sporopollen, ostracoda, and charophyte found in the Kangcun formation (Candona (Lineocypris) matures, C. (Typhlcypris) ionus, C. (Pseudocandona) subequalis, Paracandona eulectella, Cyclocypris cavernosa, and Ilyocypris Cornae) suggest that it should be assigned to the Miocene (Editing Committee of the Stratigraphy of China, 1999). Combined with the constraint provided by the mammal fossil from the Jidike formation, the Kangcun formation should be deposited in the late Miocene; (4) the Kuqa formation contains abundant ostracoda (Candona (Pseudocandona) rostrta, C. (Lineocypris) asceptis, C. (Candona) neglecta, Subulacypris subtilis, Paracandona eulectella, Cyclocypris regularis, Cypridopsis vidua, Zonocypris membranae, Limnocythere lenue, and Eucypris inflata), which are considered to be Pliocene in age (Editing Committee of the Stratigraphy of China, 1999); (5) previous magnetostratigraphic results from different parts of the depression (Charreau et al., 2006; Huang et al., 2006, 2010; Sun et al., 2009; Zhang et al., 2015, 2016); (6) the characteristics of the magnetozone sequence in this study.

Figure 7 represents our preferred correlation between the established magnetozone sequence and GPTS 2012 (Gradstein et al., 2012), which provides the best fitting age range of the Awate section, although there are some missing events (from ~1,400 to 1,940 m in depth), which are likely because of their relatively short duration and/or the low sampling density due to the coarse-grained sediments. The results suggest that the age range of the Awate section is between ~20 and ~5 Ma.
The plot of magnetostratigraphic age versus stratigraphic thickness of the section is shown in Figure 8. The results show that sedimentary accumulation rates are relatively constant at ~22–24 cm/ky, which are compatible with those of the surrounding areas during the same period (e.g., Charreau et al., 2006; Huang et al., 2006; Sun et al., 2009; Zhang et al., 2016).

4.5. Anisotropy of Magnetic Susceptibility

The AMS deciphering the preferred orientation of magnetic minerals in a rock or unconsolidated sediments is usually used to obtain information about geological processes during and/or after deposition (Hrouda, 1982, 1991; Tarling & Hrouda, 1993). AMS can be described by a symmetric second rank tensor with three principal susceptibility axes $K_1$ (the maximum), $K_2$ (intermediate) and $K_3$ (the minimum; Jelinek, 1981; Tarling & Hrouda, 1993).
In this study, the AMS results show obvious changes with stratigraphic position across the section. In order to track potential changes of AMS parameters, the plots of the parameters versus thickness are divided into two parts based on their behaviors at different levels (Figure 9). Generally, $K_m$ ranges from $56.9 \times 10^{-6}$ to $520 \times 10^{-6}$ SI, with an average value of $182 \times 10^{-6}$ SI. It remains relatively constant in the lower part (from the base to 2,900 m) and increases upward. The overall relatively low values of the bulk magnetic susceptibility $K_m$ in the lower part indicate that both the susceptibility and anisotropy of these sediments are dominated by paramagnetic and/or antiferromagnetic minerals, rather than ferromagnetic minerals (Tarling & Hrouda, 1993). However, the abrupt increase of $K_m$ above ~2,900 m may be due to an increase in supply of magnetic minerals during the denudation processes of the southern Tian Shan, mostly from new inputs of volcanic rocks to the basin.

The marked increase of the corrected anisotropy degree $P_j$ and shape parameter $T$ also occurs at the ~2,900-m level. In the lower part (0–2,900 m), sample values of $P_j$ oscillate about a relatively small mean value ($1.07$) and then increase upward. $P_j$ is believed to be an indicator of lithology (such as clay content) or strain (from compaction, etc.; Charreau et al., 2009). If they were controlled by strain (such as compaction), one would expect that $P_j$ will increase and $T$ should become more oblate as a function of depth. However, neither is observed (Figure 9). Therefore, we interpreted to be related to changes in lithology. We have other two lines of evidence: first, in a foreland basin, older strata generally have experienced stronger tectonic shortening compared with the younger strata as demonstrated in the Pyrenean foreland in Spain (Parés et al., 1999); nevertheless, our AMS record displays a different variation trend. Second, the general synchronous variation trends of the bulk magnetic susceptibility $K_m$ and that of the corrected anisotropy degree $P_j$ (please see Figure 9), suggesting a sedimentary control on the anisotropy. The shape parameter $T$ is affected not only by compaction but also by horizontal shortening. The neutral ellipsoids (with $T$ oscillating around 0) may also be the result of the interaction between compaction and horizontal shortening, though its mechanism is still unclear.
Numerous studies have shown that AMS in weakly deformed mudrocks can undergo a series of changes as the intensity of deformation increases (e.g., Mattei et al., 1997; Parés, 2004; Parés et al., 1999; Robion et al., 2007; Sagnotti & Speranza, 1993; Saint-Bezar et al., 2002). The first effect of layer-parallel shortening is the cluster of the $K_1$ perpendicular to the shortening direction, producing a magnetic foliation parallel to the flattening plane, whereas the $K_3$ directions are still normal to bedding (Borradaile & Tarling, 1981; Borradaile & Henry, 1997; Parés et al., 1999; Parés & van der Pluijm, 2002; Pueyo Anchuela et al., 2010; Sagnotti & Speranza, 1993, and references therein). With further shortening, the $K_3$ becomes distributed along a girdle that is parallel to the tectonic shorting direction, and the magnetic ellipsoid is prolate. Additionally, Parés (2004) defined a general $P_T$ $T$ path to show the development of AMS ellipsoids as the intensity of deformation increases in weakly deformed rocks.

**Figure 9.** Variations of AMS parameters ($K_m$, $P_T$, and $T$) versus thickness of the Awate section. Obvious changes of these three parameters occurred synchronously (~2,900 m in thickness). The paleocurrent directions were restored by conventional methods (cross bedding and ripple marks).
Since sediment input of the whole Baicheng-Kuqa Depression has been predominantly from the erosion of Tian Shan to the north, $K_1$ should be distributed in roughly N-S direction if the fabric was depositional. However, in this study, the magnetic lineation is parallel to the fold axis of the Awate anticline, with $K_3$ roughly perpendicular to the bedding (Figure 10). Therefore, we suggest that the fabric can be assigned to embryonic fabrics in weakly deformed sedimentary rocks implying that the sediments in this section have been subjected to incipient deformation. The angles between magnetic foliation and the bedding have been proposed as a way to determine the intensity of tectonic deformation experienced by sediments (Robion et al., 2007; Sagnotti & Speranza, 1993). In this study, the magnetic foliations are roughly parallel to the bedding indicating that the angles between magnetic foliations and the bedding are obviously <15°. Additionally, the $P_T-T$ plots also suggest that tectonic deformation was not strong (Parés, 2004). The types of magnetic fabrics can be compared with embryonic magnetic fabrics in weakly deformed sedimentary rocks (e.g., Parés, 2004; Parés et al., 1999; Robion et al., 2007) recording a signature of incipient deformation. However, the magnetic fabrics are still dominated by depositional processes, rather than being completely overprinted by tectonic deformation.

The Tian Shan range and Tarim Basin, including the studied section, have been subjected to roughly N-S directed tectonic compression due to the far-field effect caused by the collision between India and Eurasia. The weak strains caused $K_3$ directions to become perpendicular to the shortening direction. Generally, the observed AMS reflects the regional stress field caused by the far-field effects in response to the collision between India and Eurasia. It is worth noting that no directional changes in $K_3$ orientations are observed in both parts indicating that the whole section has been equally affected by NE-SW shortening (personal communication with Josep M. Parés) postdating the youngest sediments in this study (~5 Ma).

4.6. Tectonic Rotations

To evaluate the directions and magnitudes of tectonic rotations of the Awate anticline, the Fisher mean directions of the 22 paleomagnetic sampling sites were compared to the expected directions derived from the apparent polar wander path of Eurasia (Torsvik et al., 2012) for the same age range. The rotational analyses were performed using the methods of Butler (1992).

The site-mean directions before and after tilt correction, the sense and magnitudes of tectonic rotations are shown in Table 1. The results suggest that the Miocene and Pliocene sediments have been subjected to
22.6° ± 6.1° and 26.7° ± 5.5° clockwise rotations, respectively. It is obvious that the tectonic rotations postdate the youngest sampled sediments (~5 Ma).

5. Discussions

5.1. The Chronology of the Cenozoic Sediments in the FTBs of Baicheng-Kuqa Depression

To date, there have been numerous magnetostratigraphic studies across the Baicheng-Kuqa Depression (Charreau et al., 2006; Huang et al., 2006, 2010; Jing et al., 2011; Sun et al., 2009; T. Zhang et al., 2014; Z. Zhang et al., 2015, 2016, 2018). However, the age ranges of the Cenozoic sediments are still uncertain. The possible reasons are listed as follows: (1) the definitions of formations vary in each individual study, (2) the absence of diagnostic mammal fossils and volcanic rocks permitting precise stratigraphic assignments, (3) the different sampling interval due to variable lithology along the foreland basin, and (4) the different criteria used to correlate the polarity sequences with GPTS. Therefore, at this stage it is difficult to present sensible correlations of the different magnetostratigraphic columns. In particular, there is a magnetostratigraphic study about 15 km to the east of ours (Jing et al., 2011). However, unfortunately, we do not think that they got reasonable magnetostratigraphic results as their GPS location of the start and finish points of their cross section suggest that the uppermost part of the section was still in the Kuqa Formation, rather than the Xiyu Formation as claimed by the authors. We state this based on our field investigation (Figure 11) and the 1:100,000 geological map of this area (Figure 12).
Although Charreau et al. (2006) conducted a magnetostratigraphic study at the Yaha section (the same FTB as this study) in the eastern part of the depression, their assignments of the sampled column to different formations are still unclear. Sun et al. (2009) reported a mammal fossil (*Hipparion chiai*) in the upper part of the Jidike formation, which is helpful to constrain the age range of the Cenozoic sediments. In this study, the formations are assigned using the same criteria as Sun et al. (2009), and thus, the mammal fossil is used to constrain the correlation of our polarity sequence with GPTS. Although the magnetozones cannot be correlated one by one, especially in the Jidike formation, which may be due to the frequent polarity reversals in the middle-late Cenozoic and lack of absolute ages, our magnetostratigraphic correlation is considered reasonable.

### 5.2. Mechanism for the Variations of AMS Parameters

According to previous studies (Bande et al., 2017; Charreau et al., 2009; Huang et al., 2006; Sun et al., 2009; Zhang et al., 2016), significant deformation occurred since the late Miocene around the Tian Shan range. Our AMS results clearly indicate abrupt synchronous increase of parameters and that of the bulk magnetic susceptibility at the ~2,900-m level, and we suggest that tectonic uplift-induced provenance changes have played a dominant role in the above changes. We propose the following mechanism to account for the abrupt changes in AMS parameters at ~2,900 m. In the latest Miocene, the tectonic stress caused by the Indian-Eurasian collision accelerated the tectonic uplift of the southern Tian Shan. This is evidenced by considerable deceleration of strike-slip motion along the Kashgar-Yecheng Transfer System in the latest Miocene-Pliocene (Sobel et al., 2011) and the tectonic deformation in the Pamir-SW Tian Shan convergence zone at ~5 Ma (Fu et al., 2010; Thompson et al., 2015). Uplift of the mountains can accelerate bedrock denudation in response to high-energy release and thus expose new deeper bedrocks for erosion. Therefore, a larger amount of ferromagnetic minerals began to accumulate in the foreland basin, being responsible for the abrupt increase in $K_m$ and $P_j$ at the ~2,900-m level.

### 5.3. The Nature of the Kalayuergun Dextral Strike-Slip Transfer Fault

As the western boundary between the Quilitage FTB and the areas to the west, the Kalayuergun dextral strike-slip transfer fault is one of the NW trending faults in the northern Tarim Basin. Although several studies have argued about the offset and mechanism of it in recent years (He et al., 2009; Tang et al., 2010; Wang et al., 2004; Yan, 1996), some fundamental questions, such as the nature, timing, and mechanism are still unclear.

Some researchers suggested that the Kalayuergun dextral strike-slip transfer fault is a deep fault formed as early as the late Paleozoic (He et al., 2009; Yan, 1996). However, we propose that it is just a thin-skinned structure, which formed since the latest Miocene to early Pliocene in order to accommodate the
discrepancy in horizontal displacement on both sides of it, based on the following reasons: (1) the seismic reflection profile across it (Figure 13a) shows that the Pre-Cenozoic basement is roughly continuous and horizontal, whereas the overlying Cenozoic sediments have been faulted suggesting that the Kalayuergun dextral strike-slip transfer fault is just a shallow boundary fault, rather than a deep fault cross-cutting the entire upper crust; (2) another seismic reflection profile to the south of the Northern Kalayuergun anticline (Figure 13b) indicates that the Kakayuergun dextral strike-slip transfer fault is about 25 km in length and does not extend southward; (3) the discrepancy in horizontal displacement evidenced by the folded and horizontal Cenozoic strata on the eastern and western side of it, respectively (Figure 13a); (4) the variations in faulting widths (Figure 13a) suggest that the formation of the Kalayuergun dextral strike-slip transfer fault initiated at the boundary between the Kangcun and Kuqa formations (corresponding to the late Miocene to early Pliocene in age; Tang et al., 2010). Seismological records in the study area (Figure 2a) show that there were few earthquakes on both sides of the Kalayuergun dextral strike-slip transfer fault during the last several decades. In addition, no obvious offset of the post-Pliocene sediments has been found in the Digital Elevation Model (DEM) images or during our field investigation, suggesting that the activity of the Kalayuergun dextral strike-slip transfer fault weakened since the late Pliocene.

5.4. Oroclinal Test: Implications for Rotational Pattern and Mechanism

We reviewed the published paleomagnetic results and recalculated the relative rotations in different parts of the depression with respect to the Eurasian apparent polar wander paths of corresponding age (Torsvik et al., 2012). Our compilation is listed in Table 2 and is plotted in Figure 2. In order to verify whether a statistically significant rotational difference (i.e., oroclinal bending) exists at sites characterized by different structural attitude, the oroclinal test (Hirt & Lowrie, 1988; Maffione et al., 2009; Schwartz & van der Voo, 1983; Speranza et al., 1997) was performed on our paleomagnetic data and previous studies. Then we compare the mean paleomagnetic declinations (in tilt-corrected coordinates) to the local bed strikes, regarded as proxies of structural directions (Figure 14). $D_0 = 0^\circ$ and $S_0 = 90^\circ$ were adopted as referenced paleomagnetic declinations and structural directions, respectively (reference values are trivial for the test result). The best fit line was calculated by linear regression analysis to the paleomagnetic data from different parts of the Qiulitage FTB. The slope of the best fit line indicates the degree of correlation between paleomagnetic declinations and structural trending. A unitary slope and a zero slope imply that structural trend variability is and is not (respectively) related to paleomagnetic rotations. The statistical $t$ test suggested by Hirt and Lowrie (1988) was used to assess whether the slope value of the best fit line obtained from our data set is significantly different from zero (indicating no paleomagnetic versus structural trending) or unitary (indicating full correlation between paleomagnetic rotation and structural trending). The $t$ test on the slope of the regression line compared to zero slope gives $t = 7.75$, which is much greater than the critical value at the 99% significance level ($t_{99} = 2.44$). However, when the best fit line is compared to the unitary slope line, the $t$ test yields $t = 2.14$, being smaller than the critical value, implying both lines are statistically indistinguishable.

![Figure 13. Seismic reflection profile across the Kalayuergun dextral strike-slip fault to the north (a) and south (b) of the North Kalayuergun anticline (see the detailed locations in Figure 2a, modified after Tang et al., 2010).](image-url)
We suggest that twofold tectonic implications of the oroclinal test in this study can be concluded: first, the curve-shaped Qiulitage FTB with variable structural trending (i.e., E-W to NW-SE or NE-SW) has undergone local rotations in opposite directions, which mainly resulted from oroclinal bending mechanism; second, the tectonic rotations in the eastern and western ends of the Qiulitage FTB are asymmetric, with ~20° clockwise rotations and variable amounts (~1°–20°) of counterclockwise rotations in the western and eastern end, respectively.

Generally, the paleomagnetic results from the eastern part of the Baicheng-Kuqa Depression suggested counterclockwise rotations with variable magnitudes (e.g., Charreau et al., 2006, 2009; Chen et al., 1992; Fang et al., 1998; Huang et al., 2006). However, the mechanism is still unclear. For example, Chen et al. (1992) suggested that counterclockwise rotations of the Baicheng Depression are local, whereas Huang et al. (2006) proposed that the interactions between the depression and adjoining crustal block resulted in these counterclockwise rotations. Charreau et al. (2009) reviewed several paleomagnetic results on both flanks of Tian Shan and proposed that the strike-slip component in the piedmonts caused these counterclockwise rotations. However, our data from the western end of the depression document a rotational pattern different from that at the eastern part. The oroclinal test on the paleomagnetic directions from previous studies and ours suggests that bending of the Qiulitage FTB was the main reason for the rotations.

To date, numerous studies have been published dealing with the evolution and mechanism of curved FTB (e.g., Davis et al., 1983; Davis & Engelder, 1985; Hatcher, 1989; Konstantinovskaya & Malavieille, 2005; Macedo & Marshak, 1999; Lawton et al., 1994; Lawton et al., 1994; Lawton et al., 1994). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-strike variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Platt, 1986; Thomas, 1977). Based on the long history of the studies, geologists have defined a great variety of curve-forming processes in FTB. The processes can be affected by many factors, such as the along-stripe variations in the predeformational sedimentary thickness (Hatcher, 1989; Lawton et al., 1994; Maceo & Marshak, 1999; Marshak & Wilkerson, 1992; Thomas, 1977).
The leading edge of the curve coincides with the Cenozoic depocenter. The heavy dashed line represents the foreland (modified after Wang et al., 2011). Note that the apex of the curve coincides with the Cenozoic depocenter. The heavy dashed line represents the leading edge of the curve.

Figure 15. Simplified map of the curved Qiulitage fold-and-thrust belt and predeformational sedimentary thickness, southern Tian Shan foreland. The shading represents the thickness of pre-Miocene sediments across the foreland (modified after Wang et al., 2011). Note that the apex of the curve coincides with the Cenozoic depocenter. The heavy dashed line represents the leading edge of the curve.

The magnetic lineation is tilted slightly to SE in in-situ coordinates (Figure 9) indicating that the western end of the Awaite anticline has been subjected to two phases of deformation. As shown in Figure 3, the studied strata are slightly bended to the north, which can be related to the drag force caused by the Kalayuergun dextral strike-slip transfer fault. Although the orocline bending is interpreted as the primary reason for the clockwise rotations in the westernmost part of the Qiulitage FTB, the drag force resulting from the Kalayuergun dextral strike-slip transfer fault cutting the southern limb of the anticline probably increased the magnitude of clockwise rotation of the studied strata.

5.5. Tectonic Implications

Although numerous magnetostratigraphic studies have been carried out in the Baicheng-Kuqa Depression in recent years (e.g., Charreau et al., 2006; Huang et al., 2006, 2010; Jing et al., 2011; Sun et al., 2009; T. Zhang et al., 2014; Z. Zhang et al., 2015, 2016, 2018), the reactivation timing of the Tian Shan range in the Cenozoic is still uncertain. For example, Charreau et al. (2006) suggested that the uplift of Tian Shan accelerated at ~11 Ma, based on the abrupt increase of sedimentation rates; Huang et al. (2006) defined a framework for Cenozoic uplift and deformation of the Tian Shan range at 16–17 and ~7 Ma according to height-dependent changes of AMS and substantial increase in accumulation rate; Jing et al. (2011) also suggested an accelerated uplift of Tian Shan based on the abrupt increase in sedimentation rate. However, the magnetostratigraphy of growth strata in the depression indicated that the uplift of Tian Shan initiated at ~7–5 Ma (Sun et al., 2009; T. Zhang et al., 2014; Z. Zhang et al., 2016, 2018). Such differences are maybe due to the different methods and/or the uncertainty of chronology of the Cenozoic sediments used to estimate the onset of mountain building.

The paleomagnetic data from the eastern and western parts of the curved Qiulitage FTB enable us to decipher its forming mechanism. Our results suggest that the tectonic rotations were mainly the result of orocline bending (Figure 14) and the formation of the curved FTB must, therefore, have occurred after ~5 Ma,
which is roughly contemporaneous with deformation in several anticlines, such as the Yaken (Hubert-Ferrari et al., 2007), the Kuchetawu (Sun et al., 2009), the Kumugeliemu, and the Kasantuokai anticlines (Zhang et al., 2016).

Our study complements several previous analyses, which document the deformation history of the southern Tian Shan foreland since the early Pliocene providing a comprehensive view of the deformation along the range. Although the tectonic uplift of Tian Shan has been constrained as ranging from the Oligocene to Miocene (Bande et al., 2015; Charreau et al., 2006; Hendrix et al., 1994; Huang et al., 2006; Hubert-Ferrari et al., 2007; Sobel et al., 2006; Winfield, 1994; Yin et al., 1998), it is widely accepted that Tian Shan has been subjected to an episode of widespread deformation since the latest Miocene-Pliocene (Bande et al., 2017; Bullen et al., 2003; Burchfiel et al., 1999; Sun et al., 2009; Zhang et al., 2016), and the deformation might have continued to the present (e.g., Brown et al., 1998; Hubert-Ferrari et al., 2005; Saint-­Carlier et al., 2016; Scharer et al., 2004; Tang, 2017; Tian et al., 2016; Wang et al., 2001; Yang et al., 2008). For example, in the eastern Tian Shan, Brown et al. (1998) proposed a slip rate of less than ~2 mm/yr on the bounding thrust fault of the southern Tian Shan. Additionally, Scharer et al. (2004) suggested that the average shortening rates across the Kashgar-­Atushi anticline were ~5 mm/yr since ~1.2 Ma. Hubert-Ferrari et al. (2005) constrained a minimum total shortening rate (>7 mm/yr) for the last 12,500 years across the southern Tian Shan front near Aksu city. Saint-­Carlier et al. (2016) proposed that the Yaken anticline accommodates up to 25% of the total shortening currently absorbed across the whole eastern Tian Shan (~8 mm/yr). Tian et al. (2016) suggested that there was a linear increase in crustal shortening rates from the latest Miocene (~2 mm/yr) to present day (4.7 ± 1.5 mm/yr). The GPS observations showed that the present shortening rates on the central and eastern parts of the southern Tian Shan foreland are ~5 and ~2 mm/yr, respectively (Wang et al., 2001; Yang et al., 2008). Additionally, the frequent historical and modern earthquake records suggest that the current deformation is active. All the evidence indicates that the southern Tian Shan foreland has been subjected to significant deformation during the past ~5 million years and that crustal shortening is still ongoing.

6. Conclusions

1. The detailed magnetostratigraphy of the Awate anticline indicated that the age range of the coarse-grained continental sediments is from ~20 to ~5 Ma. The sedimentary accumulation rates were relatively constant at 22–24 cm/ky during this period.
2. The westernmost part of the Qiulitage fold-and-thrust belt has been subjected to ~20° clockwise rotations mostly after the early Pliocene (~5 Ma), which can be mainly attributed to the oroclinal bending during the formation of the curved belt since that time.
3. The Kalayeurgun dextral strike-slip transfer fault, delimiting the boundary between the Baicheng Depression and the areas to the west, is a thin-skinned structure, which initiated during the formation of the curved Qiulitage fold-and-thrust belt in order to accommodate the discrepancy in horizontal displacement on both sides of it.
4. The asymmetry of the curved Qiulitage fold-and-thrust belt may be related to the along-strike variations in predeformational sedimentary thickness, the strength of detachment, and/or the basal friction.
5. The southern Tian Shan piedmont has been subjected to significant deformation during the past 5 million years, and the crustal shortening is still ongoing.

References


Acknowledgments

The data for this paper are available as in the supporting information Tables S1 and S2. This study was financially supported by the Strategic Priority Research Program of Chinese Academy of Sciences (XDA20070202), National Nature Science Foundation of China (41702209, 41888101 and 41672168), and China Postdoctoral Science Foundation (2016 M601125). We are grateful to Fabio Speranza, Rosanna Maniscalco, and another anonymous reviewer for their contribution to the improvement of this manuscript. John Geissman (Editor) and Augusto Rapalini are thanked for their great help and English improvement. We also thank the constructive suggestions provided by Josep M. Pare, Shihu Li, and Qingqing Qiao.


Yu, S., Chen, W., Evans, N. J., McInnes, B. I. A., Yin, J., Sun, J., et al. (2014). Cenozoic uplift, exhumation and deformation in the north Kuqa Depression, China as constrained by (U-Th)/He thermochronometry. Tectonophysics, 630, 166–182. https://doi.org/10.1016/j.tecto.2014.05.023


