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Structural Geology of a Central Segment of the Qilian Shan-Nan Shan Thrust Belt: Implications for the Magnitude of Cenozoic Shortening in the Northeastern Tibetan Plateau

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Structural Geology of a Central Segment of the Qilian Shan-Nan Shan Thrust Belt: Implications for the Magnitude of Cenozoic Shortening in the Northeastern Tibetan Plateau

A thesis submitted in partial satisfaction
of the requirements for the degree Master of Science in Geology

By

Robin Christine Reith

2013
ABSTRACT OF THESIS

Structural Geology of a Central Segment of the Qilian Shan-Nan Shan Thrust Belt: Implications for the Magnitude of Cenozoic Shortening in the Northeastern Tibetan Plateau

By
Robin Christine Reith

Masters of Science in Geology
University of California, Los Angeles, 2013
Professor An Yin, Chair

Due to the lack of systematic geologic mapping across the Qilian Shan and Nan Shan region, the amount of crustal strain that accumulated in the northeastern margin of the Tibetan plateau during the Cenozoic Indo-Asian collision remains poorly understood. The aim of this study was to establish a structural framework of an ~15,000 km² region in the central Qilian Shan-Nan Shan thrust belt based on detailed geologic mapping. This study reveals the presence of two Cenozoic thrust systems in the region: (1) the Shule thrust system and (2) the Tuo Lai thrust system. The Shule thrust system consists of thin-skinned, north-propagating thrust faults truncated by younger, thick-skinned, south-propagating thrusts and >5000 m-thick Permian to Upper Triassic folded strata. The Tuo Lai thrust system is composed of five main thrust faults that duplicate a regional
unconformity between Carboniferous strata above and Ordovician mélange below. Based on our geologic mapping, we have constructed and restored a balanced cross-section that suggests a minimum shortening during the Cenozoic of 18 km along the Shule thrust system, which corresponds to ~45% crustal strain in this region. Interpretation of zircon U-Pb dates from a foliated granitoid yields an age of 911 ± 12 Ma, confirming previous work suggesting a phase of magmatism between 930 and 900 Ma in the Qilian Shan-Nan Shan region.
This thesis of Robin Christine Reith is approved.

Mark Harrison
Gilles Peltzer
An Yin, Committee Chair

University of California, Los Angeles
2013
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1. INTRODUCTION

A first-order problem in the study of continental tectonics is how intra-plate deformation is distributed and evolves in response to edge-driven, plate-boundary forces (e.g., McKenzie, 1972; Molnar and Tapponnier, 1975; Allmendinger et al., 1997; Atwater and Stock, 1998; McQuarrie et al., 2005; McQuarrie and Wernicke, 2005; Molnar and Stock, 2009; Yin, 2010). Arguably, the best place in the world to address this issue is the Cenozoic Indo-Asia collision zone, spanning from the Himalaya in the south to Lake Baikal in the north, over a distance of more than 3000 km (Molnar and Tapponnier, 1975). Although intense research has been carried out in this region over the past decades, where and how the Indo-Asian convergence has been accommodated remains debated with several competing hypotheses including: (a) lateral extrusion (Tapponnier et al., 1982; Avouac and Tapponnier, 1993), (b) vertically coherent deformation of the entire lithosphere (England and Houseman, 1986), (c) middle/lower-crustal flow (Zhao and Morgan, 1987; Royden et al., 1997, 2008; Clark and Royden, 2000), (d) intracontinental subduction (Tapponnier et al., 2001), and (e) mid-crustal detachment faulting (Burchfiel et al., 1989; Tapponnier et al., 1990; Gaudemer et al., 1995; Meyer et al., 1998; Wang et al., 2011a). Efforts to determine strain distribution across the Indo-Asian collision zone have led to two end-member views that either the (1) convergence was completely absorbed by deformation in the Asian plate (e.g., Tapponnier et al., 1982; England and Houseman, 1986) or (2) convergence was largely or entirely absorbed by underthrusting and/or lower-crustal injection of India into the Asian plate (e.g., Powell and Conaghan, 1973; Zhao and Morgan, 1987; DeCelles et al., 2002; van Hinsbergen et al., 2011, 2012).
The debate on the locus of strain distribution stems largely from the apparent discrepancy between the currently estimated shortening of ~1000-1500 km that is based on field studies across the Indo-Asian collision zone and the > 2500 km Indo-Asian convergence since the onset of the collision at about 50-60 Ma (e.g., Molnar and Tapponnier, 1975; Dewey et al., 1989; Burchfiel and Royden, 1991; LePichon et al., 1992; Yin and Harrison, 2000; van Hinsbergen et al., 2011). Robust determination of the magnitude of Indo-Asian convergence (i.e., > 2500 km for an initial collision at 45 Ma) (Patriat and Achache, 1984; Dewey et al., 1989; LePichon et al., 1992; van Hinsbergen et al., 2011) requires that the strain discrepancy stem either from uncertainties of field-based shortening estimates or loss of crustal lithosphere from the system.

One of the important factors in creating the aforementioned strain discrepancy is the lack of field-based geologic studies in northern Tibet. This is particularly the case for the 1500 km long and 300 km wide Qilian Shan-Nan Shan thrust belt (Taylor and Yin, 2009) (Fig. 1). Although several studies based on field mapping have been carried out in this region, they were mostly focused on determining the strike-slip magnitude of the Haiyuan fault along the eastern edge of the thrust belt (e.g., Burchfiel et al., 1991; Zhang et al., 1991; Gaudemer et al., 1995) (Fig. 2). The lack of systematic and detailed geologic mapping across the Qilian Shan and Nan Shan, in particular, in its west-central segment where thrust faults dominate, makes it impossible to currently estimate how much crustal strain has accumulated in the region during the Cenozoic Indo-Asian collision. The aim of this study is to establish the structural and stratigraphic framework of a region in the central Qilian Shan-Nan Shan thrust belt based on systematic and detailed geologic mapping. The newly acquired geologic map allows the construction of a balanced cross-
section. Based on the stratigraphic ages and their relationship to Cenozoic deformation, a minimum of 45% shortening is estimated.

2. REGIONAL GEOLOGY

The northern margin of the Tibetan plateau is marked by the Qilian Shan-Nan Shan (Shan means mountain in Chinese) thrust belt, which is comprised mostly of northwest-trending ranges and intermontane basins (Fig. 2). The regional elevation ranges from 4000-5000 m in the west and 2000-3000 m in the east. Topography of the western Qilian Shan-Nan Shan region is largely due to Cenozoic thrusts and Altyn Tagh fault (Molnar and Tapponnier, 1975; Peltzer et al., 1989), whereas that of the eastern Qilian Shan-Nan Shan is controlled by strike-slip and thrust faults (Fig. 2). The high topography of the Qilian Shan-Nan Shan ends abruptly along its northwest boundary, the Altyn Tagh fault, but in contrast, gradually decreases to the northeast (Figs. 1 and 2). Morphology across the northeastern corner of the plateau has been attributed to lower-crustal channel flow (e.g., Clark and Royden, 2000; Royden et al., 2008).

2.1 Cenozoic Structures and Deformation History

The Qilian Shan-Nan Shan thrust belt is bounded in the west by the ENE-striking Altyn Tagh fault and in the east by the Liupan Shan thrust belt (Molnar and Tapponnier, 1975; Peltzer et al., 1989; Burchfiel et al., 1991; Zhang et al., 1991; Gaudemer et al., 1995; Meyer et al., 1998). The northern margin of the Tibetan plateau is marked by an active thrust zone in the west and two prominent strike-slip faults in the east (i.e., the Haiyuan and Tianjin faults, Fig. 2). Deformation of the Cenozoic Qilian Shan-Nan Shan thrust belt began at ~50-40 Ma (Jolivet et al., 2001; Dupont-Nivet et al., 2004; Horton et al., 2004; Dai et al., 2006; Yin et al., 2008a,b; Clark et al., 2010; Duvall et al., 2011),
followed by widespread development of thrusts and thrust-bounded intermontane basins at 30-20 Ma (Fang et al., 2005; Garzione et al., 2005; Lease et al., 2007, 2011, 2012a; Jin et al., 2010; Craddock et al., 2011; Lin et al., 2011; Zheng et al., 2010; Liu et al., 2011; X. Wang et al., 2011; W. Wang et al., 2011; Zhuang et al., 2011; Z. Wang et al., 2012; Xiao et al., 2012; H.P Zhang et al., 2012; Lu et al., 2012). Late Cenozoic development of the eastern Qilian Shan-Nan Shan thrust best has been related to the initiation of the sinistral Haiyuan fault (Burchfiel et al., 1991; Meyer et al., 1998; Duvall and Clark, 2010). Estimates of total displacement along the Haiyuan fault is debated, with estimates ranging from 20-30 km (Burchfiel et al., 1991; Zhang et al., 1991) to ~120 km (Gaudemer et al., 1995). All exiting tectonic models envision that the Cenozoic thrusts in the region sole into a south-dipping low-angle (i.e., a few degrees) décollement (e.g., Burchfiel et al., 1989; Tapponnier et al., 1990; Gaudemer et al., 1995; Meyer et al., 1998; Gao et al., 1999). Specifically, Gaudemer et al. (1995) interpreted that the Haiyuan fault soles into this thrust décollement at a depth of about 20 km.

Estimated Quaternary slip rates on the Haiyuan fault vary from 19-11 mm/yr to < 4 mm/yr (Lasserre et al., 1999; Lasserre et al., 2002; Zheng et al., 2009, 2013; Jolivet et al., 2012). Whereas slip rates on active thrusts are 2-4 mm/yr (Hetzel et al., 2004; Zheng et al., 2009, 2013; Seong et al., 2011; Yuan et al., 2011; Z.Q. Zhang et al., 2012). Previous total Cenozoic shortening estimates across the Qilian Shan-Nan Shan thrust belt are largely speculative inferences. Meyer et al. (1998) proposed a total shortening of 150 km across the Qilian Shan-Nan Shan based on a balanced cross-section constructed by interpreting satellite images. The lack of field confirmation makes this estimate of uncertain value. van der Woerd et al. (2001) estimated about 9-12 km of shortening...
across the Danghe Nan Shan thrust, the southernmost thrust in the Qilian Shan-Nan Shan thrust belt, by extrapolating the Holocene slip rates of the fault over the past 8 Ma. Given that the thrust could have initiated at ≥29 Ma (Yin et al., 2002) and slip rate could have varied with time, this estimate remains poorly constrained. Gaudemer et al. (1995) constructed a cross-section across the Haiyuan fault in the eastern Qilian Shan. Using an area-balancing method, they concluded a minimum shortening of 25 km over an original section length of 100 km, implying > 25% shortening. It should be noted that their approach violates the fundamental assumption of the area-balancing method due to the presence of the left-slip Haiyaun fault (i.e., deformation is no longer plane-strain but involves out-of-the-plane mass transfer) (Dahlstrom, 1969).

2.2 Jurassic-Cretaceous Extensional Tectonics and Basin Formation

The Qilian Shan-Nan Shan thrust belt and its neighboring regions (i.e., Hexi corridor to the north, Qaidam basin to the south, and Altyn Tagh Range to the west) (Fig. 2) have experienced at least two phases of extension during the Late Jurassic and Cretaceous (e.g., Vincent and Allen, 1999; Yin et al., 2007, 2008a,b). These events are expressed by the development of east-trending Jurassic and Cretaceous basins bounded by extensional and transtensional faults (e.g., Huo and Tan, 1995; Vincent and Allen, 1999; Chen et al., 2003; Yin et al., 2008a,b). In the eastern Qilian Shan-Nan Shan, no Mesozoic extensional faults have been mapped at the surface. However, fining-upward sequences of Cretaceous strata were documented and interpreted as resulting from extensional tectonics (e.g., Vincent and Allen, 1999). It is possible that the Cretaceous extensional faults responsible for the formation of the Cretaceous basins are buried below Quaternary deposits or obscured by later Cenozoic deformation.
2.3 Early Paleozoic Qilian Orogen and its Basement Rocks

The most dominant tectonic event in northern Tibet is the Early Paleozoic Qilian orogeny (e.g., Yin and Harrison, 2000; Gehrels et al., 2003a,b; Yin et al., 2007; Xiao et al., 2009; Song et al., 2012). Current debates regarding the Qilian orogen are centered around (1) whether its formation was due to south- and/or north-dipping subduction (e.g., Sobel and Arnaud, 1999; Yin and Harrison, 2000; Gehrels et al., 2003a,b; Yin et al., 2007; Xiao et al., 2009; J. Yang et al., 2009, 2012; Yan et al., 2010; Song et al., 2012), and (2) whether the final closure of the Qilian ocean(s) occurred during the Silurian or Devonian (e.g., Xiao et al., 2009; J. Yang et al., 2012). Independent of these questions, the existing data suggest that (1) an open ocean(s) existed in the Qilian Shan-Nan Shan region prior to 530 Ma (e.g., Smith et al., 2006; Tseng et al., 2007; Xia and Song, 2010; Song et al., 2012), (2) oceanic subduction occurred at 477-490 Ma (Su et al., 2004; Zhang et al., 2007), and (3) the Qilian arc was constructed as a result of oceanic subduction (Song et al., 1991; Bian et al., 2001; Xu et al., 2008; Tseng et al., 2009) (Fig. 2).

In the western Qilian Shan, Proterozoic passive-margin sequences were intruded by 920-930 Ma plutons (Gehrels et al., 2003a,b) and were thrust under an eclogite-bearing metamorphic complex that is interpreted to be part of the Early Paleozoic Qilian arc (Tseng et al., 2006). This metamorphic complex contains 775-930 Ma orthogneiss (Tseng et al., 2006; Xue et al., 2009; Tung et al., 2007) and paragneiss with > 880 Ma detrital zircon grains (Tung et al., 2007). Arc activity at 520-440 Ma in the western Qilian Shan was followed by 435-345 Ma post-orogenic magmatism (e.g., Gehrels et al., 2003a; Su et al., 2004; Wu et al., 2004, 2006, 2010; Liu et al., 2006; Quan et al., 2006; He et al., 2007; Tseng et al., 2009; Dang et al., 2011; Song et al., 2012; Xia et al., 2012; Xiao et al., 2012; Xiong et al., 2012).
In the eastern Qilian Shan, Archean to Paleoproterozoic crystalline rocks are intruded by 750-1190 Ma plutons and overlain by Neoproterozoic cratonic sequence and Early Paleozoic arc sequences (Guo et al., 1999; Wan et al., 2001, 2003; Wang et al., 2007; Tung et al., 2007). Arc magmatism occurred in two discrete phases in the eastern Qilian Shan: first at 507-355 Ma, related to the formation of the Qilian arc, and second, at 218-245 Ma during development of the Kunlun arc to the south (Li et al., 2003; Jin et al., 2005; Pei et al., 2005, 2007a,b,c, 2009; Zhang et al., 2006; Chen et al., 2008; Yong et al., 2008; Dong et al., 2009; Guo et al., 2012; Luo et al., 2012).

It is likely that the heterogeneity of the arc basement controls the style of Cenozoic faults and is responsible for lateral variation in structural style in the Qilian Shan-Nan Shan thrust belt. In the western Qilian Shan, Cenozoic structures are predominantly thin-skinned, closely spaced (30-50 km), and laterally continuous over 100s km-long thrusts. In contrast, Cenozoic structures in the eastern Qilian Shan are dominated by short-segmented thrusts terminating at strike-slip faults (Fig. 2).

3. DATA COLLECTION AND ANALYTICAL METHODS

3.1 Field Mapping

The field area is located in the central Qilian Shan-Nan Shan thrust belt, encompassing the Tuo Lai Shan in the north and the Tuo Lai Nan Shan in the south (Figs. 2 and 3). Geologic mapping was conducted over a one month field season in 2012 (Fig. 4). Data collected in the field included bedding-attitude measurements of sedimentary rocks, lithologic descriptions, structural fabrics and lineations of metamorphic rocks, bedding truncations, structural slickenlines, cleavage relationships, fold geometry and shear-sense indicators where possible. The uncertainty of the field relationships shown in
Figure 4 depends on the coverage of the mapping and exposure. Areas without field measurements are extrapolated from interpretation of satellite images (i.e., LANDSAT and Google Earth images) (Fig. 5). A typical spacing of field observations along a traverse is approximately several hundreds of meters. The mapped structures and field measurements were projected on a LANDSAT-7 image with a spatial resolution of 30 m/pixel. Using various false color images from the LANDSAT image, we correlate and extrapolate stratigraphic units of the same composition to areas that were not accessible by foot (Fig. 5). High-resolution Google Earth images (1 to 10 m/pixel resolution) were also used to aid the interpretation.

3.2 Construction and Restoration of a Balanced Cross-section

The line balancing method, in conjunction with the kink-bend technique, was used in constructing a balanced cross-section (Dahlstrom, 1969; Suppe, 1983). The kink-bend method uses polygons to approximate a curved surface. Restoration of the cross-section was done by hand sequentially, from restoring the youngest to oldest structures until marker units tracking the deformation were all restored to their original horizontal positions. Fold restoration was done by assuming flexural slip as the only folding mechanism (Suppe, 1983) and maintaining constant bed thickness before and after deformation (Dahlstrom, 1969).

3.3 U-Pb Zircon Geochronology

U-Pb zircon geochronology was conducted following procedures reported in Quidelleur (1997) using a CAMECA ims1270 ion mass spectrometer at the University of California, Los Angeles. The zircon grains were separated using standard crushing and density methods at the Institute of Geomechanics, Chinese Academy of Geological
Sciences in Beijing. The grains were mounted with zircon standard, AS-3 (1099 Ma: Schmitz et al., 2003) into 1 in epoxy mounts, polished to 0.25 µm and then coated with ~100 Å of Au. Primary beam intensity was ~12 nA with a spot size of ~20 µm. All isotope measurements (\(^{203.5}\)Pb, \(^{204}\)Pb, \(^{206}\)Pb, \(^{207}\)Pb, \(^{208}\)Pb, Th, U, and UO) were obtained over 9 cycles. U-Pb ratios were found relative to zircon standard AS-3 (Paces and Miller, 1993) using a linear calibration curve (UO/U vs. Pb/U relative sensitivity factor) with a slope determined to be 0.204. Absolute abundances of U and Th were calculated using standard 91500 (Wiedenbeck et al., 2004). \(^{204}\)Pb intensity was measured to correct for the presence of common lead, using the anthropogenic Pb ratios of Sañudo-Wilhelmy and Flegal (1994). Data reduction was accomplished using the in-house program ZIPS 3.0.3 and concordia plots and ages were obtained using Isoplot/Ex (Ludwig, 1991a,b).

4. LITHOLOGIC UNITS

The general division and age assignments of major lithologic units follow that of Pan et al. (2004). The more detailed stratigraphic framework used in this study is based on QBMGR (1991).

4.1 Sedimentary Units

Sedimentary units with ages ranging from Carboniferous to Quaternary are distributed throughout the study area (Figs. 4 and 6). Among them, the Carboniferous to Triassic strata are marine, while the Jurassic to Quaternary sequences are non-marine. The ~400 m thick Carboniferous unit was deposited unconformably on top of mélange and metamorphic rocks (Fig. 7A, 7B, and 8A). The main lithology of the unit includes conglomerate, quartz sandstone, arkosic sandstone, and siltstone. This unit is conformably overlain by Permian strata in the northern study area (Tuo Lai Shan) and
unconformably overlain by Middle Triassic strata in the southern study area (Tuo Lai Nan Shan) (Fig. 7B).

The Permian unit is composed mostly of arkosic sandstone with a maximum thickness of ~1000 m. Permian sequences are conformably overlain by Lower to Upper Triassic strata with a combined total thickness of 4500 m. Triassic strata consist of interbedded arkosic sandstone, siltstone, and shale that define turbidite sequences (Fig. 9A).

Jurassic strata are only exposed in the central regions of the mapped area in two intermontane basins (Fig. 4). Its base is defined by a conformable contact with the underlying Upper Triassic sequences. Its top is eroded away and only a minimum thickness of ~1100 m can be estimated for the unit. The lithology of the Jurassic unit is comprised of interbedded quartz sandstone, siltstone and coal-bearing shale (Fig. 9B).

Cretaceous strata are composed of polymictic conglomerate and pebble sandstone. The unit is only exposed along the northern margin of the mapped area. Its base marks an angular unconformity on top of the Ordovician mélange and Carboniferous sequences. From the map relations, its minimum thickness can be estimated as ~270 m, as the top of the unit has been eroded away (Fig. 7A).

Neogene sedimentary rocks are exposed only in the southwestern part of the study area. They lie in the footwalls of several thrusts and are juxtaposed below Triassic units (Fig. 4). The Neogene unit contains interbedded fine-grained sandstone and muddy limestone with resistant conglomerate marker-beds (Fig. 6). Quaternary sediments are composed of fluvial, glacio-fluvial, and lacustrine deposits which were divided into
active fluvial deposition, active fan alluvium, and inactive alluvium during the geologic mapping (Fig. 4).

4.2 Metamorphic and Mélange Units

Metamorphic Rocks. A gneiss complex (unit gn in Fig. 4) consisting of quartzofeldspathic gneiss, mylonitic orthogneiss, mica and garnet schist, meta-basite, garnet amphibolite and phyllite is exposed widely in the study area (Fig. 9D). The orthogneiss component in this complex is expressed by the occurrence of foliated granitic lenses (1-5 m wide). The gneiss complex contains detrital zircon that yields a U-Pb age of 1.8 ± 13 Ga (Tung et al., 2007). The protolith of the unit has been variably assigned to be marine and littoral sandstone deposited in a passive-margin setting (Tung et al., 2007) or igneous rocks formed in an arc setting (Wan et al., 2000; Smith, 2006). White mica $^{39}$Ar/$^{40}$Ar ages and Pb-Th ages of monazites included in syn-kinematic garnets suggests metamorphism occurred 457 to 413 Ma (Liu et al., 2006; A.V. Zuza, unpublished data). A paragneiss unit (sch in Fig. 4) is mapped separately from the gneiss complex. The former consists of mica schist, garnet schist, quartzite, marble and locally phyllite and slate, whereas the latter is characterized by the presence of mylonitic orthogneiss.

Ordovician Mélange. Ordovician rocks form a heterogeneous unit consisting of flysch sequences, volcanics, and meta-sediments. Deformation in the form of penetrative cleavage and schistosity is variable throughout the unit. The main lithology consists of pillow basalt, siltstones, metapelite grading from slate to phyllite, quartzite, limestone, gabbro, and ultramafic rocks (Figs. 9C and 9E). The gabbro yields U-Pb zircon ages of 480-550 Ma (Shi et al., 2004; Song et al., 2012). We refer to this structurally complicated and lithologically diverse Ordovician unit as the Ordovician mélange (O in Fig. 4). The
Ordovician mélange mapped in this study is a segment of a previously documented ophiolite in the Qilian Shan-Nan Shan (Song et al., 2012).

4.3 Igneous Units

A foliated granitoid is exposed along the base of the northern Tuo Lai Nan Shan (unit grf in Fig. 4) and intrudes into the surrounding gneiss complex (Fig. 4). Several other non-foliated plutonic bodies (unit gr in Fig. 4) intrude into the gneiss complex and Carboniferous strata. This relationship places a lower age bound of the earliest Permian for the timing of the plutonism. The composition of the non-foliated plutons range from granite to quartz monzonite with variable amounts of alkali feldspar and plagioclase. A K-Ar age of 345 ± 17 Ma of one plutonic body (Fig. 4) was reported by QBGMR (1991).

5. GEOCHRONOLOGY

Sample AY-09-21-2011 (4) was collected from the foliated granitoid (grf in Fig. 4), from which 17 zircon grains were extracted and analyzed using the UCLA ion-microprobe laboratory (Table 1). The dated zircon grains are euhedral and display oscillatory bands indicative of magmatic origin (Fig. 10). Concordia plots of the dated zircon grains with high levels of concordance is shown in Figure 11. The slight reverse concordance is the result of a single zircon age outside the range of zircon standard ages. This does not affect the interpreted Pb-Pb age presented here. The mean weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age is 911 ± 12 Ma (MSWD=1.3) (Figs. 11 and 12). We interpret this to be the crystallization age of the granitoid.
6. STRUCTURAL GEOLOGY

The dominant structural trend in the study area is northwest-southeast, expressed by the general strike of major faults and folds (Fig. 4). All mapped brittle faults juxtapose older rocks over younger rocks. Minor thrusts and parallel asymmetric folds are closely associated with the brittle faults and found above and below the fault zones. For these reasons, we interpret all mapped brittle faults as thrusts. Some of the faults place Paleozoic and Mesozoic strata over Neogene sediments, suggesting activity in the Cenozoic (Fig. 4), while others cut and offset Quaternary alluvial fans (see details below).

The first-order structures in the mapped area consist of the (1) Shule thrust system in the south and (2) Tuo Lai thrust system in the north (Figs. 3 and 4). The fault systems are separated by the extensively exposed gneiss complex in the central study area, in which the dominant foliation trend is also to the northwest direction (Fig. 4). Below, we describe the juxtaposition relationships, map-view geometry, and kinematic indicators associated with each fault zone.

6.1 Shule Thrust System

The Shule thrust system lies along the southern edge of the Tuo Lai Nan Shan (Fig. 3). This system consists of several fault strands that merge with one another along strike or die out into folds that are parallel to the faults in the thrust system. For the convenience of description we name the faults in the Shule thrust system as Faults 1a, 1b, 1c, 1d, and 1e sequentially from north to south (Fig. 4).

Fault 1a. This fault marks the northern edge of the Shule thrust system and places the gneissic unit over Triassic strata (Fig. 4). The dip of the fault is estimated to be 50-60° using the 3-point-problem method. The fault lies within the gneiss unit in the east but cuts
up section of the footwall strata to the west, placing the gneiss unit over the Middle Triassic and Upper Triassic from east to west. This juxtaposition relationship implies that the stratigraphic throw across the fault increases from east to west. The footwall of Fault 1a exposes a westward-plunging and northward-overturned syncline (Shule syncline in Fig. 4) within the Triassic strata. The syncline is truncated in the west by Fault 1a.

Fault 1b. This is the most laterally continuous fault in the study area. In the east, the fault places a plutonic body (unit gr) and the gneiss complex (unit gn) over Permian and Triassic strata (Fig. 4). There, the fault also truncates an anticline cored by Permian strata in its footwall (Shule anticline in Fig. 4). Across the central segment, the fault places Triassic strata over Triassic and Jurassic strata. The fault also truncates a major unconformity between Carboniferous in the east and Middle Triassic strata in the west in the hanging wall (Figs. 4 and 7B). Along the western segment, the fault places Triassic strata over Jurassic and Neogene strata.

The fault zone slickenlines are dominantly down-dip and plunge 60° to the north. Cleavage is well developed in the Middle Triassic strata directly north of the fault. It dips between 20° and 60° to the north and is associated closely with minor folds (wavelength < 1 m) that have sub-horizontal fold axes and lie parallel to, and directly above, the fault zone.

Fault 1c. This fault dips generally to the south and terminates at Fault 1b in the east and west. The fault juxtaposes Triassic strata over Upper Triassic and Jurassic strata (Fig. 4). The fault zone was observed at one location where it is ~1m wide and dips about 70° to the north at its central segment (Fig. 8B). No clearly defined striations were observed at the exposed fault zone. The fault cuts down-section in the footwall from Jurassic strata
in the east to Triassic strata in the west. In contrast, the same fault cuts up-section in the
hanging wall, from Lower Triassic to Upper Triassic strata from east to west. The fault
truncates the western end of the Shule anticline in the west and the anticline’s northern
limb in the east (Fig. 4). Note that the Shule anticline is north-verging fold with its
northern limb being overturned to the south (Fig. 4). This fold vergence is consistent with
the top-to-the-north sense of motion on Fault 1c.

Fault 1d. This south-dipping fault juxtaposes south-dipping Upper Triassic strata and
north-dipping Jurassic sequences (Fig. 4). The fault zone is not well exposed but its dip
angle and dip direction can be inferred from its relationship to topography. Using a 3-
point-problem method, its dip is estimated to be 35-50° SW. Minor faults parallel to the
main thrust are present directly above and below the main fault zone. One of such zones
displays down-dip slickensides that plunge ~20° to the south. Upper Triassic strata in the
hanging wall define a syncline parallel to the fault. Both the fault and the fold are
overlain in the east by Neogene strata.

Fault 1e. This fault lies south of Fault 1d and places Triassic strata over Neogene
strata (Fig. 4). The fault splits into two branches in the east and west based on the
interpretation of satellite images. The fault zone is observed at one location at its eastern
end. It is about 20 m thick, dips to the northwest, and placing north-dipping Triassic beds
over south-dipping Neogene strata. This structure truncates the syncline common to the
hanging wall of both Faults 1d and 1e in the east.

6.2 Tuo Lai Thrust System

The Tuo Lai thrust system lies along the northern margin of the study area and is
composed of 5 main fault strands that merge with one another along strike into a single
thrust in the northwest (Figs. 3 and 4). The structural trend of Cenozoic thrust parallels the early Paleozoic contact between the gneiss complex in the south and Ordovician mélange in the north (Fig. 4). For the convenience of description we name the faults in the Tuo Lai thrust system as Faults 2a, 2b, 2c, 2d, and 2e sequentially from north to south (Fig. 4).

**Fault 2a.** This structure is the largest through-going fault in the Tuo Lai San (Fig. 3). Along its eastern segment, the fault cuts a gabbro unit in the north and the Ordovician mélange unit in the south (Fig. 4). An ~2 m wide fault zone exposes down-dip slickensides with an average plunge of 61°N. Cleavage in both the gabbro and Ordovician mélange units strike sub-parallel to the trend of the thrust system and dip at ~40-75° N-NE. Northward along strike, Fault 2a juxtaposes the paragneiss unit (sch in Fig. 4) in the hanging wall over the Ordovician mélange unit and south-dipping Cretaceous strata in the footwall (Figs. 4). Cretaceous strata in the footwall form a south-verging syncline, consistent with the thrusting direction of the Tuo Lai thrust system.

**Fault 2b.** This thrust displays a well exposed ~2 meter wide fault zone. The fault dips ~60° to the north. Fault 2b branches from Fault 2a and juxtaposes the paragneiss unit in the north against gabbro and ultramafic rocks in the south and east (Fig. 4). The continuation of this fault zone, and therefore the relationship between the paragneiss complex and ultramafic rocks, are unknown due to inaccessibility of higher elevation regions.

**Fault 2c.** This north-dipping thrust branches from Fault 2a southeastward. The hanging wall consists of the Ordovician mélange unit in which both bedding and north-dipping cleavage strike parallel to the fault zone. From west to east, Fault 2c cuts down
section through folded Cretaceous strata in the west and underlying Carboniferous strata and granitic rocks in the east. Where the fault is exposed, it dips ~50° N (Fig. 8C). Fault scarps across alluvial fans are evident in Google Earth images (Fig. 13) and suggest Holocene activity.

Fault 2d. This thrust branches off Fault 2c and extends to the southeast (Fig. 4). Measured dips range from 49-60° N. The hanging wall consists mostly of Carboniferous strata that form a southwest-verging fault-bend fold (Fig. 8D). The fold plunges 54°W and trends parallel to the thrust. In the east, Fault 2d juxtaposes granitic rocks in the hanging wall over Permian strata in the footwall. The relationship between the Permian sequences and the thrust defines a footwall ramp dipping ~50° to the north.

Fault 2e. This north-dipping thrust is truncated in the northwest by Fault 2c and in the east by Fault 2d. At its western segment, it places Ordovician mélange over gabbro and along its central and eastern segments it juxtaposes Ordovician mélange over Carboniferous through Permian sequences. In the footwall, Carboniferous strata are folded into an anticline-syncline pair with a ~500 m wavelength (Fig. 4).

6.3 Structures along Pluton Margins

A granitoid in the northeastern part of the study area is mylonitized along its southern margin (Fig. 4). The shear zone is 30 m wide and dips to the south. The stretching lineations defined mostly by hornblende plunge in the down dip direction and kinematic indicators, such as verging folds and feldspar sigma-clasts, suggest a top-to-the-north sense of shear (Fig. 4). The interior of the pluton is not deformed ductility as indicated by the lack of foliation. We relate these structures to pluton emplacement (i.e., Paterson et al., 1998).
7. CONSTRUCTION AND RESTORATION OF A BALANCED CROSS-SECTION

A balanced cross-section was constructed using the line-balancing method, which represents an admissible non-unique tectonic model (Judge and Allmendinger, 2011) (Fig. 14A). The regional stratigraphic information of QBGMR (1991) and geometrical constraints from field observations were used to infer the stratigraphic thickness of the units. Cross-section A-A’ was restored for a portion of its length (Figs. 14A and 14B). The Tuo Lai thrust system was ignored for cross-section reconstruction and shortening estimation due to the abundance of mélange units with no marker beds and few stratigraphic cutoffs. Restoring the section yields a total shortening of 18 km with an estimated original section length of 40 km (Fig. 14B). This requires 45% shortening strain to have been accommodated by the Shule thrust system. The evolution of the balanced cross-section may be achieved by four stages of crustal shortening and deformation (Fig. 15).

Stage 1 involves the formation of the Shule anticline and initiation of Fault 1c in Permian strata. The Shule anticline formed as a fault-bend fold with circular hinge-zone geometry during initial propagation of slip along Fault 1c. With continued thrusting Fault 1c cut up-section through Permian to Upper Triassic strata, and truncated the northern limb of the Shule anticline (Fig. 15).

The initiation of Fault 1b and related folds occurs during Stage 2. Fault 1b roots at depth in the gneissic complex and juxtaposes these rocks, and overlying Carboniferous strata, over Triassic-Jurassic sequences. Related to thrust propagation, Triassic sequences form a hanging wall anticline (Fig. 15). The southern limb of the anticline is truncated by Fault 1b. Thrusting on this fault produced widespread footwall deformation including the
formation of a footwall syncline, re-folding of the Shule anticline, and overturning of Fault 1c (Fig. 15).

During Stage 3, Middle Triassic and younger strata in the Tuo Lai Nan Shan developed northwest trending folds. Folding occurred as a syncline (Shule syncline in Figs. 4 and 14A) to the south and anticline in the north. This anticline-syncline pair was produced by the initiation of Fault 1a at depth during Stage 3 (Fig. 15).

In the final stage, propagation or initiation along Fault 1a truncated the northern limb of the Shule syncline (Fig. 15). As a result of thrusting on Fault 1a, the gneiss complex is emplaced over the Shule syncline. Subsequent erosion has removed both sedimentary and metamorphic rocks from the hanging walls of Faults 1a, 1b, and 1c.

8. DISCUSSION

8.1 Uncertainties and Implications of Geologic Mapping

Detailed geologic mapping in a small area of the Qilian Shan-Nan Shan thrust belt reveals two complex thrust systems: (1) the Shule thrust system in the south and (2) the Tuo Lai thrust system in the north (Fig. 3). The dominant structural trend in the study area is northwest-southeast, expressed by the general strike of major faults and folds. The Shule thrust system is composed of both north and south dipping thrusts. South-dipping faults (Faults 1c and 1e in Fig. 4) are truncated by north-dipping faults and are covered in some regions by Neogene strata (Fig. 4), requiring them to be the oldest structures in the system. South-dipping thrusts are thin-skinned, rooting in Permian or younger strata, whereas south-propagating thrust faults root into the gneiss complex (gn in Fig. 4) and are responsible for the large exposure of crystalline basement rocks.
All major thrust faults in the Shule system are associated with folds, ~4-6 km in wavelength, within Permian and Triassic strata (Fig. 4). Out-of-sequence thrusting suggests development of folding prior to thrusting (Fig. 4 and Fig 15). For example, in the hanging wall of Fault 1c, the Shule anticline is an overturned, north-verging fold, consistent with northward propagation of Fault 1c. However, Fault 1c truncates the northern limb of the anticline, cutting up section through Triassic strata from east to west. Similarly, the northwest striking Shule syncline composes the footwall of the Fault 1a, with a north-dipping and overturned axial surface in the west (Fig. 4) This geometry suggests the Shule syncline formed as a fault-propagation fold during thrusting on Fault 1a. However, the Shule syncline axis is truncated in the west by Fault 1a. From these fold-thrust relationships, we suggest that during each stage shortening was accommodated first by folding of Permian-Triassic strata during initial development of thrust faults, and second, by slip on brittle thrust faults resulting in offset of the previously formed fold limbs.

A decrease in slip from west to east along Fault 1a is deduced based on a decrease in stratigraphic offset and topographic relief along Fault 1a (Figs. 3A, 3B and 4). However, this inference is only valid if the fault dip remains constant along strike. The mapped linear trace of the fault along changing topography suggests the fault dip does not vary along strike in the east (Figs. 3 and 4). We interpret that Fault 1a is the dominant structure in the west, whereas Fault 1b accommodated most shortening in the east (Fig. 4).

The Tuo Lai thrust system is composed of both imbricated thrusts and crustal duplexes inferred from fault cutoffs and branching geometry (Fig. 4). In the northwest,
shortening was largely accommodated on one main thrust fault, Fault 2a. However, to the east shortening strain is distributed on multiple fault strands, branching off Fault 2a. This system is responsible for repeating a major unconformity between Carboniferous strata above and Ordovician mélange below (Fig. 4).

South-propagating thrusts in both the Tuo Lai and Shule thrust system place metamorphic basement rock over early Paleozoic-Mesozoic and Cenozoic strata. Similar thrusting style and structural trends suggest all laterally continuous north-dipping thrusts root into a common décollement at depth (Faults 1a, 1b, 2a, 2b, 2c, and 2d). This relationship is highly speculative since our mapped region does not contain faults that link to both the Tuo Lai and Shule thrust systems. Continued detailed geologic mapping is necessary in order to determine the exact relationship between the two thrust systems.

Middle Triassic strata unconformably overlie the gneiss complex in the center of the mapped region (Figs. 4 and 7B). This same unconformity marks the contact between Middle Triassic strata and tilted Carboniferous strata (Fig. 7B). In the Shule anticline, Permian to Upper Triassic strata comprises a continuous sequence. This observation necessitates that this unconformity roots near the base of the Permian unit in the south and migrates northward up section to the base of Middle Triassic unit. This migration is represented in Figure 14 as a buttress unconformity. In this model, Permian and Lower Triassic sediments were deposited in a topographic low against a steep slope resulting in apparent truncation along the contact with the gneiss complex (Fig. 14). The absence of an unconformity at the base of Middle Triassic strata in the Shule anticline can also be explained by Permian to Lower Triassic strata pinching out to the north. Both interpretations do not affect shortening estimates.
8.2 Cenozoic Shortening

According to the current knowledge of the regional tectonic history (e.g., Yin et al., 2007; Xiao et al., 2009; Gehrels et al., 2011; Song et al., 2012), the Qilian Shan-Nan Shan region experienced an arc-continent collision in the Devonian, followed first by shallow-marine deposition from Carboniferous to Jurassic and then Latest Jurassic-Cretaceous intracontinental extension. Assuming that our current understanding of pre-Cenozoic tectonics is accurate and no regional shortening occurred prior to the Cenozoic, all contractional structures involving Carboniferous and younger strata must have formed in the Cenozoic. Assuming these strata were not tilted in the Mesozoic during extension, our balanced restoration would yield a shortening magnitude for Cenozoic deformation.

Previous stratigraphic work has determined that the Neogene strata in Figure 4 is Miocene in age (QBGMR, 1991). These strata are truncated by two major thrusts, and related splays, in the Shule thrust system, Faults 1b and 1e (Fig. 4). We interpret north-dipping thrusts in the Shule thrust system to be kinematically linked and most likely link at depth to common décollement. This is inferred from similar stratigraphic cutoffs (Fig. 14A) and branching geometry in map view (Figs. 3 and 4). Assuming this is correct, the truncation of tilted Cretaceous and Miocene strata by the Shule thrust system also attests to shortening in our field area occurring during the Cenozoic. Based on these arguments, the 45% shortening strain estimated from restoring folded Permian through Upper Triassic strata in Figure 14 most likely resulted from Cenozoic deformation.

We interpret the estimated 45% shortening strain in the Shule thrust system (Fig. 14A) as a minimum due to erosion of hanging wall cutoffs of thrusts (Boyer and Elliot, 1982; Judge and Allmendinger, 2011). In the absence of hanging wall cutoffs, stratigraphic horizons were lined up in order to make displacement as small as possible in
the restored section (Fig. 14B). Figure 16 illustrates how line-balancing produces a minimum estimate of shortening when restoring a balanced cross-section (Judge and Allmendinger, 2011). Eroded hanging wall cutoffs are not the only potential sources of uncertainty in balanced cross-sections. For example, uncertainty in the depth to detachment, potential hidden décollements and duplex systems in the subsurface, non-consistent bed thickness, and additional internal deformation within the mapped units that occurs at a smaller scales (~>100 meters) than the map resolution (Fig. 9A) (e.g., Marrett and Allmendinger, 1992) all add error to shortening estimates. However, these uncertainties would lead to an increase in the actual magnitude of shortening.

If the 45% shortening strain obtained in this study was representative across the whole Qilian Shan-Nan Shan thrust belt (300 km wide), then the initial width of the Qilian Shan Nan Shan region would have been ~580 km, suggesting that there was 280 km of crustal shortening in the Cenozoic. This estimation is only valid if deformation was completely plane-strain and does not involve out-of-the-plane mass transfer (Dahlstrom, 1969).

Development of the western Qilian Shan-Nan Shan thrust belt is related to slip along the Altyn Tagh fault (Molnar and Tapponnier, 1975; Peltzer et al., 1989). Assuming all left-slip on the Altyn Tagh fault was absorbed in the Qilian Shan-Nan Shan thrust belt we can speculate the amount of plane-strain shortening across the thrust belt (Fig. 17). Cenozoic offset on the Altyn Tagh fault is reported to be 375-475 km (Gehrels et al, 2003b; Cowgill et al., 2003). The 280 km of crustal shortening strain in the Qilian Shan-Nan Shan thrust belt suggested here can be compared to the 290-365 km of shortening based on the amount of left slip motion along the eastern end of the Altyn Tagh fault (Yin
and Harrison, 2000) (Fig. 17A). These amounts of shortening favor ~50% shortening strain and >1500 km of shortening required by the model of distributed-strain in the Asian plate (Dewey et al., 1989).

8.3 Kinematic Evolution

Restoration of the Cenozoic shortening strain accumulated in the Shule thrust system suggests a minimum of four stages of northeast-southwest shortening: (1) initiation of north-verging folds in Permian–Upper Triassic strata and related south-dipping thrusts; (2) initiation of north-dipping thrusts and uplifting of the gneiss complex; Fault 1b and its related hanging wall anticline; (3) development of an anticline-syncline pair in the hanging wall of Fault 1b followed by (4) offset of the northern limb of the Shule syncline by Fault 1a. The sequential stages presented here are chronologically non-unique. Stage 1 is constrained as the earliest stage by the truncational relationship between Fault 1c and 1b (Fig. 15). Stage 4 must occur after Stage 3 if our interpretation of the Shule syncline truncated by Fault 1a is correct. However, the order of Stage 2 relative to Stage 3 is interchangeable.

It is important to note that the angular unconformity between Middle Triassic strata above and Carboniferous strata below implies a previous tilting event prior to Stage 1 in the proposed kinematic model (Figs. 7B and 15). This relationship implies the folding of Carboniferous strata and subsequent erosion prior to the deposition of Permian sequences. Minimal exposure of Carboniferous strata and the overlying unconformity in the Shule thrust system and abundant Ordovician mélangé exposed in the Tuo Lai thrust system preclude an estimation for the amount of, and mechanisms for, shortening during this event.
The kinematic history shown in Figure 15 is non-unique and represents only one of several possible interpretations. However, our reconstruction captures the first-order features of Cenozoic deformation in the mapped region, and the observations presented here will add to the current understanding of Cenozoic deformation in the Qilian Shan-Nan Shan thrust belt.

8.4 Foliated Granitoid Age

Ion microprobe U-Pb dating of zircon from a foliated granitoid yields a weighted mean $^{207}\text{Pb} / ^{206}\text{Pb}$ age of 911 ± 12 Ma (Fig. 12). We interpret this to correspond to the age of crystallization of the pluton body and places an upper bound for the gneiss complex protolith. Previous studies within the Qilian Shan-Nan Shan report a cluster of U-Pb zircon ages of granitic rocks ranging between ~900-930 Ma. Our results lie within this range and support a Neoproterozoic phase of magmatism within coeval passive continental margin sequences along the southern edge of the North China Craton. The contact between the foliated granitoid (unit gr in Fig. 4) and the gneiss complex is intrusive giving a Proterozoic upper bound on the age of the protolith of the gneiss complex.

9. CONCLUSION

1. Detailed field mapping reveals two Cenozoic thrust systems within a segment of the central Qilian Shan-Nan Shan thrust belt: (1) the Shule thrust system in the south and (2) the Tuo Lai thrust system in the north. Faulting involves high-grade metamorphic basement that is brought up from depth as thrust sheets along multiple north-dipping thrust faults. Both thrust systems are imbricated and also form smaller crustal duplexes.
2. Construction and restoration of a balanced cross-section perpendicular to major structures resulted in a minimum shortening estimate of 18 km, equating to ~45% shortening strain in the Shule thrust system. Reconstruction of a balanced cross-section yields a minimum of four stages of Cenozoic shortening in the Shule thrust system.

3. U-Pb dating of zircon grains from a foliated graniotid yields a mean weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 911 ± 12 Ma (MSWD=1.3) which is consistent with previous work that indicates a phase of magmatism in the Qilian Shan-Nan Shan region between ~900-930 Ma.
FIGURE CAPTIONS

Figure 1. Regional map of the Himalayan-Tibetan orogen with major tectonic boundaries showing the location of the Qilian Shan-Nan Shan thrust belt in the northeastern Tibetan Plateau. Underlying base map from Ryan et al. (2005) and (www.geomapapp.org).

Figure 2. Cenozoic tectonic map of the Qilian Shan-Nan Shan thrust belt and its adjacent regions. Major faults are denoted with black lines (thrust), blue lines (sinistral strike-slip), yellow lines (dextral strike-slip), and pink lines (normal). Underlying base map from Ryan et al. (2005) and (www.geomapapp.org).

Figure 3. (A) Shaded topographic relief map of the Tuo-Lai Nan Shan and Tuo-Lai Shan indicating the location of the Shule and Tuo Lai thrust systems. Heavy black lines mark major faults as expressed in the geomorphology. (B) Topographic profiles across the Shule and Tuo Lai thrust system (Vertical exaggeration: x8). Underlying base map from Ryan et al. (2005) and (www.geomapapp.org).

Figure 4. Geologic map of a region in the central Qilian Shan-Nan Shan thrust belt produced for this study. Location of cross-section A-A’.

Figure 5. (A) Exemplary 731 false color LANDAT-7 image used to correlate and extrapolate between field measurements and observations. Spatial resolution of 30 m/pixel. (B) Corresponding mapped region with stratigraphic contacts extrapolated from field observations using above LANDSAT-7 false color image.
Figure 6. Simplified stratigraphic column of mapped region adapted from QBGMR (1991). Asterisked thickness of units represent thicknesses obtained with field observations and cross-section construction.

Figure 7. (A) Map highlighting angular unconformites between Cretaceous strata, Carboniferous strata and Ordovician mélange. (B) Map highlighting the angular unconformity between Middle Triassic and Carboniferous strata and the nonconformity separating Carboniferous and the gneiss complex.

Figure 8. Field photos and annotations: (A) Angular unconformity between near-horizontal Carboniferous rocks and vertically foliated Ordovician rocks below (B) Overturned thrust (Fault 1c) juxtaposing Middle Triassic and Jurassic strata. (C) Ordovician meta-sediments thrust over Cretaceous non-marine sequences. (D) Hanging wall anticline above Fault 2d.

Figure 9. Field photos of representative rock units. (A) Upper Triassic siltstone with minor folds (B) Ripple marks in Jurassic sandstone indicating younging direction (C) Ordovician siltstone and shale as flysh deposits in mélange unit (D) Orthogneiss in gneiss complex with sigma clasts and top-to-the-north sense of shear (E) Pillow basalts exemplary for volcanics in the Ordovician mélange.
**Figure 10.** Cathodoluminescent images of typical zircon grains selected for U-Pb dating. Grains selected for analysis were euhedral and showed magmatic oscillatory bands.

**Figure 11.** Concordia plots of 17 zircon grains from a foliated granitoid sample, AY-09-21-2011 (4) (see Fig. 4 for location). Error ellipses represent 1σ.

**Figure 12.** $^{207}$Pb/$^{206}$Pb ages for 17 zircon grains from sample, AY-09-21-2011 (4), with 2σ error bars. Grains are displayed from left to right in increasing age. The interpreted $^{207}$Pb/$^{206}$Pb results give an age of $911 \pm 12$ Ma (MSWD=1.3).

**Figure 13.** Image of active alluvial fans from Google Earth and corresponding interpretation indicating thrusts in the Tuo Lai thrusts system controlling fan geometry.

**Figure 14.** (A) Balanced cross-section A-A’ (see Fig. 4 for location) and shortening calculations indicating 18 km of shortening and ~45% crustal strain. (B) Restored section between pins in A-A’.

**Figure 15.** Balanced restoration between pins in A-A’ and non-unique sequential development.

**Figure 16.** Schematic cross-section taken from Judge and Allmendinger (2011) illustrating “minimum shortening estimate” associated with line-balanced cross-sections.
(A) The actual section without erosion and total amount of offset. (B) Section with erosion and minimal restoration.

**Figure 17.** Shortening estimate assuming all slip from the Altyn Tagh strike-slip fault was absorbed in the Qilian Shan-Nan Shan thrust belt: ~290-365 km of shortening strain produced from 375-475 km of total slip along the Altyn Tagh fault (Gehrels et al., 2003 and Cowgill et al., 2003).

**TABLE CAPTIONS**

**Table 1:** U–Pb zircon data from a foliated granitoid sample, AY-09-21-2011 (4), taken from the gneiss complex. The interpreted $^{207}\text{Pb}/^{206}\text{Pb}$ results give an age of $911 \pm 12$ Ma (MSWD=1.3).
Figure 1
Figure 2
Figure 3
Figure 4
Figure 5
conglomerate with clay and lime matrix; quartz sandstone interbedded with mudstone; muddy limestone; gypsum
course sandstone; pebble sandstone and conglomerate
arkosic sandstone interbedded with siltstone and carbon shale; some coal; basal conglomerate
interbedded arkosic sandstone and calcareous siltstone and carbon shale; basal conglomerate
cross-bedded arkosic sandstone and quartz sandstone interbedded with siltstone; basal conglomerate bearing sandstone
arkosic and quartz sandstone; siltstone and calcareous siltstone; sandy limestone
arkosic and quartz sandstone; siltstone and calcareous siltstone; sandy limestone; calcareous sandstone
quartz sandstone and pebble sandstone interbedded with siltstone; basal conglomerate
see text for unit descriptions

Figure 6
Figure 7
Figure 8
Figure 10
Figure 11
Figure 12

Mean = 911±12
95% confidence
Figure 13
Deformed section length: 21.9 km
Restored section length: ~> 40 km
Shortening: 18.1 km
Shortening strain: 18.1/40 km = 45 %

Figure 14
Kinematic Reconstruction

Stage 1

- Initiation of Fault 1c and fault-propagation fold
- Hanging wall-foot wall ramp geometry
- Overturning of Fault 1c and related folding in hanging wall
- Footwall syncline and related parasitic folds

Stage 2

- Fault-propagation syncline-anticline pair
- Initiation of Fault 1a

Figure 15
Fault 1a propagation and offset on fold limb

Stage 3

Deformed section length: 21.9 km
Restored section length: ~> 40 km
Shortening: 18.1 km

Stage 4

Figure 15
Altyn Tagh fault
Qilian Shan-Nan Shan thrust belt
Estimated Shortening ~ 290 - 365 km

Altn Tagh offset ~ 375 - 475 km (Gehrels et al., 2003; Cowgill et al., 2003)

Estimated Shortening ~ 290 - 365 km

Figure 17
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Table 1
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