STRUCTURAL AND GEOMORPHIC EVOLUTION OF THE GONGHE BASIN COMPLEX, NORTHEASTERN TIBET: IMPLICATIONS FOR THE TIMING OF PLATEAU GROWTH

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by

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Abstract

Reconstructing the kinematic history of the outward expansion of the Tibetan plateau is central to ongoing debates over the geodynamics of continental plateau growth and the manner in which the growth of high topography shapes the earth’s climate. For the broad northeastern margin of the plateau, disagreement exists over the timing, magnitude, rate, and style of contractional deformation in the upper crust. I present four field based studies from the Gonghe basin complex in the regions interior which bear on these issues. First, I document regionally extensive contractional deformation across a broad swath of interior northeastern Tibet (the Anyemaqen Shan and west Qinling Shan) during the Cretaceous, thereby providing evidence for pre-Cenozoic crustal thickening of the region. Second, I show that although northeastern Tibet may have experienced contractional tectonism during the early Tertiary, this episode appears to be confined to regions near the plateau edge (e.g. the west Qinling fault and the western Qaidam basin) and is not apparent in the intervening region. Third, I add new evidence from interior northeastern Tibet (the Gonghe basin region) to a growing body of work that points to a rapid change in structural style and depositional patterns across the entire plateau margin during the late Miocene, from slow sedimentation in broad basins, to rapid sedimentation in narrow, structurally bounded basins. Fourth, I show that upper crustal shortening since the late Miocene has been small, on the order of 4%, along a 350 km profile in interior northeastern Tibet. Fifth, I show that fault networks in the region sole into decollements at deep levels (10s of km) in the crust, analogous to other intracontinental mountain ranges such as the Laramide ranges in the western United States, or the Sierra Pampeanas of Argentina. Finally, I reconstruct the time-transgressive incision of the Yellow River during the Quaternary. Canyon incision lagged the onset of mountain building in the Miocene by nearly ~10 Ma, and spatiotemporal patterns of incision suggest that it resulted drainage basin integration around northeastern Tibet.
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CHAPTER 1
INTRODUCTION

The evolution of high topography within the Indo-Asian collision zone lies at the center of ongoing debates over the nature of interactions between climate, erosion and tectonics. In particular, the Tibetan plateau is thought to influence global climate during the Cenozoic by acting as a major sink for carbon dioxide and by perturbing patterns of atmospheric circulation in the northern hemisphere (e.g. Raymo and Ruddiman, 1992; Zachos and Kump, 2005; Molnar, 2010 and references therein). Thus, in addition to serving as the prototype for collisional orogens, Tibet appears to be a critical site for understanding some of the controls on earth’s climate over geologic time.

Several basic aspects of the geologic history of the orogen are well established. First, the Tibetan plateau owes its existence to ongoing convergence between the Indian subcontinent and southern Eurasia, which initiated when a south-facing subduction zone between the two continents closed at ~55-45 Ma (Garzanti and van Haver, 1988; Rowley, 1996; Zhu et al., 2005; Jain et al., 2009; Najman et al., 2010). Rather than subducting below southern Tibet, buoyant, rigid India has indented into the deformable southern Eurasian lithosphere along a NNE vector (e.g. Tapponnier and Molnar, 1976). Plate reconstructions indicate 2300-3500 km of convergence between India and Eurasia since the time of collision (Molnar and Stock, 2009; Dupont-Nivet et al., 2010). About 1800 ± 700 km of which has been accommodated by contractional deformation of northern Indian crust (Dupont-Nivet et al., 2010; see also DeCelles et al., 2002 and references therein), and anywhere from 500 – 2500 km of which has been accommodated by diffuse deformation distributed over 1000s of km of southern Eurasia (Chen et al., 1993; Dupont-Nivet et al., 2010; Liebke et al., 2010). Second, most researchers accept that
prior to the onset of collisional orogenesis, thick crust and corresponding high topography extended across southern Tibet (England and Searle, 1986; Murphy et al., 1997; Kapp et al., 2003). Although the height and areal extent of this high topography is not clear, most workers envisage that southern and central Tibet were at elevations similar to the present during the Cretaceous and the early Tertiary (e.g. Kapp et al., 2005; Rowley and Currie, 2006; DeCelles, 2007), and that the marginal regions lay several km lower than today (e.g. Kapp et al., 2005; Cyr et al., 2005) (Figure 1.1). The kinematics and dynamics of the expansion of high topography to the marginal regions of the present day plateau remain controversial. Whereas some workers advocate for progressive outward expansion (England and Houseman, 1986; Royden et al., 1997; Tapponnier et al., 2001), others hold that the peripheries of Tibet experienced a rapid change in surface elevation (Molnar et al., 1993). Although outward expansion is commonly thought to occur in the Miocene (Tapponnier et al., 2001; Molnar, 2005; Royden et al., 2008), several recent studies challenge this view by suggesting that outward growth occurred, or initiated, during the Eocene (e.g. Dupont-Nivet et al., 2008a; Clark et al., 2010). Defining this kinematic history is critical for understanding the dynamics of plateau growth in Tibet.

Despite the fact that the topographic history of Tibet remains contentious, it has been invoked to explain many regional and global climate patterns during the Cenozoic. First, high topography in central Eurasia has been linked to global cooling during the Cenozoic Era because enhanced weathering of silicate minerals and enhanced burial of organic carbon during plateau growth are thought to sequester carbon from the atmosphere (e.g. Raymo et al., 1988; Zachos and Kump, 2005). Second, the Tibetan plateau is thought be sufficiently high and broad to affect the development and intensity of the east and south Asian monsoons. Although the commonly held view that high topography of Tibet drives south Asian monsoon circulation by acting as a
heat source in the atmosphere (e.g. Prell and Kutzbach, 1992) has been challenged by emerging empirical (Yanai and Wu, 2006) and theoretical work (Chakraborty et al., 2006; Boos and Kuang, 2010), it seems likely that northward expansion of Tibet during the Cenozoic would have diverted the subtropical jet stream that flows west to east across Eurasia, dictating patterns of east Asian monsoon rainfall and, along with topography in the western US, established hemispherical waves in the global circulation that influence the modern climate of both the eastern US and Europe (see Molnar et al., 2010 and references therein). Indeed, many paleoclimate proxies from eastern Asia suggest pronounced environmental change during the time that Tibet is commonly thought to have grown outward, the mid-late Miocene (Ding et al., 1999; Sun et al., 1998; Qiang et al., 2001; Guo et al., 2002, Ma et al., 1998). A detailed understanding of these enticing links between growing topography and climate awaits better constraints on spatiotemporal patterns of plateau growth around Tibet.

The northeastern margin of the Tibetan plateau represents a key region to assess the timing of growth of high topography. Although there is clear evidence for Miocene contractional deformation and attendant basin formation in the region which has been linked to northward plateau expansion (e.g. Molnar, 2005, 2010), much of the evidence that underpins suggestions of significant Eocene contractional deformation >3000 km north of the Indo-Eurasian collision zone derives from this region (e.g. Dupont-Nivet et al., 2008a,b; Clark et al., 2010; Dayem et al., 2010 and references therein). Conflicting interpretations over the kinematic history of the region underscore the need to accurately reconstruct its past.

Northeastern Tibet occupies the region north of the Kunlun fault and east of the Qaidam Basin, and it is blanketed by broad sedimentary basins, which are in turn transected by elongate, fault bounded mountains ranges (Figure 2). These basins and ranges extend over spatial scales
approaching 1000 km, such that they are uniquely well suited for reconstructing spatiotemporal patterns of plateau growth across a broad region. Many of the basins are deeply dissected by the Yellow River, which originates on the plateau and flows across the entire region. Deep exposures of basin fill reveal that Cretaceous and Cenozoic strata are mostly undeformed except near the basin margins where strata are steeply tilted and faulted (e.g. Fang et al., 2003; 2005; Dai et al., 2006).

Northeastern Tibet is situated between two disparate physiographic and geologic terranes, which appear to relate to variations in upper crustal structure and may represent differing mechanisms of plateau growth (e.g. Tapponnier et al., 1990; Burchfiel et al., 1995; Meyer et al., 1998; Clark and Royden, 2000). To the south is the broad, flat interior of the Tibetan plateau, where the lack of short wavelength topography is interpreted to reflect weak lower crust (e.g. Fielding et al., 1994) (Figure 1.1). Eastward flow of material in the lower crust beneath this region is thought to be responsible for building the 3-km topographic escarpment along the eastern and southeastern edges of Tibet (e.g. Burchfiel et al., 1995; Royden et al., 1997; Clark and Royden, 2000; Kirby et al., 2002; Clark et al., 2005). To the east of northeastern Tibet is the Qilian Shan, a mountainous region which extends over an area on the order of ~10^5 km^2 (Figure 1.1). In contrast to the low wavelength, low amplitude topography of interior Tibet, the Qilian Shan contains closely spaced, narrow, fault bounded mountain ranges and intervening sedimentary basins. Regional budgets of upper crustal shortening suggest that high topography is compensated by thickened upper crust (e.g. Meyer et al., 1998), a markedly different mode of plateau growth than the terrain to the south. Due to its transitional position between interior Tibet and the Qilian Shan, northeastern Tibet provides an opportunity to evaluate contrasting modes of plateau growth.
Regional patterns of crustal thickness and topography also reflect the transitional nature of the broad, northeastern margin of Tibet. Along a profile that extends ~SW-NE across the region, crustal thickness and topography steadily decrease from ~60 km and ~4.5 km near the plateau interior, to ~47 km and ~1.5 km in the Linxia-Lanzhou region at the edge of the plateau (Liu et al., 2006; Zhang et al., 2010) (Figure 1.2b). Farther west, in the vicinity of the Gonghe basin complex, the transitions in thickness and topography are less pronounced across northeastern Tibet, but they are abrupt at the plateau edge which is marked by the northern Qilian Shan. Crustal thickness is ~55 - 60 km beneath the Anyemaqen Shan near the plateau interior and ~50-60 km beneath the Gonghe basin region to the north (Vergne et al., 2002; Zheng et al., 2010) (Figure 1.2c). Thick crust extends to the plateau edge at the North Qilian Shan-North China boundary, and then abruptly thins to ~ 42 km in the Hexi Corridor north of the Qilian Shan (Meyer et al., 1998). Similarly, mean topography is fairly uniform across the Anyemaqen, Gonghe, and the Qilian Shan, at ~4.0-4.5 km, and it decreases abruptly at the northern plateau margin to ~1.5 km.

Several important research questions about the geologic evolution of northeastern Tibet remain unanswered. For example, stratigraphic archives from basins near the periphery of northeastern Tibet indicate slow subsidence (~10 m Ma⁻¹) across much of the region north of the West Qinling fault from the Cretaceous until at least the late Oligocene (Horton et al., 2004; Dai et al., 2006). Unfortunately, the margins of Cretaceous basins are typically poorly exposed, and their geodynamic significance is poorly known (e.g. Horton et al., 2004; Ritts and Biffi, 2002). Thus, a key issue concerning the evolution of northeastern Tibet is determining the tectonic setting of these sedimentary basins.
The early Tertiary tectonic evolution of the region is also rather enigmatic. Although there is significant evidence for rotation, cooling of ranges, and basin subsidence, evidence for this deformation derives from Linxia and Xining basins and the western Qilian Shan-Qaidam basin, leaving the early Tertiary history of vast portions of the intervening interior of northeastern Tibet uncertain. In particular, vertical axis rotation of the Xining basin by ~41 Ma, (Dupont-Nivet et al., 2004; 2008) and thrusting along a segment of the west Qinling fault at ~45-50 Ma (Clark et al., 2010; Duvall et al., in review) provide localized evidence for deformation only a few million years after the beginning of the Himalayan-Tibetan orogeny (e.g. Rowley, 1996) (Figure 1.2). Similar early Tertiary contractional tectonism has been inferred from basin subsidence and sedimentation patterns across western Qaidam (Kent-Corson et al., 2009; Bovet and Ritts, 2009; Yin et al., 2002). Although workers have linked this deformation to plate boundary stresses generated at the Indo-Eurasian collision zone (e.g. Clark et al., 2010), robust dynamic interpretations await a better understanding of the early Tertiary kinematics of the region along the structural trend that extends from the west Qinling Shan to the southern Qilian Shan (Figure 1.2)

Another critical issue in northeastern Tibet is assessing the evolving patterns of deformation around the region subsequent to the Eocene and throughout the remainder of the Cenozoic. Slow sediment accumulation was punctuated in the Linxia basin region during the mid-Oligocene by rapid, flexural subsidence, and the Laji Shan emerged from a once broad region of sediment accumulation in the early Miocene (Fang et al., 2003; Lease et al., in review). Both of these events are confined to a swath of southwestern Linxia, and the extent to which similar mid-Tertiary deformation occurred regionally is unclear. By the late Miocene, however, the depositional patterns and tectonic style across northeastern Tibet clearly differed from the
Cretaceous and early Cenozoic. Several basins, including Guide, the Hexi corridor, and northeastern Qaidam, experienced rapid, flexurally-driven sedimentation at this time (Fang et al., 2005; 2007; Bovet and Ritts, 2009). In addition, a variety of stratigraphic proxies for sediment provenance reveal the rapid unroofing of source terranes to basins in the interior of the region from ~14 – 8 Ma (Dettman et al., 2003; Fang et al., 2003, 2005; Garzione et al., 2005; Lease et al., 2007), and thermal histories from ranges along the plateau edge, the Liupan Shan and the northern Qilian Shan, record a coeval acceleration in exhumation (Zheng et al., 2006; 2010; Kirby et al., 2002; Clark et al., 2005). To date, it is unclear whether the late Miocene deformation marks a distinct change in structural style, or whether it is part of a more continuous history of deformation that is recorded in the Linxia region.

Although a growing number of studies bear on the timing of deformation around northeastern Tibet, very little research bears on the magnitude of upper crustal shortening or the architecture of major fault networks. Several attempts have been made at budgeting upper crustal shortening in the neighboring Qaidam-Qilian Shan region to the west (e.g. Meyer et al., 1998; Yin et al., 2008). These studies suggest that fault networks which extend deep into the crust accommodated a large magnitude of shortening across the Qilian Shan and western Qaidam during the late Cenozoic (Tapponnier et al., 1990; Meyer et al., 1998). Although serial cross sections across Qaidam imply an eastward decrease in the amount of contractional deformation across the region, the degree to which this pattern extends into northeastern Tibet is unclear.

A final, outstanding issue is the pronounced, regional transition from Cretaceous-Tertiary basin filling to Quaternary excavation by the Yellow River. The youngest strata in a basin near the plateau edge (Linxia basin), suggest that lacustrine deposition persisted until at least ~1.7 – 1.8 Ma. Along the margin of the plateau, the onset of basin excavation is tightly bracketed by
fluvial terraces related to the Yellow River that are dated to ca. 1.7 Ma (e.g. Li et al., 1997). Based on the age of the youngest preserved lacustrine sediments farther upstream (Guide basin) (~1.7 Ma), many have inferred that incision began simultaneously along this reach of the Yellow River and attribute this to widespread surface uplift of the northeastern Tibetan Plateau (Li et al., 1991, 1997; Fang et al., 2005). However, the absence of direct age control on fluvial terraces in the upstream basins allows that this incision may have been time transgressive, such that incision need not reflect widespread mountain building across the region.

This dissertation addresses several first-order questions regarding the late Mesozoic and Cenozoic evolution of northeastern Tibet. What is the tectonic setting of Cretaceous basins that are exposed around the region? To what extent did early Tertiary deformation extend interior to the Linxia-Lanzhou region? Did mountain building occur steadily and continuously throughout the Cenozoic, or does a distinct acceleration in contractional tectonism occur in the Miocene? How much upper crustal shortening has occurred across the region, and what is the architecture of major fault networks? How and why did the Yellow River incise a ~500-700 m deep canyon across the region? In order to address these, I focus on the interior region of northeastern Tibet, in the vicinity of the Gonghe basin complex and the Anyemaqen Shan (Figure 1.2). The Gonghe basin extends across a region that is ~200 x 200 km, and lies at the transition between the plateau interior and the easternmost Qilian Shan (Figure 1.1). South vergent networks of imbricate thrust faults, including the Qinghai Nan Shan (QNS) and Gonghe Nan Shan (GNS), override the northern and southern margins of the basin, respectively. This basin is situated along the structural grain which extends from the southwestern Qilian Shan to the west Qinling Shan, two of the primary sites with evidence for early Tertiary contractional deformation in the region (e.g. Clark et al., 2010; Yin et al., 2008). Although most of the vast interior of the basin is poorly
exposed, the Yellow River carves a 500 – 700 m deep canyon across a N-S striking corridor in the center of the basin. Within this canyon, broad expanses of undeformed Cenozoic basin fill are exposed. Additional exposures of the basin fill occur at the basin margins, where it is folded over the bounding mountain ranges. The Gonghe basin is bordered to the south by a broad region of mountainous topography associated with the Kunlun fault, called the Anyemaqen Shan. Little is known about the history of this range (c.f. Kirby et al., 2007; Harkins et al., 2010), but narrow, fault bounded basins which are scattered across the region contain important archives of mountain building (Figure 1.2).

To address the research needs outlined above, I present four field based studies, which focus on the evolution of the Gonghe basin complex and the Anyemaqen Shan, but bear on the evolution of northeastern Tibet as a whole. The first study aims to reconstruct the evolution of the network of basins that is distributed across the Anyemaqen Shan, and it is presented as Chapter 2. The study is currently accepted for publication at Basin Research, pending some revisions, and Eric Kirby (The Pennsylvania State University), Zheng Dewen (China Earthquake Administration), and Liu Jianhui (Chinese Academy of Geological Sciences) are coauthors. By exploiting regional lithostratigraphic observations, cross-cutting volcanic rocks, and regional biostratigraphy, I demonstrate that the basins are Cretaceous in age. Although previous inquiries into the significance of these basins have suggested that some formed in a transtensional setting (e.g. Horton et al., 2004), clear associations between basin bounding faults and sedimentary strata show that they developed during an episode of contractional deformation. I interpret this deformation to correspond to an episode of right lateral shear on a proto-Kunlun and proto-Qinling fault. Importantly, this study shows that a potentially significant amount of crustal
thickening and topographic growth may have occurred in interior northeastern Tibet prior to the Cenozoic, although budgets of this deformation remain elusive.

The second study concerns the timing of contractional deformation around the Gonghe region and is presented as Chapter 3. This study is in preparation for peer review and will be submitted to Tectonics. It is coauthored with Eric Kirby, and Huiping Zhang (China Earthquake Administration). We combine a variety of geochronologic techniques, including magnetic reversal stratigraphy, palynology, cosmogenic burial dating, and lithostratigraphic correlation in order to precisely determine the age of Cenozoic strata in the southern portion of the Gonghe basin complex. Results of this work indicate the initiation of sediment accumulation in the Gonghe region by around ~20 Ma. By integrating this chronologic information with new regional geologic mapping that focuses on the relationship between the GNS fault networks with basin fill, we show that the GNS fault network initiated during the late Miocene, from ~10-7 Ma. In light of other studies from around the region, our work suggests that whereas slow (~10m/Ma) sediment accumulation near the exterior of northeastern Tibet occurred throughout the late Cretaceous and early Tertiary, sediment accumulation did not begin in the region’s interior until Miocene time, from ~20 Ma, and it accumulated rapidly (~100m/Ma). Moreover, this work adds to a growing body of evidence that points to a rapid change in structural style across the breadth of northeastern Tibet in the mid-late Miocene (ca. 10 Ma, see Molnar, 2010 for a review), from sediment accumulation in a broad, connected foreland basin, to rapid sediment accumulation in structurally isolated basins, as narrow mountain ranges emerged from the early Tertiary basins (e.g. Lease et al., 2007).

The third study, Chapter 4, concerns the magnitude, style, and rates of upper crustal deformation in the Gonghe region. The study is currently in preparation for peer review, and it
will be submitted to Geological Society of America Bulletin for consideration. In it, I present a synthesis of the stratigraphy and timing of deformation in the Gonghe basin complex and an analysis of the architecture of the fault networks that deform the margins of the basin. Various stratigraphic and geologic data suggest that like the southern margin of central Gonghe, the other margins of Gonghe basin experienced contractional deformation beginning in the Late Miocene. Detailed structural reconstructions suggest ~1-2 km of upper crustal shortening across the QNS, and ~5-7 km of upper crustal shortening across the GNS, both since the Late Miocene. Moreover, thermal histories from these ranges corroborate the geologic evidence pointing to emergence of these ranges since the Miocene, and place an important bound on the development of structural relief in these ranges (no more than 1 or 2 km). Given that geologic shortening rates for these ranges can be confidently calculated, I compare these rates to late Quaternary shortening rates for the NW QNS and show the rates of contractional deformation appear to have been steady since the pronounced change in structural style in the mid-late Miocene. Finally, by integrating detailed geologic observations from the QNS and GNS with reconnaissance observations from around the region, I construct a budget of late Cenozoic upper crustal shortening along a profile that extends ~350 km, from the Anyemaqen Shan to the northern edge of the Qinghai Lake basin. I find only about 4% shortening along this cross section, such that late Cenozoic thickening of the upper crust does not appear to be a likely mechanism for constructing high topography in interior northeastern Tibet, given isostatic considerations.

The final study concerns the pronounced transition from basin filling to basin excavation that occurred along the Yellow River during the Quaternary. This study was published in Nature Geoscience in 2010, and was coauthored with Eric Kirby (The Pennsylvania State University), Nathaniel Harkins (The Pennsylvania State University and ExxonMobil Upstream Research),
Huiping Zhang (Institute of Geology, China Earthquake Administration), Liu Jianhui (Institute of Geology, Chinese Academy of Geological Sciences), and Xuhua Shi (The Pennsylvania State University). By employing a variety of geochronologic techniques, including magnetic reversal stratigraphy, luminescence dating, radiocarbon dating, and cosmogenic burial dating, we place constraints on both the depositional surface that marks the cessation of sediment accumulation in Gonghe, and the strath terraces which record the history of canyon incision by the Yellow River. This work brackets this transition to ~0.5 Ma, and in a regional context, it reveals a time-transgressive pattern of canyon incision which initiated at the plateau periphery at ~1.8 Ma and swept headward throughout the Quaternary. The incision lags the onset of upper crustal shortening in the region by nearly 10 Myr, and as such, appears to reflect drainage capture by the Yellow River, rather than surface uplift (Li et al., 1991; 1998; Fang et al., 2005).

In sum, the four studies described above represent a significant advancement in our understanding of the geologic and topographic history of the northeastern Tibetan plateau. I summarize the key findings below.

1. Significant contractional deformation occurred across a broad swath of the Anyemaqen Shan and west Qinling Shan during the Cretaceous. This finding is direct evidence for the possibility that significant crustal thickening and attendant surface uplift of northeastern Tibet predates the Cenozoic Era. Importantly, pre-Cenozoic crustal thickening and plateau growth need not be mutually exclusive of other mechanisms during the Cenozoic.

2. Although northeastern Tibet may have experienced contractional tectonism during the early Tertiary (e.g., Clark et al., 2010), this appears to be confined to the vicinity of the West Qinling fault and the western Qaidam basin (e.g. Yin et al., 2008) and is not apparent in the intervening region.
3. A growing body of evidence points to a rapid change in structural style and depositional patterns across northeastern Tibet during the late Miocene (e.g., Molnar, 2010), from slow sediment accumulation in broad sedimentary basins, to rapid sediment accumulation in narrower, fault bounded basins. Moreover, a progressive outward pattern of deformation in northeastern Tibet is not apparent during this time. If the upper crust was coupled to the lower lithosphere during plateau growth (e.g. Medvedev and Beaumont, 2006), then progressive outward thickening of the crust at any depth appears to be incompatible with geologic observations from northeastern Tibet (e.g. England and Houseman, 1986; Royden et al., 1997). Kinematic observations from around the region seem to be compatible with geodynamic models that predict a rapid (~2-5 Myr), outward expansion of contractional deformation across a region that extends for hundreds of km (e.g. Molnar et al., 1993).

4. Upper crustal shortening since the late Miocene has been relatively small, on the order of 4% in interior northeastern Tibet. This suggests that late Cenozoic thickening of the upper crust is not a viable explanation for the construction of high topography in northeastern Tibet. Additionally, the fault networks in the interior of the region sole into decollements at deep levels (10s of km) in the crust, analogous to other intracontinental mountain ranges such as the Laramide ranges in the western United States, or the Sierra Pampeanas of Argentina.

5. Time-transgressive incision of the Yellow River in the Quaternary lagged the onset of mountain building in the Miocene by nearly ~10 Ma. This difference in timing suggests that canyon cutting is not a good proxy for surface uplift in northeastern Tibet, as it is for other marginal regions of Tibet to the south (e.g. Kirby et al., 2002; Clark et al., 2005). Rather, the spatiotemporal pattern of Yellow River incision suggests that canyon cutting was the result of
drainage basin integration around northeastern Tibet, which began at the beginning of the Quaternary.
Figure 1.1. Topography, active faults, and major rivers of the Tibetan plateau. Faults from Molnar and Tapponnier, 1978, Tapponnier and Molnar, 1979. Background is GTOPO30 digital topography with 30 arcsecond resolution. Northeastern Tibet is outlined in black. ATF = Altyn Tagh fault, HF = Haiyuan fault, HFT = Himalayan frontal thrust, KF = Kunlun fault, KRF = Karakorum fault, RRF = Red River fault, SF = Saigang fault, XF = Xianshuihe fault
Figure 1.2. a) Quaternary faults and Cretaceous and Cenozoic basins in northeastern Tibet. Faults compiled from Tapponnier and Molnar, 1977 and Molnar and Tapponnier, 1978. Background is 90-m Shuttle Radar Topography Mission data draped over a hillshade image. b and c) Maximum, minimum, and mean swath topography, derived from GTOPO-30 data, which has a nominal resolution of 1 km. Moho depths are also shown. For b, moho depths are from Liu et al., 2006. For c, Moho depths from the Anyemaqen and Gonghe are from Vergne et al., 2002, and depths from the Qilian Shan are from Meyer et al., 1998.
CHAPTER 2

TECTORIC SETTING OF CRETACEOUS BASINS IN NE TIBET: INSIGHTS FROM THE JUNGONG BASIN

2.1 Abstract

Quantifying the Cenozoic growth of high topography in the Indo-Asian collision zone remains challenging, due in part to significant shortening that occurred within the Eurasian tectonic collage prior to collision. A growing body of evidence suggests that regions far removed from the suture zone, in present-day NE Tibet, experienced deformation immediately prior to and during the early phases of Himalayan orogenesis. Widespread deposits of Cretaceous sediment attest to significant basin formation; however, the tectonic setting of these basins remains enigmatic. We present a study of a regionally extensive network of sedimentary basins that are spatially associated with a system of SE-vergent thrust faults. We focus on a particularly well-exposed basin, herein referred to as Jungong basin, located ~20 km north of the Kunlun fault in the Anyemaqen Shan of NE Tibet. The basin is filled by ~900 m of fluvial and alluvial fan sediments that become finer-grained away from the basin-bounding fault. These facies associations, progressive, up-section shallowing of beds adjacent to the surface trace of the fault, and the presence of a progressive unconformity all suggest that sediment accumulated in the basin during fault growth. Regional constraints on the timing of sediment deposition are provided by lithostratigraphic correlation to basins with 1) fossil assemblages from the Early Cretaceous, and 2) volcanic rocks dated with new and existing K-Ar geochronology that floor and cross-cut sedimentary fill. We argue that during the Cretaceous, interior NE Tibet experienced NW-SE directed contractional deformation similar to that documented throughout the Qinling-Dabie orogen to the east. The Songpan-Ganzi terrane apparently marked the
southern limit of this deformation, such that it may have been a rigid block in the Tibetan
lithosphere during the Cretaceous time that separated regions experiencing deformation north of
the convergent Tethyan margin from regions deforming inboard of the east Asian margin.

2.2 Introduction

Contrasts in lithospheric strength in the complex tectonic collage of central China may
exert a first-order influence on the topography, deformational style, and the distribution of strain
across the region (Clark and Royden, 2000; Shen et al., 2001; Cook and Royden, 2008; Dayem et
al., 2009). For example, spatial variations in elastic thickness (Jordan and Watts, 2005), modern
thermal structure (Wang, 2001), and lower crustal P-wave velocity (Li et al., 2006) across the
region indicate that northeastern Tibet may be relatively weak compared to the adjacent, strong
Sichuan and Tarim basin lithosphere. Tectonic quiescence in the Sichuan and Tarim basins since
Mesozoic time likewise indicates that they have been persistently rigid blocks in the central
Chinese lithosphere over time scales of $10^8$ Ma (England and Houseman, 1985). The role of
more subtle strength contrasts remains in controlling the tectonic evolution of the region,
however, is poorly understood.

Pioneering research on the Cretaceous and early Tertiary tectonics of central China
provides a hint that higher order strength contrasts may control the distribution of strain prior to
the onset of the Himalayan orogeny. The Qinling-Dabie orogen contains a rich record of
Cretaceous to early Paleogene deformation (e.g. Ratschbacher et al., 2000, 2003; Enkelmann et
al., 2006). To the west, mineral cooling ages and the absence of late Mesozoic/early Cenozoic
sedimentary basins indicate that the Songpan-Ganzi terrane and the Yangtze craton were mostly
tectonically quiescent between the late Triassic Qinling-Dabie Shan orogeny and the onset of
Himalayan orogeny-related deformation, which occurred locally in the mid-Neogene (Figure 2.1,
e.g. Burchfiel et al., 1995; Kirby et al., 2002; Reid et al., 2005). On both the south and the north side of the Songpan-Ganzi terrane, however, recent studies reveal widespread deformation during the Cretaceous and Paleogene. Importantly, this pre- and early Himalayan deformation may have played a role in the topographic evolution of the modern Tibetan plateau; such a case has been made for southern and central Tibet, south of the Songpan-Ganzi terrane (e.g. Kapp et al., 2005).

Northeastern Tibet is a particularly interesting location because it bridges the Qinling-Dabie orogen of east China with the high topography of central Tibet. Recent biostratigraphic and magnetostratigraphic analysis from an extensive network of basins distributed across the region indicates that many are Cretaceous in age, suggesting that northeastern Tibet may be the site of protracted pre-Himalayan deformation (Figure 2.1b, Ritts and Biffi, 2001; Horton et al., 2004). However, the nature of deformation along the margins of these basins, and thus their geodynamic significance, remains relatively unknown. We present new structural and lithostratigraphic observations from one particularly well exposed basin in this region, locally named the Jungong basin, and combine these with regional geologic observations to provide new insight into the tectonic history of interior northeastern Tibet.

2.3. Background

2.3.1. Paleozoic through Triassic assembly of central China

We briefly review the tectonic evolution of central China, as well as the studies that have provided important insight into the Cretaceous tectonics of the region. For a more complete discussion of the evolution of the northern Tibetan plateau and the Qinling-Dabie orogen, readers are directed to reviews by Yin and Nie (1996), Yin and Harrison (2000), Ratschbacher and others (2003), and Hacker and others (2004).
During the Paleozoic, the core of the modern Eurasian continent expanded as numerous exotic terranes were accreted to its southern margin. Beginning in the early Ordovician, the terranes near the modern Qilian Shan, which are the North China Terrane, the North Qilian Arc, the Central Qilian Terrane, the South Qilian Terrane, and the Kunlun-Qaidam Terrane, were separated by a series of subduction zones (Figure 2.1). To the east, near the modern Qinling-Dabie Shan of east and central China, North China was separated from the Erlangping Arc and the Qinling terrane, also by a series of subduction zones (Yin and Harrison, 2000; Ratschbacher et al., 2003; Xiao et al., 2009). From the Ordovician to the Devonian, the Kunlun-Qaidam Terrane and the various Qilian Shan terranes accreted to the southern margin of North China, while the Erlangping arc and the Qinling Terrane accreted to North China farther east (Ratschbacher et al., 2003; Xiao et al., 2009) (Figure 2.1). By the early Devonian, the southern margin of Eurasia was marked by the Kunlun-Qaidam terrane in the west and the Qinling terrane in the east, and Paleotethyan ocean lithosphere was subducted to the north below both (Yang et al., 1996). Northward subduction of Paleotethyan/Songpan-Ganzi oceanic lithosphere persisted through the remainder of the Paleozoic and into the late Triassic. During this episode, exhumation of the Qinling-Dabie orogen provided a source of sediment to the Songpan-Ganzi ocean basin (Weislogel et al., 2006). In the late Triassic, renewed accretion along the southern margin of the Eurasian continent was marked by the collision of the Qiangtang terrane, and the South China – North China collision along the Qinling-Dabie orogen (Yin and Harrison, 2000; Ratschbacher et al., 2003).

2.3.2. Cretaceous and early Tertiary deformation in central China

The southern and eastern margins of Eurasia were characterized by active plate boundaries throughout the Cretaceous and early Tertiary Periods. In the late Jurassic and early
Cretaceous, the Lhasa terrane was sutured to the southern margin of Asia along the Banggong-Nujiang suture (BNS) (Şengör and Natal'in, 1996; Yin and Nie, 1996). Following that episode, the southern margin of Asia was characterized by a north dipping subduction zone, until it closed along the Indus-Yalu Suture (IYS) in the early Tertiary, as the Indian craton collided with Eurasia at ca. 51 Ma (Garzanti and Van Haver, 1988; Rowley, 1996; Zhu et al., 2005).

Cretaceous deformation within the physiographic boundaries of the modern Tibetan plateau is well documented in the Lhasa and Qiangtang terranes (Murphy et al., 1997; Guynn et al., 2006; He et al., 2007; Kapp et al., 2007; Leier et al., 2007). Workers here envisage an Altiplano-like ancestral Tibetan plateau related to Cretaceous and early Paleogene tectonics of the southern margin of Asia (England and Searle, 1986; Kapp et al., 2003; DeCelles et al., 2007). During the Paleogene, the Lhasa and southern Qiangtang block experienced continued upper crustal shortening and magmatism (e.g. Kapp et al., 2005). The deformation extended NE to the northeastern portion of the Qiangtang terrane, where the Hoh Xil, Nangqian-Yushu, and Gonjo basins all contain archives of early Paleogene subsidence related to contractional deformation (Liu and Wang, 2001; Horton et al., 2002; Liu et al., 2003; Spurlin et al., 2005; Studnicki-Gizbert et al., 2008). Similar records can be found in the Lanping and Jianchuan basins in southeast Tibet (Wang and Burchfiel, 1997). Importantly, the Songpan-Ganzi basin seems to delineate the northern boundary for much of the recognized deformation in Cretaceous and Paleogene time.

An equally rich history of Cretaceous deformation occurred in northern and eastern China, and this history has been attributed to the combined effects of plate boundary activity along the southern and eastern margins of Eurasia, and also to closure of the Mongol-Okhost Sea (Ratschbacher et al., 2003). Thermal histories from across the Qinling-Dabie orogen indicate
extensive exhumation and basin formation during the mid-late Cretaceous (Ratschbacher et al., 2000, 2003; Enkelmann et al., 2006). The orientation of stresses responsible for this deformation is inferred from mesoscale geologic structures indicating NW-SE compression and NE-SW tension (Ratschbacher et al., 2003). These stresses have been linked to dextral shear on ESE-striking faults (Webb et al., 1999; Ratschbacher et al., 2003).

In contrast to these regions experiencing Cretaceous deformation inboard of either the southern or eastern Asian margins, the Songpan-Ganzi terrane exhibits little evidence of deformation at this time. Much of the Songpan-Ganzi terrane is characterized by a broad, low relief, erosion surface that now sits at high elevation in eastern and southeastern Tibet (Kirby et al., 2002; Clark et al., 2005). Cooling ages from low-temperature thermochronometers sampled from this surface (e.g., (Arne et al., 1997; Clark et al., 2005) yield Mesozoic ages, suggesting limited exhumation. Moreover, adjacent to the Sichuan Basin, thermal histories derived from multiple thermochronologic systems (Kirby et al., 2002) suggest very slow rates of exhumation throughout the Late Jurassic and Cretaceous.

In the region north of the Songpan-Ganzi terrane, in present-day northeastern Tibet, the history of Cretaceous deformation is less-well understood. In this region, the eastern Kunlun Shan and the Anyemaqen Shan of interior northern Tibet merge eastward with the Qinling-Dabie orogen. Extensive sedimentary basins of Cretaceous and Early Tertiary age are present in this region (e.g. Horton et al., 2004; Dai et al., 2006), but their tectonic setting is somewhat uncertain. In particular, whether they represent deformation of the Eurasian tectonic collage far inboard of the Tethyan margin or whether they are controlled by the same tectonic regime driving deformation in the Qinling (e.g Ratschbacher et al., 2003; Horton et al., 2004), is uncertain. The answer to this question is not only of local interest, but is also of critical importance for
elucidating the spatial distribution of deformation attributable to the Indo-Asian collision in the Cenozoic (Tapponnier et al., 2001; Yin et al., 2002; Dupont-Nivet et al., 2008a,b; Clark et al., 2010).

Cretaceous sedimentary basins are distributed across much of present-day NE Tibet, indicating that the region experienced extensive, if subtle, deformation well before the initiation of the Himalayan orogeny (GBGMR, 1989; QBGMR, 1991) (Figure 2.1). Cretaceous basins of NE Tibet may be divided into two categories based on their geographic distribution. North of the South Qilian Suture (SQS), the Xining-Minhe basin complex (Horton et al., 2004) crops out over a broad region (~20,000 km²) (Figure 2.1b). Recent study of several stratigraphic sections from the southeastern portion of this basin complex indicates slow and continuous sedimentation beginning in the late Jurassic and persisting to the present (Horton et al., 2004; Dai et al., 2006). Although various thermochronologic studies suggest that the Qilian Shan terranes, which form the floor of the Xining-Minhe basin complex, were tectonically quiescent during the Cretaceous (Jolivet et al., 2001; Sobel et al., 2001), workers infer a transtensional tectonic setting during Cretaceous time, largely based on subsidence rates and paleocurrent analysis (Vincent and Allen, 1999; Horton et al., 2004). Because the structures bounding these basins are not particularly well-exposed, the direct mechanisms responsible for creating accommodation space are uncertain.

South of the SQS, in the high elevation regions of the northeastern Tibetan Plateau, a series of basins were deposited atop the extensive erosion surface beveled across the Songpan-Ganzi terrane and west Qinling Shan of interior northeastern Tibet (Figure 2.2). Only one of these basins, the Dangchang basin, has been studied previously; this basin is Aptian-Albian in age based on a combination of palynology and magnetic reversal stratigraphy (Horton et al.,
Dangchang has been inferred to have formed in a transtensional setting because of its spatial association with the Xining-Minhe basin complex (Horton et al., 2004); structures along the basin margin are again poorly exposed. West of this basin, however, numerous small basins are distributed over a broad region of the west Qinling Shan and Anyemaqen. We informally refer to the collection of these basins as the Jungong-Dangchang basin complex (Figure 2.2). Most of the basins within the Jungong-Dangchang complex are poorly exposed on the high elevation, grass-covered plateaus of northeast Tibet. Little is known about their tectonic significance. A survey of the rich fossil assemblages from several of the basins (presented herein), however, provides moderately good age control for several of these basins.

We focus on the Jungong basin, a narrow, NE-trending elongate basin that is well-exposed in the deep canyons of the Yellow River through the Anyemaqen Shan (Figure 2.1, 2.2). Exposure of at least ~900 m of basin fill provides a superb view of both the basin architecture and of the basin-bounding structures along the northwestern basin margin. Moreover, the basin fill is readily correlated to other basins at the NE and SW ends of the Jungong basin proper (Figure 2.2), and they provide reasonable chronologic control on the timing of sediment accumulation. Together with structural and lithostratigraphic observations of the basin fill, the chronology from surrounding basins provides new insight into the tectonic evolution of the region.

2.4. Structural geology of the Jungong basin

Detailed structural mapping was conducted along several well exposed transects orthogonal to the basin margins, and in other key locations near the basin margins (Figure 2.3). Because of the markedly different spectral character of the brick red basin strata and the grey Triassic basement rocks, we have used ASTER Level 1a imagery to guide our interpretations of
the locations of key geologic contacts between structural transects (Figure 2.3a). In regions beyond Jungong basin, information about the geometry and architecture of the other Cretaceous in the region derives from regional 1:200,000 Chinese geologic maps (QBGMR, 1991).

The northwestern edge of the basin is marked by a SE-vergent thrust fault that places Triassic rocks on top Cretaceous basin fill (Figure 2.3, 2.4). In the western part of the basin, the fault splays into two distinct strands (Figure 2.3). Both the basin bounding fault and the splay fault in the footwall break the surface. The bounding fault is continuous for ~50 km along the basin margin and it dips 30° NNW along much of this distance. In the western part of the basin, the bounding fault dip steepens to ~60-65° NNW. The intrabasin fault in the western part of the basin daylights and abuts a lense of conglomerate beds immediately to the south. On the northern border of the town of Jungong, it has a dip of ~30° NNW (Figure 2.3). A few kilometers along strike to the west, the fault steepens to 81° NNW (Figure 2.3). Between the two faults, the basin strata are folded into a doubly plunging anticline. Most of the rock exposed in this fold is Cretaceous conglomerate except for a window of Triassic rocks at the apex of the fold. Near the western edge of the basin, the footwall of the southern fault is characterized by a tight anticline-syncline pair (Figure 2.3). The basal contact here is steep, dipping between ~30-50° NNW (Figure 2.4d). A few kilometers to the east, the folds open abruptly into a broad, open, asymmetric syncline in the footwall of the basin bounding thrust fault (Figure 3). Most of the muddy strata in the center of the basin dip gently (~10 - 25°) to the NNW, whereas beds along the NW margin dip steeply (~20-90°) SSE. In the center of the basin, the SSE dipping beds that define the NW limb of the syncline are overthrust, and therefore not exposed.

2.5. Character and significance of Jungong basin fill
Below, we describe the deposits found in Jungong basin, interpret their environment of deposition, and we relate the stratigraphic units in Jungong to those found across the region.

2.5.1 Description and interpretation of lithofacies

Lithofacies 1 - Basal granule and cobble orthoconglomerate - Lithofacies 1 blankets the basin floor; it is found at all outcrops of the lowest basin fill. At our measured section, it is ~100 m thick. Lithofacies 1 is brick red to purple, granule to cobble orthoconglomerate with a matrix of mud-size particles (Figure 2.5, 2.6). The clasts are angular to subangular, and they comprise sandstone, shale, phyllite, and quartz. Although the unit appears fairly massive, variations in color suggest that the beds are ~1 m thick and tabular/sheet-like, with sharp unchannelized bases. The beds are ungraded. Small lenses of silt and/or sand up to ~30 cm in thickness are present are interbedded within the conglomerate.

We interpret lithofacies 1 to be a debris flow-dominated, alluvial fan deposit. The muddy matrix, large clast size, and lack of grading within beds suggest that the clasts were supported during transport by a high viscosity flow (Nemec and Steel, 1984). No imbrication was observed within the conglomerate beds suggesting that the clasts are not traction deposits, however, silt and sand lenses may be traction deposits formed during waning flows. The lack of channelized basal contacts and the sheet-like geometry of the beds suggest that the unit was deposited by unconfined, rather than channelized, flows. Clast composition is similar to the surrounding bedrock, suggesting a local origin for these deposits.

Lithofacies 2 - Mudstone with fine-medium sand lenses - Lithofacies 2 is widespread throughout Jungong basin. It is found in the distal portions of the basin, away from the basin bounding faults. In our measured section, it comprises the middle ~700 m of basin fill. Mudstones are massive, brick red/maroon and unindurated. The mudstones are abundant in clay
minerals and reduction spots. Branches and roots are present within the individual beds; in some places, beds are characterized by prismatic structure.

Sandstone lenses are composed of reddish-tan muddy medium to fine, compositionally immature sandstone (Figure 2.5, 2.6). Dense intervals of lenticular sandstone beds are separated by mudstone intervals that are ~a few tens of meters in thickness. They are 10-40 cm thick and can be traced laterally over hundreds of meters. The lenses exhibit parallel planar laminations and trough cross stratification (Figure 2.5). Channel scours along the base of beds have a few cm to a few tens of cm of relief. Bi-directional paleocurrent measurements made on channel scours in unit 2 in the central part of the basin indicate flow along azimuths of 155/355 and 198/18. Some sandstone beds exhibit burrowing and bioturbation. Burrowing/bioturbation is more common near the top of the unit where the unit is generally coarser.

We interpret lithofacies 2 to represent fluvial floodplain deposits. The muds and the bioturbated sands represent floodplain and crevasse splay deposits, and the other sands represent channel deposits. The massive texture of the mudstone indicates that grains settled out of suspension in still water. The prismatic texture, abundance of clay minerals, massive appearance, and presence of roots and branches indicate paleosol formation occurred in these beds during periods of subaerial exposure between large floods. The brick red color of the mud beds may have developed as a result of oxidation during subaerial exposure (Figure 2.6). The reduction spots suggest that floodplains may have been inundated with stagnant waters, with little fresh water input. We interpret the broad sand lenses to be crevasse splay deposits based on their lenticular geometry and the presence of burrows and bioturbation. We interpret the trace fossils to be beetle trace fossils, which form as beetles rework crevasse splay sediment between flood events (Hasiotis, 2002).
The trough cross stratification and planar lamination observed in the lenses formed in a unidirectional current, in lower and upper flow regime transport conditions, respectively (Boggs, 2001). The lenticular geometry, as well as the presence of channel scours, indicate transport and deposition of the beds by a channelized flow.

Lithofacies 3-Upper pebble to cobble orthoconglomerate—Lithofacies 3 is spatially associated with the major thrust faults in Jungong basin. It comprises the upper ~100 of the measured stratigraphic section.

Lithofacies 3 is buff-colored, pebble to cobble ortho- and paraconglomerate with a muddy and sandy mud matrix (Figure 2.5, 2.6). Clasts comprise sandstone, shale and quartz. Although the unit is fairly massive in appearance, variations in color and weathering resistance suggest that the beds are ~1 m scale in thickness. Individual beds are ungraded, and they are tabular/sheet-like. The unit is more massive near the fault. Lithofacies 3 grades laterally and interfingers into lithofacies 2 (Figure 2.6a).

We interpret lithofacies 3 to be a debris flow dominated alluvial fan deposit, which was sourced off the thrust sheet west of the basin margin. The muddy matrix, abundance of matrix in paraconglomerate beds, and lack of grading/sorting within beds suggests that the clasts were supported during transport by a high viscosity flow (Nemec and Steel, 1984). The relatively sharp, and unchannelized, basal contacts of the beds suggest that the unit was deposited by an unconfined flow. Moreover, no cross beds or imbricated clasts were observed, suggesting that clasts are not traction deposits. The proximal alluvial fan interpretation is consistent with the clast composition of this unit, which suggests that these sediments were derived locally.

2.5.2 Growth strata and progressive unconformity at northern basin margin
Having described the structural architecture and the lithostratigraphic character of the basin, we present key structural observations that bear on the relationship between the basin strata and the basin-bounding fault. At several locations immediately south of the basin bounding thrust fault, beds in the footwall exhibit shallowing dips upsection (Figures 2.3, 2.7). At the best exposed of these outcrops, which is located at the northern edge of cross-section A-A’, slightly metamorphosed and very fractured beds dip sub-vertically (Figure 2.3, 2.7a,b). These beds are pebble orthoconglomerate with a muddy matrix, and they grade laterally into the mudstones in the central part of the basin. Dips shallow progressively upward, such that at the top of the outcrop, beds are sub-horizontal. The lower, steeply dipping beds are in progressively unconformable contact with the gently dipping strata near the top of the section, such that the unconformity diminishes up section and towards the center of the basin (Figure 2.7a,b,c). Strata that are correlative with the lower, steeply dipping beds along the northern flank of the basin are found in the distal portions of the basin, in conformable contact with overlying beds. A similar pattern of fanning dips was observed a few kilometers to the west, where dips vary from 63° to 17° in a vertical section abutting the thrust fault (Figure 2.3). Additionally, in the far southeastern portion of the basin at the base of the measured section, the basal strata thin toward the southern margin of the basin. At this location, the dips fan slightly such that beds near the base of the section are slightly steeper than those near the top (Figure 2.7d).

2.5.3 Stratigraphic units and regional lithostratigraphic correlation

We subdivide Jungong basin into 2 stratigraphic units. Unit 1 is equivalent to lithofacies 1. It is separated from the Triassic bedrock that floors Jungong basin by an angular unconformity. Unit 2 corresponds to lithofacies 2 and 3. The upper conglomerate (lithofacies 3) is spatially associated with the fault bounding the northwest margin of the basin and the
intrabasin thrust (Figure 3). The conglomerate beds fine laterally and interfinger with the mudstone (lithofacies 2), which occupy the distal portions of the basin (Figure 6a). The interfingering relationship indicates chronostratigraphic equivalence between the two. We attribute lithostratigraphic changes across unit 2 to the proximity of the depositional site and the basin bounding faults. The nature of the contact between units 1 and 2 is uncertain because it is concealed.

Based on field observations and a review of existing geologic maps (GBGMR, 1989; QBGMR, 1991), we correlate units 1 and 2 from Jungong basin with the early Cretaceous Hekou and Wanxiou Groups (Figure 2.8--hereafter referred to solely as the Hekou Group). Regionally, the Hekou Group is well indurated, brick-red to purple fluvial and alluvial fan conglomerate which underlies well indurated, brick-red fluvial red beds (mudstone with sandstone lenses). Thicknesses of these deposits range from a few hundred meters to a few km (Halim et al., 1998; Horton et al., 2004). Various geologic, biostratigraphic, and geochronologic evidence supports our regional lithostratigraphic correlation, and we present this evidence in section 7.

2.6. Synthesis and tectonic significance of Jungong basin structure and stratigraphy

Overall, the architecture of Jungong basin appears to be relatively simple. A single basin bounding thrust fault along the northeastern edge branches into two imbricate thrusts toward the southwest. Wedge-shaped packages of conglomerate extend basinward from the surface trace of the faults. The conglomerate beds fine away from the faults and interfinger with mud and sandstone, which occupy the distal parts of the basin. In the east, the basin strata are folded into an open syncline (Figure 2.3). Along strike to the west, the syncline is tighter and is eventually folded into an anticline-syncline pair. The structural block between the two thrust faults in the
western part of the basin is folded into a doubly plunging anticline. The entire basin is floored by a ~100m thick package of conglomerate.

One of the more structurally complex portions of the basin is located where the projections of each of the mapped faults overlap. Studies of the architecture of thrust faults have demonstrated that where thrusts overlap, they may be linked by splay faults (Boyer and Elliot, 1982). Based on the anomalous, ~E, strike of the basin bounding fault at this location and the anomalous ~E to NE dips of basin strata in this area, we interpret the geologic observations in north central Jungong basin to be consistent with the presence of a network of splay faults, linking overlapping basin bounding faults (Figure 2.3).

Unit 2 is interpreted to represent a system of alluvial fans and rivers that grew off the evolving topography in the hanging wall of the basin bounding faults. The presence of coarse grained alluvial fan deposits adjacent to the basin bounding thrust faults which fine towards the center of the basin (Figure 2.6a) and the similarity in clast composition between alluvial fans of Unit 2 and the rocks in the hanging wall of the fault support this interpretation. Alluvial fans are not only spatially associated with the faults along the north side of the basin (Figure 2.3), but also with the intrabasin fault that is exposed in the west half of the basin, suggesting the sediment accumulation in western Jungong was coeval with growth of both faults. Because the contact between the Unit 2 and the basal conglomerate is concealed, the tectonostratigraphic significance of the basal unit remains unclear. There is no angular unconformity between the basal conglomerate and the sediments above, indicating that no tilting of the basin occurred between deposition of the two units.

Key structural observations unequivocally demonstrate a temporal correspondence between the growth of the SE-vergent basin bounding faults and basin sedimentation, such that
the basin formed at a time of NW-SE contractional deformation. At two outcrops near the top of the measured stratigraphic section we have observed a progressive upward shallowing of dips and intraformational unconformities with stratigraphic Unit 2, associated with the surface trace of the basin bounding fault. The observations indicate progressive tilting of the proximal footwall strata by growth of the eastern bounding fault during sediment accumulation (Figure 2.3, 2.7).

Palinspastic restoration of two cross sections perpendicular to the structural fabric of Jungong basin permits us to estimate the minimum magnitude of Cretaceous upper crustal shortening here (Figure 2.9). Admissible, deformed state cross sections were generated from surface geologic observations. We have constructed cross sections in a way that minimizes the shortening of the beds; because hanging-wall cutoffs are not exposed, the amount of shortening could be greater. Folds along the southern edge of the basin are kinematically incompatible with faults mapped within the basin, so we speculate that additional structures must exist to the south and east of the basin. Indeed, several faults are mapped within the Triassic rocks to the south of Jungong basin that could be candidates for such a structure (QBGMR, 1991) (Figure 2.2). Moreover, the fanning dips found along the southeastern basin margin suggest that growth of faults to the SE of the basin may have been coeval with the filling of Jungong basin.

Restoration of the sections was conducted using the Move software package (available from Midland Valley Exploration) and indicates that the deformed state cross sections are viable. The restored length for section A-A’ is ~5000 m, and the deformed state length is ~4200 m, such that the section experienced ~17% shortening (Figure 2.9). Along cross-section B-B’, the basin strata were eroded in a hanging wall anticline of the intrabasin thrust fault (Figure 2.3, 2.9). The geometry of the fold is constrained by an up plunge projection of beds into the line of cross
section. The restored length for section B-B’ is ~11400 m, and the deformed state length is ~6100 m. Additionally, along this section, the basal contact of the basin does not intersect the surface at the northern boundary of the basin. In order to estimate the minimum magnitude of shortening here, we also restore the basal contact to the surface along the basin bounding fault, which adds an additional 1300 m of shortening (Figure 2.4). Thus the minimum shortening along B-B’ is 52%

2.7. Age of Jungong basin fill

Mudstone samples were collected at 12 sites within Jungong basin for the purpose of palynologic dating. Unfortunately, the samples yielded almost no identifiable organic material because of their high degree of oxidation. Therefore, we rely on correlation to nearby basins for information about the age of sediments in Jungong basin proper.

2.7.1 Local and regional early Cretaceous fossil assemblages

At Jungong, the basin-bounding fault forms a distinctive lineament on ~90-m Shuttle Radar Topography Mission (SRTM) digital topographic data. The lineament can be traced ~15 km to the SW, where it intersects a series of narrow basins containing Hekou Group sediments near the town of Dawu. Toward the northeast, it can be traced ~130 km NE where it is also associated with Hekou Group deposits near the towns of Tongren and Xia He (Figure 2.2, 2.8). Near Dawu, E-W striking faults, subparallel to the Kunlun fault, truncate both the Jungong fault network and Hekou Group deposits (QBGMR, 1991), but the presence of Hekou Group deposits of similar lithology on both sides of the Kunlun fault has been used to infer relatively little separation across this Cenozoic structure (Fu and Awata, 2007). A type of bivalve Musculus sp., (Table 2.1--QBMGR, 1991) was found in the upper part of the fill near Dawu, on the north side of the Kunlun fault (location shown in Figure 2.3), suggesting that these deposits are of
Cretaceous age. To the NE of Jungong basin, an abundant and diverse Early Cretaceous fossil assemblage has been found in Hekou Group deposits near the towns of Xia He and Tongren (Table 2.1, QBGMR, 1991). Additionally, Neogene fossils have been found in the upper part of the Tongren basin fill, suggesting the presence of a second generation of Tertiary fill, at least in the northern part of this region. Farther to the east, an E-W elongate sliver of Hekou Group deposits on the north side of the Kunlun fault near the town of Langmu Si, also contains Early Cretaceous fossils (Table 2.1, GBGMR, 1989). Farther still to the east, an Aptian-Albian age has been inferred for the Dangchang basin complex based on extensive palynologic and magnetostratigraphic analysis (Horton et al., 2004). Overall, the regional consistency in the age of fossil assemblages is rather remarkable and suggests that many of the small basins exposed throughout the Jungong-Dangchang basin complex share a common origin.

2.7.2 Age and association of volcanic rocks

The Songpan-Ganzi flysch deposits that mantle the Kunlun-Qaidam terrane are pervasively intruded by granitoid plutons (Figure 2.2). A narrow sliver of Hekou Group deposits south of Dawu sits unconformably atop one of these plutons (QBGMR, 1991). Existing K-Ar ages for plutons around the region indicates that they date to the Late Triassic – Early Jurassic, and thus place a rough upper limit on the age of the Hekou Group (Figure 2.2). Between Jungong and Tongren along the strike of the Jungong fault network, poorly exposed terrestrial deposits lithologically similar to the Hekou Group near the town of Zeku sit directly upon volcanic rocks (Figure 2.2, 2.10). Whole rock, K-Ar dating of 3 samples from these basal volcanic units (analytical procedures are described in Supplementary Information section) yields a mean age of \( \sim 217 \pm 20 \) Ma, and a minimum age of \( \sim 202 \pm 2 \) Ma, (Table 2.2, Figure 2.2). Thus, it seems clear that Hekou Group sediments are younger than the Late Jurassic. Relationships with
volcanic rocks in the Tongren region provide a younger bound on the age of Hekou Group sediments. In this region, fossil bearing Hekou Group sediments north and east of the town of Tongren are cut by a basaltic dike (Figure 2.2, 2.10). Three whole-rock samples from this dike yielded a mean K-Ar age of 102.9 +/- 3.6 Ma (Table 2.2). Thus, the depositional age of the Hekou Group is tightly bracketed by fossils and radiometric constraints to the Aptian-Albian Periods between ~125 Ma (base of Albian) and ~103 Ma.

This result is consistent with recent 40Ar/39Ar dating of volcanic deposits intercalated with terrestrial sediments near Duofutun (a village between Tongren and Jungong, Figure 2.2), where ages ranged from ~80 Ma – 100 Ma (Zhao, 2009). However, the ages of these volcanic rocks have recently been called into question, as xenocrystic zircons yielded Miocene U-Pb ages (Zheng, et al., 2010). Similarly, alkalic volcanic rocks of early-mid Miocene age floor narrow basins immediately NE of Dangchang (Dupont-Nivet et al., 2004; Yu et al., 2006). Thus, although it is clear that some of these basins continued to accumulate sediment during the Late Tertiary, this appears to represent a second generation of basin fill. Overall, the fossil assemblages and cross-cutting dike near Tongren strongly suggest that initial basin development took place during the Late Cretaceous.

2.8. Discussion

2.8.1 Kinematic setting of Cretaceous basins in the Jungong-Dangchang region

The observation of syn-contractional sediment deposition demonstrates that the Jungong basin formed during NW-SE crustal shortening. The basin itself is rather narrow, and the association with similar basins throughout present-day northeastern Tibet suggests that these basins may have been controlled by the spacing of thin-skinned thrust faults, similar to Tertiary basins in east-central Tibet (Horton et al., 2002). Whether these basins were part of a larger
foreland basin complex, or whether they were developed within a deforming orogenic wedge is uncertain. However, the fact folding along the southern margin of the Jungong seems to require additional thrust faults Jungong below the basin lends a degree of support to the latter interpretation.

The system of thrust faults that marks the northwestern margin of the Jungong basin appears to be a regionally extensive system. Apparent lateral continuity between the Jungong basin bounding fault and basin bounding faults at Dawu, and near Zeku, Tongren, and Xia He, suggests that this fault system was at least ~200 km long. Additional NE-striking structures that bound narrow sedimentary basins between 102° and 103° E seem to be imbricates of the same fault network (Figure 2.2). Although exposure of these basins is often limited, reconnaissance level field observations and mapping by previous authors (QBGMR, GBGMR) suggests that many of these basins contain Hekou Group sedimentary deposits, including the Dangchang basin (Horton et al., 2004). Although these authors inferred a transtensional setting, our regional synthesis suggests that basin formation in the Creteaceous throughout the region south of the SQS involved NW-SE directed contraction. We note that this direction is similar to the deformation field in the West Qinling orogen at this time (Ratschbacher et al., 2003; Enkelmann et al., 2006).

Elongate basins distributed along the trace of the Kunlun and the West Qinling fault represent a second group of basins of Late Cretaceous age. Given the orientation of the mid-late Cretaceous strain field in the Qinling Shan (e.g. Ratschbacher et al., 2003), ESE-striking structures, such as the Kunlun fault/Anyemaqen Suture and the West Qinling fault/South Qilian suture would have accommodated dextral shear at this time. Thus, we infer that elongate basins most likely formed as pull-apart basins associated with regions of dextral shear. Notably, a
similar tectonic regime is well-documented in the eastern parts of the Qinling-Dabie orogen, where a broad dextral shear belt was activated at 75 Ma in the Tongbai Shan (Webb et al., 1999, 2001), and where the Xiaotian-Mozitang fault zone of the Dabie Shan accommodated dextral shear beginning at 110Ma (Ratschbacher et al., 2000). Our study suggests that many of the western portions of the Qinling orogen experienced Late Cretaceous deformation of similar kinematics.

The inference of significant crustal shortening and dextral shear in the Jungong-Dangchang region presents at least two possible mid-Cretaceous tectonic interpretations, given the orientation of the strain field at the time. First, it is possible that this deformation reflects a time period of continued contraction throughout eastern China. Although there are extensive Cretaceous deposits in the Sichuan Basin, most of the accommodation space developed during the Late Triassic – Jurassic Indosinian orogeny (Chen et al., 1994); Late Cretaceous deposits are largely absent (Burchfiel et al., 1995). Likewise, minimal exhumation in the Longmen Shan region during this time (Kirby et al., 2002) suggests relative tectonic quiescence. Alternatively, the Qinling-Dabie appears more likely to have been zone of intracontinental dextral shear (Ratschbacher et al., 2003). In this scenario, the Jungong-Dangchang basins may represent a distributed step-over between zones of dextral shear along the Anyemaqen suture zone and the West Qinling fault system (Figure 2.11). Thus, crustal shortening may have been a relatively local feature associated with the geometry of deforming strike-slip zones (e.g. Ratschbacher et al., 2003).

### 2.8.2 Implications for topographic evolution of interior NE Tibet

The West Qinling fault appears to represent a prominent boundary in present-day northeastern Tibet. To the south, high, mountainous topography (~4 km) is developed within
Songpan-Ganzi basin deposits of the West Qinling Shan and the Anyemaqen Shan. To the north of Qinling fault/SQS, the mountain ranges and basins are lower (~2.5 km), and the region is mantled by broad trapezoidal Cenozoic sedimentary basins, and transected by narrow Cenozoic mountain ranges (e.g., Fang et al., 2003). Given the apparent Cenozoic age of most of many of the major mountain ranges to the north of this boundary, the physiographic transition across the West Qinling fault may represent the pre-Himalayan edge of high topography in Tibet. Although we cannot say when and how this proto-plateau margin was constructed, it is clear that during the Cretaceous, the terrain south of the SQS and north of the AS experienced upper crustal shortening and dextral shear, such that significant pre-Himalayan topographic growth/crustal thickening may have occurred. Line length shortening estimates across Jungong basin indicate at least ~1-6 km of shortening locally. A more regional budget of Early Cretaceous shortening is elusive, though, due to the limited exposure of Cretaceous basins and the distinct possibility that not all Cretaceous basins are preserved.

2.8.3 Distribution of Cretaceous, pre-Himalayan deformation in central China

Our results suggest that the Anyemaqen Suture may have remained a relatively weak feature of the Asian continental lithosphere, accommodating deformation during the Late Cretaceous. Pervasive deformation occurred within the Kunlun-Qaidam Terrane, throughout the Qinling Dabie orogen, and in southern and central Tibet (e.g. England and Searle, 1986; Kapp et al., 2005), even while the eastern Songpan Ganzi-Hoh Xil terrane and the Yangtze craton were tectonically quiescent (Figure 2.11). Thus, to first order, these differences in the tectonic history of blocks within the Asian collage seem to require strength contrasts within the lithosphere beneath the present-day Tibetan Plateau. We attribute contrasts in mechanical behavior to the history of terrane accretion that characterized this region during the Paleozoic.
Moreover, mid-Cretaceous deformation in eastern China seems likely to reflect the accretion of the West Philippines block to east China along a SE facing subduction zone (Ratschbacher et al., 2003), whereas Cretaceous deformation in Lhasa and Qiangtang are the consequence of subduction zone tectonics along the southern margin of Lhasa (e.g. England and Searle, 1986, Ratschbacher et al., 2003). Because of similarities in the timing and style of deformation in the Jungong-Dangchang region and the Qinling-Dabie Shan during the Cretaceous, and because of the continuity between the two mountain belts, we infer that deformation of interior northeastern Tibet was likely driven by stresses generated along the eastern margin of China. Thus, the interplay between deformation within Eurasia driven from the convergent boundary to the south and that driven from the east appears to have been as significant during the Cretaceous as it is today.

2.9. Conclusions

New mapping, stratigraphy and structural analysis of the Jungong basin lead us to the following conclusions.

1. A new regional correlation of basins between the Anyemaqen and South Qilian suture zones in interior NE Tibet, based on lithostratigraphic, biostratigraphic, structural, and geochronologic evidence, during Aptian-Albian times, at ca. 125 to 100 Ma.

2. Facies associations and growth strata along the northwestern margin of the Jungong basin indicate that basin sedimentation was synchronous with slip on a SE-vergent thrust fault. Regionally, basin development appears to have been related to numerous individual thrust faults in a broad zone of NW-SE shortening.

3. Mid-late Cretaceous deformation in interior northeastern Tibet is kinematically consistent with deformation in the Qinling-Dabie orogen and appears to be related to a zone of
intracontinental, right-lateral shear. Basins and thrust faults in the Jungong-Dangchang region, probably reflect shortening in a broad restraining step between the Anyemaqen and South Qilian suture zones, and this deformation may have contributed significantly to crustal thickening in interior NE Tibet prior to the Indo-Asian collision.

4. Finally, the Songpan-Ganzi terrane appears to have remained relatively undeformed during the Late Cretaceous, such that it may have separated the Asian tectonic collage into domains deforming in response to plate boundary stresses generated along the southern and eastern margins of Asia.

2.10 Supplementary Information

K–Ar dating was conducted in the State Key Laboratory of Earthquake Dynamics, Institute of Geology, China Earthquake Administration. Fresh volcanic rocks were crushed and sieved to a size range between 60 mesh and 80 mesh. Mineral crystals, such as olivine and plagioclase, which may contain excess argon, were removed. The sieved samples were washed with water, ethanol and acetone. Argon was analyzed using the isotope dilution technique with a 99.98% pure $^{38}$Ar spike, on a MM1200 mass spectrometer connected to a purification and extraction system. The spike was calibrated and corrected with standard argon minerals: B4M and ZBH. The samples were enclosed in a “Christmas tree”-shaped holder and heated to about 200 °C for more than 10 h, then separated and placed in molybdenum crucibles surrounded by a titanium crucible. Argon was extracted using an electron bombardment furnace with two vacuum systems: the inner vacuum system, including the Christmas tree, was connected to the purification system and mass spectrometer. The outer vacuum system was evacuated with a diffusion pump and used for electron bombardment heating. The samples were then heated to 1320 °C. Released gases were purified, first by a cold trap to remove CO$_2$ and H$_2$O, then by a
titanium sponge to separate other active gases, and finally by Zr–Al getters. The purified noble gases were introduced into the mass spectrometer to measure the argon isotopes; K content was analyzed by a HG-5 flame photometer with Li inner standard. Age calculation was based on 40K decay to 40Ar gas with decay and abundance constants $\lambda_{\beta} = 4.962 \times 10^{-10}$ a$^{-1}$; $\lambda_{e} = 0.581 \times 10^{-10}$ a$^{-1}$; $^{40}\text{K}/K = 1.167 \times 10^{-4} \text{ mole}^{-1}$ (Steiger & Jager, 1977). $\lambda_{s} = 5.543 \times 10^{-10} /a$ , $\lambda_{e} = 0.581 \times 10^{-10} /a$, $\lambda_{\beta} = 4.962 \times 10^{-10} /a$, $^{40}\text{K}/K = 1.167 \times 10^{-4}$ mole/mole.
Figure 2.1. (a) Towns, major rivers and active faults in Central China. Inset shows terranes of China. SGT = Songpan-Ganzi terrane, KQT = Kunlun-Qaidam terrane. (b) Cretaceous and Paleocene/Eocene basins of central China. The focus of this study is the basins distributed between Jungong and Lintan. Active faults are shown in light grey for reference. Terrane boundaries are shown in heavy dashed line. Figure is adapted from Horton et al., 2004 and Enkelmann et al., 2006.
Figure 2.2. Geology of the Jungong-Dangchang region. K-Ar ages of plutons and fossil assemblages are compiled from QBGMR, 1991. Dates of Miocene volcanic rocks near Dangchang basin from Dupont-Nivet et al., 2004. Dates of Miocene volcanics near Zeku from Zheng et al., 2010.
Figure 2.3. (a) ASTER Level 1a image of Jungong basin with key geologic contacts. (b) Detailed geologic map of Jungong basin. Mapping is based on field observations and interpretation of ASTER imagery. Locations of cross sections A, and B are shown (see Figure 2.9), as well as the location of the stratigraphic section (see Figure 2.5). For location of Jungong basin, see Figure 2.2.
Figure 2.4. Key geologic contacts in Jungong basin. (a) View to the north of the basin bounding thrust fault carrying Triassic rocks over the basin fill. The Yellow River is in the central portion of the picture. Jungong town is in the lower left. (b) Close-up view of the basin bounding fault near Jungong. (c) Basin bounding thrust in the central portion of the basin, where the road crosses the fault (see Figure 2.3). (d) Unconformity at the base of Jungong basin. Contact is delineated with black line. Shrubs in foreground are ~50 cm tall. The photo was taken south of Jungong town.
Figure 2.5. Stratigraphy of Jungong basin. Location of section shown on Figure 2.3.
Figure 2.6. (a) Panorama showing the upper ~800 m of the stratigraphic section in Figure 2.5. Resistant, sandstone lenses thin toward the southern part of the basin. (b) Lithofacies 1. Purple to brick red basal orthoconglomerate. (c) Lithofacies 2. Brick red mud interbedded with sandstone lenses. Resistant bed in the middle of the photograph is ~40 cm thick. (d) Lithofacies 3. Buff colored orthoconglomerate.
Figure 2.7. (a) Progressively shallowing dips observed near the northern edge of the basin at cross section A (see Figure 2.9). In foreground, beds dip subvertically. In background, beds progressively shallow and become subhorizontal in far background, as shown by the black lines delineating various bedding planes. (b) Close-up view of subvertical bedding adjacent to basin bounding fault. (c) View of the outcrop in the far background of part a, showing progressively shallowing dips near the northern edge of Jungong basin. Truck for scale. (d) Progressive shallowing and thinning of basin fill near the southern edge of section A. Black lines delineate bedding planes. Cliff face is ~50 m tall at the right edge, above the tree line.
Figure 2.8. Generalized stratigraphy and example outcrops the Hekou and Wanxiou Groups in the vicinity of Jungong basin (compiled from field observations; GBGMR, 1989; QBGMR, 1991). (a) In general, the Cretaceous deposits are 1 -3 km thick, and they consist of a basal, well indurated, brick red to purple fluvial and alluvial fan conglomerate, interbedded with brick-red, fluvial-floodplain mudstone and well indurated lenticular sandstones. Deposits may be comparatively thin in some locations, on the order of a few hundred meters. The deposits sit unconformably atop Songpan-Ganzi flysch deposits, and in some places, the deposits also sit unconformably atop plutons that have pervasively intruded the Songpan-Ganzi flysch. (b) Outcrop near Dawu. Hill is ~8 m tall. (c) Outcrop south of Zeku. Largest clasts b-axis ~2 - 4 cm. (d) Outcrop near Tongren. Relief on cliff in background is on the order of ~80 m. Note the brick red to purple color, and the degree of induration at each outcrop. See Figure 2.2 for locations of b,c,d.
Figure 2.9. Deformed and restored cross sections across Jungong basin. Restorations were done using Move software package. Lengths of deformed and restored sections are shown above each section. Along B-B’, the geometry of the eroded beds was constrained by projecting beds up the plunge of the fold. Black unit is lithofacies 1, grey unit is lithofacies 2, white unit is lithofacies 3.
Figure 2.10. Field photographs documenting geologic relationships that help us constrain the age of the Hekou Group, at Jungong and beyond. (a) Volcanic rocks flooring a basin filled with Hekou Group sediments near the town of Zeku. The volcanics were sampled in three places and dated using whole-rock, K-Ar geochronology, and yield an average age of 217 ± 20 Ma. (b) A dike cross-cutting the Hekou Group deposits within Tongren basin. Three samples yielded a mean, whole-rock, K-Ar age of 103 ± 4 Ma. Dike outcrop is ~ 10 m high.
Figure 2.11. Inferred tectonic setting of mid-late Cretaceous basins in Jungong-Dangchang region. NE-striking structures accommodated shortening and E-striking structures accommodated dextral shear. This is related to NW-SE compressional-NE-SW extensional stress field of eastern China during the mid-late Cretaceous (Ratschbacher et al., 2003).
<table>
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<th>Region</th>
<th>Fossil Pollen Assemblages</th>
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<td>Tongren</td>
<td>1. Podozomite sp. 5. Eosestheria sp. 1. Onchiopsis sp. 2a. Schizaea sp. 2b. Cicacas sp. 2b. Cicatricosupcrites sp.</td>
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<td></td>
<td>2b. Schizaea sp. 2c. Pagiophyllum sp. 2. Equisetites sp. 2. Equisetites sp.</td>
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<td></td>
<td>2. Carpolithus sp. 2. Elatocladus sp. 1. Cyathidites 1. Todisporites</td>
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<td>Dawu</td>
<td>3. Musculus sp. 2. Elatocladus sp.</td>
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<td>Table 2.1</td>
<td>Fossil pollen assemblages in Jungong region.</td>
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Table 2.2. K-Ar samples from dikes cross cutting basin fill in Jungong region.
CHAPTER 3

LATE MIOCENE – PLIOCENE RANGE GROWTH IN THE INTERIOR OF NORTHEASTERN TIBET

3.1. Abstract

Characterizing the spatio-temporal pattern of deformation throughout the Indo-Asian collision zone can place constraints on the processes responsible for the development of high topography. In northeastern Tibet, the history of deformation appears to have been quite complex, with a clear episode of mountain building in the Late Miocene, but also involving subtle deformation as early as the Eocene. Here, we present new constraints on basin evolution and the timing of thrust faulting in southern Gonghe basin complex, which occupies an interior portion of NE Tibet between the southern Qilian Shan and the central Tibetan Plateau. Lithostratigraphic sections record the onset of basin sedimentation and the evolution of depositional environments across the region. Deformation of basin strata, lithostratigraphic patterns, and paleocurrent shifts are used to constrain the onset of growth of faults in the Gonghe Nan Shan (GNS), a major mountain range which occupies the south-central portion of the basin complex. Magnetostratigraphy and cosmogenic burial ages show basin initiation at ca. 20 Ma and that the GNS became active during the late Miocene, between ~10 and 7 Ma. The timing of basin initiation in Gonghe is similar to other interior portions of northeastern Tibet, and appears to correspond to the onset of significant Himalayan era tectonism in the region. The timing of the GNS illustrates a distinct change in structural style in northeastern Tibet during the late Miocene, in which previously vast and unbroken foreland basins were segmented by the emergence of narrow mountain ranges. Late Miocene compartmentalization of foreland basins in interior northeastern Tibet was synchronous with the rise of mountain ranges along the periphery of
northeastern Tibet and the Qilian Shan, suggesting a clear increase in contractional tectonism over length scales of thousands of kilometers at the time.

3.2. Introduction

The space-time patterns of deformation throughout the Indo-Asian collision zone provide important bounds on the processes responsible for the development of thickened crust and high topography that characterize the modern-day Tibetan Plateau (Molnar et al., 1993; Royden et al., 2008). Initial collision of India with Eurasia occurred between 45 – 55 Ma (Garzanti and van Haver, 1988; Rowley, 1996; Najman et al., 2010) and near ~20° north latitude (Patriat and Achache, 1984; Dupont-Nivet et al., 2010), implying nearly ~3000km of post-collision convergence (Dupont-Nivet et al., 2010). How this convergence has been accommodated, both within the Himalaya (e.g., DeCelles et al., 2002), and throughout the Eurasian lithosphere (e.g., Lippert et al., 2010) has driven a generation of research into the timing and magnitude of upper crustal shortening. Collectively, these studies suggest that the southern margin of Eurasia was characterized by thickened crust (e.g., Murphy et al., 1997; Kapp et al., 2005) and high elevation (DeCelles et al., 2007; Rowley and Currie et al., 2006) prior to collision, and that deformation and crustal thickening progressed outward to the north and east from this elevated core (Meyer et al., 1998; Tapponnier et al., 2001; Royden et al., 2008).

Within this regional context, the history of deformation and topography in northeastern Tibet plays a crucial role in two regards. First, although many workers held that growth of the ranges north and east of the Qaidam Basin (Qilian Shan and associated ranges, Figure 1) was largely confined to the late Miocene and Pliocene (e.g., Burchfiel et al., 1989; Tapponnier et al., 1990; Meyer et al., 1998; Tapponnier et al., 2001), recent studies suggest that the history of mountain building may have been more complex. In particular, evidence for deformation
initiating in the early-mid Tertiary along the margins of the western Qaidam basin (e.g., Ritts et al., 2004; Yin et al., 2008) and within northeastern Tibet (Fang et al., 2003; Dupont-Nivet et al., 2004; Clark et al., 2010; Duvall et al., in press) have led some to consider that the present-day boundaries of the Tibetan Plateau were largely established at or near the onset of collision (Clark et al., in review). Second, geologic studies using proxy data provide abundant evidence for a pulse of mountain building throughout northeastern Tibet in the Late Miocene (e.g., Zheng et al., 2003; 2006; Fang et al., 2005; Lease et al., 2007; 2010), apparently coincident with the onset of shortening at the present-day northern (Zheng et al., 2011) and southern (Fang et al., 2007) margins of the Qilian Shan. Widespread shortening across the peripheral regions of Tibet has been suggested to reflect a fundamental change in the geodynamics of the collision zone (Molnar et al., 2005), perhaps related to an increase in potential energy associated with removal of mantle lithosphere beneath the plateau (Molnar and Stock, 2009).

These studies leave open several outstanding questions regarding the growth of northeastern Tibet. First, how widespread was the region of early Tertiary deformation? Were early Tertiary depocenters in the western Qaidam linked with early Tertiary basins in northeastern Tibet (e.g., Clark et al., 2010), or did these evolve as separate tectonic elements (e.g., Wang et al., 2006). Second, does the episode of mountain building in the Late Miocene exhibit space-time patterns that may yield insight into proximate causes? Was range growth coordinated, either reflecting northward propagation of deformation from the plateau (e.g., Meyer et al., 2006), or did ranges grow asynchronously (e.g., Sobel et al., 2003), reflecting the disruption of a previously-unbroken foreland (e.g., Strecker et al., 2007). Here, we address these through detailed study of range growth along the southern margin of the Gong He basin, in interior northeastern Tibet. The Gong He basin complex is located along the axis separating
early Tertiary depocenters in northeastern Tibet with the Qaidam basin, and thus the Tertiary depositional history of this region provides a key test of the regional extent of early Tertiary sedimentation. Similarly, the basin sits today at a physiographic transition between the basin-and-range topography of the Qilian Shan to the north, and the highland of the Tibetan Plateau to the south, and is known to have been a significant depocenter in the Plio-Quaternary (e.g., Craddock et al., 2010). We combine geologic mapping of deformed Tertiary sediments, with stratigraphic and sedimentologic observations to place constraints on the relative timing of range growth and sediment accumulation along this margin of the Gong He basin. Absolute chronology is established using magnetic reversal stratigraphy, correlated to the geomagnetic polarity timescale using cosmogenic burial ages in alluvial sediment (e.g., Granger, 1996). This novel approach to establishing a radiometric basis for reversal stratigraphy allows us to bracket the onset of deformation along the fault systems bounding the Gong He basin.

3.3. Background

3.3.1. Basin formation in NE Tibet

Northeastern Tibet, which is covered by broad sedimentary basins and transected by narrow elongate mountain ranges, has been the site of protracted basin formation since at least the Early Cretaceous (Figure 3.1). Stratigraphic archives from Xining and Lanzhou basins, near the periphery of northeastern Tibet, indicate slow subsidence (~10 m Ma^{-1}) from the Hauterivian until the mid-late Tertiary (Horton et al., 2004; Dai et al., 2006). Whereas some have interpreted the subsidence history in the Linxia-Lanzhou region to reflect thermal subsidence following extensional deformation during the early Cretaceous (e.g. Horton et al., 2004), others have invoked vertical axis rotation of the Xining basin by ~41 Ma, (Dupont-Nivet et al., 2004; 2008a), thrusting along a segment of the west Qinling fault at ~45-50 Ma (Clark et al., 2010; Duvall et
al., 2011), and proxies for paleoaltitude (Dupont-Nivet et al., 2008b) to suggest that contractional deformation and surface uplift began in northeastern Tibet only a few million years after the beginning of the Himalayan-Tibetan orogeny (Clark et al., in review). Unfortunately, the margins of Cretaceous and Paleogene basins are typically poorly exposed, such that their structural setting is equivocal (e.g. Horton et al., 2004; Ritts and Biffi, 2002). Thus, a key issue concerning the evolution of northeastern Tibet is determining the tectonic setting of these sedimentary basins.

Although it is clear that sediment accumulation was widespread across northeastern Tibet during the late Cenozoic, the evolving locations of key depocenters remains uncertain due to inadequate study of some key regions. Slow, early Tertiary sediment accumulation was punctuated in the Linxia basin region during the mid-Oligocene by rapid, flexural subsidence (e.g Fang et al., 2003). This event appears to correspond with the emergence of the Laji Shan from a once broad region of sediment accumulation, thereby separating Xunhua basin from western Linxia (Hough et al., 2011; Lease et al., in review). The extent to which similar mid-Tertiary deformation occurred regionally is unclear, although, sedimentary archives indicate slow subsidence in the Xining region persisted through the middle part of the Tertiary (Dai et al., 2006).

By the late Miocene, the depositional patterns and tectonic style across northeastern Tibet clearly differed from the Cretaceous and early Cenozoic. Several basins, including Guide, experienced rapid, flexurally-driven sedimentation at this time (Fang et al., 2005). In addition, a variety of stratigraphic proxies for sediment provenance reveal the rapid unroofing of source terranes to basins in the interior of the region from ~14 – 8 Ma (Dettman et al., 2003; Fang et al., 2003, 2005; Garzione et al., 2005; Lease et al., 2007), and thermal histories from ranges along the plateau edge at the Liupan Shan record a coeval acceleration in exhumation
Whether the late Miocene deformation marks a distinct change in structural style, or a single event within a more continuous history of deformation is a critical unknown in terms of the kinematic history of the region.

3.3.2. Sedimentary records in Qaidam and growth of the Qilian Shan

The history of Cenozoic basin evolution and tectonism in western Qaidam has some broad similarities to northeastern Tibet (Figure 3.1). First, sediment accumulation initiated by the Paleocene, and persisted across the region for the duration of the Cenozoic. In the Paleocene, marine strata were deposited around the southwestern margin of the basin (e.g. Ritts et al., 2008 and references therein). Subsequently, during the Eocene, sedimentation expanded northward and eastward, as far east as ~Golmud, and became dominated by nonmarine deposition, with infrequent episodes of shallow marine sedimentation (e.g. Yin et al., 2008b; Ritts et al., 2008). Paleogene sedimentation appears to correspond to tectonism on three major structures. 1) Sedimentary archives from near the margins of the basin indicates that the southwestern Qilian Shan-Nan Shan was likely to be a topographic high and a sediment source during the Paleogene(Yin et al., 2008b Bovet et al., 2009). 2) A variety of thermal and geologic proxies suggest that the Altyn Tagh fault was active during this time (Jolivet et al., 2001; Yin et al., 2002). 3) Later, during the early Oligocene, the Qimen Tagh/western Kunlun Shan emerged along the southern margin of the basin and isolated western Qaidam from Hoh Xil basin to the south (Mock et al., 1999; Yin et al., 2008a, b).

Similar to northeastern Tibet, the Miocene appears to herald an important change in the style of basin evolution in the Qaidam region. First, depocenters expanded eastward in the mid-late Miocene, and periodic shallow marine sedimentation ceased (e.g. Metivier et al., 1998, Fang
et al., 2007; Ritts et al., 2008; Yin et al., 2008b). Second, stable isotopes preserved in lacustrine carbonates and paleosols suggest significant surface uplift of the region during the Neogene (Kent-Corson et al., 2008). In the late Miocene, thermal proxies and stratigraphic proxies from the northern Qilian Shan and the Hexi Corridor suggest that the Qilian Shan-Nan Shan expanded outward and attained their present-day tectonic configuration by ~8 – 10 Ma (George et al., 2001; Jolivet et al., 2001; Bovet et al., 2009; Zheng et al., 2010). The southeastern flank of the Qilian also experienced a rapid pulse of sedimentation during this time (Fang et al., 2007). The continuity of both early and late Tertiary mountain building and basin evolution in northern Tibet remains undefined because the geologic evolution of interior northeastern Tibet is not well understood.

3.3.3. Geologic setting of the Gonghe basin

In order to address the linkages between Qaidam and northeastern Tibet throughout the Cenozoic, it is critical to reconstruct the depositional and structural history of the broad interior portion of northeastern Tibet. Much of the region is occupied by the Gonghe basin complex, which is a broad, ~200 x 200 km, region that bridges Qaidam basin with the outer reaches of northeastern Tibet. Narrow, ~ESE-striking mountain ranges override the margins of the basin and merge along strike with major structural grains in the region. The GNS is one of these mountain ranges (Figure 3.2). It occupies the south-central portion of the basin complex, and it merges with the west Qinling Shan and the Kunlun Shan (Figure 3.1). The mountain range rises ~1500 m above the adjacent basin floor, and it divides the greater Gonghe basin complex into two subbasins, Gonghe basin (sensu stricto) to the north, and Tongde basin to the south.

Perhaps due to the fact that most of the Gonghe basin is poorly exposed, the timing of Cenozoic mountain building around the margins of the basin is not very well defined, with one
key exceptionalong the northwestern basin margin. There, magnetostratigraphic dating of growth
strata and paleocurrent reversals in the northern subbasinindicates that the Qinghai Nan Shan
emerged from a broader foreland that connected Qinghai Lake basin and northwestern Gonghe,
at 5-8 Ma (Zhang et al., in review). Several kilometers to the east, the northwestern Qinghai Nan Shan merge with the Ela Shan, a series of NNW striking mountain ranges related to the right lateral Ela Shan fault and associated flower structures. Recent analysis of displaced bedrock piercing points and Quaternary slip rates indicates that the Ela Shan fault initiated at 6-12 Ma (Yuan et al., 2011), although uncertainties inherent to this analysis are large. Immediately to the east of Gonghe basin, changing sediment provenance in Guide basin suggests that the Laji Shan emerged at ca. 8 Ma, hinting that structures activated along much of the northern Gonghe basin margin during the late Miocene (Fang et al., 2005; Lease et al., 2007)

Gonghe basin is flanked to the south by the Anyemaqen Shan, whichis a broad region of high elevation and high relief that is thought to be related to diffuse crustal thickening at the eastern terminus of the Kunlun fault (Harkins et al., 2010) (Figure 3.1). Analysis of longitudinal river profiles in the region suggests that the development of relief in the mountain range occurred during the late Cenozoic, although the precise timing remains unknown (Harkins et al., 2007). Virtually the only age constraint that bears on the GNS itself is the association with late Cenozoic basin fill. Age constraints on the southern Gonghe basin fillare not absolute and are based solely on regional lithostratigraphic correlation.

A few pioneering studies provide useful regional lithostratigraphic context for the southern Gonghe foreland basin. First, Cenozoic strata of the Gonghe basin and surrounding basins are deposited unconformably atop the multiply deformed Triassic flysch deposits that blanket the region (Figure3.2). The only existing information concerning the deeper levels of the
Gonghe basin fill derives from the Chaka subbasin, in the northwestern corner of greater Gonghe. Recent magnetostratigraphic and lithostratigraphic analysis of exposures of deep basin fill in Chaka basin indicates that Cenozoic sediment accumulation began at Chaka at ca. 10.5 Ma (Zheng et al., in review). Chaka is floored by ~700 m of muddy, fluvial and lacustrine deposits which date to 7.6 – 10.2 Ma (Zheng et al., in review). Overlying the basal deposits is an 800 m thick package of fluvial deposits which coarsens upward into >300 m thick package of alluvial fan deposits. The two packages of sediment date to 4.4 – 7.6 and 4.4 to <3.0 Ma, respectively (Zheng et al., in review), and they record emergence of the northwestern QNS in the late Miocene.

Chaka subbasin appears to be somewhat younger than western Qaidam, to the west, or Guide to the east. Both of these basins initiated in the early part of the Miocene, between ~15 – 20 Ma, and were dominated by fluvial-floodplain sedimentation since that time. Additionally, both basins record coarsening upward stratigraphic sequences beginning in the late Miocene, and various stratigraphic proxies suggest that this pattern relates to emergence of mountain ranges along the northern basin margins (Fang et al., 2005, 2007; Lease et al., 2007).

Although the age and stratigraphic character of the upper basin fill in Gonghe is relatively well studied, the tectonic significance of these strata, and their relationship to the GNS, is not known. In southern Gonghe and Tongde, the fill is capped by a ~300 - 600 m succession of gravels, sands and muds (Zheng et al., 1985; Craddock et al., 2010). Faunal assemblages from within the canyon indicate that the sediments that cap the basin are Plio-Pleistocene in age (e.g. Zheng et al., 1985). Recent cosmogenic burial age dating and magnetostratigraphy from the upper Tongde basin fill provides the first absolute age control for the strata and indicates that the fluvial sands and gravels that cap the Tongde basin are ~3.4 – 0.5 Ma (Craddock et al., 2010).
3.4. Regional structure and stratigraphy of southern Gonghe basin complex

The GNS occupies the southern portion of the Gonghe basin complex and relates to a network of S-vergent imbricate thrust faults (Figure 3.2). The fault network outcrops along the southern flank of the range in the Yellow River canyon, although along most of the range front, it is blind or buried (QBGMR, 1991). Exposures of basin fill within the Yellow River canyon in Tongde basin and southern Gonghe basin, on either side of the range, indicate that deep Cenozoic basin strata are folded over the GNS and that relatively shallow strata abut or onlap the range. The gross regional architecture of the foreland basin strata indicates that the range comprises a series of broad, asymmetric anticlines in the hangingwall of the GNS fault system (QBGMR, 1991). The range is cored by Triassic Songpan Ganzi flysch deposits, which are thought to represent a remnant ocean basin (e.g. Weislogel, 2008).

3.4.1 Cenozoic foreland basin stratigraphy

In order to describe the stratigraphic architecture of the foreland basin, we present a 1350 m measured section that spans the Cenozoic basin fill (Figure 3.3). The section was measured in a network of deep canyons on the northern flank of the range (Figure 3.2), and it is a composite of a deep section that extends to the basin floor and an upper section that extends to the top of the Gonghe basin fill. A small river valley and an adjoining small alluvial fan complex cover the top of the deep section, and the lower section is floored by an unconformity. We estimate that at most, a few hundred meters of basin fill is unexposed or eroded in southern Gonghe. A variety of distinctive lithofacies are present in southern Gonghe (Table 3.1) and assemblages of these facies define three lithostratigraphic units (Table 3.2). In the following section, we describe the stratigraphy the southern Gonghe stratigraphic section. Field photo graphs of each stratigraphic unit are provided in Supplementary Information section 1.
At the base of the lower section, a ~200 m thick accumulation of lithostratigraphically
distinctive sediments that bears Jurassic-Cretaceous pollen sits unconformably below our
measured section. Detailed stratigraphic observations and Jurassic-Cretaceous pollen
assemblages are presented in Supplementary Information section2.

3.4.1 Distribution, description, and interpretation of lithofacies assemblages

*Lithostratigraphic Unit1 (L1)-lenticular, trough cross-bedded siltstone and sandstone
interbedded with brick red mudstone*-L1 is found in the southern part of the Gonghe basin
complex, where it is folded over the GNS. It is ~600 m thick. It is also found in discontinuous
patches along the southern flank of the GNS (Figures 3.2). L1 consists of tabular intervals
of brick-red mud interbedded with lenses of sandstone (Figure 3.3). The sandstone lenses may
exhibit curved basal contacts, but most are sharp and planar. The lenses are cm to 1.5 m thick and
continuous over 10s to 100s of m. Dense intervals of 3-10 lenses are a few m thick, and are
separated by a few m of mud. They exhibit trough cross-bedding and horizontal lamination.
Sandstone may be very fine- to coarse-grained, and sometimes exhibits pebble lags. Mud beds
are typically massive, but occasionally exhibit planar parallel lamination. Mudstones commonly
show grey and orange mottling, often along laminae surfaces. Muddy intervals contain carbonate
nodules and concretions, burrows, and roots, and they exhibit a prismatic texture. The muds
contain abundant red, grey, and white laterally continuous, ~cm-thick calcite cemented bands.
Red and white bands are also found in massive and cross bedded sandstone beds. A variation of
L1, which we dub L1a, is identical to L1 except that it is dominated by amalgamated siltstone
and sandstone beds that are up to 10s of m thick with thin mud interbeds or parting (Figure 3.3).
We interpret this facies assemblage to represent a fluvial floodplain environment, on the basis of
indicators for channelized, unidirectional flow and well developed paleosol horizons.
Lithostratigraphic Unit 2 (L2)-horizontally laminated orange and tansheet like siltstone-L2 is found extensively across the southern Gonghe basin complex. At our measured section it is ~250 m thick before it is covered in the river valley at the top of our lower section. Correlative outcrops appear in the base of deep canyons in central Tongde (Craddock et al., 2010; Chapter 5). The unit comprises light red-orange or buff colored very fine- to coarse-grained sandstone (Figure 3.3). Finer parallel laminated sandstones and mudstones are interbedded with 1-3 m thick laterally continuous, clast supported conglomerates, or trough cross-bedded or horizontally laminated medium to coarse sandstone. Conglomerate beds exhibit irregular basal contacts, with at most, a few tens of cm of relief. The conglomerate clasts are imbricated, and they appear to be sheetlike. Clasts are poorly sorted and subrounded to angular. The unit is also interbedded with laterally continuous bands of grey, and brownish- and greenish-gray massive mudstone and siltstone. Red and white ~cm-thick calcite cemented bands are often present in this unit. Similar to L1, we interpret the abundance of indicators for channelized, unidirectional flow and paleosol development to indicate deposition in a fluvial-floodplain environment. L2 differs from L1 on the basis of 1) the relatively low degree of oxidation in L2 indicated by the orange color of the mudstones, 2) the presence of m-thick conglomerate bands, and 3) the lateral continuity of the beds. Although L2 beds appear to be sheet-like at the scale of an outcrop, they pinch out laterally, over length scales of hundreds of m. This bedding geometry indicates sediment deposition by relatively unconfined flows, such that L2 may contain a relatively high abundance of overbank, levee or crevasse splay deposits in comparison to L1.

Lithostratigraphic Unit 3 (L3)-imbricated conglomerate-L3 caps the Gonghe basin complex and along the southern margin of Gonghe, it is ~450 m thick (Figure 3.2). L3 strata can
be traced to the south through a structural saddle in the GNS and into Tongde basin, such that the upper basin fill in southern Gonghe is correlative to gravels that cap Tongde. In northern Tongde, L3 is ~400-500 m thick. These strata can be traced south along outcrops within the Yellow River canyon for tens of kilometers. Moving south, the beds interfinger with sandier deposits, and they are directly correlative with the Plio-Quaternary gravels and sands in central Tongde basin (Craddock et al., 2010; Chapter 5). The unit is composed of pebble or cobble clast supported conglomerate with a buff colored muddy matrix and with buff, sandy and silty lenses (Figure 3.3). The bedding is lenticular and laterally continuous over 1s to 100s of m. Beds are 10s of cm to a few m thick. Conglomerates are clast supported, and clast sizes range from 20-200 mm. The clasts are unsorted to poorly sorted, and subangular to subrounded. Sand lenses are separated by a few meters, they exhibit decimeter scale thickness, and they are a few m to tens of m wide. Sedimentary structures include imbrication, channel scours, large (10-100 cm high) cross bedding within gravel beds, and parallel or cross lamination in finer lenses.

This facies assemblage is interpreted to represent braided channel alluvial fan deposits. The presence of imbrication, cross bedding, and parallel bedding indicates that the unit was deposited by a unidirectional flow. The lenticular geometry of the beds, and the channel scours along the base of the beds indicate that the flow was channelized. The distribution of L3 along the GNS range front suggests that it is an alluvial fan deposit. The absence of a) matrix support, b) inverse grading, and c) tabular geometry distinguish these beds from debris flow dominated alluvial fan deposits.

3.4.2. Paleocurrent analysis

In order to begin to understand the relationship between basin evolution in southern Gonghe and the GNS, paleocurrent indicators were measured along stratigraphic sections in
southern Gonghe and Tongde (Figure3.3). Whereas some intervals are only sparsely sampled, most contain 20-40 individual measurements. In order to measure paleocurrents, we sought unidirectional paleocurrent indicators, particularly trough cross-bedding and imbricated conglomerate clasts. Given the uncertainty imposed by a lack of 3-d exposure in most outcrops, we ascribe an uncertainty to our measurements of ~±30. Due to the fact that measurements from the lower Gonghe section were taken from tilted beds, the true paleoflow direction will differ slightly from our measured direction. Given that deformation of the lower Gonghe strata can be approximated by 30° clockwise rotation about a horizontal axis, the maximum post-depositional deflection is ~15°. This error is well within the uncertainty we ascribe to our measurements and not sufficient to bias first order paleocurrent patterns that we observe.

In L1, paleoflow directions are dominantly ~ESE in the lower ~400 m of the stratigraphic section (Figure 3.3). Between 400 and 600 m, paleocurrents were either ESE or WNW, but above 600 m, paleocurrent directions return to ESE in L1 and L2. In L3, paleocurrents flow NNE, indicating a ~90° paleocurrent shift between the lower and upper sections, which is significant even within the uncertainties on our data.

3.4.3. Clast provenance

In an effort to place further constraints on the relationship between basin evolution in southern Gonghe and the GNS, we conducted clast counts at 4 different sites within L3 in southern Gonghe, and within a lag deposit on one modern Yellow River terrace (Figure3.3). At each site, we counted ~100 clasts. In general, the clast composition of L3 is fairly monotonous—most sites are dominated by sandstone and quartz. Two sites exhibit a much more diverse assemblage of clasts compared to the others. The first is a terrace a lag deposit on a modern Yellow River terrace near Gonghe city. The second is the highest measurementsite from intact
fill in southern Gonghe, ~100 m below the top of the basin (at 358 m). The sites contain a lower overall percentage of sandstone and quartz, and they contain a high percentage of granite (compared to a few percent granite at other sites), as well as a variety of rocks such as volcanic rocks and green and red matrix supported pebble conglomerates with well-rounded clasts. The southern Gonghe site marks the base of a pronounced transition in the clast composition of L3. Below this horizon, clast composition is similar to other L3 outcrops around southern Gonghe and Tongde. Above this horizon, however, the clast assemblage becomes much more diverse and the clasts comprise a higher percentage of granite as well as significant amounts of other rock types.

3.4.4. Stratigraphic proxies for GNS range growth

The combination of evolving depositional environments and shifting sediment dispersal directions in southern Gonghe suggests that the GNS emerged during sediment accumulation in the foreland basin. The entire late Cenozoic stratigraphic package in southern Gonghe exhibits a coarsening upward sequence, suggesting the development of high topography adjacent to southern Gonghe during basin filling (Figure 3.3). In particular, the prominent unconformity separating L2 and L3 and the onset of alluvial fan sedimentation above this unconformity, along both the northern and southern flank of the GNS, appear to be directly linked to the emergence of the range. The pronounced paleocurrent shift between L1/L2 and L3 also suggests an important paleogeographic transition during foreland basin sedimentation. Although many mountainous regions are characterized by strike parallel sediment transport in the adjacent foreland basin (e.g. Bentham et al., 1992; Heermance et al., 2007), transverse rivers are typically located kms away from the rangefront. The ESE-oriented paleocurrents in L2 and L3 derive from a structurally high position on the northern flank of the GNS and suggest the range post-dates deposition of
these units. Moreover, the shift to NNE paleocurrents during L3 deposition suggests that the unit accumulated in the shadow of a topographic high to the south.

The provenance of conglomerate clasts in L3 is also consistent with syntectonic deposition. Lower strata within L3 are dominated by sandstone and quartz which are common to the Triassic flysch strata that comprise the core of the GNS. Importantly, the Triassic rocks are regionally ubiquitous (e.g. QBGMR, 1991), such that their presence in L3 strata does not uniquely link the deposits to the GNS. Nonetheless, in the context of the evolving depositional environments and paleocurrents in southern Gonghe, the composition of the clasts in L3 gravels lends an additional measure of confidence to the interpretation of syntectonic sedimentation. The appearance of a diverse assemblage of clasts, which mirrors the composition of Yellow River terrace lags, suggests that an exotic sediment source was involved in basin filling, and suggests that the upper Yellow River antecedes the development of the GNS in southern Gonghe. We explore this idea further in section 3.8.

3.5. Structural evidence for growth of the GNS

In an effort to directly link the evolution of the Gonghe basin to the GNS, we conducted detailed geologic mapping along the well-exposed northern and southern flanks of the range, in the Yellow River canyon. Specifically, we augment existing 1:200,000 geologic maps (QBGMR, 1991) with a) detailed, 1:24,000-scale mapping at key sites along basin margins where strata are highly deformed (Figures 3.4, 3.5) and b) reconnaissance level mapping in the relatively undeformed expanses between the fault networks. Our mapping indicates that the broad expanses of central Gonghe and Tongde are mostly undeformed, so we reserve detailed structural descriptions for the highly deformed basin margins.
At the junction of the western and eastern GNS ranges, deep levels of the Gonghe basin fill are exposed in the Yellow River canyon, and these outcrops provide excellent control on the structural architecture of the range (Figures 3.4, 3.5). In the canyon on the southern side of the range, the range-bounding fault outcrops and patchy exposures of L1 red beds line the southern range front (Figure 3.2). The L1 red beds are overturned and dip ~30° ENE. The L1 outcrops merge along strike with discontinuous patches of black and grey sandstones and mudstone, which we tentatively correlate to L1 on the basis of their continuity with L1 strata.

In the western Yellow River canyon wall, L3 conglomerate beds are in unconformable contact with the highly deformed metasedimentary rocks that core the GNS. Within this canyon wall, L3 strata progressively shallow upward along the ~150 m high canyon wall; the beds are subvertical just above the basal unconformity and they are subhorizontal at the basin surface (Figures 3.5b). Moreover, in this outcrop, L3 exhibits prominent intraformational unconformities which can be traced into correlative conformities only a few hundred meters to the south. Importantly, the duration of the unconformity diminishes to the south, orthogonal to the GNS range front. The fanning pattern of L3 strata suggests rotation of the northern margin of Tongde basin during sediment accumulation in the basin. The intraformational unconformities within L3 unequivocally demonstrate that beds within the unit were uplifted, rotated and truncated, and then covered by additional L3 strata. Given that L3 strata can be traced directly into southern Gonghe through the structural saddle in the central portion of the GNS (Figures 3.2, 3.4), the age of L3 in southern Gonghe should correspond to the timing of GNS fault growth.

In the canyon on the north side of the range, mapping along multiple structural transects indicates that the two stratigraphic units, L1 and L2, are folded into a 20 - 30° NNE dip panel (Figure 3.4). In contrast, L3 exhibits relatively shallow dips in this region, from ~0 - 15°.
Importantly, immediately north of the small village of Yangqu, a 450 m thick exposure of L3 gravels crops out on the western Yellow River canyon wall. Bedding dips in this outcrop progressively shallow upward in this canyon wall; they dip 11° NE at the base of canyon and 0° at the top (Figures 3.5c). This package of strata with fanning dips corresponds to the section of L3 strata that exhibit a pronounced stratigraphic transition to alluvial fan deposition and a ~90° paleocurrent shift. Furthermore, at several outcrops in this area L3 is in angular unconformable contact underlying L1 and L2 deposits, and the unconformity diminishes moving away from the GNS (Figures 3.5d, e). Similar to the southern GNS front, the fanning L3 strata and the diminishing unconformity between L3 both indicate tilting of the GNS during L3 deposition, and possibly during the late stages of L2 deposition. This interpretation is further supported by the correspondence between stratigraphic proxies for range growth and fanning dips on the northern side of the range.

South of the village of Yangqu, L1 and L2 strata define a NW-plunging anticline, which tips out only a few km to the west (Figure 3.4). The trace of this inferred fault coincides with a prominent topographic break in the central GNS, and remnants of L1 or L2 strata are preserved in the structural and topographic saddle in the middle of the range (QBGMR, 1991) (Figure 3.2), suggesting the presence of a range-scale thrust fault in the interior portion of the GNS.

The western GNS overlaps the eastern part of the range in the Yellow River canyon area (Figure 3.4). Near the small village of Jiala, the fault outcrops and dips 27° N. North of Jiala, the western range is covered by basin fill, but ~10 km to the east, the eastern terminus of the range outcrops in the deepest part of the Yellow River canyon (Figures 3.2, 3.4). At both sites, L3 deposits bury the western GNS range front, indicating that, at least locally, rates of deposition exceed fault slip.
3.6. Chronology of basin fill

In order to place absolute constraints on the timing of the GNS on the basis of linkages between foreland basin strata and the range, we employ a combination of regional lithostratigraphic correlation, magnetostratigraphy, and cosmogenic burial age dating. Two stratigraphic sections in the Gonghe basin complex suggest that L3 strata in southern Gonghe are Plio-Quaternary in age. First, L3 can be traced into relatively fine-grained strata in central Gonghe which contain abundant Plio-Quaternary fossils (Zheng et al., 1985). Second, L3 conglomerates on the margin of the GNS can be traced southward into a stratigraphic section in central Tongde that was dated by biostratigraphy, cosmogenic burial age dating, and magnetostratigraphy to ~3.3 – 0.5 Ma (Craddock et al., 2010). We seek to improve upon this regional L3 chronology, and to constrain the history of basin evolution in southern Gonghe by directly dating L3 growth strata, and the underlying units in southern Gonghe. Due to the notoriously low resolution of magnetostratigraphy derived from coarse-grained alluvium, we take the novel approach of complimenting the data with a suite of cosmogenic burial age dates.

3.6.1 Cosmogenic burial dating

In order to place absolute constraints on the age of the L3 gravels in southern Gonghe, six new samples of coarse fluvial sand were collected from L3 exposures for analysis of in-situ cosmogenic $^{26}$Al and $^{10}$Be inventories (Figure3.2, Tables3.3) (Granger and Muzikar, 2001; Granger, 2006). Samples were collected from quartz-rich sand lenses intercalated with the fluvial gravel basin fill. Because the concentrations of cosmogenic $^{26}$Al and $^{10}$Be are strongly dependent on the history of post-burial production within ~10 m of the earth’s surface (Granger, 2001), we targeted samples from the base of modern roadcuts (Supplementary Figure 3.3), where we are able to geometrically constrain sample depth prior to historic road construction. Sampled
roadcut exposures are unweathered and clearly exhibit original sedimentary structures of the basin fill. In general, we selected samples that were buried by ~15 – 30 m of sediment prior to road construction. As a result, we are confident that the samples remained in-situ since the time of their deposition and were only recently exposed in the roadcuts.

Samples were collected from quartz rich sand lenses intercalated with the fluvial gravel basin fill at sites from 82 to 317 m below the top of the basin fill (Supplementary Information section 3). Samples were collected at the base of modern roadcuts that are sufficiently deep (>20 m) such that post-burial muonogenic production is minimal (see Supplementary Information section 3 for full site descriptions). Depth to the pre-road land surface was measured with a laser rangefinder and used in subsequent age calculations. Four of the samples were collected from road cuts located less than a kilometer from the upper stratigraphic section, only a few kilometers to the north of the GNS. As such, the samples are derived from syntectonic L3 strata and they are readily correlated into the southern Gonghe stratigraphic section. A fifth sample was gathered from a 5-m deep cave within the upper southern Gonghe stratigraphic section. A sixth sample was collected from a deep roadcut, about 300 m below the basin fill top, in south-central Gonghe about 20 km to the northeast of the GNS and the other cosmogenic samples (Figure 3.2, Table 3.3). Quartz from these samples was isolated and purified using physical and chemical techniques at the Purdue Rare Isotopes MEasurment (PRIME) lab, following standard procedures. Extraction of $^{10}$Be and $^{26}$Al was accomplished using cation-exchange chemistry, and the resultant isotope concentrations were measured on the accelerator mass spectrometer at PRIME, following standard protocols (see Supplementary Information section 3 for a complete description of laboratory methods).
For buried sediment that is derived from a steadily eroding source, the concentrations of $^{26}\text{Al}$ and $^{10}\text{Be}$ will evolve through time as a function of two unknown variables, the pre-burial cosmogenic inventory and the time since burial (Granger, 2006). Additionally, for samples exhumed to within a few tens of m of the surface, post-burial cosmogenic isotope production will occur (e.g. Granger and Smith, 2000; Wolkowinsky and Granger, 2004). In order to calculate burial ages, we assume that samples are instantaneously buried to their stratigraphic depth below the basin fill top. Given that sedimentation rates for Plio-Quaternary gravel sections from Tongde, Chaka, and Guide are $\sim$50-100 m/Ma (e.g. Craddock et al., 2010; Zheng et al., in review, Fang et al., 2005), ingrowth of cosmogenic isotopes during burial is negligible (see Craddock et al., 2010). Importantly, this scenario neglects post-burial production, which may be a reasonable assumption given that all samples were buried by 10 – 30 m prior to recent road construction. We model all reasonable combinations of pre-burial cosmogenic inventory and burial time and use a least squares optimization to determine a best fit burial age. Uncertainties on reported burial ages reflect both analytical uncertainty, and uncertainty related to the burial history of the sample. Recent application of burial age dating to correlative L3 deposits in central Tongde basin showed that muonogenic production following basin excavation by the Yellow River in the late Quaternary could potentially increase observed burial ages by as much as $\sim$30% (Craddock et al., 2010). A similar potential for post-exhumation muonogenic production exists in the Yellow River canyon of southern Gonghe, such that our ages should be considered as minimum bounds on the true depositional age of L3 strata by as much as a few hundred thousand years.

3.6.2. Magnetostratigraphy
In order to refine and expand our chronology for the basin fill in central Gonghe, we collected magnetostratigraphic samples along both stratigraphic sections in southern Gonghe basin (Figures 3.2, 3.4). The two sections were selected because they offer relatively continuous exposure of key stratigraphic intervals of the Gonghe basin fill, which is uncommon within Gonghe basin. Furthermore, the sections were uncomplicated structurally. These sections complement existing magnetostratigraphic sections from around Gonghe basin complex in central Tongde (Craddock et al., 2010; Chapter 5) and in Chaka basin (Zheng et al., in review).

As is typical of many stratigraphic and magnetostratigraphic studies, the sections represent composites of several shorter sections. We correlated subsections in the field, typically over distances on the order of ~100 m, and we measured and sampled overlapping stratigraphic intervals for all of the subsections. We sought to collect samples at 2-5 m intervals, in order to obtain a high resolution magnetostratigraphy. At each sample site, we collected 3-5 specimens. Most specimens were collected in-situ as 2.5-cm-diameter cores using a gas-powered drill, and in general, we sought to sample mudstones and siltstones. The lack of mudstones in the upper basin occasionally required the sampling of relatively friable fine and medium sandstones. Often, sandstones were too friable to sample by drill, and in these instances, we collected oriented block samples, approximately 125 cm$^3$ in volume. For the lower section, we sampled at a total of 218 sites. Given that the section is 670 m thick, our average sample spacing was 3.1 m. The upper section consists almost entirely of conglomerate and we exploited sandstone lenses for sampling. Therefore, our sample density was limited by the spacing of sandstone lenses, which in many instances are separated by several tens of meters. For the upper section, we attempted to collect at 45 sites, although the sandstone lenses were too friable to sample at seven sites. The upper ~40 m of section was impossible to sample because of a lack of sandstone lenses, meaning that our
magnetostratigraphy does not extend to the highest levels of Gonghe basin fill. In light of the difficulty associated with sampling the upper section, our sample mean sample spacing is 11.2 m.

Remnant magnetizations were measured using a 2G Enterprises DC SQuID three-axis cryogenic magnetometer housed in a magnetically shielded room at the California Institute of Technology. The magnetometer has a background noise of <1 pAm² and is equipped with computer-controlled AF demagnetization coils and an automated vacuum pick-and-put sample changer (Kirschvink et al., 2008). Thermal demagnetization was performed in a magnetically shielded ASCTM oven in a nitrogen atmosphere. We applied a combination of 6 alternating field (AF) steps and 15 thermal (TT) demagnetization steps during magnetic cleaning in order to isolate the characteristic remnant magnetization (ChRM) of each specimen, and least-squares, principal component analysis was conducted in order to measure the ChRM directions (Kirschvink, 1980; Jones, 2002). Magnetic cleaning was performed on 1 specimen from each sample site, and on a second specimen from 20 sites. The AF regimen contained six equal steps between 0 and 126 gauss, and was designed to remove any low coercivity magnetization. TT steps were closely spaced immediately below the unblocking temperatures of magnetite (570 °C) and hematite (680 °C), and more sparsely distributed at lower temperature ranges. For 20 sites that exhibited low unblocking temperatures (ca. 350 °C), we conducted magnetic cleaning on duplicate specimens, using a scheme five AF steps, up to 100 guass, and 15 TT steps below 570 °C. Importantly, the polarity for all sites with duplicate specimens was identical between specimens. Additionally, we conducted a full battery of rock magnetic experiments, including isothermal remnant magnetization (IRM) acquisition, IRM backfield demagnetization, and
anhydric remnant magnetization (ARM) acquisition routines, on a subset of 19 samples. All field and laboratory techniques are described in detail in Supplementary Information section4.

3.6.3. Results

3.6.3a Burial ages

Calculated burial ages from the near the stratigraphic section are in remarkable stratigraphic order (Figures 3.6, 3.7). Furthermore, by linking the burial age samples to the upper stratigraphic section in southern Gonghe, the burial ages provide a loose constraint on the deposition age of L3. The highest sample, which is 82 m below the modern basin fill top, yields a best fit age of 0.5 +/-0.3 Ma. The age is consistent with recent magnetostratigraphy, biostratigraphy, and cosmogenic burial dating in central Tongde basin, which shows that basin filling persisted until ~0.5 Ma (Craddock et al., 2010). Only ~19 m below the upper sample, sample CT7-11 yields a significantly older burial age, of ~2.5 Ma. Due to the fact that this sample was buried to ~5 m, and therefore poorly shielded, we place a low degree of confidence in this particular burial age. Progressively deeper samples from southern Gonghe yield burial ages of 3.0 +/-2.0 Ma (187 m depth), and 3.9 +/- 1.3 Ma (271 m depth). The deepest sample, from 310 m below the fill top, yields an age of 5.8 +/- 0.5 Ma. In light of the half lives of $^{10}$Be and $^{26}$Al, we consider any best fit burial age greater than 5 Ma to indicate only that the sample was buried prior to 5 Ma. As such, the deepest sample should be considered only a minimum bound on the age of sediment burial. Collectively, the burial age samples suggest that in southern Gonghe, the upper part of L3 dates to ~>5 – 0.5 Ma. Extrapolating the burial ages to the base of the L3 deposits suggests the onset of L3 deposition occurred locally at ~6-7 Ma.

Sample CT7-11 was collected ~20 km to the northeast of the others, in the south-central part of Gonghe basin (Figure 3.2). The sample was collected 301 m below the basin fill top, and it
yields a burial age of 3.8 +/- 0.6 Ma. Given that the sample located in the middle part of the Plio-
Quaternary gravel section, the sample implies that the onset of L3 deposition was fairly
synchronous across southern Gonghe, at least over length scales of a few tens of km.

3.6.3b Rock magnetism

Based on several observations from demagnetization and rock magnetic experiments, we
subdivide our paleomagnetic specimens into three basic categories, which reflect the magnetic
mineralogy of the specimens. We interpret the samples to contain a diverse assemblage of
magnetic minerals, including hematite, magnetite, titanomagnetite and/or maghemite and
goethite in varying proportions. Although specimens appear to contain diverse assemblages of
magnetic minerals, the measured natural remnant magnetism of the samples was typically
restricted to $\sim 10^5 - 10^6$ emu cm$^{-3}$.

For many specimens (Type 1), hematite appears to be the dominant carrier of the ChRM
(Figure 3.8). Although alternating fields of up to 1T are insufficient to remove the NRM of all of
the specimens (see Supplementary Information section 4), these specimens preserved a relatively
high percentage of their NRM during AF demagnetization (e.g specimen 509.2, Figure 3.8).
Likewise, these samples do not reach saturation during IRM acquisition up to 1000 G. In general,
inability to clean the NRM in high alternating fields indicates the presence of antiferromagnetic
minerals, such as hematite or goethite, thereby restricting the possible magnetic mineralogy of
these specimens. For these specimens, the NRM was not appreciably diminished by AF steps up
to 126 G followed by thermal steps up to 570 (e.g specimen 509.1, Figure 3.10). Magnetization
was abruptly removed between TT steps 570 to 680 °C. Given that goethite has an unblocking
temperature of $\sim 150^\circ$ C and hematite has an unblocking temperature of $\sim 680^\circ$ C, hematite is
most likely to carry the ChRM. Commonly, these specimens display stable magnetization
directions on Zijderveld and equal area plots to thermal steps between 570 and 680 °C. Moreover, the low coercivity overprint that these samples exhibit is typically removed at low AF steps (below 126 G), suggesting that the overprint is carried by low coercivity magnetic minerals, such as magnetite, titanomagnetite, and/or maghemite. For a subset of type 1 specimens, the magnetization is carried dominantly by magnetite, rather than hematite (specimen 918.1, Figure 3.8). Compared to the specimens described above, AF demagnetization up to 126 G removes a relatively high proportion of the NRM, and the specimens tend to show that a high percentage of the NRM is removed between 500 and 570 °C. Furthermore, the specimens often exhibit unstable behavior above thermal steps of 555-570 °C. Similar to other type 1 specimens, alternating fields of 60-100 mT or temperatures of 70-150 degrees typically remove a viscous overprint from these samples, suggesting that low coercivity minerals, such as magnetite and/or goethite to carry the VRM.

The ChRM of a second type of specimens (Type 2) appears to be carried by titanomagnetite and/or maghemite (Figure 3.8). These samples exhibit much lower coercivities during AF demagnetization in fields up to 1T and during IRM acquisition in fields up to 1T, suggesting a relatively low proportion of antiferromagnetic minerals (e.g. sample 702, Figure 3.8). During thermal demagnetization, these specimens exhibit a pronounced drop in magnetization at ~200 – 350 °C (Figure 3.8), which is similar to the unblocking temperature of titanomagnetite and maghemite (~200 - 350° C). Often, this type of specimen displays unstable demagnetization behavior at thermal steps at or above 200 - 350 °C (e.g. sample 702, Figure 3.8), suggesting that minerals with relatively high unblocking temperatures, such as magnetite and/or hematite, do not carry the ChRM. Importantly, this type of samples exhibits a pronounced drop in the NRM at low AF levels, indicating that the minerals that carry the ChRM also carry a
viscous overprint. Therefore, the stability spectrum of the VRM partially or completely overlaps the ChRM. In light of this observation, the ChRM of Type 2 specimens which exhibited overlapping stability spectra was determined using the magnetization circle method described by McFadden and McIlhenny (1988).

To more closely examine the demagnetization behavior of samples exhibiting overlapping coercivity spectra for the ChRM and the VRM, we conducted an additional set of thermal demagnetization experiments, with a high concentration of thermal steps between 70 and 350 degrees. Every one of these samples indicates 1) stable, interpretable magnetization directions below 200 - 350 ºC, 2) a sharp drop in magnetization around 350 ºC, and 3) unstable directions above 200 - 350ºC (see specimen 702.4, Figure 3.8). The clear demagnetization behavior of these samples substantiates our inference that the ChRM of these specimens is carried by titanomagnetite and/or maghemite, and that the stability spectra of the ChRM and VRM should overlap for these samples.

We did not attempt to interpret the polarity for a small subset of specimens (Type 3) for two reasons. 1) The stepwise demagnetization behavior for a few sites was too erratic to interpret for some sites. 2) The magnetic inclination and the magnetic declination suggested opposing polarities for some sites. In sum, 14% of specimens from the lower section were uninterpretable and 13% of specimens from the upper section were uninterpretable. The proportion of uninterpretable specimens is fairly typical for magnetostratigraphic analysis of terrestrial strata.

3.6.3c Magnetostratigraphy of upper section

The magnetostratigraphy for the upper section exhibits 7 normal polarity magnetozones and 6 reverse polarity magnetozones (Figure 3.7). Due to the thick gaps between suitable sample sites from this section, several magnetozones are defined by a single site. 33 interpretable sites
define the magnetostratigraphy, and an additional 5 sites were uninterpretable (Type 3). In geographic coordinates the mean normal direction has a declination and inclination of 4.4 and 46.4 (k = 8.47), and the mean reverse direction has a declination and inclination of 150.6 and -44.6 (k = 5.15) (Figure 3.9). The section is only slightly tilted, and only at its base, so the mean normal and reverse direction declination and inclination are similar, 5.2, 40.0, and 153.9, -40.2 respectively. Although several magnetostratigraphic studies discard single site polarity reversals, we retained such sites in our magnetostratigraphies because coarse-grained stratigraphic intervals, particularly in the upper section, were sparsely sampled. The section narrowly fails the reverse test, although this is probably an artifact of the low number of reversed polarity sites causing an imprecise determination of the mean reverse polarity direction.

### 3.6.3d Magnetostratigraphy of lower section

The magnetostratigraphy from the lower section exhibits 26 normal polarity zones and 25 reversed polarity zones. The magnetozones appear robust in that they are typically defined by several sites (Figure 3.7). 188 interpretable sites from the lower section define the magnetostratigraphy. 30 additional sites (Type 3) did not yield interpretable polarities. In geographic coordinates, the mean normal direction has a declination of 310.1 and an inclination of 63.3 (k = 6.6) (Figure 3.9). The mean reversed direction has a declination of 147.3 and an inclination of -55.3 (k = 4.91). In tilt corrected coordinates, the mean normal and reverse polarity directions are 359.3, 51.6 (k = 6.5), and 179.0, -41.4 (k = 4.31), respectively. In tilt corrected coordinates, the lower section passes an indeterminate reversal test, such that the groups would not share a common mean at the 75.22 % confidence level (see Supplementary Information section 4). A similar result is obtained in geographic coordinates. To assess the potential value of additional sampling, we apply a jackknife test to the lower section (Figure 3.10). The lower
section has a jackknife statistic of -0.49, which indicates that our sample density recovers at least 95% of the actual magnetozones, and that additional sampling would not reveal a significant number of additional polarity reversals (Tauxe and Gallet, 1991). We note that the inclusion of single site polarity reversals in our magnetostratigraphy lowers our jackknife statistic. We did not perform the jackknife test on the upper section because the stratigraphic character of the section prohibits additional sampling.

3.7. Tectonic history of southern Gonghe

3.7.1. Correlation of magnetostratigraphy to the GPTS

Although the magnetostratigraphy from the upper section has a low resolution, the combination of burial ages and regional biostratigraphy indicates that the sediments date to the Plio-Pleistocene and provides the basis for correlating the section to the GPTS (Figure 3.7). In particular, the two deepest burial age samples, which date to ~4-5 Ma and correlate to the section at ~150 m height suggest that the basal, normal polarity-dominated interval correlates to the upper part of Chron 3, which also dates to ~4-5 Ma. This correlation implies that locally, gravel deposition in southern Gonghe began at ~7 Ma, during Chron 3a. The shallowest burial age sample dates to ~0.5 Ma, and suggests that, like other sections from the region, the top of the section correlates to the upper part of Chron 1. The remaining polarity zones in the magnetostratigraphy in the upper part of the Gonghe section were correlated to the GPTS in a way that minimizes erratic fluctuations in sediment accumulation rates and honors thick zones of normal or reversed polarity. Given these guidelines, the thick reversed polarity magnetozones in the middle of the section appear to correlate to the long reversed polarity intervals at the base of Chron 1, Cron 2 and Chron 2a. The well-ordered stratigraphic succession of burial ages indicates that they are reliable as guides for our magnetostratigraphic correlation.
Although the lower section must be older than the overlying L3 deposits, we lack information that directly anchors the correlation to a particular interval of the GPTS. On a regional scale, stratigraphic similarity to Guide basin suggests that the deepest sediments are likely to date to the Neogene (Fang et al., 2005). Given that the average duration of Neogene polarity chron is ~0.25 Ma, and that we observe 25 normal and 26 reversed polarity intervals, we sought correlations that span ~13 Ma (e.g. Lowrie and Kent, 2004). Due to the fact that we incorporated single site polarity reversals in our magnetostratigraphy, we accepted correlations that implied deposition over slightly shorter time scales as well. Given the lack of local constraints on the depositional age of the lower levels of southern Gonghe, we also sought correlations to older intervals of the GPTS (Figure 3.7). Importantly, the mean chron duration for the Paleocene and Eocene is longer than remainder of the Tertiary, such that we considered longer windows of sediment accumulation for correlations extending into the early Tertiary (Lowrie and Kent, 2004).

We identified two possible correlations to the GPTS, one that spans much of the Miocene and another that spans much of the late Paleogene (Figure 3.7). In the Miocene correlation, we link the magnetostratigraphy to the GPTS from C6A to C4. This correlation implies a depositional age of 19.722 - ~8 Ma, a duration of ~12 Ma. The correlation implies relatively steady sedimentation rates of ~56 m/Ma. A slight variation of this correlation, in which the long normal polarity zone at the top of our section corresponds to the long normal polarity subchron in C5, is also possible. In this scenario, the top of the section dates to 9.987 Ma, which implies a mean sedimentation rate of ~70 m/Ma and sediment accumulation over ~10 Ma. In the Paleogene correlation, we match our magnetostratigraphy to the portion of the GPTS that spans C20 to C9. This implies a depositional age of 42.774 to 26.714 Ma and the persistence of
sediment accumulation for ~16 Ma. The implied mean sedimentation rate is ~42 m/Ma. The two long reversed polarity zones near the base of the section and the long normal polarity zone at the top of the section appear to prohibit additional alternative correlations. Although both the middle and late Tertiary correlations to the GPTS may be reasonable, multiple lines of evidence indicate that the Miocene correlation is robust and preferable to the late Paleogene option.

1) Although a progressive unconformity separates the lower and upper section, this contact becomes conformable in the basin center. A similar transition, from fluvial sandstones and mudstones to gravels, is visible in the center of Tongde basin (Craddock et al., 2010). Here, abundant evidence, including fossils and burial ages (Zheng et al., 1985, Craddock et al., 2010; Chapter 5) indicates that the base of L3 is mid-Pliocene, ca. 3.3 Ma. Specifically, a Pliocene fossil was identified just below this stratigraphic transition. The conformable contact between implies relatively steady sedimentation leading up to gravel deposition, and therefore, that the younger correlation is better.

2) The Miocene correlation implies that sediment accumulation rates were relatively steady, with no major permanent changes in sedimentation rates. Importantly, correlation 1a implies a duration of sediment accumulation of ~12 Ma, which is more consistent with the expected duration of ~13 Ma, than the duration implied by correlation 1b. In contrast, the late Paleogene correlation requires pronounced and erratic fluctuations in sediment accumulation rates, particularly between C18 and C17 and C13 and C11. Although the Paleogene correlation requires sediment accumulation for ~16 Ma, this may be reasonable given that the mean duration of polarity chron is longer in the Paleogene than in the Neogene (Lowrie and Kent, 2004).

3) Although anomalous polarity zones exist for both correlations, they are less abundant, and defined by fewer sites in the Miocene correlation. Anomalous polarity chron for the
Miocene correlation are typically defined by a single site, whereas five out of eight of the anomalous polarity chronis are defined by multiple sites in the middle Tertiary correlation. Interestingly, the anomalous polarity zones in the Miocene correlation are nearly identical to anomalous polarity zones observed in eastern Qaidam (Fang et al., 2007). Both correlations also omit polarity chronis recorded in the GPTS. In general, due to the depositional hiatuses and minor unconformities that are characteristic of fluvial deposits, magnetostratigraphies of such deposits often omit some short polarity chronis. The Miocene correlation omits 3 - 5 polarity microchrons (duration of ~0.1 Ma), which appears to be reasonable given the unsteady nature of sedimentation in fluvial settings. The middle Tertiary correlation omits no polarity chronis, which is to be expected given that the duration of polarity chronis is relatively long in the Paleocene and Eocene (Lowrie and Kent, 2004).

3.7.2 Timing of growth strata and Initiation of the GNS

Structural and stratigraphic observations from the Yellow River canyon in southern Gonghe are internally consistent, and they indicate the emergence of a high, eroding source terrane due to contractional deformation on the GNS fault network well after sediment accumulation began in the region. Our basin fill chronology suggests the pre-GNS L1 and L2 deposition prior to 10 Ma, and synorogenic, L3 deposition since ≥7 Ma, thereby bracketing the timing of range growth. Below, we discuss the three key lines of evidence for the initiation of the GNS during the late Miocene, after L1 and L2 deposition, and during L3 deposition.

First, paleocurrent measurements from within the alluvial fan strata on the north side of the range indicate a distinct change in paleoflow directions between the end of L2 deposition and the beginning of L3 deposition. During deposition of L1 and L2, a variety of paleoflow directions are recorded. Given that fluvial depositional systems often exhibit a wide array of
paleoflow directions because of the meandering nature of channels, the variety of paleoflow directions is not surprising. On the whole, however, L1 and L2 strata contain a majority of ~SE, paleocurrent indicators, suggesting the dominant early Miocene paleoflow direction was parallel to the strike of the GNS (Figure 3.4). In contrast, the overlying strata are dominated by NE-directed paleoflow indicators, suggesting flow away from the present day range. The transition from range parallel to basinward paleocurrents suggests the development of a topographic high to the south of the measured stratigraphic sections during the time spanned by the unconformity that separates the lower and upper sections.

Second, lithostratigraphic analysis of southern Gonghe basin suggests the development of an alluvial fan complex along the northern margin of the GNS during the late Miocene. Early and middle Miocene strata adjacent to the modern GNS are relatively fine grained, and suggest that southern Gonghe was not proximal to a sediment source at this time. A distinct coarsening upward sequence that begins ~11 – 13 Ma, and perhaps more importantly, an alluvial fan complex develops during the late Miocene unconformity that separates the lower and upper sections. The coarsening upward sequence in the mid-Miocene could be attributed to a variety of factors, and it is difficult to link specifically to the emergence of the GNS. However, the presence of an alluvial fan along the northern flank of the GNS after ~7 Ma, seems to indicate the presence of a proximal sediment source by this time, similar to the patterns of paleoflow.

Third, and most importantly, the structural architecture of the southern Gonghe basin unequivocally demonstrates the growth of the GNS during L3 deposition and after L1 and lower L2 deposition. L1 and the oldest preserved L2 sediments are uniformly folded over the GNS, indicating that the two units pre-date the emergence of the range (Figure 3.4). In contrast, the overlying gravels exhibit fanning dips that show progressive rotation of the northern limb of the
GNS starting at ~7 Ma. Furthermore, the progressive unconformity, which diminishes basinward, away from the GNS, requires the onset of GNS mountain building between the age of the youngest preserved L2 strata and the oldest preserved L3 strata.

3.7.3 Timing and source of gravel deposition around Gonghe basin complex

Although global climatic instability has been invoked to explain the deposition of broad sheets of conglomerate in basins around central Eurasia (e.g. Zhang et al., 2001), the time transgressive nature of gravel deposition in Gonghe suggests a local tectonic control on conglomerate deposition. The oldest burial ages from southern Gonghe are ≥5 Ma, and in tandem with our magnetostratigraphy, they imply the gravel deposition began at ~7 Ma locally. In contrast to southern Gonghe, burial ages, magnetostratigraphy and fossils from the lowermost gravel deposits in central Tongde basin date to 3.3 Ma. The differences between southern Gonghe and central Tongde suggest that conglomerate deposition initiated earlier on the flanks of the GNS than it did in the more distal sections of the basin. Importantly, sediments from ~301 m below the basin top, ~20 km north of the GNS exhibit mid-Pliocene burial ages, and also imply latest Miocene or early Pliocene gravel deposition. Although this is similar to what is observed along the northern range front, broad uplift in the hangingwall side of the GNS fault network may have limited rates of accommodation over large length scales, leading to widespread deposition of coarse-grained sediments (e.g. Burbank et al., 1988).

Although the volume of sediment in southern Gonghe appears to be large in comparison to the amount of erosion implied by our structural cross sections, conglomerate clast counts imply that a Neogene precursor to the upper Yellow River may have contributed a significant amount of sediment to the basin. Whereas bedrock exposed in the ranges that surround southern Gonghe and Tongde comprises weakly metamorphosed sedimentary rocks, dominantly slate and
sandstone/quartzite (QBGMR, 1991), many of the Pliocene gravel deposits adjacent to the modern Yellow River display a wide variety of clast compositions. A local source for many of these clasts cannot be identified. The close spatial association of these deposits with the present day Yellow River implies that the upper part of the river delivered exotic sediment to the basin, at least during the Pliocene. Later, in the late Pleistocene, the Neogene upper Yellow River was involved in an upstream sweeping wave of headward erosion and basin integration that swept across much of northeastern Tibet and established external drainage in many of the basins in the region (Craddock et al., 2010).

3.7.4 Techniques for dating syntectonic strata in young foreland basins

Magnetostratigraphy is frequently the only geochronologic technique that is available for dating coarse-grained, non-marine strata in intramontane foreland basins (e.g. Burbank and Johnson, 1983; Bullen et al., 2001; Charreau et al., 2008). This approach can be hindered in at least two ways. First, many intramontane sedimentary basins are characterized by relatively coarse-grained strata, such that the technique is often challenging to apply. Moreover, proximal strata, which often preserve the most accurate archive of the timing of structural development along a basin margin, are often the coarsest strata in a basin (e.g. Burbank and Raynolds, 1984). Second, independent archives of the depositional age of syntectonic strata, such as faunal assemblages or volcanic ash horizons are often absent, or characterized by low precision. We have demonstrated that by complimenting magnetostratigraphy with cosmogenic burial age dating, the confidence and resolution underlying foreland basin chronologies for faulting may be greatly enhanced.

3.8 Tectonic implications for northeastern Tibetan plateau

3.8.1 Initiation of basins interior to northeastern Tibet
A fundamental result of this study is that similar to other basins within the interior of northeastern Tibet, southern Gonghe basin initiated in the early Miocene, at ~20 Ma. This timing is remarkably similar to the development of Guide basin at 20 Ma and northeastern Qaidam at ~16 Ma (Fang et al., 2003, 2005). However, not all basins interior to northeastern Tibet initiated in the Early Miocene; sedimentation in the northwestern part of Gonghe does not appear to begin until ~12 Ma. Moreover, east of Guide, the Xunhua basin initiated at ca. 29 Ma (Lease et al., in review; Hough et al., in press). Overall, it appears that beginning in the early Oligocene, a network of disconnected basins began to develop around the interior portion of northeastern Tibet in an asynchronous fashion. Importantly, the lack of preserved early Cenozoic sediments in southern Gonghe contrasts with basins closer to the margin of northeastern Tibet, where clear evidence shows steady sediment accumulation during the Cretaceous and Tertiary (e.g. Horton et al., 2004; Dai et al., 2006). The contrasting histories of basin development suggest that any early Tertiary mountain building in northeastern Tibet may have been confined to the more exterior portions of the region (e.g. Dupont-Nivet et al., 2008a, b; Clark et al., 2010). As such, it appears that early Tertiary tectonic elements in the western Qaidam region and in the Xining Lanzhou region evolved as separate entities, until Qaidam basin expanded eastward in the Miocene (e.g. Métivier et al., 1998; Yin et al., 2008b) and middle Tertiary basins developed in interior northeastern Tibet.

### 3.8.2 Cenozoic range growth in NE Tibet

Another key finding of this study is that similar to other narrow mountain ranges that transect interior northeastern Tibet, contractional deformation and topographic growth of the GNS began at 7 – 10 Ma (Figure 3.11). This timing is similar to what has been observed along the northwestern margin of greater Gonghe, in the vicinity of Chaka. There, stratigraphic
archives and progressively tilted strata suggest that the NW Qinghai Nan Shan emerged along the margin of Chaka subbasin at ~6 – 8 Ma (Zhang et al., in review). A few tens of kilometers to the west, comparison of Quaternary fault slip rates to displaced ancient geologic markers has been used to suggest a similar late Miocene initiation age for the right lateral Ela Shan fault, although large uncertainties are inherent to this analysis (Yuan et al., 2011). Ranges bounding nearby basins record a similar episode of deformation. Sediment accumulation curves from NW Qaidam suggest renewed mountain building along the basin margin at ~8 Ma and provenance analysis of detrital zircons suggests emergence of the Laji Shan along the northern margin of Guide basin at ca. 8 Ma. This episode of late Miocene mountain building was not restricted to the interior of northeastern Tibet. Mineral cooling ages and stratigraphic archives record accelerated exhumation within the northern Qilian Shan and the Liupan Shan, between ~8 – 10 Ma (Bovet and Ritts, 2009; Zheng et al., 2006, 2010; Jolivet et al., 2002). In general, the late Miocene appears to herald a distinct paleogeographic change across northeastern Tibet, from broad, unbroken foreland basins, to compartmentalized foreland basins developing within a growing chain of mountain belts along the plateau perimeter (Figure 3.11). This transition may be somewhat analogous to the transition from the Sevier to the Laramide Orogeny along the eastern flank of the North American Cordillera (e.g. DeCelles et al., 2004). During the early Cenozoic Laramide orogeny, basement-cored mountain ranges emerged from previously vast and unbroken regions of sediment accumulation within the Sevier foreland basin. Thus, Gonghe basin illuminates an episode of intensified mid-late Miocene tectonism in northeastern Tibet, in which segmentation of late Cenozoic sedimentary basins interior to the region by narrow, elongate mountain ranges was accompanied by growth of ranges along the periphery of the Tibetan plateau, several hundreds of kilometers away.
The Late Miocene kinematic history of northeastern Tibet, in which mountain building occurred synchronously over vast spatial scales on the order of 100’s – 1000’s of km, places important limits on interpretations of the geodynamic processes that drive plateau growth in the region. First and foremost, end-member geodynamic models that invoke crustal thickening as the key process for building high topography predict progressive outward plateau growth (e.g. Clark and Royden, 2000; Tapponnier et al., 2001). Given synchronous deformation from southern Gonghe to the plateau margins, progressive outward growth cannot resolved at the scale of the entire northeastern Tibetan plateau. However, new, or renewed, contractional tectonism across the entirety of northeastern Tibet is consistent with geodynamic models that invoke a pulse-like episode of mountain building associated with convective removal of the mantle lithosphere and increased topographic stresses on the plateau foreland. Alternatively, the kinematic history of northeastern may be more complicated than the framework provided by any simple, end-member geodynamic model. For example, recent basin analysis suggests that Qilian Shan has expanded outward from a once narrow core, since the late Oligocene (e.g. Bovet et al., 2009). Late Miocene mountain building along the northern plateau margin and along the margins of Gonghe may simply reflect the expansion of the broader Qilian Shan. Regardless of whether the late Cenozoic history of northeastern Tibet is interpreted as pulse-like, or part of some more complex deformational episode, the timing of mountain building in the GNS provides a fundamental constraint on viable kinematic and dynamic interpretations for the northeastern Tibetan plateau by suggesting that plateau growth did not occur in a progressive outward fashion.

3.9 Conclusions

The stratigraphic record preserved in the Gonghe basin complex sheds important light on the Cenozoic geologic evolution of the interior part of northeastern Tibet, in the transitional
region between the high core of the Tibetan plateau and the eastern Qilian Shan and outer northeastern Tibet.

1. Late Cenozoic sedimentation initiated in the Gonghe region at ca. 20 Ma, and occurred at rates of ~50 – 70 m/Ma. This timing is similar to adjacent basins from interior northeastern Tibet, particularly Guide basin to the east, which initiated at ~21 Ma. In contrast, sedimentation near the exterior part of the northeastern Tibetan plateau has been persistent and slow throughout the late Cretaceous and Cenozoic. Any early Tertiary deformation in northern Tibet appears to be confined to the exterior parts of the region, in the Xining-Lanzhou-Linxia region and in western Qaidam. Intervening regions do not contain evidence for such an episode of deformation, even along the broad structural grain that links the two regions of early Tertiary deformation.

2. Evolving depositional environments, changing paleocurrents, foreland basin structural architecture, and the presence of growth strata and progressive unconformities reveal the emergence of the GNS, which transect southern Gonghe basin, at 7 - 10 Ma. This is part of a broader episode of late Miocene deformation, in which narrow, elongate mountain ranges emerged from the broad foreland that extended across interior northeastern Tibet, and compartmentalized the region into many smaller basins. The timing of compartmentalization is similar to the timing of range growth around the peripheries of northeastern Tibet (Zheng et al., 2006) the Qilian Shan (Jolivet et al., 2002; Bovet and Ritts, 2009; Zheng et al., 2010), and eastern Tibet (Kirby et al., 2002; Clark et al., 2010), suggesting an acceleration in contractional tectonism during the late-Miocene. Although this kinematic history cannot yet be reconciled with any one simple end-member geodynamic model for the expansion of high topography in Tibet, it appears to be consistent with models invoking regionally synchronous deformation.
3. Plio-Quaternary gravel deposition was time-transgressive around the Gonghe basin complex of northeastern Tibet. In Gonghe basin, structural and stratigraphic relationships tie conglomerate deposition to the rise of basin-bounding mountain ranges. Enhanced climate variability since the onset of icehouse conditions in the Pliocene does not explain the widespread gravel sheets that blanket Gonghe basin.

4. The resolution and quality of intramontane forelandbasin chronologies may be greatly enhanced by complementing magnetostratigraphy with cosmogenic burial age dating, particularly in settings lacking independent age archives contained within faunal assemblages or volcanic ashes. The application of magnetostratigraphy may be extended to thick successions of gravel found along faulted basin margins. Given the time-transgressive nature of conglomerate deposition in foreland basins, this stands to greatly enhance chronologic frameworks for basin margin faulting.

3.10 Supplementary Information

Section 1-Field photos of Cenozoic lithofacies in central Gonghe basin

Lithofacies 1, 1a, 2, and 3 are shown in Supplementary Figure 3.1.

Section 2-Lithostratigraphy and biostratigraphy of additional geologic units

*Facies J1-grey slate pebble conglomerate interbedded with red mudstone*- Discontinuous outcrops of facies 1 floor the basin near the town of Yangqu.

Facies J1 is composed of dark red mudstone with grey lenses of pebble conglomerate. The conglomerate lenses comprise angular-subangular slate clasts, and a quartz sand matrix. They display imbrication, and they are typically clast supported. The beds contain small, silty or fine sand lenses, that are ~3-4 cm thick and a few tens of cm wide. Near the base of the unit, gravel lenses are 2-15 cm thick and laterally continuous over 1 to a few m. Upsection, the gravel
beds amalgamate. The mudstone comprises silty clay, and the beds appear to be internally structureless. Beds are laterally continuous, a few tens of cm thick, and the sand lenses are inset in the muddy matrix. Although most of the beds are dark red, near the base of the unit, mud beds may be grey, buff or red.

The presence of imbrications and the lenticular geometry of the beds indicates that facies 1 was deposited by a channelized unidirectional flow. The red color of the mudstone may indicate subaerial exposure of muddy units, if the oxidation of the sediments is a primary feature. We interpret the beds to be fluvial-floodplain deposits, such that the gravel lenses are channel deposits and the mud beds are overbank deposits.

_Facies J2-grey, greenish grey, red and buff banded mudstone_- Facies 2 sits conformably atop facies J1. Like facies J1, it only outcrops in discontinuous patches in the southern part of Gonghe basin, near the town of Yangqu.

Facies J2 is composed of bands of grey, greenish grey, tan, brown and maroon silty clay mudstone, interbedded with a few lenses of buff colored, fine-medium sandstone. The mud beds are tabular, and laterally continuous. Each is a few tens of cm to a few m thick. They exhibit maroon or orange mottling. One ~0.8 m thick coal bed is present. Sandstone lenses exhibit planar parallel laminations, trough cross stratification and planar cross stratification. Sandstone beds are 0.1 – 0.5 m thick, and laterally continuous over tens of m.

The fine grain size and lack of sedimentary structures of facies J2 indicate deposition in relatively still water. The green and grey color of the many of the muds suggest subaqueous deposition. The red muds suggest periods of subaerial exposure. Trough cross-stratification and lenticular geometry of the sandstone lenses indicates that they were deposited by a
unidirectional, laterally confined current. We interpret these to be interbedded fluvial (red muds, sandstones lenses) and lacustrine (grey green massive muds) deposits.

In order to develop a chronology for J1 and J2 deposits, we collected pollen samples from a coal bed at the base of J2 deposits, found 74 m above the basin floor, and about 100 m below the Cenozoic section. The bed contains a rich, well-preserved pollen assemblage, with a relatively low diversity of species. Although many of the species are long-ranging, from the Jurassic to the Cretaceous, the abundant Corollinas and *Quadraeculina anellaeformis* indicate the basal deposits are Early Jurassic in age. The coordinates of the pollen sample site are 35.70496° N, 100.16672° E, 3099 m.

**Section 3-burial dating methods**

**A. Field methods**

To constrain the burial age of fluvial sediment from the Tongde basin, we collected 5 new samples of coarse fluvial sand or gravel for analysis of in-situ, cosmogenically produced $^{26}$Al and $^{10}$Be inventories in quartz (Granger and Muzikar, 2001; Granger, 2006). 4 of these samples were derived from the Yellow River canyon immediately north of the Gonghe Nan Shan (sample locations shown in Figure3.2 and Supplementary Figure 3.2), and a 5th was collected ~20 km north of the range (Figure3.4). Because the concentrations of cosmogenic $^{26}$Al and $^{10}$Be are strongly dependent on the history of post-burial muonogenic isotope production within ~10 m of the earth’s surface, we targeted samples from the base of modern roadcuts (Supplementary Figure3.3), where we are able to geometrically constrain sample depth prior to historic road construction. Sampled roadcut exposures are unweathered and clearly exhibit original sedimentary structures of the basin fill. As a result, we are confident that the samples
remained in-situ since the time of their deposition and were only recently exposed in the roadcuts.

**B. Laboratory techniques**

Samples were subjected to several physical and chemical treatments designed to purify the raw material to pure quartz. First, samples were crushed and sieved, in order to obtain a desirable grain size for the remaining treatments. In order to remove carbonates and minor metals, the crushed material was leached in nitric acid and aqua regia. Next, the sample was subjected to a suite of physical separation steps which were: froth flotation, magnetic separation, and following a purification bath in a hydrofluoric acid/nitric acid solution, heavy liquid separation. The remaining material was soaked for a second time in a hydrofluoric acid/nitric acid solution to remove any remaining feldspars. During this step, the outermost layers of the quartz grains were dissolved to remove meteoric $^{10}$Be. After completing this routine, Al concentrations were measured on an inductively coupled plasma optical emissions spectrometer (ICP-OES) to assess the purity of the remaining quartz. If the measured Al concentration (which signifies the presence of residual feldspars) exceeded 200 ppm, the final step was repeated as necessary.

In order to extract Be and Al isotopes from the purified quartz samples, a second series of chemical treatments was applied. After adding Be and Al carriers, quartz was dissolved in concentrated hydrofluoric acid. Following dissolution, an Al aliquot was extracted from the solution and prepared for precise measurement on the ICP-OES. The volume of the solution containing the dissolved sample was reduced and the hydrofluoric acid was removed by a series of evaporation and fuming steps. The residual material was taken up in a sodium hydroxide solution, centrifuged, and decanted in order to separate Fe and Ti ions from the solid residual
Next the pH of the remaining solution was adjusted to ~8 to precipitate the Al and Be out of the solution as hydroxides (Ochs and Ivy-Ochs, 1997). After dissolving the remaining hydroxides in oxalic acid, cation and anion columns were used to removed residual Na, Fe, and other undesired ions, and to isolate Be and Al. The samples were dried and fired in an oven, and then loaded into a cathode for accelerator mass spectrometry (AMS). AMS was conducted at PRIME lab at Purdue University, following standard protocols.

Section 4-Supplementary paleomagnetic information

A. Field methods

For both sections, 3-4 specimens were collected from each bed, using a gas-powered drill with a 2.5 cm diameter core bit. The core-plate orientation was measured using a magnetic compass. Bedding dip of sites was also measured using a magnetic compass. In certain stratigraphic intervals, the sediment was too friable to be sampled with a drill. At these sites, oriented block samples ~2 cm³ in volume, were collected. These samples were carved into smaller cores in the laboratory.

B. Laboratory techniques

Magnetic cleaning was conducted at The California Institute of Technology. Magnetization was measured using a three-axis DCSQUID moment magnetometer in a magnetically shielded μ-metal room. The background noise of the magnetometer is <1 pA m². It is equipped with a vacuum pick and put, computer-controlled sample changing system. After measuring the natural remnant magnetism of each specimen, most samples were subjected to up to 20 steps of alternating field (AF) and thermal (TT) demagnetization. Five or six evenly spaced AF steps, between 0 and 100 or 120 gauss were first applied in order to remove low coercivity viscous remanent magnetization (VRM). AF demagnetization was conducted with a computer-
controlled, three-axis coil system. Following that, thermal steps at 70, 150, 250, 350, 450, 500, 530, 555, 570, 600, 635, 655, 670, and 680 °C were applied. Thermal demagnetization was conducted with a commercially built, magnetically shielded furnace. The treatment was designed to have a high density of steps leading up to the unblocking temperatures of magnetite and hematite (570 and 680 °C, respectively) For the lower section, 238 specimens were demagnetized, one from each of the 218 sites, and duplicate specimens from 20 sites. The duplicate measurements were made in order to obtain a higher density of thermal steps for sites with overlapping characteristic remnant magnetization (ChRM) and VRM (see below). For these samples, we implemented thermal steps at 70, 100, 120, 150, 190, 230, 260, 290, 320, 350, 425, 500, 530, 550, 560, 570, 615, and 665 °C. If the magnetization measurement following an AF or TT step yielded a circular standard deviation (CSD) of >15°, the measurement was repeated up to two times. If a CSD of ≤15° could not be obtained, the data for the AF/TI step were discarded.

ChRM directions were determined using a least-squares fit, principal component analysis (Kirschvink, 1980; Jones, 2002). In general, we sought to perform the least squares, principal component analysis on ≥ 4 TT/AF steps. Commonly, more steps were incorporated, however, in a few cases, only three TT/AF steps were used. For magnetization components that were believed to be characteristic, regression included the origin and were forced through the origin. We consider mean angular deviations for regression that exceed 15° to indicate a poor quality regression, although this did not apply to any of our interpretable samples. Although some authors filter data with VGP latitudes of <30° (19 sites in the lower section), we choose not to do so simply because our magnetostratigraphy is unaffected by applying this filter. The inclusion of these low VGP latitude sites has no effect on our magnetostratigraphy. If brief polarity reversals occurred where segments overlapped, the stratigraphic height of specimens from a segment was
adjusted by up to 30 cm in order to eliminate reversals. This correction was applied to three samples. In order to apply this correction, the slight adjustment in height was forbidden if it altered the stratigraphic order of samples within a single segment. Because some stratigraphic intervals were only sparsely sampled, we did not reject single site reversals, although this filter is sometimes applied to other magnetostratigraphic data. After determining the orientation of the ChRM for each sample, the declination and inclination were used to calculate the VGP position. Northern hemisphere poles were assigned normal polarity and southern hemisphere poles were assigned reversed polarity.

A small number of samples were subjected to a full battery of rock magnetic experiments. First, the NRM of samples was removed in alternating fields up to 1000 G. Subsequently, isothermal viscous remanent magnetizations (IRM) and anhysteric remanent magnetizations (ARM) were imparted and removed by IRM backfields and/or by AF cleaning.

C. Rock Magnetism: Alternating field removal and acquisition of magnetization

Supplementary Figure 3.4 shows the end-member NRM-removal and IRM-acquisition behaviors in alternating fields for samples subjected to the full battery of rock magnetic experiments. Samples that retain a relatively high proportion of NRM in an alternating field, and that develop a relatively small IRM during IRM acquisition are interpreted to have a high proportion of hematite.

D. Reversal test

Lower section- In tilt corrected coordinates, the lower section passes an indeterminate reversal test, such that the groups would not share a common mean at the 75.22 % confidence level. The critical angle, or the angle at which the two distributions become significantly different is 23.3°, and the angular difference between the mean normal polarity direction and the
antipode of the mean reversed polarity direction is $10.19^\circ$. The results indicate that the mean normal polarity and reversed polarity directions are antipodal, but also that data are widely scattered around mean polarity directions. The boundary between a “C-quality” reversal test and an indeterminate reversal test is a critical angle of $20^\circ$ (McFadden and McElhinny, 1990), so the section nearly passes a C-quality reversal test.

*Upper section*-The upper section fails the reversal test. The critical angle is $17.27^\circ$ and the difference between the mean normal polarity direction and the antipode of the mean reverse polarity direction is $23.82^\circ$. Given that this section comprises only 33 samples, we speculate that additional samples would better define mean polarity directions, and that this section would pass a reverse test given better defined mean polarity. In particular, the mean reversed polarity direction is much different from the mean normal direction, as well as the mean normal and reverse polarity direction for the lower section.
Figure 3.1. a) Quaternary faults and Cenozoic basins in northern Tibet. Inset shows GTOPO-30 digital topography of Tibetan plateau and Quaternary faults, adapted from Tapponnier and Molnar, 1977; Molnar and Tapponier, 1978. Grey dashed lines are terrane boundaries. JS = Jinsha suture, AS = Anyemaqen suture, SQS = South Qilian suture, DHS = Danghe Nan Shan suture, NQS = North Qilian suture, NCS = North China suture. Adapted from Yin and Harrison, 2000 and Xiao et al., 2009 and references therein. b and c) Maximum, minimum and mean swath topography, derived from GTOPO-30 data, which has a nominal resolution of 1 km. Moho depths are also shown. For b, moho depths are from Liu et al., 2006. For c, moho depths from the Anyemaqen and Gonghe are from Vergne et al., 2002, and depths from the Qilian Shan are from Meyer et al., 1998.
Figure 3.2. Geology and topography of the Gonghe Nan Shan region. Geologic map is draped by a hillshade image generated from 90-m SRTM digital topography. Open circles show burial age sites. Sites labeled P exhibit progressive unconformities, sites labeled G exhibit growth strata, sites labeled A exhibit angular unconformities, and sites labeled O exhibit an onlapping relationship between basin fill and bedrock.
Figure 3.3. Lithostratigraphy, paleocurrents, and clast composition of lower and upper stratigraphic sections in southern Gonghe. In rose diagrams, the radius is equal to the number of measurements in the largest bin. The petal width is 20°. S = sandstone, Q = quartz, G = granite, X = no clast, V = volcanics, T = schist, M = marble, C = matrix supported conglomerate with green or red matrix and rounded clasts, L = shale, R = chert. Locations of sections shown in Figure 3.4.
Figure 3.4. Detailed map of the GNS in the Yellow River canyon area.
Figure 3.5. Geologic cross section and supporting photographs. Location of cross section shown on figure 4. Relationships between basin strata and faults. b) Growth strata and progressive unconformity on the southern flank of the Gonghe Nan Shan in the Yellow River canyon. Cliff is ~200 m high. c) Growth strata on the northern flank of the Gonghe Nan Shan, in the Yellow River canyon. Cliff is ~500 m high. d and e) Progressive unconformity on the northern flank of the Gonghe Nan Shan near the Yellow River canyon. C is located near the base of the lower stratigraphic section in southern Gonghe, whereas d is located at the base of the upper stratigraphic section, such that the unconformity between L3 and underlying rocks diminishes away from the GNS.
Figure 3.6. Banana diagrams and contour plots of chi-squared statistics for combinations of burial age and erosion rate. Both burial age and erosion rate misfits are normalized such that an error of 1 is the largest error for a model iteration. Each contour represents a 1-order of magnitude decrease in the size of the chi squared statistic. Reported uncertainites represent an average of upper and lower bounds on burial ages/erosion rates.
Figure 3.7 Possible correlations of observed magnetostratigraphy to the Geomagnetic Polarity Time Scale (GPTS) of Ogg and Smith (2004). Cryoconths are included in the GPTS. Age in Ma is shown to the right of the GPTS and chron number is shown to the left of the GPTS.
Figure 3.8 Example orthogonal vector, equal area, and magnetization intensity plots for representative samples from the southern Gonghe stratigraphic sections.
Figure 3.9. Equal area plots for lower and upper sections in southern Gonghe, showing a, b) declination and inclination of all specimens in geographic coordinates, c, d) declination and inclination for all specimens in tilt-corrected coordinates, and e, f) mean normal and reversed polarity directions in tilt-corrected coordinates. Stars in e, f indicate present day declination and inclination at sample sites.
Figure 3.10. Jackknife test for lower Yangqu section. The jackknife statistic, or the slope of the regression line, is better than the recommended limit of -0.5., indicating that additional sampling would not significantly increase the observed number of polarity zones.
Figure 3.11. Distribution of late Tertiary deformation around northeastern Tibet. 1 = Zheng et al., in press; 2 = Zheng et al., 2006; 3 = Wang et al., in review; 4 = Lease et al., in press; 5 = Zhang et al., in review; 6 = Fang et al., 2007; 7 = Lease et al., 2007; 8 = Yuan et al., in press; 9 = This study.
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<th>Facies</th>
<th>Structures</th>
<th>Interpretation</th>
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<tr>
<td>Fl</td>
<td>Paleosol, calcite cemented sandstone or mudstone</td>
<td>massive or fine lamination</td>
<td>suspension settling in overbank deposits</td>
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<td>P</td>
<td>Pedogenic features</td>
<td>soil formation on floodplain</td>
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<td>Sr</td>
<td>v. fine to coarse-grained sandstone</td>
<td>ripple crosslamination</td>
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<td>St</td>
<td>v. fine- to coarse-grained sandstone</td>
<td>solitary or grouped trough crossbeds</td>
<td>sinuosly crested and linguoid 3-D dunes</td>
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<tr>
<td>Sp</td>
<td>v. fine to coarse grained sandstone</td>
<td>planar stratification</td>
<td>upper plane bed</td>
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<td>Cci</td>
<td>conglomerate, stratified, clast supported</td>
<td>imbrication</td>
<td>longitudinal bedforms, lag deposits</td>
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Table 3.1. Lithofacies and interpretations used in this study. Modified after Miall, 2000.
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Table 3.2. Lithostratigraphic units and associated facies codes.
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<th>longitude(°)</th>
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<th>depth (m)</th>
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<th>1σ</th>
<th>$^{26}$Al (atom/ g qtz)</th>
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Table 3.3. Burial age sample information.
Supplementary Figure 3.1. Stratigraphic units in lower levels of southern Gonghe subbasin. a, b) Lithostratigraphic units 1. c,d) Lithostratigraphic unit 1a. e,f) Lithostratigraphic unit 2. g, h) Lithostratigraphic unit 3. Cliffs in b, d, and f are on the order of ~200 m high.
Supplementary Figure 3.2. Cosmogenic burial age sample sites. Sample locations marked with open stars.
Supplementary Figure 3.3. Schematic diagram showing depth of L3 burial age samples.
Supplementary Figure 3.4. Alternating field removal of NRM and IRM acquisition curves. Y-axis represents the proportion of the total NRM, or the Saturation IRM, respectively.
CHAPTER 4

TIMING, MAGNITUDE, AND RATE OF UPPER CRUSTAL SHORTENING IN THE GONGHE BASIN REGION OF INTERIOR NORTHEASTERN TIBET

4.1. Abstract

Reconstructing the kinematic history of the Tibetan plateau is central to ongoing debates over the geodynamics of continental plateau growth and the manner in which the growth of high topography shapes the earth’s climate. For the broad northeastern margin of the plateau, disagreement exists over the timing of plateau growth and the magnitude and rates of contractual deformation. We combine detailed structural mapping of foreland basin strata and analysis of digital topography in order to reconstruct the deformational history of two major fault networks in the interior of the region, the Qinghai Nan Shan (QNS) and the Gonghe Nan Shan (GNS). Balanced cross-sections suggest ~1 - 2 km and 6 - 8 km on the QNS and GNS, respectively, and growth strata in the Gong He basin imply that this shortening initiated during the Late Miocene. (U-Th)/He and apatite fission track thermochrology indicate slow cooling rates on the order of ~1°C/Ma in the QNS and GNS during the Late Cretaceous and Paleogene and limit the magnitude of Late Cenozoic exhumation. We estimate mean shortening rates across the QNS since the initiation of the fault network in the late Miocene, and we compare these estimates to new late-Quaternary shortening rates derived from high resolution topographic surveys and cosmogenic dating of a deformed alluvial fan surface. Shortening rates across the QNS have likely remained steady since the Miocene, on the order of 0.1 – 0.4 mm/yr. Finally, we integrate detailed mapping of the QNS and GNS with various geologic, geophysical, and geomorphic observations from the surrounding region in order to construct and restore a regional geologic cross section across northeastern Tibet. This exercise reveals limited upper crustal
shortening since the late Miocene, on the order of ~4%, such that the budget of upper crustal shortening is apparently insufficient to account for the topographic growth of the region.

4.2. Introduction

The evolution of high topography within the Indo-Asian collision zone lies at the center of ongoing debates over the manner in which processes in the lithosphere build high topography at earth’s surface and shape global climate (e.g. Royden et al., 2008; Molnar et al., 2010 and references therein). For example, contrasting views of the geodynamic mechanisms underlying growth of the Tibetan plateau invoke drastically different patterns of plateau expansion (e.g. Tapponnier et al., 1982; England and Houseman, 1986; Molnar et al., 1993; Royden et al., 1997). Broadly, these competing geodynamic models for plateau growth can be divided into two end-member categories. One holds that high topography expanded progressively outward through time in response to crustal thickening (e.g. England and Houseman, 1987; Royden et al., 1997; Clark and Royden, 2000; Tapponnier et al., 2001; Medvedev and Beaumont, 2006). The opposing view is that convective removal of the mantle lithosphere changed the buoyancy of the continental crust beneath Tibet. In this view, the kinematic history of the plateau is marked by a punctuated change in surface elevation and rapid outward expansion corresponding to the timing of this event (e.g. Molnar et al., 1993). The elevation history of Tibet has also been linked global environmental change. First, enhanced weathering of silicate minerals and enhanced burial of organic carbon related to the rise of Tibet during the Cenozoic are thought to drawdown atmospheric CO$_2$ and drive global cooling (e.g. Raymo et al., 1988; Zachos and Kump, 2005). Second, Tibet is thought to influence patterns of northern hemisphere atmospheric circulation and climate through its role as a large scale obstruction to the jetstream (Boos and Kuang, 2010; Molnar et al., 2010).
Despite the importance of the topographic evolution of the Tibetan plateau, the timing, magnitude, and style of plateau growth remains only loosely bracketed across vast regions. Most workers envisage that prior to the onset of collisional orogenesis, thick crust and corresponding high topography extended across southern Tibet (England and Searle, 1986; Murphy et al., 1997; Kapp et al., 2003). Although the height and areal extent of this high topography is not clear, most workers hold that southern and central Tibet were at elevations similar to the present during the Cretaceous and the early Tertiary (e.g. Kapp et al., 2005; Rowley and Currie, 2006; DeCelles, 2007), and that the marginal regions lay several km lower than today (e.g. Kapp et al., 2005; Cyr et al., 2005) (Figure 4.1). Although outward expansion is commonly thought to occur in the Miocene (Tapponnier et al., 2001; Molnar, 2005; Royden et al., 2008), the precise patterns of outward expansion remain unclear. Moreover, several recent studies challenge the view of Miocene plateau expansion by suggesting that outward growth occurred, or initiated, during the Eocene (e.g. Dupont-Nivet et al., 2008a). Thus, a key goal of current geologic research in Tibet is to define the patterns of outward plateau expansion, particularly along the plateau margins, where geodynamic models make differing kinematic predictions.

Given that competing mechanical models for plateau growth invoke differing mechanisms for building and supporting high topography, a second key goal of geologic research in Tibet is reconstructing the evolution of fault networks across the orogen. Plate reconstructions indicate 2300-3500 km of total convergence between India and Eurasia since the time of collision (Molnar and Stock, 2009; Dupont-Nivet et al., 2010), and anywhere from 500 – 2500 km of this budget has been accommodated by contractional deformation across 1000s of km of southern Eurasia (Chen et al., 1993; Dupont-Nivet et al., 2010; Liebke et al., 2010, also see DeCelles et al., 2002 and references therein). Although regional shortening budgets are emerging
for many regions (e.g. Murphy et al., 1997; Kapp et al., 2003; Taylor et al., 2003; Yin et al., 2008), entire physiographic provinces of the plateau have not been analyzed. Moreover, by reconstructing the evolution and architecture of major fault networks, important constraints may be placed on the mechanical structure of the Tibetan crust (e.g. Masek et al., 1993).

The northeastern margin of the Tibetan plateau represents an important region to assess the mechanisms and timing for the growth of high topography. Whereas clear evidence indicates Miocene contractional deformation and attendant basin formation in the region (e.g. Fang et al., 2005; Zheng et al., 2006; Lease et al., 2007; Hough et al., 2011), much of the evidence that underpins suggestions of significant Eocene contractional deformation >3000 km north of the Indo-Eurasian collision zone derives from this region (e.g. Dupont-Nivet et al., 2008a,b; Clark et al., 2010; Dayem et al., 2010 and references therein). Specifically, basins and ranges near the exterior of region contain archives that indicate slow, continuous late Cretaceous and Tertiary subsidence (Horton et al., 2004; Dai et al., 2006), and vertical axis rotations of sedimentary basins and localized thrust faulting from 50-40 Ma (Dupont-Nivet et al., 2008; Clark et al., 2010). A similar episode of early Tertiary deformation is recorded in the vicinity of western Qaidam (e.g. Yin et al., 2002, 2008). Outer northeastern Tibet and western Qaidam are linked by an elongate, WNW-striking structural trend, but little evidence bears on the kinematic history of the interior portion of this structural grain. Although the region is transected by several elongate, narrow mountain ranges that are flanked by well exposed foreland basins, virtually no analysis bears on the architecture of major fault networks in the region, or the amount of shortening they have accommodated.

In this study we investigate the timing, magnitude, and style of Cenozoic upper crustal deformation and basin evolution in the vicinity of the Gonghe basin complex in the interior
portion of northeastern Tibet (Figure 4.1). The basin complex is overridden at its northern and southern margins by the Qinghai Nan Shan (QNS) and the Gonghe Nan Shan (GNS) respectively, two narrow elongate mountain ranges related to south-vergent networks of imbricate thrust faults (Figure 4.2). The ranges merge along strike with sites with evidence for Paleogene contractional tectonism, such that the study area constitutes a key site to evaluate the extent of early Paleogene contractional tectonism (e.g. Fang et al., 2003; Clark et al., 2010; Duvall et al., in review; Lease et al., in review). We organize our study around three key research problems. First, in order to address the timing of foreland basin development in interior northeastern Tibet, we augment well-dated, stratigraphic sections with reconnaissance level stratigraphic observations from around the basin complex. The observations provide the basis for a regional synthesis of the depositional history of the region that bears on the timing of plateau growth in the region. Second, in order to address the timing, style, and magnitude of contractional deformation along the faulted margins of the basin complex, we construct serial cross sections through the QNS and the GNS. These cross-sections incorporate data derived from mapping of foreland basin strata, regional geomorphic observations, and thermochronologic constraints on the development of structural relief during the Cenozoic. Third, in order to assess temporal changes in rates of contractional deformation, we determine Late Pleistocene rates of deformation along the Qinghai Nan Shan derived from cosmogenic dating of a deformed alluvial fan surface. We compare the late Quaternary shortening rates to geologic shortening rates derived from line-length shortening measurements and the initiation age of the QNS. In the context of northeastern Tibet as a whole, our study places important kinematic constraints on the geologic evolution of one of the broad marginal regions of the Tibetan plateau.

4.3 Tectonic setting of the Gonghe basin complex
Northeastern Tibet, the area bounded to the south by the Kunlun fault and to the west by the Qilian Shan-Qaidam region, sits between two disparate physiographic and geologic terranes, which appear to relate to variations in upper crustal structure and may represent differing mechanisms of plateau growth (e.g. Tapponnier et al., 1990; Burchfiel et al., 1995; Meyer et al., 1998; Clark and Royden, 2000) (Figure 4.1). To the south is the broad, flat interior of the Tibetan plateau, where the lack of short wavelength topography is interpreted to reflect weak lower crust (e.g. Fielding et al., 1994). Eastward flow of material in the lower crust beneath this region is thought to be responsible for building the 3-km topographic escarpment along the eastern and southeastern edges of Tibet (e.g. Burchfiel et al., 1995; Royden et al., 1997; Clark and Royden, 2000; Kirby et al., 2002; Clark et al., 2005). To the west of northeastern Tibet is the Qilian Shan, a mountainous region which extends over an area on the order of $\sim10^5$ km$^2$ (Figure 1.1). In contrast to the low wavelength, low amplitude topography of interior Tibet, the Qilian Shan contains closely spaced, narrow, fault bounded mountain ranges and intervening sedimentary basins. Regional budgets of upper crustal shortening suggest that high topography is compensated by thickened upper crust (e.g. Meyer et al., 1998), a markedly different mode of plateau growth than the terrain to the south.

Due to inadequate study of vast tracts of the region, the tectonic evolution of northeastern Tibet remains less clear than adjacent the adjacent provinces. The region constitutes a broad, topographic ramp, over which mean topography and crustal thickness decrease from the interior plateau to the foreland (e.g. Clark and Royden, 2000), and key changes in crustal thickness and mean elevation are correlated to the position of major geologic terrane boundaries (Figure 4.1). The southernmost boundary is the Kunlun-Anyemaqen-Qaidam Suture (KAQS), which separates the Songpan-Ganzi terrane from the Southern Qilian Shan block. In interior northeastern Tibet,
crustal thickness decreases from 60-65 km south of the KAQS to ~50-55 km (Vergne et al., 2002; Zhang et al., 2010—Figure 4.1). A corresponding decrease in mean topography occurs across this boundary, from high plateau elevations of ~4500m, to elevations of ~3500m in interior northeastern Tibet. Both crustal thickness and mean topography decrease again, along the physiographic margin of the plateau. Across the northern Qilian Shan and the northern Liupan Shan, which sit at the northern and northeastern edge of the region respectively, crustal thickness decreases to ~40-45 km and mean topography decreases to elevations of ~1500 m north and east of the plateau.

The Gonghe basin complex sits astride the northern flank of the physiographic transition between the high and greater northeastern Tibet (Figure 4.1). The northern margin of the basin complex roughly corresponds with a second, major Paleozoic suture, called the South Qilian Suture (SQS), which separates the South Qilian Shan arc from the Kunlun-Qaidam block. Much of the basin complex is floored by the multiply deformed and weakly metamorphosed Triassic Songpan-Ganzi flysch (Figure 4.2). The rocks extend to the south and cover much of central Tibet; they are thought to represent a remnant ocean basin (Weislogel, 2008). North of the suture zone, the QNS mountain range is cored by Paleozoic and Mesozoic plutonic rocks, and Paleozoic sedimentary strata (QBGMR, 1990).

Cenozoic sedimentation appears to begin diachronously across the basin complex, beginning at ca. 20 Ma in southern Gonghe and at ca. 15 Ma in the northwestern Chaka subbasin. To the east, sedimentation began across Guide basin at ca. 20 Ma (Fang et al., 2005), and to the west, sedimentation initiated in northeastern Qaidam at ca. 15 Ma (Fang et al., 2007). The growth of the bounding mountain ranges appears to lag the onset of sediment deposition in region by several millions of years. In southern Gonghe, progressive unconformities, growth
strata, and several stratigraphic proxies indicate the emergence of the GNS from amidst a broader Gonghe basin complex at 7 – 10 Ma (Craddock et al., in review, Chapter 3). Along the northwestern flank of the basin, stratigraphic archives and possibly structural relationships suggest the emergence of the NW QNS at 6 – 8 Ma (Zhang et al., in review). Moreover, a comparison of late Quaternary slip rates to offset ancient geologic markers suggests that the right lateral Ela Shan fault system activated in the late Miocene (Yuan et al., 2010). The presence of deformed, late Quaternary alluvium along the southern flanks of both ranges, and historic earthquakes associated with the GNS, attest to recent and even ongoing mountain building in both ranges. To the east, the QNS merge with the Laji Shan, and detrital zircon provenance studies in Guide basin suggest the emergence of the western part of the range at ~8 Ma (Fang et al., 2005; Lease et al., 2007). To the west, the QNS merge with the southern Qilian Shan, and analysis of sediment accumulation rates suggests renewed tectonism along the eastern portion of this range, during the middle and late Miocene (Fang et al., 2007).

4.4. Depositional history of the Gonghe foreland basin complex

The depositional history of the Gonghe basin complex may reveal critical information about the emergence of mountainous topography in interior northeastern Tibet. In order to construct a regional synthesis of the depositional history of the Gonghe, we augment existing stratigraphic observations from around the basin complex with new, reconnaissance-level stratigraphic characterization. Overall, we identify five distinct Cenozoic lithostratigraphic units (Figure 4.3). We briefly describe and interpret each, and we describe the distribution, thickness, and depositional age of each unit across the Gonghe basin complex. These observations provide the basis for constructing a regional model of the stratigraphic architecture of the Gonghe basin.
complex (Figure 4.4). For more detailed lithostratigraphic description, readers are referred to Zeng et al., 1985; Craddock et al., 2010, in review; Zhang et al., in review; and Chapters 3 and 5.

Unit M1-M1 consists of lenticular, trough cross-bedded siltstone and sandstone interbedded with brick red massive or planar parallel laminated mudstone (Craddock et al., in review; Chapter 3). The lenses are cm to ~1.5 m thick and continuous over 10s to 100s of m. Sandstone lenses may be spaced by a few cm to ~1.5 m, or they may be amalgamated. Mud prone intervals contain carbonate nodules and concretions, burrows, and roots, and they exhibit a prismatic texture. The muds exhibit red, grey, and white, laterally continuous bands with a high degree of calcite cementation. M1 strata accumulated in a fluvial/floodplain environment.

The only exposed M1 strata in Gonghe basin complex are located along the southern margin of Gonghe basin, where the unit floors the basin is folded over the GNS (Figures 4.2, 4.4). Locally, the unit is ~600 m thick, and magnetostratigraphic analysis of the strata suggest a depositional age of ~20 – 10 Ma (Craddock et al., in review; Chapter 3).

Unit M2-M2 comprises planar parallel laminated, orange and tan, sheet-like siltstone (Craddock et al., in review; Chapter 3). In places, the unit may also be grey or brownish- and/or greenish-gray. The siltstones are interbedded with a variety of coarser grained strata, including decimeter to meter thick, trough cross-bedded sandstone lenses, and meter thick, sheet-like orthoconglomerates. Red and white bands exhibiting a high degree of calcite cementation are often present in this unit. M2 strata were deposited in a fluvial/floodplain environment. M2 differs from M1 on the basis of 1) the orange color of the unit, 2) the thick interbedded conglomerate bands (mostly present in southern Gonghe, but not elsewhere), 3) a lack of brick red mud interbeds or mud partings, and 4) the lateral continuity of the beds.
M2 is commonly found in relatively deep exposures along the margins of the basin complex (Figures 4.2, 4.4). In southern Gonghe, the unit exhibits a gradational contact with underlying M1 (see Craddock et al., in review; Chapter 3). Locally, the deposits are interbedded with sheets of grey conglomerate and orange sandstone. The upper part of the unit is truncated by an erosional unconformity, but the stratigraphic thickness must be in excess of 100 m. Magnetostratigraphic analysis of the strata suggest a depositional age of ~10 – 8 Ma. M2 is also well exposed in the proximal footwall of the QNS, where it floors the northern part of Chaka subbasin (see Zhang et al., in review) (Figures 4.2-4.4). Locally, it is relatively fine grained, and beds may be buff, orange, yellow, greenish grey, or whitish tan. Magnetostratigraphic analysis of the strata suggest a depositional age of 11.6 – 8.6 Ma.

Hundreds of km to the east, we identified new M2 outcrops on the flanks of a deeply eroded, fault-bounded uplift to the east of Gonghe City (Figures 4.2, 4.4). The unit floors NE Gonghe basin. We did not measure a precise stratigraphic thickness, but the unit appears to be tens to ~100 m thick on the flanks of the fault bounded uplift. Locally, the interbedded coarse grained sandstones and gravels are relatively thin and lenticular in geometry. Coarse grained lenses may be cm to decimeters thick, and laterally continuous over 10s of m. The age of other M2 deposits suggest that the strata in NE Gonghe date to the Late Miocene. Moreover, M2 strata in NE Gonghe can be traced laterally into the Herjia formation of Guide basin, which was deposited in a fluvial/floodplain environment and dates to ~7.8 – 3.6 Ma (Fang et al., 2005) (Figure 4.2).

We also identified new exposures of M2 within central and southern Tongde basin (Figures 4.2, 4.4). Along the southern basin margin, the unit appears to floor the basin, although
exposure is limited. Given that the unit is poorly exposed, and truncated by an erosional unconformity along the southern margin of Tongde, the stratigraphic thickness is unknown.

*Unit PQcgl* - PQcgl is composed of brown and grey pebble or cobble orthoconglomerate with fine grained lenses (Craddock et al., in review; Chapter 3). Bedding is lenticular, laterally continuous over a few to 100s of m, and decimeters to meters thick. Conglomerates are typically clast supported, and clast sizes range from pebble to gravel to cobble around the Gonghe region. The clasts are unsorted to poorly sorted and subangular to subrounded. Sedimentary structures include imbrication, channel scours, large (10-100 cm high) cross bedding within gravel beds, and parallel or cross lamination in the fine grained lenses. PQcgl is interpreted to be a braided channel alluvial fan deposit (see Craddock et al., in review; Chapter 3).

Regional mapping indicates that PQcgl is ubiquitous along the fringes of the GNS (Figures 4.2, 4.4). In southern Gonghe, PQcgl is separated from underlying M2 by a progressive unconformity. The unit is ~450 m thick. A combination of cosmogenic burial ages and magnetostratigraphy suggests that in southern Gonghe, the unit dates to ~0.5 – 7 Ma (Craddock et al., in review; Chapter 3). The unit is also exposed in the Yellow River canyon on the southern flank of the GNS, where it is at least a few hundred meters thick and the basal part of the unit is not exposed. PQcgl can be traced within the Yellow River canyon for tens of km north and south of the GNS range front. Cosmogenic burial dating, magnetostratigraphy, and fossil assemblages suggest deposition during the Plio-Quaternary in central Gonghe and deposition between ~0.5 – 3.3 Ma in central Tongde (Craddock et al., 2010; in review, Chapters 3, 5).

PQcgl also fringes the entire southern margin of the QNS (Figures 4.2, 4.4). In Chaka subbasin, exposed PQcgl strata are 300 m thick, and the upper part of the unit is not exposed. It dates from 4.6 Ma to the present (Zhang et al., in review). Locally, the conglomerate beds occur
in 1-5 m thick intervals and are separated by 1-3 m thick, massive siltstone beds. We also
identified new exposures of PQ_{cgl} in various small exposures along the south-central QNS and
the southeast QNS. Notably, along the southern margin of the QNS, PQ_{cgl} deposits appear to be
confined to a narrow swath along the range front, and interfinger with fine grained deposits to
the south.

*Unit PQfl*-PQfl comprises buff colored trough cross bedded sandstone interbedded with
imbricated conglomerate lenses and mudstone (e.g. Zeng et al., 1985; Craddock et al., 2010).
Beds are lenticular, laterally continuous over 1-100s of m and 10 cm to 2 m thick. Sedimentary
structures include imbrication, trough cross bedding, parallel bedding and lamination, and
channel scours. Ginger colored concretions and calcite cemented horizons are commonly found
in this unit. PQfl is a fluvial floodplain deposit. We differentiate it from M1 and M2 because it is
stratigraphically higher and unoxidized.

PQfl interfingers with basin margin PQ_{cgl} deposits (Figure 4.4). It is widely distributed
across central Gonghe, and southern and central Tongde. For example, PQfl interfingers with
PQ_{cgl} in central Tongde, and in aggregate, it is ~100 m thick. Where the basal part of the unit is
exposed in the interior portions of Gonghe and Tongde, it is typically in conformable contact
with underlying M2. In Chaka subbasin, PQfl conformably overlie basal M2 deposits. Locally,
the unit is ~800 m thick. Magnetostratigraphy of the strata suggest deposition of these rocks from
~8.6-4.6 Ma.

We identified new exposures of PQfl in northeastern Gonghe, in relatively close
proximity to the QNS range (Figure 4.2, 4.4). Locally, the strata are relatively fine grained,
sheet-like and internally massive or planar parallel laminated. Although we have no direct age
constraints, association with other well-dated PQ strata suggest that they date to the Plio-
Quaternary.

*Unit PQ₁-PQ₃* comprises tabular, light tan or yellow, siltstone and sandstone. The deposits
are compositionally mature, and they are interbedded with light red or gray-green sandy muds
(e.g. Zeng et al., 1985). Sedimentary structures include parallel laminations, and asymmetrical
wave ripples, but most beds are massive. Beds are tabular, and range from a few to several tens
of cm thick. These strata are interpreted to be shallow, near-shore lacustrine deposits (Zeng et al.,
1985; Craddock et al., 2010; Chapter 5), and they are confined to central Gonghe basin, south of
Gonghe city.

**4.4.1 Architecture of Qinghai Lake basin**

Seismic reflection profiles reveal the presence of up to ~500 – 1000 m of Tertiary
sediment in Qinghai Lake basin (An et al., 2006). Due to the fact that almost no exposure of the
deep levels of basin fill is available for the Qinghai Lake basin to the north, we are unable to
obtain descriptions of these rocks. However, several seismic reflection profiles across the basin
reveal the presence of several packages of Tertiary strata. Importantly, the basin contains
evidence for a prominent regional unconformity between the upper levels of basin fill and the
lower levels. Strata that floor the basin are tilted gently northward in the vicinity of the QNS and
warped over a broad, bedrock high in the central part of the basin. Overlying units truncate these
deposits, and the upper part of the fill in Qinghai lake basin is generally flat-lying. We tentatively
correlate the strata overlying the unconformity to the PQ units in Gonghe basin, and the strata
underlying the unconformity, to the older, Miocene strata in Gonghe (Figure 4.4).

**4.4.2 Regional lithostratigraphic correlation and basin architecture**
By synthesizing the various observations presented above, we are able to construct a depositional model for the Gonghe basin complex, which reveals three stages of sedimentation in the basin (Figures 4.3, 4.4). First, during the early Miocene, fluvial floodplain deposition occurred with a confined area in the southern part of Gonghe basin. It appears that the initial, relatively confined early Miocene depocenters expanded outward during the middle Miocene. We consider the later episode, of spatially extensive sediment accumulation across the greater Gonghe region to mark a second, distinct episode of basin formation. From ~12 – 10 Ma, the area of sediment accumulation expanded to encompass the Chaka subbasin and northeastern Gonghe. This expansion is also reflected in an episode of fluvial floodplain deposition in Guide basin. Finally, starting as early as ~8.5 Ma, the stratigraphic character of Gonghe basin gradually changed as the QNS and GNS emerged along the margins of the basin complex. At this time, the fringes of the mountain ranges were marked by aggregated alluvial fans, which are laterally correlative with fluvial sandstones and shallow lacustrine mudstones in the distal parts of the basin. During this stage of basin evolution, the previously continuous region of sediment accumulation was segmented by the emerging topography along the basin margins.

4.5 Structural evolution of bounding ranges

In order to reconstruct the magnitude and style of shortening along the flanks of the Gonghe basin complex, we constructed serial cross sections through both of the ranges (Figures 4.5 – 4.8). In order to constrain the structural architecture of the ranges, we integrated geologic mapping of deformed strata along the flanks of the ranges, geomorphic observations, and thermochronologic constraints on the development of structural relief. In order to conduct detailed geologic mapping, we augment existing 1:200,000 geologic maps (QBGMR, 1991) with 1:24,000-scale mapping at key sites along basin margins where strata are highly deformed.
We employ 90-m Shuttle Radar Topography Mission (SRTM) digital topography and field observations in order to conduct geomorphic analysis of the two major fault networks in the Gonghe region, the QNS and GNS (Figures 4.5, 4.6). Thermochronologic constraints derive from a combination of Apatite Fission Track (AFT) and Apatite (U-Th)/He thermochronology.

4.5.1 Structural mapping

4.5.1a Qinghai Nan Shan Overview

The QNS, along the northern edge of Gonghe basin, comprises a series of range-scale asymmetric anticlines with broad, gently dipping north limbs, and steep, narrow south limbs (Figures 4.5, 4.7). The folds are related to a network of imbricate S-vergent thrust faults. The fault network merges along the strike of the SQS with the Laji Shan to the east and the southeastern Qilian Shan to the west (Figure 4.1). Along most of the QNS range front, the fault network is blind, although faults daylight and displace Quaternary alluvium in the proximal footwall of the northwestern QNS (Figure 4.5). Cenozoic strata are not present along the top of the range, and the QNS is cored by plutonic rocks of the Paleozoic South Qilian Shan arc (Yin and Harrison, 2000; Xiao et al., 2009).

4.5.1b Observations bearing on S-vergent structural asymmetry

Regional trends in the architecture of deformed foreland basin strata provide insight into the architecture of the range. Steeply tilted panels of M2 strata abut the southern front and indicate that the beds were uplifted and folded during range growth and subsequently stripped away by erosion (Figure 4.7). We describe the structure of beds at several locations along the southern range front, including the western QNS, the eastern QNS, and the fault bounded structural high at the southeastern terminus of the QNS. Although poor geologic exposure
prevents geologic mapping along the northern side of the QNS range, recent seismic surveys reveal the architecture of the deeper levels of the basin fill (An et al., 2006). The surveys show that the ~1 km thick package of strata dip gently to the north across the southern part of the basin, and the pattern suggests that the north limb of the QNS extends far to the north, beneath the Qinghai Lake basin. The thick dip panels on either side of the QNS indicate that strata predate range growth, and that the Gonghe basin and Qinghai Lake basin were connected prior to the late Miocene (Chapter 3; Zheng et al., in review).

Faults in the proximal footwall of the NW QNS in the Chaka region provide excellent exposure of foreland basin strata, which we exploit for two structural transects that bear on the local architecture of the range (Figures 4.7a,b, 4.9). Along the eastern transect, the basal M1 and PQn strata dip 30° SW, except for a narrow, ~1 km-wide syncline near the QNS front (Figures 4.7b, 4.9). Along the western transect, the correlative beds dip steeply, ~50-60° to the SW, and they are folded into a small anticline that sits above a bedrock sill near the range front (Figure 4.7a, 4.9). Upsection, the deposits grade into a >1 km-thick package of PQ sandstones and gravels (Figure 4.9). Along the eastern structural transect, the mid to upper levels of PQ exhibit progressively shallowing dips, such that the highest exposed PQcgl are nearly flat-lying (Figures 4.7a, b, 4.9). Although the structure of the western transect is complicated slightly by a fault that strikes nearly orthogonal to the range front, the overall structural pattern is the same.

To the east, much of the QNS range front is poorly exposed and inaccessible (Figure 4.5). However, the Yellow River canyon provides excellent exposure of strata on the flanks of the far eastern QNS near Gonghe City (Figures 4.5, 4.7f,g). In a tributary canyon north of Gonghe city, PQcgl are in unconformable contact with the bedrock core of the QNS, and dip 45° S (Figure 4.5). Only hundreds of meters to the south, the PQcgl strata are flat lying, implying a fairly narrow,
synformal hinge along the southern range front. We did not observe progressive unconformities within PQcgl at this location, so interpretation of the local timing of range growth is equivocal, but folding of PQcgl strata along the eastern QNS front indicates range growth during sediment deposition during the Pliocene and Quaternary. Moreover, similarities in the geometries of deformed strata at this location, and near Chaka to the west, suggest that the Qinghai Nan Shan is developed in the hanging wall of a regionally-extensive, S-vergent thrust system.

Deep M2 strata are well exposed on the flanks of a pluton that outcrops ~20 km east of Gonghe city (Fig. 4.5, 4.10). The pluton is bounded by a NE-vergent thrust fault along its eastern edge, which dips ~60° W (Figures 4.7g, 4.10). Along the west side of the pluton, M2 strata onlap the pluton and dip gently west, ~5-10°. Along the east side of the pluton, M2 strata dip 20°-50° E in several outcrops in the proximal footwall of the fault. In one ~80 m-thick exposure, east dipping strata exhibit progressively shallowing dips. Moreover, near the top of the outcrop, one bed dipping 31° SE truncates into a shallower, 22° SE dipping bed. A progressive unconformity and fanning dips on the eastern side of the pluton and onlapping strata on the western side of the pluton, indicate that the fault along the eastern edge of the pluton was active during M2 deposition. Similar growth strata are associated with mesoscale faults that cut M2 strata nearby. Although these deposits have not been dated directly, their lithostratigraphic similarity to the Chaka section (Zhang et al., in review—Figure 4.3) suggests activity along the eastern range front during the past 5-7 Ma.

4.5.1c Quaternary mapping of deformed alluvium

Preserved along the southern flank of the QNS is a sequence of well exposed, aggregated alluvial fan deposits that extend south from the range front (Figure 4.11). In the proximal footwall of the QNS, multiple generations of these alluvial surfaces are associated with well
defined fault scarps that strike subparallel to the range front. Because displaced depositional surfaces can serve as structural markers for reconstructing fault slip rates, we present detailed observations about several generations of alluvial surfaces at a site that is ~5 km from the western structural transect along the western QNS range. We identify at least four generations of alluvial fan surfaces on the basis of field observations and satellite imagery (Figure 4.11). These generations include 1) the abandoned depositional surface in the uplifted hangingwall of the fault, 2) several generations of strath terraces in the hangingwall of the fault, 3) active channels that cross the fault, and 4) the depositional surface onto which the narrow channels spill after crossing the fault trace. Of particular interest is preserved depositional surface in the hangingwall of the fault, because the abandonment age should provide a maximum bound for the onset of faulting in the proximal footwall of the QNS.

The abandoned depositional surface that is uplifted in the hangingwall of the thrust fault has a distinct appearance, both in satellite imagery and in the field. In satellite images, the surface is relatively dark, and sits atop undissected portions of the uplifted alluvium (Figure 4.11). Near the surface trace of the fault, the surface appears to be slightly degraded, and in satellite imagery, the areal distribution of surface degradation appears to be defined by a slight color change near the fault trace. Field observations indicate that the highest preserved surface in the hangingwall of the fault is dark in appearance, and characterized by a lack of textural roughness. The color and the texture of the surface are very distinct from lower terraces in the hangingwall of the fault, from active channels, and from the depositional surface in the footwall of the fault.

A 3-m soil pit dug into this surface reveals the stratigraphic character of the soil and alluvium underlying this abandoned surface. The stratigraphy of the sample pit is relatively
simple (Figure 4.11). The upper 20 cm of the pit are 75% silt, and pebble clasts float in the matrix. Between 20 and 25-35 cm, the proportion of matrix decreases. At 25-35 cm depth a 1 cm thick, unbroken, wavy band of pedogenic carbonate extends across the pit wall. Clasts beneath this horizon exhibit carbonate coatings to a depth of ~50 cm. Below this band, the lower 260-270 cm of the pit contains clast supported, conglomeratic alluvial deposits, with about 35-50% matrix material. Clasts are granules and pebbles and b-axes may be as large as a few tens of centimeters. Given the well developed caliche horizon, and the significant inflation of the surface by loess, the surface appears to be relatively old, although an absence of soil chronologies for the region prohibits age correlation solely on the basis of soil character.

4.5.1d Gonghe Nan Shan overview

The GNS is related to a second, S-vergent network of imbricate thrust faults that break upward through the south-central part of the basin (Figures 4.6, 4.8). The high topography of the range divides the eastern part of the broader Gonghe basin complex into two subbasins, the Gonghe basin (*sensu stricto*) to the north, and the Tongde/Xinghai basin to the south (Figure 4.2). The GNS fault network merges to the west into the ~NNW-striking Ela Shan mountain ranges (Figure 4.1). The primary fault bounding the southern side of the GNS is well exposed in the Yellow River canyon where it crosses the southern boundary of the range, but to the east and west along the range front, the GNS fault network is either blind or buried by upper Cenozoic basin fill. Exposures on either side of the range indicate that Cenozoic basin strata are folded into a broad anticline across the GNS. Similar to the QNS, the range comprises a series of broad, asymmetric anticlines in the hangingwall of the GNS fault system (Figure 4.6, 4.8). The range is cored by Triassic flysch of the Songpan Ganzi flysch terrane (e.g. Weislogel, 2008).

4.5.1e Folding of foreland basin strata along the flanks of the GNS
At the junction of the western and eastern GNS ranges, deep levels of the Gonghe basin fill are exposed in the Yellow River canyon, and these outcrops provide excellent control on the structural architecture of the range (Figures 4.6, 4.12). Along the southern range front, the range-bounding fault daylights (Figure 4.12) and patchy exposures of M1 redbeds line the southern range front (Figure 4.6). The M1 redbeds are overturned and dip ~30 ENE. The M1 outcrops merge along strike with discontinuous patches of grey sandstones and mudstone, which we tentatively correlate to M1 on the basis of their continuity with M1 strata. Along the Yellow River canyon, PQcgl beds are in unconformable contact with the highly deformed metasedimentary rocks that core the GNS and become progressively shallower upsection toward the top of the basin fill (Figure 4.8a); beds are subvertical just above the basal unconformity and they are subhorizontal at the basin surface (Figure 4.8c). Moreover, in this outcrop, PQcgl exhibits prominent intraformational unconformities which can be traced into correlative conformities only a few hundred meters to the south. The progressive tilting and intraformational unconformities in PQcgl strata indicate that the unit accumulated during fault slip.

In the canyon on the north side of the range, mapping along multiple structural transects indicates that both M1 and M2, are folded into a 20 - 30° NNE-dipping panel (Figure 4.8, 4.12). In contrast to the relatively uniform dips in the lower part of the section, PQ exhibits relatively shallow dips in this region, ranging from ~0 - 15°. Importantly, immediately north of the small village of Yangqu, a ~450 m thick exposure of PQ gravels is well-exposed along the western canyon wall. Bedding dips in this outcrop progressively shallow upward from ~ 10-15° NE at the base of canyon to subhorizontal at the top of the section (Figures 4.8b, 4.12). Furthermore, unit PQ in this region is in angular unconformable contact underlying M1 and M2 deposits, and the degree of angular discordance diminishes toward north, away from the range and from the axis.
of the anticline (e.g. Figure 4.12). Similar to the southern GNS front, upstream variations in dip of the PQ strata and the variable degree of angular discordance imply ongoing tilting of the GNS during PQ deposition, that is attributed to deformation in the synformal hinge along the backlimb of the GNS fault-fold system.

South of the village of Yangqu, M1 and M2 strata define a NW-plunging anticline, which plunges shallowly to the west (Figure 4.8). The fold is several kilometers wide, and it appears to extend to the SE into the range, although the Neogene strata which define the architecture of this fold are eroded. The structural high in the core of this fold projects along strike into the topographic high to the north of the saddle interior to the range, and the structural low along southern front of the fold projects along strike into the topographic saddle that is interior to the GNS.

The western GNS overlaps the eastern part of the range in the Yellow River canyon area (Figure 4.12). Near the small village of Jiala, the fault that bounds the western subrange daylights and dips 27° N (Figure 4.12). North of Jiala, the western range is covered by basin fill, but ~10 km to the east, the eastern terminus of the range outcrops in the deepest part of the Yellow River canyon (Figures 4.6, 4.13). At both sites, PQ\textsubscript{cgl} deposits bury the western GNS range, indicating that range growth pre-dated PQ\textsubscript{cgl} deposition locally.

A narrow bedrock sill crops out near the SW margin of Gonghe basin, and bedded PQ silts are tilted up to 20° along the strike of this sill (Figure 4.6), suggesting late Pleistocene deformation occurred a few tens of km N of the Gonghe Nan Shan. The 1990, a M\textsubscript{w} = 6.9 earthquake is located below this sill. Pre- and post earthquake line leveling surveys indicate up to ~0.25 m of coseismic surface uplift to the south of the sill, and up to ~0.04 m of coseismic
subidence to the north, suggesting that the feature is related to a steeply dipping 60° backthrust to the GNS (Figure 12a,b, see Chen et al., 1996; Chen and Xu, 2000).

4.5.1f Deformation interior to Gonghe and Tongde subbasins

We conducted reconnaissance level mapping across Gonghe and Tongde sub-basins, in order to evaluate the possibility of contractional deformation interior to the basins. Mapping efforts were focused within the Yellow River canyon, which is ~600 m deep in Gonghe and ~500 m deep in Tongde. In general, almost no deformation is recorded within the exposed strata along the Yellow River corridor. One exception is in central Tongde, where a small monocline with a few tens of m of structural relief was identified in the lower part of the PQfl section (shown in Figure 4.8). The fold appears to be buried by the youngest sediments in the basin. The finding of relatively little shortening across Gonghe and Tongde basins is consistent with observations of relatively undeformed strata across the vast expanses of the deeply exhumed basins to the east, at Guide, Xunhua, and Linxia (e.g. Fang et al., 2003, 2005; Lease et al., in review).

4.5.1g Northern flank of the Anyemaqen Shan and Ela Shan

In general, a lack of well exposed basin fill along the flank of the Anyemaqen makes it difficult to assess the Cenozoic history of the range. However, we have identified two exposures of fill in southern Tongde that may provide limited information about this history. First, in the Yellow River canyon at the boundary between Tongde basin and the Anyemaqen Shan, subhorizontal PQ sands and gravels onlap multiply deformed Triassic Songpan-Ganzi flysch beds. Notably, there are no redbeds underlying the PQ gravels at this site. Second, to the west, where a small tributary to the Yellow River provides limited exposure of the lower M2 beds south of the town of Xinghai, U2 strata dip gently, ~17° to the NNE (see angular unconformity in Figure 4.8). The strata dip uniformly, and they are in angular unconformable contact with the
overlying, subhorizontal PQ sands and gravels (Figure 4.13). The moderate dips along the
northern flank of this range suggest that it may be structurally similar to GNS, which is also
characterized by a ~20-30° dip panel along its northern limb. The angular unconformity between
M2 and PQ indicates that tilting must have occurred along the southern margin of Tongde basin
between deposition of the two units, near the Miocene-Pliocene boundary.

4.5.1h Interior Qinghai Lake basin

Seismic reflection data reveals that in the central part of Qinghai Lake basin, lower basin
strata are warped over a broad, E-W striking structural high that divides the basin into southern
and northern subbasins. Across the structural high, lower basin strata are truncated by the
relatively thin, flat-lying strata that cap the basin. These observations suggest the presence of an
additional fault network across the central part of the Qinghai Lake basin which was active
during basin filling. At the northern basin margin, the lower strata are folded gently on top the
broad range along the northern margin of Qinghai Lake basin. Deeper beds are truncated by
upper levels of basin fill, and the relationship suggests that the eastern Qilian Shan was active
during Qinghai Lake basin infilling locally.

4.5.2 Topographic analysis of fault networks

4.5.2a QNS

The topography of the QNS exhibits pronounced asymmetry, with a gently sloping,
relatively un-eroded north limb, and a steep, narrow, highly dissected south limb (Figure 4.14).
Along the north limb of the range, the topographic slope is low, typically on the order of a few
degrees. Similarly, topographic relief measured over a 1 km window is also low, typically on the
order of a couple of hundred meters or less. Furthermore, topographic profiles show that
maximum, minimum, and mean topography converge along the northern flank of the range,
indicating that little erosion of the northern limb of the range has occurred since it was uplifted. In contrast, the topographic slope of the narrow southern flank of the range is on the order of a few tens of degrees and the relief is on the order of 1000-2000 m. Maximum and minimum topography diverge across the narrow southern flank of the range, indicating relatively deep excavation of the southern range front. Similarities in dip between the strata in southern Qinghai Lake basin and the low-relief, gently dipping topographic surface on the northern limb of the QNS range suggests that the surface represents the early Miocene basin floor. Given this observation, we use the well preserved erosion surface along the northern side of the QNS as a structural marker that can be used to describe the highly asymmetric architecture of the range (Figure 4.7).

### 4.5.2b Topography of the GNS

Both the northern and southern flanks of the GNS appear to be characterized by steep slopes and relatively high local relief compared to the QNS (Figure 4.15). One of the most prominent features of the morphology of the range is the E-W striking topographic saddle that divides the eastern half into two parts. Serial N-S topographic profiles reveal that the saddle is ~500 m lower than adjacent peaks to the north and south (Figure 4.15). Although the topography of the GNS is more rugged than the QNS, topographic profiles of the range suggest that it exhibits similar N-S asymmetry. In particular, the saddle in the center of the range and a steep topographic escarpment immediately to the north of it seem to divide two shallowly north-dipping surfaces in the range. The southern range front is very steep, and relief within a 1-km window can be as high as ~1 km.

The correspondence between structural relief and topography suggests that the topographic character of the GNS may contain important information about the architecture of
the range. We interpret that the saddle in the central part of the GNS reveals that the eastern GNS comprises two, imbricated south vergent anticlines (Figures 4.6, 4.8, 4.15). In support of this view, remnants of M1 or M2 strata are preserved in the structural saddle in the middle of the range, indicating that the topographic low is also structural low, which must extend at least several tens of km to the SE of the town of Yangqu (QBGMR, 1991) (Figure 4.15). Using the topography as a proxy for structural relief, we extend this fold across the eastern GNS, such that we interpret the fold to extend over hundreds of km.

4.5.2c Interior Anyemaqen Shan

Previous investigators indentified a broad zone of high channel steepness and high Quaternary fluvial incision rates located in the Anyemaqen Shan mountains and centered over the Kunlun fault (Harkins et al., 2007; 2010). They inferred that this region of anomalously oversteepened channel profiles relates to a zone of late Cenozoic uplift due to distributed deformation at the tip of the Kunlun fault (e.g. Kirby et al., 2007; Harkins et al., 2010). The distribution of structures that accommodate this deformation in the upper crust is not clear based on surface geology (Harkins et al., 2010), and we attempt to identify possible structures based on 90-m Shuttle Radar Topography Mission (SRTM) digital topography. A prominent, E-W striking lineament in the topography of the Anyemaqen Shan (Figure 4.16). Elevation, topographic slope, and relief within a 1 and 5 km radius are markedly higher on the south side of this lineament in comparison to the north side. Although fluvial terraces drape the trace of this feature and do not record any surface deformation since ~25-30 ka, the pronounced differences in topography across the lineament suggest that it may be related to a fault at depth. We place a lower bound on the amount of vertical rock uplift that this putative structure could facilitate by comparing the elevation of peaks on either side of the lineament (Figure 4.16). Analysis of two different
topographic profiles indicates that this offset is ~250 – 500 m. Due to the fact that topographic differences across this lineament may be reduced by erosion, we consider ~250-500 m to represent a minimum vertical offset for a potential fault.

4.5.3 Thermochronologic constraints on structural relief

4.5.3a Methods

In order to reconstruct the thermal evolution of the ranges, vertical transects were collected from the southern flank of the QNS and GNS ranges, where the maximum structural relief is located (sample locations shown in Figures 4.5-4.8). We collected granitic and granodioritic rocks along two vertical transects spanning ~1 km of elevation from plutons in the Qinghai Nan Shan. Although the QNS plutons are not directly dated, pluton emplacement in the QNS likely occurred during oceanic and continental subduction beneath the South Qilian terrane from the Cambrian to the Silurian (Xiao et al., 2009). Due to the lack of plutonic rocks in the Gonghe Nan Shan, we collected sandstone samples from the Triassic Songpan-Ganzi flysch along a short (~600 m) vertical transect.

After rock samples were crushed and sieved, we selected 2-5 from each granite sample for (U-Th)/He (AHe) analysis. Using an optical microscope, we selected grains exhibiting equant, euhedral geometry and a diameter in excess of 80μm, and lacking mineral inclusions. Due to the fact that the apatites from the flysch samples were small (<80μm in diameter), frosted, and broken, only AFT cooling ages were measured for flysch samples. AHe cooling ages were measured at the California Institute of Technology (Farley, 2002), and AFT cooling ages were measured at Apatite to Zircon, in both cases following standard protocols (Farley, 2002; Donellick et al., 2005).

4.5.3b Data
Apatite fission tracks from the QNS exhibit relatively little scatter, ranging from 11 – 16 μm, and exhibiting peaks at 14 ± 1 μm (Figure 4.17). The lowest sample, CT8-2 passes the $\chi^2$ and visual inspection of ages and standard errors on a Galbraith plot shows that the cooling ages derive from a homogeneous population (Figure 4.18). In contrast, the upper samples, CT8-4 and CT8-6 do not pass the $\chi^2$ test. Visual inspection of Galbraith plots for these samples suggests that a few grain ages vary from the central age by > 2 standard errors.

Apatite fission track lengths from the detrital rocks that core that core the GNS exhibit significantly more variation. Track lengths range from 4 – 17 μm and the mode track length is 10 – 13 μm (Figure 4.17). All samples from the GNS fail the $\chi^2$ test, suggesting the presence of multiple kinetic populations (e.g. Figure 4.18).

AHe cooling ages from the central QNS are Cretaceous and Paleogene (Table 4.1, Figure 4.19). The slope of the age-elevation array indicates that Cretaceous and Paleogene erosion rates of the QNS were low, on the order of ~10-20 m/Ma, during this time window. Interestingly, the cooling ages of each replicate from the lowest sample of this transect, CT8-6, correlate strongly to their effective uranium concentration ([eU]) (Figure 4.19) (Flowers et al., 2007, 2009). Paired apatite fission track samples along this vertical profile exhibit mid-Cretaceous pooled cooling ages, and they appear to indicate comparatively rapid erosion rates, on the order of ~90 m/Ma.

For the western QNS vertical transect, mean AHe cooling ages are in excess of 146 Ma, (Figure 4.19) such that the samples do not record information relevant to the Cenozoic or the Cretaceous time-temperature history of the western QNS.

Pooled cooling ages from the GNS vertical profile are Cretaceous and Paleocene in age, and range from 51.6 ± 3.5 Ma to 126.9 ± 5.0 Ma (Figure 4.19). The vertical profile of AFT ages from the GNS implies a mean erosion rates on the order of ~5 m/Ma
4.5.3c Structural depth

In neighboring Linxia basin, AHe samples distributed over ~1 km of structural relief exhibit rapid Paleogene erosion rates of ~200 m/Ma, and they are located only 850 m below an inferred Cretaceous erosion surface (e.g. Clark et al., 2010). In order to reconcile the rapid Paleogene cooling rates with the apparent structural depth, the presence of a thick (1-3 km) package of late Cretaceous sediments was inferred. In order to investigate this possibility, and in order to verify the consistency between our structural interpretations and the thermochronologic data, we address the pre-exhumation burial depth of our samples.

Given that most samples from the central QNS exhibit Cretaceous or Paleogene cooling ages, and geologic evidence suggests that structural relief developed on the range during the late Miocene, it does not appear that a large amount of structural relief has developed on the range during late Cenozoic contractional deformation. For example, given that the PRZ for helium in apatite (HePRZ) is from ~40 – 85 °C (Wolf et al., 1998; Stockli et al., 2000), and given a typical continental geothermal gradient of 15 - 30 °C/km, the thermochronologic data imply less than ~1 - 2 km of structural relief has developed. However, closer examination of the central QNS data suggests that the deepest sample may have been exhumed from the shallow levels of the HePRZ during the Miocene. First, although the deepest sample from the central QNS exhibits a late Eocene mean AHe cooling age, individual replicates are as young as 23 Ma. Second, the sample exhibits a remarkably strong cooling age vs. [eU] correlation, which implies long residence in the HePRZ. Overlying samples do not exhibit such young cooling ages for replicate samples, or such strong cooling age vs. [eU] concentrations, such that it appears that the deepest sample may have resided in the shallow part of the HePRZ. Our structural interpretation suggests that the sample is characterized by a structural depth of ~1 km, which is entirely consistent with the
likely depth of the HePRZ (Figure 4.7). The consistency between the observed structural relief and the thermal history of the deepest sample from the central QNS does not seem to require the presence of a thick late Cretaceous or early Paleogene overburden above the range, as has been suggest for the west Qinling.

For the western QNS, samples from structurally deep portions of the range yield early Mesozoic and late Paleozoic cooling ages, whereas geologic evidence implies the development of structural relief during the Miocene. Given the same assumptions about the geothermal gradient and HePRZ as above, the thermochronologic data imply less than ~1 - 2 km of structural relief has developed in the western part of the range since the Miocene. Importantly, this finding is consistent with our structural interpretation for the western part of the range (Figure 4.7a,b). Similarly, cooling ages from the GNS are Cretaceous and Paleocene whereas geologic evidence suggests the rapid development of structural relief for the GNS beginning in the late Miocene. Given a nominal partial annealing window of ~60 - 120° for apatite fission track, and typical continental geothermal gradients, this implies that less than ~1.6 – 3.3 km of structural relief has developed along the southern GNS range front during the late Cenozoic. Again, this finding is internally consistent with our structural interpretation of the GNS.

4.5.4 Shortening estimated from balanced cross sections

In order to begin to address the magnitude of late Cenozoic upper crustal shortening around northeastern Tibet, we calculate line length shortening measurements for the serial cross sections that derive from the geologic and topographic observations for the QNS and GNS presented above. For each cross section, we combine minimum shortening accommodated by both folding and faulting. Below, we walk through the key assumptions involved in each cross
section and describe shortening budgets for both. Additionally, we use the cross sections to illustrate information about the architecture of the fault networks at depth.

**4.5.4a Key assumptions for cross section construction**

One of the largest challenges involved in constructing geologic cross sections across the QNS is constraining the architecture of the north limb of the range, where geologic exposure is very limited. Fortunately, we are able to use the erosion surface that extends across much of the north flank of the range (Figure 4.7). Where the erosion surface is buried by Qinghai Lake basin fill, we project the dip of the north limb of the QNS beneath the basin to a depth of ~500 m thick, the lower bound for the thickness of sediment beneath southern Qinghai Lake (An et al., 2006). At that point, we assume that the north limb passes through a kink band, and becomes flat. Projecting to a greater depth would slightly increase the shortening recorded by the cross section.

In order to constrain the architecture of the southern limb of the range, we use detailed geologic mapping where possible. For several cross sections in the central QNS, however, no strata are exposed in the proximal footwall of the range. In these instances, we assume that the forelimb of the range is tilted ~45° S because that is the dip of the strata made near the eastern tip of the range near the town of Gonghe. It seems likely that an even higher degree of tilting has occurred in the central portion of the range in comparison to the eastern terminus, such that we consider 45° to be a conservative estimate for the degree of tilting along the range front.

In the absence of other information, we assume that bedrock-basin fill contact approximately marks the position of the pre-faulting basin floor, and we require the steep forelimb of the QNS to pass through this point. For the cross sections adjacent to thermochronologic transects, this assumption is incompatible with age-elevation pattern of the
cooling ages, and requires us to shift the position of the basin floor to the south by a few kilometers (e.g. 4.7d vs. 4.7e).

In contrast to the QNS, deep exposures of foreland basin strata provide at least local constraints on the architecture of both the forelimb and the backlimb of the range along the GNS (Figure 4.8). Numerous measurements of bedding dip extending over several kilometers indicate that strata on the northern side of the range are folded into a ~20-30° N dipping panel. Given that this dip is consistent with focal mechanisms for earthquakes located beneath the QNS, we infer that the regional dip of GNS fault network is within this range. Following this inference, we assume that for structures where we lack direct data bearing on the degree of backlimb rotation, beds dip ~20°. Similarly to the QNS, the position of the back limb fold hinge for the GNS is difficult to define on the basis of surface observations. In order to minimize our shortening estimate, we assume that the hinges are located immediately below the outcropping sections of rock (e.g. fold in central portion of Figure 4.8b).

In the absence of a clearly defined broad erosion surface along the north side of the GNS, the most interpretive part of the cross sections through the range involves the projection of backlimb dips to the south, towards the fault plane along the southern margin of the range. Although the cross sections would record less shortening the base of M1were projected along the envelope defined by the maximum topography of the range, we prefer to project the moderate dip of the backlimb into the fault plane along the southern side of the range because outcrop data across the central part of the range suggest a south vergent asymmetry to the structure.

4.5.4c Line-length shortening

Line-length measurements of shortening for each of the QNS cross sections reveals that shortening across the QNS has been minimal, on the order of 1-2 km (Table 4.2). Given that the
width of the range is ~30 - 40 km, our budgets imply less than 10% shortening has occurred across the range since the late Miocene. Although these are mimimum estimates, the extensive erosion surface and the moderately good exposure of deformed foreland basin strata in the proximal footwall of the range suggest that the cross sections are accurate.

In contrast to the QNS, the backlimb of the GNS appears to have accommodated a relatively large degree of shortening. Line length estimates range from ~5 – 7 km (Table 4.2). The differences between the two ranges appear to be mostly controlled by high degree of fault slip that is recorded in the GNS cross sections, typically 3-4 km. This, in turn, is a function of the way we interpreted the backlimb architecture of the two ranges, but our interpretations seem justified by geologic and topographic observations from both ranges. Moreover, the emergent, or shallowly buried fault along the southern front of the GNS seems to imply a relatively high degree of shortening compared to the QNS, where thrust faults are blind.

4.5.4d Fault architecture and decollement depth

In the absence of geophysical data, it is difficult to interpret the geometry of the QNS and the GNS fault networks. However, basic architectural features of both ranges seem to provide some insight. First of all, the steep forelimbs of the range, and the broad, gently dipping backlimbs is kinematically compatible with a curviplanar fault ramp (e.g. Amos et al., 2007). Second, it appears that both fault networks penetrate to depths of at least 1 – 2 km (the approximate thickness of sediment in Gonghe basin), because basement rocks are exposed in the core of the fault related folds. Further information about the decollement depth of the two fault networks may be calculated with the regional fault dip and the width of an individual fold. Although the width of both ranges is subject to uncertainty, our cross sections appear to permit a reasonable estimate of range width. Across the central part of the range, where the erosion
surface is well preserved, the QNS appears to be ~30-40 km wide. Although we have no data that bears on the regional dip of the QNS fault network, the large width of the range implies that for a reasonable range of fault dips (i.e. ~20-60°), the regional decollement depth is no less than ~10 km, but perhaps tens of km deeper. In contrast, the decollement depth of the GNS seems to be more shallow. Given a regional fault dip of ~20-30°, and a characteristic fold width of ~10 km, the GNS is more likely to sole into a decollement at ~4 – 7 km.

Information about the decollement depth of a fault network may also be inferred from the geometry of fault related folds (see Woodward et al., 1985). If cross-sectional area is conserved during fold growth, then the area within a fault-related fold must be equivalent to the product of horizontal shortening and decollement depth. We exploit this fact in an attempt to constrain the depth of brittle faulting in the QNS and the GNS (Table 4.2). Again, uncertainties in the position of the backlimb hinge introduce uncertainty into our analysis, although fold area is not extremely sensitive to small changes in the position of the backlimb hinge because most of the area of a fold corresponds to the area of maximum structural relief.

On the basis of this calculation, the QNS fault network appears to extend to at least 30 km depth (Table 4.2). In the eastern sub-range of the QNS, however, where the structural relief is low in comparison to the western and central parts of the range, calculated decollement depths are somewhat shallower, ~15 – 20 km. In contrast, the calculated fault depths for the GNS are shallow, between 7 and 9 km. Importantly, both means of calculating decollement depth provide overlapping estimates of fault depth, and in tandem the results imply that the decollement depth beneath the QNS is 10 – 30 km, or possibly deeper, and the decollement depth beneath the GNS is ~3 – 9 km.

4.6 Rates of deformation
Much of the evidence for early and mid Tertiary deformation in northeastern Tibet comes from the structural corridor that extends from the western Qilian Shan to the west Qinling (e.g. Yin et al., 2008; Clark et al., 2010; Lease et al., in review). Given that the Gonghe basin complex occupies a broad swath of the interior of this structural trend, we search for the possibility of early to mid Tertiary deformation in this region. To do so, we exploit age-elevation relationships of vertical transects, inverse models of thermochronologic data, and geologic constraints (Chapter 3). This analysis provides the basis for reconstructing geologic shortening rates across both ranges. Moreover, by dating displaced alluvial landforms along the QNS range front, we are able to compare late Quaternary deformation rates to late Cenozoic rates and to provide a local evaluation of secular variation in deformation rates in northeastern Tibet.

4.6.1 New thermal constraints on the onset of Cenozoic deformation

Because the oldest preserved strata along the QNS range front date to the mid-Miocene, the earlier tectonic history of the range must be investigated with other data. Two data sets from the central Qinghai Nan Shan vertical transect appear to contain information about the Paleogene history of the range. First, the age-elevation relationships imply that during the Paleocene and Eocene, erosion rates in the QNS were on the order of ~10-20 m/Ma. For any reasonable geothermal gradient, this erosion rate implies a cooling rate on the order of 1°C/Ma. Thus, the age-elevation data suggest that the QNS was tectonically quiescent, during the Paleogene.

Recent studies show that He retentivity in apatite increases with radiation damage (e.g. Flowers et al., 2007), particularly for samples with long residence in the HePRZ. Importantly, such changes in retentivity cause slight differences in the closure temperature of an apatite for the AHe system. As such, samples yielding large cooling age-[eU] spreads may effectively act as paired thermochronometers, each with slightly different closure temperatures. Given that
replicates from the deepest sample in the central QNS span much the Paleogene, (24.5 Ma - 49.2 Ma), and that the samples exhibit a strong age-[eU] correlation, we build an inverse model on the basis of these data, in an effort to provide an additional constraint on the early Cenozoic history of the range (Figure 4.20).

In the model, we impose a surface temperature of 10°C. Given the likely Cambrian-Silurian timing for pluton emplacement (Xiao et al., 2009), we constrain the time-temperature history of the rocks by requiring that during the Devonian, following pluton emplacement, the minimum temperature was 45 °C, the lower limit of the partial retention zone for He in apatite. We impose a maximum Devonian temperature of 250°C, a reasonable maximum temperature for a pluton emplaced in the middle to upper crust. An additional constraint on the time-temperature history of the pluton comes from the pooled AFT cooling age of sample CT8-6, which is 99.6 +/- 4 Ma. We require the sample to be within the AFT partial annealing zone (60 - 120 °C) during this time window. The model shows that from the late Paleozoic to the Late Cretaceous, a fairly wide range of cooling histories is consistent with the AHe data, although all of the histories are characterized by slow cooling (Figure 4.20). However, during the Paleogene, over the interval encompassed by the various CT8-6 replicates, the cooling history is well constrained. Any given path suggests cooling rates on the order of ~1 °C/Ma, which is consistent with slow erosion rates, on the order of a few tens of m/Ma.

In tandem, the age-elevation gradient of the vertical transect data and the inverse model based on the data from sample CT8-6 imply slow cooling in the central QNS during the Paleogene. When combined geologic observations along the QNS range front implying that range growth began in the late Miocene, the thermal modeling indicates that the QNS was tectonically quiescent for most of the Tertiary, until range growth initiated at ~9 – 6.5 Ma.
Fission track cooling ages from the GNS are older than AHe ages in the central QNS, and therefore the age-elevation data is not directly relevant to much of the Paleogene history of the range. However, populations of fission track lengths may contain information about the thermal history of rock subsequent to its transit through the partial annealing zone. In order to investigate the possibility that fission track length distributions contain information about the Paleogene time-temperature history of the range, we use the fission track length distribution of each sample to construct a series of inverse models (Figure 4.21).

We place the following constraints on the thermal history of the sample. 1) Between 220 and 200 Ma, the sample must be between 60 °C, the upper limit of the PAZ and 250 °C. This implies relatively deep burial of sediments following deposition in the Songpan-Ganzi basin, but it is consistent with the metamorphic grade of sediments. 2) The sample must be in the PAZ during the cooling age of the sample. Because the observed cooling ages may represent prolonged periods in the PAZ (on the order of 10s or 100s of Ma), it is important to note that samples may have actually cooled beyond the PAZ later than the measured cooling age, but in this case, the sample would still be required to reside in the PAZ during the time window suggested by the cooling age. 3) The surface temperature is fixed at 10 °C. Inverse models for all of the samples, even those with mid-Cretaceous cooling ages, suggest that rapid cooling must have begun between ~50 and 0 Ma (Figure 4.21), but it does not appear possible to glean any additional information from the inverse models. Due to the fact that thermochronologic data from the GNS simply appears to limit the timing of cooling to the Cenozoic, the primary evidence for the timing of contractional tectonism in the GNS derives from geologic observations, which bracket the timing of range growth between ~10 and 7 Ma (Chapter 3).

4.6.2 Mean geologic rates
By combining line length shortening measurements from the QNS and GNS with aforementioned constraints on the initiation ages of the ranges, we are able to assess mean shortening rates for the ranges since the time of initiation in the late Miocene. We have derived line-length shortening estimates for the QNS based on serial, deformed state cross sections (Figure 4.7). Given that our structural analysis implies 1 – 2 km of shortening across the range, and that geologic and thermochronologic analysis of QNS imply the initiation of contractional tectonism at 6.5 – 9 Ma (Fang et al., 2005; Lease et al., 2007; Zheng et al., in review), late Cenozoic shortening rates appear to be 0.1 - 0.3 mm/yr. The finite strain accommodated by the GNS appears to be relatively large compared to the QNS, about 5-7 km. Given an initiation age for the range of 10-6 Ma, this implies a geologic shortening rate of 0.5-1.2 km/Ma.

### 4.6.3 Late Quaternary deformation rates

Detailed topographic surveying and cosmogenic burial dating of the abandoned depositional surface that is uplifted in the hangingwall of the faults along the proximal QNS range front provides the basis for estimating late Pleistocene slip rates. Our sample site is located only a few km to the west of structural transects along the range front, such that geologic shortening rates should be directly comparable to late Pleistocene deformation rates. Surveying was conducted with a differential global position system with sub-centimeter precision along two parallel transects separated by a few meters (Figure 4.22). In order to reconstruct fault slip from topographic data, we developed a script that accounts for uncertainties in both the surface slopes and fault dip using monte carlo statistics. The script calculates vertical displacement and assumes a range of fault dips in order to calculate horizontal and total fault displacements (e.g. Thompson et al., 2002). This script is described in detail in the supplementary information section.
Well developed pedogenic carbonates and thick accumulations of loess suggest that the abandoned depositional surface in the hangingwall of the fault is likely to date to the late Pleistocene, and as such, cosmogenic dating is a useful way to constrain its abandonment age. We sampled along a depth profile (location shown in Figure 4.11), and from seven different 5 cm-thick stratigraphic horizons, we collected >250 clasts, with b-axis diameters of 0.5 – 1.5 cm. The highest sample was collected at 40 cm, just below the calcrete band at the top of sample pit. Samples were collected to a depth of ~3 m, because at this depth spallogenic cosmogenic isotope production occurs at negligible rates (e.g. Anderson et al., 1996; Hancock et al., 1999) (Figure 4.23), and measured cosmogenic inventories reflect inherited $^{10}$Be inventory, prior to sediment deposition. We measured $^{10}$Be inventories for these samples following standard protocols at PRIME lab at Purdue University. Laboratory techniques are described in detail in the supplementary information section.

4.6.3a Measured cosmogenic inventories

Measured $^{10}$Be inventories range from 49.9 x $10^5$ atoms of $^{10}$Be per gram of quartz at the shallowest sample in the depth profile (40 cm), to 2.8 x $10^5$ atoms of $^{10}$Be per gram of quartz at the base of the profile (300 cm). Visual inspection of the depth profile suggests an exponential decay in cosmogenic $^{10}$Be concentration with depth (Figure 4.23). At the lowest part of the profile, $^{10}$Be concentration is nearly invariant with depth, so we take the 2.8 x $10^5$ atoms of $^{10}$Be per gram of quartz to represent the inherited component of cosmogenic $^{10}$Be.

4.6.3b Abandonment age calculation

In order to calculate the surface abandonment age, we use a new, freely available program which employs monte carlo statistics to calculate the abandonment age of a depositional surface (Hidy et al., in press). The program is capable of implementing a variety of production
rate schemes (e.g. Stone, 2000), surface slope and topographic shielding corrections, and it accounts for uncertainties in several variables, including post-deposition erosion rate, the soil bulk density, and inherited cosmogenic isotope concentrations.

In order to implement the abandonment age calculator, we gathered several observations in the field and in the laboratory. Visual inspection of the depth profile reveals an exponential decay in cosmogenic $^{10}$Be concentration with depth (Figure 4.23). At the lowest part of the profile, $^{10}$Be concentration is nearly invariant with depth, so we take the $^{10}$Be concentration from the lowest part of the profile to represent the inherited component of cosmogenic $^{10}$Be. We allow inheritance to vary between the measured concentration of the lowest sample, $28 \times 10^5$ atoms of $^{10}$Be per gram of quartz to half of that value. The lower inheritance limit is somewhat arbitrary, but several preliminary iterations of the program revealed that the calculated abandonment age is not highly sensitive to reasonable changes in the lower bound for inheritance. We did not measure bulk material density in the field, and we assume a density of 2.2 g/cm$^3$, a typical bulk density value for unconsolidated alluvial material. We assume that post-depositional surface lowering rates may be as rapid as 0.2 cm/kyr. Due to the proximity of the sample site with an adjacent mountain range front, we measured the angle to the horizon, at 15° intervals, for a 360° spectrum. The topographic shielding factor is $\sim$0.99, such that it is nearly negligible.

The critical difficulty for determining the abandonment age of the surface of interest is accounting for the effect of periodic loess exposure. The depositional surface is clearly inflated by loess at present. Based on the difference in matrix abundance between the upper and lower part of the profile, we estimate that the upper ~30 cm has been inflated by 20 cm of loess input, such that immediately following surface abandonment, the uppermost stratigraphic interval was only 10 cm thick. Although we have no direct constraints on the age of the loess, elsewhere in
the Qilian Shan optical dating of m-thick loess deposits that cap multiple generations of late-
Quaternary river terraces indicates that loess accumulation began in the early Holocene, and
persisted through the Holocene (Stokes et al., 2003; Küster et al., 2006). The onset of loess
deposition in the region is attributed to warming temperatures and wetter climate following the
last glacial period (see Küster et al., 2006 and references therein). Given that preliminary
iterations of the calculation indicate for any reasonable loess accumulation scenario, that the
surface is likely to be ~100 kyr, and given that surface inflation by loess began approximately at
10 ka, we calculate a coverage factor, which is a number from 0 to 1 that is multiplied by the
production rate (e.g. Gosse and Phillips, 2001). Our calculated coverage factor of 0.987,
suggesting that loess burial has a minimal effect on the abandonment age.

Chi-squared optimization of various synthetic abandonment histories indicates that
optimal histories range between 109 – 142 ka (Figure 4.23b). The optimal abandonment age
appears to be 114 ka, although this number is indistinguishable from the range present above on
the basis of chi-squared statistics.

4.6.3c Magnitude of displacement and slip rates

We calculate vertical offsets of 21.7 +/- 1.9 m for scarp profile 1 and 19.3 +/- 1.7 m for
scarp profile 2. Reported uncertainties are 1 standard deviation. By assuming fault dips of 20-
45°, we estimate 24.9 +/- 1.6 m of total slip and 12.4 +/- 0.3 m of horizontal slip along profile 1
and 22.3 +/- 1.4 m of total slip and 11.1 +/- 0.3 m of horizontal slip along profile 2. By
integrating these displacement calculations with the abandonment age calculation, we find
minimum rock uplift rates in the hangingwall of the fault scarp of ~0.12 – 0.22 m/ka. Integrating
the surface abandonment age with the estimated horizontal fault displacement yields minimum
shortening rates across the western QNS of 0.08 – 0.12 m/ka.
4.7 Discussion

4.7.1 Temporal variations in shortening rates across the QNS

Before drawing tectonic inferences on the basis of the measured late Quaternary shortening rates for the northwestern QNS, it is important to consider possible sources of uncertainty in this result. In one sense, the late Quaternary slip rates we measured could be considered a minimum rate because deformation of the surface may have significantly lagged depositional abandonment of the surface. However, the uplifted surface appears to be significantly older than those in the footwall of the fault. Whereas the uplifted surface is dark, smooth, and lacking protruding clasts, the adjacent surfaces in the footwall of the fault appear rougher. Moreover, textural variations on the surface in the footwall of the fault seem to suggest several generations of alluvium, including active channels, but also older generations. In other words, there seems to be evidence for a protracted depositional history on top of the non-uplifted surfaces, such that abandonment was likely to correspond to the timing of deformation.

In another sense, assumptions inherent to our abandonment age calculation may tend to deflate the abandonment age, and inflate the slip rate, such that our estimate is a maximum. For example, we make the simplifying assumption of loess accumulation on the surface at 10 kyr, on the basis on chronologies of loess accumulation around the Qilian Shan (Stokes et al., 2003; Küster et al., 2006). If loess accumulation began earlier, it would tend to reduce production rates within the depth profile, and increase the abandonment age. Moreover, given the significant age of the uplifted surface and the well developed pedogenic carbonate horizon, it is possible that some stripping of the surface has occurred. This would tend to erase cosmogenic inventories, and reduce the observed abandonment age.
With the caveats associated with late Pleistocene slip rates in mind, the correspondence between the late Quaternary rates and the geologic shortening rates since the late Miocene indicate that rates of contractional deformation have been steady across the range over the last ~10 Ma. Mean geologic shortening rates across the range from 0.1 – 0.3 mm/yr. The range agrees remarkably well late Quaternary shortening rates of ~0.1 mm/yr that we have measured along the QNS range front near the town of Chaka.

Recent studies cast the Elashan fault and the Riyueshan as right lateral, antithetic faults that accommodate shear between the Kunlun and Haiyuan faults (Duvall et al., 2010; Yuan et al., in press). In this view, the QNS and GNS fault networks accommodate shortening related to block rotation between the Elashan and Riyue Shan. Late Quaternary rates of strike slip for the Riyue Shan and Elashan are 1.2 ± 0.4 mm/yr and 1.1 ± 0.3 mm/yr, respectively. By dividing the total offset by the late Quaternary slip rate for these faults, an initiation age of 9.5 ± 3.5 Ma for these two structures has been determined. First, if the QNS is indeed kinematically linked to the adjacent strike slip faults, then the observations of steady deformation along the QNS provides some support for the late Miocene initiation age for the Riyue Shan and Elashan. Second, it appears that shortening along the QNS has been ~an order of magnitude slower than the adjacent strike slip faults, although shortening across the GNS may occur at a rate that is comparable to strike slip rates along the bounding structures.

4.7.2 Magnitude and style of upper crustal shortening in Gonghe region

We synthesize our detailed structural observations from the QNS and the GNS, with the various regional constraints on the structural evolution of interior northeastern Tibet presented above, in order to construct a regional, crustal-scale geologic cross section. In addition to the various constraints on the structure of the upper crust and the deep architecture of major regional
fault networks, we have incorporated Moho depths, which are inferred from analysis of receiver functions (Vergne et al., 2002) and other geophysical proxies (see Meyer et al., 1998).

In this cross section, the architecture of the QNS and GNS is adapted from our detailed cross sections presented in section 4.5. From south to north, other structures that record shortening within this cross section are: 1) a blind thrust fault in the interior portion of the Anyemaqen Shan, 2) a gentle fold along the southern margin of Tongde basin, 3) a broad gentle fold in interior Qinghai Lake basin, and 4) a broad, shallowly dipping fold across northern Qinghai Lake basin that may be related to geologic structures further north in the Qilian Shan (Figure 4.23). Observations that form the basis for interpreting these structures are presented in detail in 4.6.

Summing all of the shortening on the various structures around the region indicates at least 10 km (Figure 4.24). Importantly, we may not account for all of the shortening in the Anyemaqen Shan, and we have constructed cross sections in the QNS and GNS in a way that minimizes the shortening that they record. Given that the restored length of the cross section is 237 km, this represents 4% shortening of the upper crust in the late Cenozoic.

Detailed reconstructions of geologic shortening on the QNS and GNS indicate that the QNS and the GNS of interior northeastern Tibet accommodate less upper crustal shortening than the neighboring ranges in the Qilian Shan. Whereas the QNS and GNS appear to accommodate up to 2 and 7 km of late Cenozoic shortening, cross sections through major geologic structures in the Qilian record ~10-20 km of shortening for individual structures (Meyer et al., 1998). To the extent that the QNS and GNS are representative of other mountain ranges around northeastern Tibet, these differences suggest different that plateau building mechanisms between northeastern Tibet and the Qilian Shan may differ, despite the proximity of the two regions.
The paucity of shortening recorded in our regionally integrated budget for late Cenozoic structures across the northern Anyemaqen, Gonghe basin, and Qinghai Lake basin indicates that thickening of the upper crust since ~10 Ma did not drive surface uplift of the region. We calculate about 4% shortening, or ~10 km of shortening, distributed along a ~230 km long transect. Our budget is similar to recent reconstructions of upper crustal shortening for the eastern Qaidam basin, which find about 2 km of shortening along a 52 km section across eastern Qaidam 12 km shortening along a 40 km section across the southwestern shortening, or about 15% shortening regionally. Given that the east-west distance between eastern Gonghe and eastern Qaidam is ~350 km, the combined results of these studies imply small amounts, typically on the order of a few percent of Cenozoic upper crustal shortening have occurred within a radius that is centered over Gonghe basin and extends ~115 – 175 km in any direction. Importantly, the breadth of this area of low shortening is approximately large enough to be isostatically compensated (Watts, 2001). Without thick crust below the Gonghe basin prior to the Miocene, this deformation certainly does not appear sufficient to elevate the broad floor of Gonghe basin to ~3 – 3.5 km. In contrast to the region centered over Gonghe basin, shortening budgets extending over a similar spatial scale in the Qilian Shan imply ~100 – 200 km, again suggesting that upper crustal thickening plays a relatively important role in building high topography in the Qilian compared to northeastern Tibet.

In the context of regional geologic observations, the decollement depths of the QNS and GNS fault networks may provide insight into the crustal structure beneath the Gonghe basin region. Although the uncertainties on calculated decollement depths are large and overlapping, the calculations suggest that the GNS may sole into a relatively shallow decollement, perhaps around 10 km. The relatively shallow decollement depths could be reconciled in the context of
the bedrock geology of the Gonghe basin complex. The QNS, which is located north of the South Qilian suture, is cored by the plutonic rocks of the Paleozoic South Qilian shan arc terrane. In contrast, the GNS is cored by Triassic basin strata that blanket the Kunlun-Qaidam terrane. Although the precise thickness of the Triassic sedimentary fill in southern Gonghe is not well known locally, regional map patterns suggest that the thickness is between 5 and 15 km in the interior parts of northeastern Tibet (GBGMR, 1991). Recent active source seismic profiling suggests that the Songpan-Ganzi flysch may be as thin as ~2km beneath Tongde basin, and possibly as thick as ~10 km (Zhang et al., 2010). The correspondence between these thicknesses and the calculated depth of the GNS fault system suggests that the mechanical discontinuity at the base of the pre-Cenozoic sedimentary strata may be the detachment horizon for the GNS fault network (Figure 4.24).

Although the uncertainties on decollement depth calculations are large, 10 km seems to be a reasonable minimum depth. In tandem with the broad backlimbs of the ranges which extend over 10s of km and suggest a curviplanar fault ramp geometry, the fault networks appears to be structurally analogous to the Laramide structures found in the western United States (e.g. the Wind River range) or the Sierras Pampeanas of Argentina (e.g. Brewer et al., 1982; Snyder et al., 1990). In the Wind River range, for example, seismic reflection profiles reveal thrust fault that dip shallowly (~20°) down to ~30 km depth (e.g. Lynn et al., 1983). The backlimb of this structure dips no more than a few degrees and extends over several tens of kilometers. The fault architecture of the Wind River range is attributed to the strength of the upper and middle crust and Wyoming, and, if our analogy is applicable, implies strong upper and middle crust in northeastern Tibet beneath the Gonghe basin region.

4.7.3 Paleogeographic reconstruction of Cenozoic northeastern Tibet
The combination of mass accumulation curves (Figure 4.25) and geologic observations on the timing of key fault networks reveals three key phases of mountain building in northeastern Tibet: early Tertiary (50 – 30 Ma) slow subsidence near the plateau margins, middle Tertiary rapid subsidence in interior northeastern Tibet (30-10 Ma), and late Tertiary (10 – 2 Ma) emergence of basement-cored mountain ranges within broad regions of sediment accumulation (Figure 4.26).

Recent suggestions of widespread contractional deformation across the northern margin of Tibet (e.g. Yin et al., 2002; Clark et al., 2010) appear to be at odds with space-time patterns of basin subsidence in the region. The depositional evolution of the Gonghe basin complex reveals that early Tertiary depocenters near the northeastern plateau margin and in western Qaidam, were not linked. Rather, these regions evolved as distinct tectonic elements that were separated by a largely quiescence swath of terrane interior to northeastern Tibet. Moreover, it is difficult to link basin subsidence to any geodynamic process. One possible subsidence mechanism is flexural loading of the lithosphere. However, regional patterns of early Tertiary sediment thickness are unclear making it difficult to tie early Tertiary basins to a specific topographic load. With the exception of evidence for rotation of Xining basin at ~41 Ma (Dai et al., 2006; Dupont-Nivet et al., 2008) and unroofing of a segment of the west Qinling fault beginning at 45 – 50 Ma (Clark et al., 2010; Duvall et al., in review), there is a dearth of empirical evidence that points to the growth of a major mountain belt in the region during the early Tertiary. Finally, the margins of the broad early Tertiary basins in the Xining-Lanzhou region are poorly exposed, such that basin fill cannot be directly linked with a specific geologic structure. One aspect of early Tertiary mass accumulation that is clear is that it occurred slowly, at rates of ~10 m/Ma (Figure 4.25). Moreover, it is part of a longer-lived episode of slow sediment accumulation that began with
rapid subsidence in the latest Jurassic/earliest Cretaceous, followed by slow subsidence during the early Cretaceous and Paleogene (e.g. Horton et al., 2004). By invoking this subsidence history, as well as paleocurrent data, previous workers have interpreted these patterns to reflect late Jurassic extensional deformation, followed by post-rift thermal subsidence in the Cretaceous and early Tertiary (Horton et al., 2004). These authors have suggested that post-rift thermal subsidence persisted until the onset of contractional deformation, the later part of the Eocene (Horton et al., 2004).

Basin initiation from 30 – 10 Ma across the interior of northeastern Tibet seems to be more definitively linked to contractional tectonism. Middle Tertiary basin formation is somewhat asynchronous across the region, but it overlaps with evidence for emergence of the Laji Shan and Jishi Shan fault networks, to the east of Gonghe basin (Hough et al., 2011; Lease et al., in review) (Figure 4.25). Similar to early Tertiary basin evolution, the mechanism for accommodation creation is somewhat enigmatic. There is not clear evidence for the development of a broad topographic load at this time. One possible topographic load is the broad region of mountainous topography that extends from the west Qinling Shan to the Anyemaqen Shan to the Kunlun Shan (Figure 4.1). Although a segment of the west Qinling fault records evidence for early Eocene deformation (Clark et al., 2010) and a segment Kunlun Shan records evidence for late Eocene-early Oligocene accelerated exhumation (Mock et al., 1999), the growth of this mountain chain is poorly defined. A second possible topographic load is the Qilian Shan to the northwest. Again, although there is evidence for late Eocene-Oligocene basin development on the flanks of the Qilian Shan (e.g. Bovet and Ritts, 2009; Kent-Corson et al., 2010), the development of the region as a whole is poorly known. To further complicate the issue, regional patterns of Tertiary sediment thickness in interior northeastern Tibet are not well known, such
that it is difficult to link depocenters to topographic loads. The staggered pattern of basin
initiation across the region may suggest that subsidence did reflect the development of a broad,
coherent orogenic belt, but instead reflected the growth of individual, relatively small mountain
ranges, and the development of localized depocenters, which slowly coalesced through time. An
alternative explanation for driving mechanism is the development of a topographic barrier
around interior northeastern Tibet, such that the region became internally drained. This scenario,
however, is somewhat difficult to reconcile with the asynchronous pattern of basin initiation
during the middle Cenozoic.

The period since 10 Ma appears to represent a distinct tectonic episode in the history of
northeastern Tibet, during which narrow, elongate mountain ranges emerged from previously
broad regions of sediment accumulation. Herein, we have presented evidence for thrusting along
the margins of Gonghe basin complex during the late Miocene. Ranges bounding nearby basins
record a similar episode of deformation. Sediment accumulation curves from NW Qaidam
suggest renewed mountain building along the basin margin at ~8 Ma and provenance analysis of
detrital zircons suggests emergence of the Laji Shan along the northern margin of Guide basin at
c. 8 Ma. This episode of late Miocene mountain building was not restricted to the interior of
northeastern Tibet. Mineral cooling ages and stratigraphic archives record accelerated
exhumation within the northern Qilian Shan and the Liupan Shan, between ~8 – 10 Ma (Bovet
and Ritts, 2009; Zheng et al., 2006, 2010; Jolivet et al., 2002). The late Miocene change in
structural style around the region may be somewhat analogous to the transition from the Sevier
to the Laramide Orogeny along the eastern flank of the North American Cordillera (e.g. DeCelles
et al., 2004).
The timing, magnitude and style of mountain building in northeastern Tibet place important limits on interpretations of the geodynamic processes that drive plateau growth in the region. First and foremost, end-member geodynamic models that invoke crustal thickening as the key process for building high topography predict progressive outward plateau growth (e.g. Clark and Royden, 2000; Tapponnier et al., 2001). Given synchronous deformation from southern Gonghe to the plateau margins, progressive outward growth cannot resolved at the scale of the entire northeastern Tibetan plateau. Moreover, structural analysis of the Gonghe region suggests a relatively modest amount of upper crustal shortening in the region since the Miocene. Although quantitative budgets of shortening remain elusive for other structures in northeastern Tibet, the presence of preserved AFT partial annealing zones in the Liupan Shan (Zheng et al., 2006) and partial retention zones in the Laji Shan (Lease et al., 2010) suggests that structural relief in other parts of the region is also relatively modest. Thus, if thickening of the upper crust has been a major driver of plateau growth, then this process must have occurred primarily in the lower crust (e.g. Clark and Royden, 2000).

New, or renewed, contractional tectonism across the entirety of northeastern Tibet, over length scales of 100s to 1000s of km, since 10 Ma is consistent with suggestions that late Miocene deformation in the region was pulse-like. If so, this would bolster interpretations of an increase in potential energy associated with removal of mantle lithosphere beneath the plateau during the late Miocene (Molnar and Stock, 2009). Certainly, other marginal regions of the plateau exhibit evidence for topographic growth since ~8 – 10 Ma. Rapid unroofing of the Longmen Shan along the eastern margin of the plateau (Kirby et al., 2002) occurred during the late Miocene and surface uplift of the broad, gently dipping southeastern plateau margin (Schoenbohm et al., 2004; Clark et al., 2005, 2006) occurred at a similar time. Moreover, this
time appears to correspond with the initiation of new structures beyond the plateau margins, from the Indian Ocean to the Tian Shan (see Molnar, 2005 and references therein). Importantly, geodynamic interpretations invoking convective removal of the mantle lithosphere suggest a relatively modest amount of crustal shortening associated with the rise of the interior portions of the plateau. Certainly, the structural reconstructions from the interior of Gonghe basin are consistent with this prediction.

4.8. Conclusion

We have presented important new constraints on the style, timing, rates, and magnitude of upper crustal shortening during the Himalayan orogeny in interior northeastern Tibet.

1) Structural and geomorphic markers facilitate detailed analysis of the structural architecture of two south vergent networks of imbricate thrust faults along the margins of the Gonghe basin complex, the QNS and the GNS. The QNS has accommodated ~1-2 km of upper crustal shortening since the late Miocene, and the GNS has accommodated ~5-7 km over a similar time frame. Geologic cross sections, as well as apatite fission track and (U-Th)/He thermochronometers reveal no more than 1 – 2 km of structural relief has developed on the QNS and no more than 3-4 km has developed on the GNS during this time.

2) The structural architecture of the QNS and the GNS suggests that they sole into decollements in the upper or middle crust. The decollement depth for the QNS seems to be at least 10 km, although it may be tens of km deeper. In tandem with the pronounced south-vergent asymmetry of the QNS, the architecture of the fault network appears analogous to other mountain ranges in intracontinental settings, such as the Laramide ranges of western North America. The decollement depth of the GNS is shallower, perhaps 4 – 7 km. This is similar to
the thickness of Triassic Songpan-Ganzi flysch deposits in the vicinity of southern Gonghe, and suggests that the basal contact of the flysch deposits may set the decollement depth of the range

3) By integrating apatite fission track and (U-Th)/He thermochronology with geologic constraints on the timing of deformation around the margins of Gonghe (Chapter 3), we evaluated the tectonic history of the QNS and GNS during the Cenozoic. Age-elevation relationships as well as inverse modeling of thermal data suggest slow cooling rates in the QNS during the Paleogene, on the order of 1°C/Ma. In tandem with geologic evidence suggest Miocene range growth, it appears that the range was quiescent throughout much of the Tertiary, and emerged at 6 - 10. Although thermal constraints on the Paleogene history of the GNS are relatively weak, geologic evidence from that range suggests that is has a similar history to the QNS. Any episode of significant early Tertiary deformation in northeastern Tibet appears to be confined to the periphery of the region.

4) Although they are subject to uncertainty, late-Quaternary slip rates across the western QNS are ~0.1 mm/yr, and mean geologic shortening rates across the range are ~ 0.1 -0.4 mm/yr. The correspondence between the two rates suggests steady rates of contractional deformation during the late Cenozoic.

5) By integrating our detailed structural analysis of the QNS and GNS with reconnaissance level geologic mapping, seismic reflection surveying in Qinghai Lake basin, and topographic analysis of a possible large, blind thrust fault in the Anyemaqen Shan, we are able to assemble a budget of late-Cenozoic upper crustal shortening over a spatial scale of 250 km. We find evidence for about 4% upper crustal shortening since the late Miocene across the region. The modest shortening that occurred across interior northeastern Tibet during the late Cenozoic can not account for the topographic growth of the region, and requires either significant pre-
Cenozoic crustal thickness or a mechanism to compensate high topography in the lower lithosphere.

**Supplementary information**

**Reconstructing fault slip based on topographic surveys of fault scarps**

In order to assess the magnitude of displacement recorded in the fault scarp, it is necessary to precisely determine the slope of both the upper and lower surfaces. Small differences in surface slope can result in significantly different slip estimates when restoring the upper surface down the fault plane to its initial position. The estimated fault slip is also sensitive to the assumed fault dip. In order to obtain a fault slip estimate that accounts for uncertainties in both the surface slopes and fault dip, we developed a script that employs monte carlo statistics to calculate horizontal, vertical and total fault displacements, based on topographic survey data and incorporating uncertainties in surface slope and fault dip (e.g. Thompson et al., 2002). We assume that fault dip may be anywhere from 20° to 45°, and that any fault dip within this range is equally probable. This range of values is typical of thrust faults. It is also consistent with earthquake focal mechanisms from thrust faults in the GNS the region to the south (Figure 2), and with observed fault dips in the GNS. Linear regression analysis of the topographic survey data was used in order to determine the slope of the upper and lower surfaces. Goodness of fit ($r^2$) statistics are used to ascribe an uncertainty to our regression. We assume that possible surface slopes will be distributed normally about the best-fit slope, and we calculate the standard deviation of the slope is determined using the $r^2$ statistic.

**Cosmogenic laboratory methods**

Samples were subjected to several physical and chemical treatments designed to reduce the raw material to pure quartz, and to extract Be isotopes from the purified quartz. First,
samples were crushed and sieved, in order to obtain a desirable grain size for the remaining treatments. In order to remove carbonates and minor metals, the crushed material was leached in nitric acid and aqua regia. Next, the sample was subjected to a suite of physical separation steps which were: froth flotation, magnetic separation, and following a purification bath in a hydrofluoric acid/nitric acid solution, heavy liquid separation. The remaining material was soaked for a second time in a hydrofluoric acid/nitric acid solution to remove any remaining feldspars. During this step, the outermost layers of the quartz grains were dissolved to remove meteoric $^{10} \text{Be}$. After completing this routine, Al concentrations were measured on an inductively coupled plasma optical emissions spectrometer (ICP-OES) to assess the purity of the remaining quartz. If the measured Al concentration (which signifies the presence of residual feldspars) exceeded 200 ppm, the final step was repeated as necessary.

In order to extract Be and Al isotopes from the purified quartz samples, a second series of chemical treatments was applied. After adding Be and Al carriers, quartz was dissolved in concentrated hydrofluoric acid. Following dissolution, an Al aliquot was extracted from the solution and prepared for precise measurement on the ICP-OES. The volume of the solution containing the dissolved sample was reduced and the hydrofluoric acid was removed by a series evaporation and fuming steps. The residual material was taken up in a sodium hydroxide solution, centrifuged, and decanted in order to separate Fe and Ti ions from the solid residual sample. Next the pH of the remaining solution was adjusted to ~8 to precipitate the Al and Be out of the solution as hydroxides (Ochs and Ivy-Ochs, 1997). After dissolving the remaining hydroxides in oxalic acid, cation and anion columns were used to removed residual Na, Fe, and other undesired ions, and to isolate Be and Al. The samples were dried and fired in an oven, and
then loaded into a cathode for accelerator mass spectrometry (AMS). AMS was conducted at PRIME lab at Purdue University, following standard protocols.
Figure 4.1. a) Quaternary faults and Cenozoic basins in northern Tibet. Inset shows GTOPO-30 digital topography of Tibetan plateau and Quaternary faults, adapted from Tapponnier and Molnar, 1977; Molnar and Tapponier, 1978. Grey dashed lines are terrane boundaries. JS = Jinsha suture, AS = Anyemaqen suture, SQS = South Qilian suture, DHS = Danghe Nan Shan suture, NQS = North Qilian suture, NCS = North China suture. Adapted from Yin and Harrison, 2000 and Xiao et al., 2009 and references therein. b and c) Maximum, minimum and mean swath topography, derived from GTOPO-30 data, which has a nominal resolution of 1 km. Moho depths are also shown. For b, moho depths are from Liu et al., 2006. For c, moho depths from the Anyemaqen and Gonghe are from Vergne et al., 2002, and depths from the Qilian Shan are from Meyer et al., 1998.
Plio-Quat. - Late Miocene - fluvial-floodplain
Early Miocene - fluvial floodplain
Cenozoic strata-undiff.
Cretaceous intramontane basins
Cretaceous strata
early Mesozoic sediments
Jurassic strata
Triassic-Songpan-Ganzi flysch
Permian-Songpan-Ganzi flysch
Paleozoic island arc terrane
Carboniferous strata
Devonian strata
Silurian strata
Ordovician strata
Triassic backarc
Permian-backarc
Pz - e. Mz plutons
Paleozoic south Qilian arc
Precambrian massifs

Figure 4.5
Figure 4.6

Figure 4.2. Geologic map of the Gonghe basin complex. Geology adapted from QBGMR, 1991 and field observations.
Figure 4.3. Existing constraints on the age of the Cenozoic basin fill in the Gonghe basin complex, and a comparison to the stratigraphic units in Guide basin and northeastern Qaidam basin, adapted from Fang et al., 2005, 2007, Chapter 3.
Figure 4.4. Lithostratigraphic architecture of the Gonghe basin complex. The upper picture strikes N-S, and extends from the southern edge of Tongde basin to the northern edge of Qinghai Lake basin. The lower picture strikes E-W and extends from the Chaka subbasin in the west to Guide basin in the east. Stratigraphic sections in central Gonghe subbasin modified from Zheng et al., 1985. Stratigraphic section in Chaka subbasin modified from Zheng et al., in review. Stratigraphic sections in Qinghai Lake basin modified from An et al., 2006. Stratigraphic section in Guide basin modified from Fang et al., 2005.
Figure 4.5. Geology and topography of the Qinghai Nan Shan. See Figure 4.2 for explanation of tectonostratigraphic units. Geologic map is draped by a hillshade image generated from 90-m Shuttle Radar Topography Mission (SRTM) digital topography. Sites labeled G exhibit growth strata, sites labeled P exhibit progressive unconformities, and sites labeled O exhibit an onlapping relationship between basin fill and bedrock.
Figure 4.6. Geology and topography of the Gonghe Nan Shan region. See Figure 4.2 for explanation of tectonostratigraphic units. Geologic map is draped by a hillshade image generated from 90-m SRTM digital topography. Open circles show burial age sites. Sites labeled P exhibit progressive unconformities, sites labeled G exhibit growth strata, sites labeled A exhibit angular unconformities, and sites labeled O exhibit an onlapping relationship between basin fill and bedrock.
Figure 4.7. Serial deformed state cross sections through the Qinghai Nan Shan. Location of cross sections is shown in Figure 4.5.
Figure 4.8. Serial deformed state cross sections through the Gonghe Nan Shan. The location of the cross sections is shown on Figure 4.6. The location of thermochronology sites is shown by closed black circles.
Figure 4.9. Detailed geologic map of the QNS range front in Chaka subbasin.
Figure 4.10. Detailed map of the eastern QNS range, east of Gonghe city.
Figure 4.11. a) Deformed alluvial fan surfaces along the Chaka range front. Although several generations of surfaces are visible, we only delineate the well preserved portions of the uplifted alluvial surface, and the active channel. Open rectangle shows the location of topographic profiles through the scarp. b) Stratigraphy of the upper 3 m of the uplifted deposition surface along the QNS range front. Most of the pit is clast supported alluvial fan conglomerate. The upper 30 cm contains loess, which we interpret to be the result of surface inflation. A wavy, caliche band, with an amplitude of 5-10 cm is located at ~30 cm depth. We show the positions of cosmogenic depth profile samples to the right of the stratigraphic column.
Figure 4.12. Detailed map of the GNS in the Yellow River canyon area.
Figure 4.13. Angular unconformity between M2 and PQ in southern Tongde basin, south of the town of Xinghai. See Figure 4.7 for location of outcrop.
Figure 4.14. Geomorphic analysis of the Qinghai Nan Shan. a) Shuttle Radar Topography Mission (SRTM) 90-m digital elevation model. Locations of topographic profiles A-D is shown. b) Map of surface slope derived from digital topography. c) Topographic relief measured within a moving window with a 1 km radius. d) Topographic profiles A-D. The profiles show maximum, minimum, and mean topography measured over 30 km wide swaths. Areas highlighted in grey show the preserved erosion surface along the north limb of the range, where maximum, minimum, and mean topography are nearly coincident.
Figure 4.15. Geomorphic analysis of the Gonghe Nan Shan. a) Shuttle Radar Topography Mission (SRTM) 90-m digital elevation model. Locations of topographic profiles A-D are shown. b) Map of surface slope derived from digital topography. c) Topographic relief measured within a moving window with a 1 km radius. d) Topographic relief measured within a moving window with a 5 km radius. e) Topographic profiles A-D. The profiles show maximum, minimum, and mean topography measured over 30 km wide swaths. The vertical black bar in B-D show the location of the prominent topographic break in the central part of the range.
Figure 4.16. Geomorphic analysis of the Anyemaqen Shan, to the south of Tongde basin. a) Shuttle Radar Topography Mission (SRTM) 90-m digital elevation model. Locations of topographic profiles A and B are shown. b) Map of surface slope derived from digital topography. c) Topographic relief measured within a moving window with a 1 km radius. d) Topographic relief measured within a moving window with a 5 km radius. e) Topographic profiles A and B. The profiles show maximum, minimum, and mean topography measured over 30 km wide swaths. The vertical dashed line marks the trace of the prominent lineament that is visible in the various map view images. We infer this lineament to be a blind fault, and estimate the minimum vertical displacement on the fault using the topographic profiles.
Figure 4.17 Fission track length histograms for samples from the GNS and QNS.
Figure 4.18. Galbraith plots for fission track samples from the GNS and QNS.
Figure 4.19. Age-elevation profiles for thermochronologic data from the Qinghai Nan Shan and the Gonghe Nan Shan. Closed circles show AFT cooling ages and open circles show AHe cooling ages. AFT ages are pooled ages and error bars represent 2σ uncertainties. AHe ages are mean ages and error bars show the range of the data. Effective Uranium concentration plotted against mean AHe age for sample CT8-6 from the central Qinghai Nan Shan vertical transect.
Figure 4.20. Inverse model of the time-temperature (tT) history for sample CT8-6 from the central Qinghai Nan Shan vertical transect. See text for a discussion of geologic constraints imposed on the thermal history. Modelling was conducted using HeFTy (Ketcham et al., 2005). Black curves represent both acceptable and good fits to the thermochronologic data, in a least squares sense.
Figure 4.21. Inverse models of time-temperature (tT) history of the Gonghe Nan Shan, in which 10000 random thermal histories are realized and evaluated using least squares statistics. Modelling was done in the program HeFTy (Ketcham et al., 2005). Each segment is divided into 4 subsegments. Monotonic consistent paths between segments.
Figure 4.22. High resolution topographic profiles measured across deformed alluvial fan surface in the proximal footwall of the Qinghai Nan Shan near the town of Chaka. The location of the topographic surveys is shown in Figure 4.9.
Figure 4.23. Measured concentration depth profile and least squares optimization of synthetic surface abandonment scenarios. a) Measured cosmogenic $^{10}$Be inventories are shown in black. Horizontal bars represent $1\sigma$ uncertainties. Grey curves represent synthetic surface abandonment scenarios, which are evaluated using least squares statistics. b) Chi-squared statistics for various abandonment ages. The grey envelope represents a cloud of points, with each point corresponding to a synthetic surface abandonment scenario as in part a. Although a 114 ka abandonment age is optimal, is only slightly better than abandonment ages as low as 109 ka and as high as 142 ka. c) Chi-squared statistics for various surface lowering rates. The grey envelope represents a cloud of points, with each point corresponding to a synthetic surface abandonment scenario as in part a. The optimal surface abandonment scenario does not appear to be sensitive to the range of surface erosion rates that we allow.
Figure 4.24. Crustal scale cross section extending from the Anyemaqen Shan, across the Gonghe basin complex, into northern Qinghai Lake basin.
Figure 4.25. Compacted sediment thickness versus depositional age for stratigraphic sections from around Northeastern Tibet.
Figure 4.26. Paleogeographic reconstruction of northeastern Tibet throughout the Cenozoic Era.
Table 4.1. Locations, cooling ages and uncertainties for thermochronology samples.

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*Pooled age is reported for AFT data

**2σ standard error reported for AFT data
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Table 4.2. Shortening budgets and decollement depth calculations.
Rapid fluvial incision along the Yellow River during headward basin integration

The topographic growth of the Tibetan Plateau is central to ongoing debates over the linkages between mountain building and global climate (Molnar, 2005). Recent studies utilize the erosional response to surface uplift as a proxy for the development of high relief along the eastern margins of the plateau (Kirby et al., 2002; Schoenbohm et al., 2004; Clark et al., 2005). However, processes that retard incision, such as spatially-variable rock uplift (Zeitler et al., 2001), orographic changes in rainfall (Sobel et al., 2003), and/or damming by glaciers or landslides (Ouimet et al., 2007; Korup and Montgomery, 2008) may lead to exhumation that significantly lags surface uplift (van der Beek et al., 2009). Here we evaluate the timing, rates and patterns of fluvial incision along the Yellow River, in northeastern Tibet using a combination of stratigraphic, geochronologic, and geomorphic data from Plio-Quaternary basins along the present-day river. Our results reveal that the transition from basin filling to excavation significantly postdates Late Miocene range growth in the region (Lease et al., 2007). Moreover, new constraints on the timing of this transition in basins interior to the plateau reveal that the onset of incision was time-transgressive along the Yellow River, initiating near the plateau margin at ca. 1.8 Ma (Li et al., 1997) and progressing upstream (Harkins et al., 2007) at ~350 km Ma⁻¹. Rather than a response to surface uplift of the plateau (Li, 1991; Fang et al., 2005), our work implies that fluvial incision was the result of climatically-
driven expansion of lake systems (Fang et al., 2003; Wu et al., 2007) that led to spillover and integration of the modern Yellow River.

Along the northeastern margin of Tibet, the Yellow River is deeply incised (>600m) into a series of intermontane basins and bedrock ranges (Figure 5.1). The space-time patterns and ultimate cause of incision, however, remain unknown. Basin stratigraphy suggests slow subsidence (~10 m Ma⁻¹) across much of the region from the Cretaceous until at least the late Oligocene (Horton et al., 2004), followed by an abrupt increase during the mid to late Miocene (Fang et al., 2003). Concomitant changes in sediment provenance (Garzione et al., 2005; Lease et al., 2007), paleosol isotopic composition (Dettman et al., 2003), and an increase in exhumation rate within the bounding ranges (Zheng et al., 2006) attest to the onset of significant mountain building beginning at ~14 – 8 Ma. Subsequent sediment accumulation was dominated by local sources (Lease et al., 2007), and basins were hydrologically isolated by intervening bedrock sills (Métevier et al., 1998).

A pronounced transition from basin filling to excavation (Li et al., 1997) appears to be linked to the establishment of external drainage along the modern Yellow River. The youngest strata in both the Linxia (Fang et al., 2003) and Guide (Fang et al., 2005) basins suggest that lacustrine deposition persisted until at least ~1.7 – 1.8 Ma. Along the margin of the plateau, the onset of basin excavation is tightly bracketed by fluvial terraces related to the Yellow River that are dated to ca. 1.7 Ma (Lanzhou and Linxia basins, Figure 5.1) (Li et al., 1997; Zhu et al., 1995). Based on the age of the youngest preserved lacustrine sediments farther upstream in the Guide basin (~1.7 Ma) (Fang et al., 2005), many have inferred that incision began simultaneously along this reach of the
Yellow River and attribute this to widespread surface uplift of the northeastern Tibetan Plateau (Li, 1991; Li et al., 1997; Fang et al., 2005). However, the absence of direct age control on fluvial terraces in the Guide region allows the possibility that incision began more recently. Indeed, dating of high terraces in the Xunhua (>1.1 Ma) (Pan et al., 1996) and Xining (>1.2 Ma) (Zeng et al., 1995) basins (Figure 5.1) hint that incision upstream of the plateau margin may be somewhat younger.

Recent analysis of longitudinal river profiles along tributaries in the headwaters of the Yellow River reveals an ongoing pulse of transient fluvial incision across the Anyemaqen Shan (Harkins et al., 2007) (Figure 5.1). Whether this increase in incision rate is linked to basin excavation downstream or whether it is related to local deformation (Harkins et al., 2007), however, remains uncertain. To elucidate the time-space patterns of incision along the course of the Yellow River in northeastern Tibet, we conducted new stratigraphic, geomorphic, and chronologic studies of the transition from basin filling to excavation in the Gonghe basin complex (Figure 5.1). Gonghe basin is located approximately 600 km upstream from the plateau margin, and thus provides a critical link between incision in the headwater reaches (Harkins et al., 2007) and basins near the plateau margin (Li, 1991). The Gonghe basin complex is comprised of southern and northern sub-basins (herein referred to as the Tongde and Gonghe basins, respectively), separated from one another by an actively-growing E-W mountain range. The modern course of the Yellow River flows through a water gap in this range that forms a bedrock sill (Figure 5.1b).

We measured and logged stratigraphic sections exposed along the Yellow River in both the Tongde and Gonghe basins (see Supplementary Figure 5.1) in order to
investigate the character of the uppermost basin fill. Our results indicate that the final, Plio-Quaternary, stages of basin filling were dominated by aggradation along a proto-Yellow River system. In Tongde, the basal strata consist of mudstone with lenses of sand and gravel (Figure 5.2) that coarsen upward to dominantly gravel, with imbricated clasts and cross-stratification. We interpret these deposits to be fluvial. Moreover, the paleodepositional surface atop the basin fill is on grade between the Tongde and Gonghe basins (Figure 5.1c), suggesting that both sub-basins shared a common baselevel. Fluvial deposits grade laterally into nearshore lacustrine and lacustrine strata in the center of the Gonghe basin (sensu stricto) (Zheng et al., 1985), suggesting that the bedrock sill between the Gonghe and Guide basins was a barrier to transport (Métevier et al., 1998).

To determine the age of these deposits, we use a combination of biostratigraphy, magnetostratigraphy, and cosmogenic isotope burial ages (Figure 5.2). Details of the analytical procedures are presented in the Supplementary Information. Although previous studies of faunal mammal assemblages assign an age range of Pliocene – Middle Pleistocene (Zheng et al., 1985), a precise chronology of basin fill is unknown (see Supplementary Information and Supplementary Table 5.1). At the base of the Tongde section, a newly discovered assemblage of mammalian fossils, including a well-preserved fossil rodent skull identified as an adult male member of the species *Allosiphneus teilhardi* [Chinese zokor], place the base of the section in the Late Pliocene, ~3.4-2.0 Ma. Near the top of the section, we obtained two samples of quartz-rich fluvial sand from road cuts for burial dating using the cosmogenic radionuclides $^{10}$Be and $^{26}$Al. We restricted our sampling to deep road cuts where geometry of the relict deposits shows that samples were covered by >10 m of basin fill prior to road construction.
(Supplementary Figure 5.2). Possible combinations of burial age and pre-burial inheritance were used to generate numerous synthetic, present-day $^{26}\text{Al}/^{10}\text{Be}$ ratios and $^{10}\text{Be}$ concentrations. Synthetic values were compared to measured values using a least squares optimization in order to determine the burial age for the samples (see Supplementary Information). The resultant ages of $\sim0.7 \pm 0.3$ Ma and $\sim1.2 \pm 0.4$ Ma account for muonogenic isotope production (Granger, 2006) and the uncertainties reflect the range of possible burial/exhumation histories for the samples. The ages constrain the top of the basin fill to the Middle Pleistocene, and suggest an average sedimentation rate of $\sim80 – 90$ m Ma$^{-1}$ (Figure 5.2c-see Supplementary Table 5.2).

We resolved the chronology of basin fill using magnetic reversal stratigraphy recorded in the deposits. Our preferred magnetostratigraphic correlation spans the Brunhes, Matuyama, and upper part of the Gauss polarity chrons (Figure 5.2a). The stratigraphic interval between reversals indicates that temporally averaged sedimentation rates were $\sim104$ to $125$ m Ma$^{-1}$ (Figure 5.2b), consistent with the sedimentation rate derived from the burial age samples (see Supplementary Figures 5.3 and 5.4, and Supplementary Table 5.3). Projecting sediment accumulation rates from the base of the Jaramillo subchron to the top of the section yields an age of $\sim0.35$ Ma. Given that the upper burial age is $0.7 \pm 0.3$ Ma, a reasonable estimate for the age of the top of the section is $\sim0.5 \pm 0.2$ Ma.

To elucidate the subsequent history of incision along the Yellow River in the Gonghe region, we studied fluvial terrace deposits below the top of the Tongde basin fill. Terraces are strath surfaces inset into older basin fill, and characterized by 1-3 m of unconsolidated deposits of imbricated gravel, that overlie indurated sandstones and
conglomerate and are, in turn, mantled by overbank silt and eolian loess. Often, fluvial terraces are associated with tributary alluvial fan complexes that grade to the terrace level and suggest a protracted history of lateral erosion and incision. We obtained an optically-stimulated luminescence (OSL) age of 141 ± 9 ka from loess accumulated on a 10 m thick channel lag deposit atop a fluvial strath terrace of the Yellow River (Harkins et al., 2007) (Figure 5.3). The strath beneath this terrace is ~140 m above present river level, indicating an average incision rate of ~1 m ka⁻¹. Assuming this represents the mean incision rate, the position of the terrace ~370 m below the top of the basin fill implies that incision initiated at ~500 ka, a result consistent with the chronology of the basin fill.

We also obtained radiocarbon ages from charcoal and freshwater shell material in floodplain deposits associated with terrace complex 2 (Figure 5.3, Supplementary Table 5.4). Results suggest that the higher terrace, ~44 m above river level, is early Holocene and that incision following terrace abandonment was rapid, ~4 m ka⁻¹. The lower terrace (~24 m above river level) dates to the Late Holocene, and suggests extremely rapid incision (~9-12 m ka⁻¹). Although these rapid incision rates are consistent with transient incision, we caution that these are upper bounds, in that they only reflect abandonment of the terrace tread, and do not reflect the time required to bevel the terrace tread itself (Wegmann and Pazzaglia, 2002).

When considered in a regional context, our results have several important implications. First, our new chronology of Gonghe basin fill reveals a pronounced diachroneity in the behavior of the Yellow River during the Quaternary. Basin filling in the Gonghe region appears to have been driven by aggradation along a proto-Yellow River that persisted to ~500 ka. However, aggradation in the Gonghe basin was coeval
with fluvial incision and excavation of the Linxia and Xunhua basins near the plateau margin\textsuperscript{11}, and thus our results contradict the notion that incision began simultaneously across northeastern Tibet at ca. 1.8 Ma (Li, 1991; Li et al., 1997; Fang et al., 2005).

Second, our results indicate that excavation of basins along the Yellow River progressed systematically upstream. A regional synthesis of stratigraphic and geomorphic constraints demonstrates that the onset of fluvial incision swept progressively upstream in a remarkably steady fashion, at a rate of $\sim$350 km Ma$^{-1}$ (Figure 5.4). Topographic reconstruction of the amount of sediment excavated from the greater Gonghe and Guide basins provides additional confirmation of this pattern, demonstrating that the volume of eroded material, normalized by stream length, decreases upstream (Supplementary Figure 5.5). Notably, the rate at which this wave of incision migrated upstream seems to have remained steady, even along bedrock reaches of the river in the Anyemaqen Shan (Harkins et al., 2007), upstream of the Gonghe basin. Thus, it appears that the Yellow River, once established, rapidly excavated through both basin sediments and bedrock mountain ranges across northeastern Tibet, implying that negative feedbacks such as landslide dams or glaciers were not a significant impediment to headward erosion as they may have been in other parts of Tibet (Ouimet et al., 2007; Korup and Montgomery, 2009).

Third, fluvial incision along the Yellow River significantly lags the growth of high topography in northeastern Tibet. Proxy data reveal that individual ranges bounding the Gonghe, Guide and Linxia basins began to grow at 14 – 8 Ma (Dettman et al., 2003; Garzione et al., 2005; Zheng et al., 2006; Lease et al., 2007), and subsequent sediment accumulation is consistent with models of isolated basins ponded behind rising ranges.
These basins appear, however, to have remained isolated throughout the final stages of basin filling (Métevier et al., 1998) and hosted a chain of shallow lakes into the Late Pliocene and Early Pleistocene (Li et al., 1997; Fang et al., 2003, 2005). Final integration of the Yellow River, and excavation of these basins, postdated the initiation of widespread mountain building across the region by nearly 10 Ma.

Finally, our results suggest linkages between climate and erosion that may control the dissection of high topography along orogenic plateaus. Although the exact mechanism for triggering basin integration at 1.8 Ma remains uncertain, the temporal coincidence of the onset of fluvial incision along the Yellow River and the onset of northern hemisphere glaciations (Molnar, 2004) is striking. Cooling and aridification of the northeastern Tibetan plateau, inferred from fossil pollen assemblages (Wu et al., 2007), is coincident with widespread lacustrine deposition in Guide and Linxia basins, from ~2.6 – 1.8 Ma (Fang et al., 2003, 2005). We suggest that cooler climates likely facilitated lake expansion, and that at ~1.8 Ma this expansion culminated with spillover into downstream basins that initiated fluvial incision and established the course of the modern Yellow River.

5.1 Supplementary Information

5.1.1. Lithostratigraphy

5.1.1.1. Lithofacies description and interpretations

*Lithofacies A-Mudstone with lenses of cross-beded siltstone and sandstone-

Lithofacies A is restricted to the lower portions of the Plio-Quaternary basin fill in the Gonghe basin complex (Supplementary Figure 5.1).
The coarser beds are lenses of light tan or reddish-tan siltstone and fine to medium sandstone. The lenses are 0.3 - 1.5 m thick and continuous over 10s to 100s of m. 1-2 meter thick intervals of amalgamated or closely spaced sand lenses are separated by muddy intervals that are several meters thick. The beds exhibit erosional bases such that they are inset into underlying beds. Relief along the base of these beds is typically on the order of several cm to several decimeters. Sand beds exhibit parallel bedding and lamination and trough cross bedding. Coarse or granular sand is found along the base of beds or draping cross beds. Orange mottling is commonly found along planar laminae or cross strata.

We interpret these beds to be channel deposits associated with fluvial transport systems. The prevalence of cross-bedding and parallel bedding/lamination indicates transport by a unidirectional current. The lenticular geometry of the beds and the channel scours indicate that the flow was channelized.

The finer deposits are tabular intervals of reddish-tan mud, surrounding the coarse lenses. Mud beds are typically massive in appearance, but occasionally exhibit parallel lamination. Orange mottling is found along laminae. Mud prone intervals contain horizons that are characterized by bands of siltstone lenses and balls. The beds contain carbonate nodules and burrows.

We interpret these beds to be overbank deposits. The presence of carbonate nodules indicates pedogenesis occurred during periods of subaerial exposure on floodplains. Burrowing also indicates prolonged periods of subaerial exposure. The fine grain size of these beds is consistent with the inference that the sediments settled out of suspension on an inundated floodplain. Parallel lamination may have formed as the
sediments settled from suspension. Silt balls and lenses in mud prone intervals are interpreted to be crevasse splay deposits that were broken into discontinuous balls and lenses during burial and compaction.

**Lithofacies B-Interbedded sandstone pebble conglomerate with trough cross-bedding and imbrication** - Lithofacies B is distributed widely across the Gonghe basin complex, particularly in the lower Tongde basin fill and across much of the central part of Gonghe basin.

The coarse grained beds are light tan, silty, fine and medium sandstone with pebble orthoconglomerate lenses and mud lenses. Sand and gravel beds are lenticular and laterally continuous over 1-100s of m and 10 cm to 2 m thick. Sand beds occur in intervals that are several m thick, and the beds may be amalgamated or have mud partings. The sand intervals are separated by mud intervals that are several tens of cm thick. 1 or 2 gravel lenses are typically interbedded with the sands. Beds have curved, erosional basal contacts with a few to several tens of cm of relief. Clasts in conglomerate beds are dominantly sandstone and quartz. Sedimentary structures include imbrication, trough cross bedding, parallel bedding and lamination, and channel scours. Orange mottling is common in finer beds, and it is found along fractures, laminae, or randomly.

We interpret these beds to be channel deposits. The imbrication, cross bedding, and parallel bedding/lamination formed in a unidirectional current. The lenticular geometry, as well as the presence of channel scours, indicates transport and deposition by a channelized flow.

Finer grain beds are composed of light tan siltstone and fine to medium sandstone. Beds are massive or parallel laminated and tabular. Ginger colored concretions and
calcite cemented horizons are commonly found in this unit. Orange mottling is prevalent in the fine layers.

We interpret these beds to originate as floodplain or crevasse splay deposits. Parallel lamination may indicate that the sediments settled out of still water, perhaps on a floodplain. The tabular geometry of the unit is also consistent with deposition on a broad floodplain. The calcite cemented horizons and ginger colored concretions formed as a result of pedogenesis during periods of subaerial exposure.

*Lithofacies C-Imbricated, clast supported, pebble and cobble conglomerate-

Lithofacies C outcrops along the flanks of the Gonghe Nan Shan and along the southern edge of the Qinghai Nan Shan.

Lithofacies C is composed of pebble or cobble clast supported conglomerate with sandy and silty lenses. Bedding is lenticular and laterally continuous over a few to 100s of m, and tens of cm to a few m thick. Sand lenses are separated by a few meters, they exhibit decimeter scale thickness, and they are a few m to tens of m wide. The matrix is brown and the clasts are grey. Clasts comprise sandstone and quartz. In certain locations near the modern Yellow River, the clasts comprise a variety of rocks, including granite, sandstone, quartz, metamorphic rocks, and pebble conglomerates with a green matrix. Sedimentary structures include imbrication, channel scours, large (10-100 cm high) cross bedding within gravel beds, and parallel or cross lamination in finer lenses. Relief on channel scours is on the order of 10s of cm.

This lithofacies is interpreted to be a braided channel alluvial fan deposit. The presence of imbrication, cross bedding, and parallel bedding indicates that the unit was deposited by a unidirectional flow. The lenticular geometry of the beds, and the channel
scours along the base of the beds indicate that the flow was channelized. The outcrop distribution of lithofacies C also suggests that it is an alluvial fan deposit. The deposits line the flanks of the mountain ranges in the region, forming a broad alluvial apron. The absence of a) matrix support, b) inverse grading, and c) tabular geometry distinguish these beds from debris flow dominated alluvial fan deposits.

Lithofacies D-Tabular, light tan or yellow, siltstone and sandstone—Exposures of lithofacies D are limited to the north-central portion of Gonghe basin, near Gonghe city. It can be found in the Gonghe north section.

Lithofacies D comprises light tan or yellow, compositionally mature, siltstone and sandstone interbedded with light red or gray-green sandy muds. Sedimentary structures include parallel laminations, and asymmetrical wave ripples. Most beds are massive. Beds are tabular, with parallel contacts, and on the order of a few to several tens of cm thick.

Lithofacies D is interpreted to be a shallow, near-shore lacustrine deposit. The massive bedding and fine grain size of these deposits indicates that the sediments settled out of suspension. Wave ripples imply subaqueous deposition above the wave base. The compositional maturity of the wave-rippled sands implies that muddy sediments were winnowed from these deposits by wave action. The lack of oxidation in this lithofacies is consistent with subaqueous deposition. The gravel interbeds found in the Gonghe north section are attributed to rare, catastrophic events such as floods, turbidites and/or debris flows.
Lithofacies E-Imbricated, clast supported, cobble conglomerate atop fluvial terraces-This lithofacies drapes the Yellow River strath terraces inset into the basin fill in the Gonghe basin complex. It caps the Gonghe north stratigraphic section.

Lithofacies E is composed of imbricated cobble orthoconglomerate. The clasts comprise a variety of rocks, including granite, sandstone, quartz, metamorphic rocks, and green conglomerates. This unit contains sand lenses that are ~1m wide and ~10-30 cm thick.

Lithofacies E is interpreted to be a channel lag deposit, draping the Yellow River terrace surfaces, based on the presence of imbrication and the spatial distribution of the unit.

Lithofacies F-Massive siltstone-This lithofacies drapes the top of most of the Gonghe basin complex. It is found on top of all of the stratigraphic sections presented herein. Lithofacies F is composed of light tan, very well sorted, siltstone. We interpret the massive appearance, sorting, and fine grain size to indicate that these are air fall deposits, or loess.

5.1.1.2. Stratigraphic units

Based on the lithofacies interpretations presented above, we subdivide the Plio-Quaternary stratigraphic units in the Gonghe basin complex into 6 distinct stratigraphic units (see Supplementary Figure 5.1). Unit 1 comprises the lower portion of the Tongde stratigraphic section as well as the central Gonghe stratigraphic section. This unit encompasses lithofacies A, B, and C. Unit 2 consists of most of the braided river dominated alluvial fan deposits (lithofacies C) found in the upper portion of the Tongde stratigraphic section and the lower portion of the of the southern Gonghe stratigraphic
section. Small outcrops of Unit 2 are located along the flanks of the Gonghe Nan Shan and the Qinghai Nan Shan. Unit 3 is located in the upper portion of the southern Gonghe stratigraphic section. It is distinct from Unit 2 because it contains a much wider array of clast types. The clast composition of Unit 3 is identical to the clast composition of modern Yellow River terraces, thus we interpret this unit to be the location of the proto-Yellow River that drove the latest stages of sediment accumulation in the Gonghe basin complex. Unit 3 is found only in outcrops near the modern Yellow River. Units 4-6 are identical to lithofacies D-F.

5.1.2. Age of basin fill

5.1.2.1. Synthesis of existing biostratigraphy

For a synthesis of existing biostratigraphic work on the Gonghe basin complex, please see Supplementary Table 5.1.

5.1.2.2. New biostratigraphy in Tongde basin

A Chinese Zokor skull was found in situ, in the lower part of the Tongde stratigraphic section (Figure 5.2, Supplementary Figure 5.1). Based on the suture morphology around incisive foramen area, nasal width, size of postorbital process, narrowness of postorbital constriction, and the reentrant angle in upper cheek teeth, this skull should belong to an adult male member of the genus *Allosiphneus*. Two species of *Allosiphneus* are presently known: *A. arvicolinus* (Nehring, 1885) and *A. teilhardi* (Kretzoi, 1961). Both are found in and around the Loess Plateau area of central China. Although the Tongde skull is not strictly comparable to *A. teilhardi*, it is referred to this species mainly because of its comparable size. The holotype of *A. teilhardi* was collected from the Jingle area in Shanxi Province, and was from the red loam capped by the Malan...
loess. In Chinese chronology, this is usually in the late Pliocene to early Pleistocene (~2 – 3) Ma. The Jingle Formation has not been precisely dated, but is believed to be around ~2.5 - 3 Ma (Zhang and Liu, 2005). In Lingtai, Gansu Province, *A. teilhardi* occurs in the late Pliocene around 3.4 Ma. Thus, we assign an age of ~2-3.4 Ma to this skull.

### 5.1.2.3. Cosmogenic burial dating

#### 5.1.2.3a. Field techniques

To constrain the burial age of fluvial sediment from the Tongde basin, we collected 2 samples of coarse fluvial sand for analysis of in-situ, cosmogenically produced $^{26}$Al and $^{10}$Be inventories in quartz (Granger and Muzikar, 2001; Granger, 2006). Samples were collected from quartz-rich sand lenses intercalated with the fluvial gravel basin fill. Because the concentrations of cosmogenic $^{26}$Al and $^{10}$Be are strongly dependent on the history of post-burial production within ~10 m of the earth’s surface (Granger, 2001), we targeted samples from the base of modern roadcuts (Supplementary Figure 5.2), where we are able to geometrically constrain sample depth prior to historic road construction. Sampled roadcut exposures are unweathered and clearly exhibit original sedimentary structures of the basin fill. As a result, we are confident that the samples remained in-situ since the time of their deposition and were only recently exposed in the roadcuts.

Both of the samples from Tongde were extracted from the extensive package of intact basin fill alluvium preserved in the central region of the basin. Basin fill alluvium at both sites consists of poorly cemented, predominantly pebble to cobble gravels with minor intercalated lenses of cross-bedded sand. The uppermost sample at Tongde (NHKCOS-YT2) was collected ~14 m below the top of basin fill. Prior to excavation
along the road, this sample was buried to a depth of 13 m below the top of the gravel, which is in turn buried by ~1 m of loess. The top of the loess at this site is on grade with the extensively preserved depositional surfaces that mark the top of the basin fill; we therefore interpret the top of the gravel at this site to represent the uppermost strata within the basin. Undisturbed gravels are preserved on the opposite side of the road from the sample exposure, providing an estimate of the topography of this location prior to road construction. We surveyed these exposures with a laser rangefinder, and the geometry of the preserved basin fill indicates that the sample was shielded by >10.5 m from the canyon sidewall (Supplementary Figure 5.2).

The second sample at Tongde (NHKCOS-YT3) was collected from an elevation ~56 m below the top of the basin fill. Basin fill is horizontally stratified between the two sites, and so we interpret the lower sample site to be ~42 m below sample NHKCOS-YT2. Prior to excavation of the roadcut, this sample resided at least 15 m below the modern land surface. Undisturbed basin-fill gravel on the opposite side of the road from the sampled exposure indicates a nominal shielding depth for this sample of >13 m (Supplementary Figure 5.2).

5.1.2.3b. Sample processing and burial age calculations

Quartz from these samples was isolated and purified using physical and chemical techniques at the Purdue Rare Isotopes MEasurement (PRIME) lab, following standard procedures. Extraction of \(^{10}\)Be and \(^{26}\)Al was accomplished using cation-exchange chemistry, and the resultant isotope concentrations were measured on the AMS, following standard protocols (Supplementary Table 5.2).
For buried sediment that is derived from a steadily eroding source, the concentrations of unstable cosmogenic isotopes (in this case, $^{26}{\text{Al}}$ and $^{10}{\text{Be}}$) will evolve through time as a function of two unknown variables, the pre-burial cosmogenic inventory and the time since burial (Granger and Muzikar, 2001).

\[
(1) N_{Al}(t) = N_{Al}(0)e^{-\frac{t}{\tau_{Al}}} + P_{Al}(d)\tau_{Al}(1 - e^{-\frac{t}{\tau_{Al}}})
\]

\[
(2) N_{Be}(t) = N_{Be}(0)e^{-\frac{t}{\tau_{Be}}} + P_{Be}(d)\tau_{Be}(1 - e^{-\frac{t}{\tau_{Be}}})
\]

where $N_{Al}$ is the number of $^{26}{\text{Al}}$ atoms per gram of quartz, $N_{Be}$ is the number $^{10}{\text{Be}}$ atoms per gram of quartz, $t$ is time, $P_{Al}$ is production rate of cosmogenic $^{26}{\text{Al}}$ in atoms per gram per year, $P_{Be}$ is the production rate of cosmogenic $^{10}{\text{Be}}$ in atoms per gram per year, $d$ is depth in cm, $\tau_{Al}$ is the radioactive mean-life of $^{26}{\text{Al}}$ (1.02 * $10^6$ a) (Norris et al., 1983), and $\tau_{Be}$ is the radioactive mean-life of $^{10}{\text{Be}}$ (1.93 * $10^6$ a) (Nishiizumi et al., 2007). The first term on the right hand side of (1) and (2) describes the post-burial decay of the isotopes and the second term describes the post-burial production of cosmogenic isotopes. For sediment that is buried to a sufficient depth, a few tens of meters, the second terms in (1) and (2) may be considered negligible, however, post-burial production significantly contributes to the inventory of cosmogenic isotopes in shallowly buried sediment (Granger and Muzikar, 2001; Granger and Smith, 2000; Wolkowinsky and Granger, 2004).

Under the assumption that a sample acquired its pre-burial inventory of cosmogenic isotopes as it was advected to the surface of a steadily eroding landscape, the
initial concentration of a given cosmogenic isotope will simply be a function of the erosion rate.

\[
N_{Al}(0) = \frac{A_0}{1 + \frac{E}{\tau_{Al}}} + \frac{A_1}{1 + \frac{E}{L_0}} + \frac{A_2}{1 + \frac{E}{L_1}} + \frac{A_3}{1 + \frac{E}{L_2}}
\]

\[
N_{Be}(0) = \frac{B_0}{1 + \frac{E}{\tau_{Be}}} + \frac{B_1}{1 + \frac{E}{L_0}} + \frac{B_2}{1 + \frac{E}{L_1}} + \frac{B_3}{1 + \frac{E}{L_2}}
\]

Where \( E \) is erosion rate (cm yr\(^{-1} \)) and \( L_j \) refers to the attenuation length for a cosmogenic isotope production reaction (cm g cm\(^{-3} \)). \( L_0 \) is the attenuation length for spallogenic production reactions; \( L_1 \) and \( L_2 \) are the attenuation lengths for negative muon capture production reactions; and \( L_3 \) is the attenuation length for fast muon production reactions.

We assign values of \( L_0 = 160/\rho \), \( L_1 = 738/\rho \), \( L_2 = 2688/\rho \), and \( L_3 = 4360/\rho \), where \( \rho \) is the density of the rock covering the sample in g cm\(^{-3} \) (Granger and Muzikar, 2001). We assume that prior to erosion, the target minerals were covered by rocks with a density of 2.6 g cm\(^{-3} \) (equations (3) and (4)), and after burial, the target minerals were covered by sediment with a density of 2.0 g cm\(^{-3} \) (equations (5) and (6)). The other constants, \( A_j \) and \( B_j \), have units of atoms year\(^{-1} \) gram\(^{-1} \) of quartz. We assign sea level high latitude values of \( A_0 = 30, A_1 = 0.72, A_2 = 0.16, A_3 = 0.19, B_0 = 5, B_1 = 0.09, B_2 = 0.02, \) and \( B_3 = 0.02 \), such that sea level high latitude production rates for \(^{26}\text{Al} \) and \(^{10}\text{Be} \) are 30.07 and 5.13 atoms gram\(^{-1} \) year\(^{-1} \) (Granger and Muzikar, 2001). The values of \( A_j \) and \( B_j \) are scaled to the field area using a latitude and altitude dependent correction for spallogenic
production, and an atmospheric pressure dependent correction for all muonogenic
production reactions (Stone, 2000). We note that the production rates and atmospheric
pressure scaling for fast muon reactions remain uncertain (Wolkowinski and Granger,
2004; Braucher et al., 2003), but because the fast muon terms are small, we neglect any
uncertainty that they introduce into our calculations.

For sediment that has not been buried to a sufficient depth to be completely
shielded from cosmogenic radiation, post-burial production is described by the following
relationships:

\[
(5) P_{26\text{Al}}(d) = A_0 e^{-d/L_0} + A_1 e^{-d/L_1} + A_2 e^{-d/L_2} + A_3 e^{-d/L_3}
\]

\[
(6) P_{10\text{Be}}(d) = B_0 e^{-d/L_0} + B_1 e^{-d/L_1} + B_2 e^{-d/L_2} + B_3 e^{-d/L_3}
\]

If the present day concentration of \(^{26}\text{Al}\) and \(^{10}\text{Be}\) is measured, and the burial depth is
measured, then equations (1)-(6) can be combined to form a system of two equations with
two unknowns, \(t\) and \(E\). The equations may be solved graphically, or numerically. We
present a graphical solution for these equations in Figure 3c, but in practice, we have
solved the equations numerically. This was done by forward modeling the range of
possible combinations of \(t\) and \(E\), and then identifying the \(t\) and \(E\) pair that best
reproduces the modern day \(^{26}\text{Al}/^{10}\text{Be}\) ratio and \(^{10}\text{Be}\) concentration in a least squares
sense.

Our calculations rely on two simplifying assumptions. First, we scale production
rates of \(^{26}\text{Al}\) and \(^{10}\text{Be}\) to the sea level high latitude values using the present-day latitude
and altitude of the samples. Because the quartz sand in each of the samples was eroded in a different location than it was deposited, this assumption introduces some error into our calculations. Second, we assume that uncertainties in the reported burial age reflect only the analytical uncertainties on the concentrations of $^{26}$Al and $^{10}$Be. We neglect uncertainties in the production rates and mean-lives of the isotopes.

We consider two different scenarios for sample burial. In the first case, we consider instantaneous and complete burial to present-day depths. Because the upper of our samples, NHKCOS-YT2, is only buried to a depth of 14 m, we account for post-burial production via muonogenic reactions. This approach yields burial ages of 0.726 $^{+0.322/-0.329}$ Ma and 1.234 $^{+0.449/-0.424}$ Ma for NHKCOS-YT2 and NHKCOS-YT3, respectively. The initial concentrations imply average erosion rates in the source areas of $0.0037 \pm 0.0007$ cm yr$^{-1}$ and $0.0048^{+0.0015/-0.0012}$ cm yr$^{-1}$, consistent with modern erosion rates (Harkins et al., 2001). If we neglected muonogenic production, we would underestimate the burial age of the shallow sample by $\sim$18% and the deeper sample by $\sim$5% (ages of 0.603 $^{+0.269/-0.266}$ Ma and 1.173 $^{+0.434/-0.346}$ Ma, respectively.)

We also consider the effect of protracted exposure during slow burial and sediment accumulation. We modeled burial histories wherein the burial depth is time dependent and is based on the magnetostratigraphically determined sediment accumulation rates. Our magnetostratigraphy indicates that sedimentation rates were at the least $\sim$100 m Ma$^{-1}$. At this relatively rapid burial rate, the results are negligibly different from the instantaneous burial case. The burial ages for the upper and lower sites are 1.271 $^{+0.452/-0.423}$ Ma and 0.731 $^{+0.324/-0.328}$ Ma. Only as sedimentation rates
become significantly slower, on the order of \(~10\, \text{m Ma}^{-1}\) is there a significant effect on the burial age. Thus, we are confident that this effect is minor for our samples.

Finally, we explored the influence of post-burial exposure during basin excavation. Although samples were buried to minimum depths of 10.5 and 13 m, prior to road construction, there may have been a minor amount of production through the canyon sidewall. We performed a set of calculations that assumes instantaneous burial, followed by exhumation to the minimum burial depth of each sample as described in the field techniques section. We consider that the maximum duration of shallow burial (\(~10-15\, \text{m}\)) could be \(~0.7\, \text{Ma}\) (representing a case where the Yellow River began incising immediately after deposition of the uppermost basin fill). Given that our magnetostratigraphic section extends well into the Brunhes normal polarity chron, this is a very conservative estimate of the timing of basin excavation. Moreover, rapid retreat of the angle-of-repose hillslopes during incision of the canyon implies only recent exhumation to shallow depth. For the shallower sample, NHKCOS-YT2, we find a burial age of \(0.908 \pm 0.399/0.414\, \text{Ma}\). For the deeper sample, the burial age of NHKCOS-YT3 could be as old as \(1.676 \pm 0.634/-0.671\, \text{Ma}\). We find it likely that the samples were deeply buried for most of their lifetime, and that it is only within the last few tens of thousands of years that the samples were excavated to a depth of 10-15 m. Therefore, we prefer a burial age of \(~0.7 \pm 0.3\, \text{Ma}\) for the upper sample and \(~1.2 \pm 0.4\, \text{Ma}\) for the lower sample.

In order to compare the \(^{26}\text{Al}/^{10}\text{Be}\) ratio and \(^{10}\text{Be}\) concentration between sites, the abscissa in Figure 5.2c is normalized by the product of the \(^{10}\text{Be}\) production rate at sea level and high latitude and the inverse of the local production rate (Granger, 2006 and
references therein). Solid curves represent $^{26}$Al and $^{10}$Be concentrations for various burial times. It can be seen from the zero burial age curve that $^{26}$Al and $^{10}$Be are produced at a ratio of ~6 or slightly lower. Prior to transport, deposition, and burial, samples are eroded to earth’s surface at some unknown rate. The sub-vertical, stippled lines represent the burial history for samples with various inherited erosion rates. Dashed curved represents $^{26}$Al and $^{10}$Be concentrations for samples experiencing constant exposure at earth’s surface. Because unstable cosmogenic isotopes tend towards a secular equilibrium, there is a finite limit to the $^{10}$Be concentration, which is located on the far right hand side of the burial time = 0 curves.

5.1.2.4. Magnetostratigraphy

5.1.2.4a. Field methods

In the field, we attempted to collect samples at 2-5 m intervals. In certain stratigraphic intervals, such a high density was not possible, because a) the sediments were too coarse for paleomagnetic analysis and/or b) the sediments were too friable to sample. If possible, 3 specimens were collected from each bed, using a gas-powered drill with a 2.5 cm diameter core bit. The core-plate orientation was measured using a magnetic compass. Bedding dip of sites was also measured using a magnetic compass. In certain stratigraphic intervals, the sediment was too friable to be sampled with a drill. At these sites, oriented block samples ~5 cm$^3$ in volume, were collected. These samples were carved into smaller cores in the laboratory.

5.1.2.4b. Laboratory methods

The laboratory methods used to measure the characteristic remnant magnetism are broadly similar to Heermance et al., 2007, with only slight differences. Stepwise
demagnetization experiments were conducted at The California Institute of Technology. Magnetization was measured using a three-axis DCSQUID moment magnetometer in a magnetically shielded μ-metal room. The background noise of the magnetometer is <1 pA m². It is equipped with a vacuum pick and put, computer-controlled sample changing system, which is capable of measuring batches as large as 180 specimens. A 20 or 21 step demagnetization scheme was developed. After measuring the natural remnant magnetism of each specimen, it was subjected to six alternating field (AF) demagnetization steps between AF = 0 gauss and AF = 100 gauss (AF = 0, 20, 40, 60, 80, 100 gauss). AF demagnetization was conducted with a computer-controlled, three-axis coil system. Next the samples were subjected to 15 thermal demagnetization (TT) steps between 80° and 680° C (TT = 80°, 150°, 250°, 350°, 450°, 500°, 530°, 555°, 570°, 600°, 635°, 655°, 665°, 673°, and 680°). The 80° temperature step was omitted for approximately one-third of the samples. The thermal demagnetization scheme was designed to have a high density of steps just below the Curie temperatures of magnetite (570° C) and hematite (680° C). Thermal demagnetization was conducted with a commercially built, magnetically shielded furnace. If the magnetization measurement following an AF or TT step yielded a circular standard deviation (CSD) of >10°, the measurement was repeated up to two times. If a CSD of ≤10° could not be obtained, the data for the AF/TT step were discarded.

The natural remnant magnetism of the samples was on the order of $10^5 – 10^6$ emu cm⁻³. In general, stepwise demagnetization revealed two components of magnetization, a low coercivity component and high coercivity component (Supplementary Figure 5.3). The low coercivity component was typically removed between AF = 0 and 100 gauss -
150° C. The higher component was either fully or partially removed between TT = 80° - 250° and 570°. For a few samples, the high coercivity component was not removed until the specimen was heated to 680°, the Curie temperature of hematite. Because we were able to remove a component of magnetization that we infer to be post-depositional, because the sites pass a C-quality reversal test (see below), and because the magnetostratigraphy is a good fit to the independent age constraints that we have on the section, we interpret the high coercivity component of magnetization to be characteristic remnant magnetization (ChRM). Because the high coercivity component of magnetization is typically removed at ≤570°, we interpret magnetite to carry the ChRM. Hematite is also likely to carry ChRM at sites where the high coercivity component is not removed below 680°.

ChRM directions were determined using a least-squares fit, principal component analysis. In general, we sought to perform the least squares, principal component analysis on ≥ 4 TT/AF steps. Commonly, more steps were incorporated, however, in a few cases, only three TT/AF steps were used (Supplementary Table 5.3). The principal component analysis was performed using the program PaleoMag v3.1 d36 (Kirschvink, 1980; Jones, 2002). For magnetization components that were believed to be characteristic, regression included the origin and were forced through the origin. If the mean angular deviation of the regression exceeded 15, the ChRM direction for the specimen was considered unreliable and discarded. Moreover, specimens with VGP latitudes of <45° were discarded. Samples not meeting these two quality criteria are shown as open circles in Figure 5.2a. Our data pass a C-quality reversal test (McFadden and McElhinny, 1990).
(Supplementary Figure 5.4), indicating that the measured ChRM directions have been stable since the time of deposition.

5.1.3. Chronology of fluvial terraces

5.1.3.1. Geomorphic mapping

Six terrace levels and the top of the Tongde basin fill were identified based on a combination of field observations, surveys, 90-m Shuttle Radar Topography Mission (SRTM) digital topography, ASTER imagery and Google Earth imagery. Because the terraces are composed of material with a similar degree of weathering and induration as the intact basin fill, and because the terrace material is less weathered/indurated than the terrace lag deposits, we consider these to be strath terraces. In the field, six distinct terrace levels were identified and surveyed using a laser range finder with centimeter scale resolution. Using ASTER imagery in a GIS and Google Earth, the spectral character of each terrace level was identified, and used to extend terrace mapping along the length of the Yellow River canyon in Tongde basin. Terrace correlations were vetted by comparing digital elevation data to field survey data. Many of the Yellow River terraces in Tongde basin are covered by broad, aggregated alluvial fans that imply a protracted history of incision in the basin. The fan deposits grade to the paleo-river level that corresponds to their depositional age (i.e. the strath terrace level). Alluvial fan deposits are placed into our relative chronology by identifying the terrace level to which they grade.

5.1.3.2. Radiocarbon constraints

5.1.3.2a. Field Methods
Charcoal and fresh water shell samples were collected from low Yellow River terraces from the base of very thinly bedded and laminated silt deposits that cover the terrace lag deposits. We interpret these to be overbank deposits, and deposition of the beds from which the samples were derived may have lagged strath terrace development. Thus, incision rates calculated from the age of these units are upper bounds on the true Holocene incision rate. The elevation of each sample above the modern river level was measured locally using a laser range finder.

5.12.3b. 14C Dating and calibration

14C dating was performed by the University of Arizona and Beta Analytic Radiocarbon Dating Laboratory. Calibrated ages were determined using Calib 5.0 (Stuvier and Reimer, 1993), using the IntCal04 calibration curve19 (Supplementary Table 5.4).

5.1.2.3c. Young 14C ages

Samples MRTC-05-03 and MRTC-05-04 yielded anomalously low 14C ages compared to the other samples. We attribute this to mixing of “modern” charcoal during post-depositional bioturbation of the upper ~0.5m of loess atop these terraces.

5.1.2.4. Volumetric denudation calculation

We reconstruct the total amount of material excavated from the Tongde, Gonghe and Guide basins to help elucidate the spatial and temporal patterns of basin dissection along the upper reaches of the Yellow River. Using 90-m SRTM digital topography, 717 points from the Gonghe basin surface and 529 points from the Tongde basin surface, were selected adjacent to the canyon of the modern Yellow River. Using a simple triangular irregular network (TIN) interpolation, we stretched a surface through these
points and across the modern canyon of the Yellow River. Differencing the reconstructed surface and the modern topography reveals that the total volumetric denudation in Gonghe basin is approximately 802 km$^3$, whereas it is only 202 km$^3$ in Tongde basin (Supplementary Figure 5.5).

In Guide basin, the basin surface is relatively poorly preserved, and we relied heavily on SRTM topography, ASTER imagery, and direct field observations to identify remnants of the basin surface. Flat surfaces atop interfluves in the northeast corner of the basin, which may represent basin fill top terraces or high strath terraces, are also used to reconstruct the pre-erosion surface. These surfaces were extracted from the 90-m topographic grid and a high (10$^\text{th}$) order polynomial was interpolated through them. The root mean square error for the reconstructed surface and the control surfaces is 16 m. We find that the total denudation here has been 691 km$^3$. Because the pre-incision surface is tied to topography in the northeast corner of the basin which may be lower than the true pre-incision surface, we consider this estimate to be a minimum.

Because the streamwise distance in each of the three basins is quite different (Figure 5.4), it is not appropriate to compare total volumetric denudation measurements from each basin. Instead, we divide the total volumetric denudation of each basin ($V$), by the streamwise length in each basin in order to obtain the normalized volumetric denudation ($V^*$). Streamwise distance was measured by tracing the channel along 90-m SRTM digital topography. A large reservoir presently occupies central Gonghe basin, and that portion of the digital topography was replaced with 90-m DTED digital topography that predates reservoir construction. From Guide to Tongde, we find $V^*$ values of 14.7, 6.6, and 3.3 km$^3$ km$^{-1}$, respectively. The spatial trend in denudation per unit stream length
suggests longer duration of erosion in Guide (downstream) than in Tongde (upstream). This finding is consistent with the interpretation of headward erosion that is driven by regional stratigraphic and geomorphic data. A downstream increase in the amount of eroded material may also reflect downstream increases in discharge, meaning that the pattern is a necessary, but not sufficient, condition for documenting headward erosion. However, the relatively arid climate and the small number of adjoining tributaries suggest that downstream increases in discharge are minimal along this reach of the Yellow River, and we suggest that much of the pronounced downstream increase in basin excavation relates to headward erosion by the Yellow River.
Figure 5.1. Cenozoic sedimentary basins of the northeastern Tibetan Plateau. a) Major basins along the Yellow River in the northeastern Tibetan plateau and constraints on the transition from filling to excavation. 1 = Li et al., 1997; 2 = Fang et al., 2003; 3 = Pan et al., 1996; 4 = Harkins et al., 2007; 5 = Fang et al., 2005; 6 = Zeng et al., 1995; 7 = Zhu et al., 1995. Grey lines denote basin boundaries. b) Gonghe basin complex with locations of stratigraphic sections, including Tongde North (1) and Tongde South (2) (see Supplementary Information). Burial age samples NHKCOS-YT2 and NHKCOS-YT3 are abbreviated YT2 and YT3. c) Maximum and minimum topography along a 20 km wide swath, derived from Shuttle Radar Topography Mission (SRTM) 90-m data. In minimum topography, spikes represent SRTM noise and blue segments coincide with Yellow River.
Figure 5.2. Chronology of the upper basin fill in the Gonghe basin complex. a) Lithostratigraphy and magnetostratigraphy of Tongde basin (see Supplementary Information). The Tongde section correlates to the geomagnetic polarity time scale (GPTS) between ~3.3 - 0.5 Ma (Ogg and Smith, 2004). The correlation is constrained by burial and fossil ages. Open circles indicate a sample that was discarded based on quality criteria. b) Sediment accumulation rates. c) Cosmogenic burial ages plotted on a graph of $^{26}$Al/$^{10}$Be ratio versus $^{10}$Be concentration (see Supplementary Information). Uncertainties in burial ages reflect both analytical uncertainty and uncertainty in the burial history.
Figure 5.3. Distribution and age of Yellow River terraces in Tongde basin. In addition to the basin fill top, we mapped 6 different terrace levels based on field surveys, SRTM digital topography, and ASTER imagery. Locations of dated terraces are shown with circles (grey = OSL, white = 14C). Radiocarbon ages are presented as calibrated years BP. Thin grey lines mark the base of major terrace risers. Background is ASTER L1a spectral data, VNIR bands 1, 2, and 3, with a nominal 15-m resolution, draped over a hillshade generated from 90-m SRTM digital topography. See text and Supplementary Information for details.
Figure 5.4. Constraints on timing of incision along the Yellow River. Symbols explained in Figure 5.1. G = Guide, L = Lanzhou, Li = Linxia, SQ = Sai Qe, X = Xunhua, ZQ = Zhe Qu. Uncertainties at Tongde and Linxia reflect bounding ages of basin fill and fluvial terraces. Regression of data, excluding Guide, yields a headward erosion rate of ~350 km Ma⁻¹.
Supplementary Figure 5.1. Gonghe basin complex lithostratigraphy. Central and northern Gonghe stratigraphic sections are modified from Zheng et al., 1985 and field observations. Stratigraphic sections are hung to the broad geomorphic surface across the Gonghe Basin complex, which represents the maximum extent of basin filling.
Supplementary Figure 5.2. NHKCOS-YT2 and NHKCOS-YT3 sample sites. The samples were collected from deep roadcuts where post-burial cosmogenic inheritance was small and possible to calculate.
Supplementary Figure 5.3. Examples of stepwise demagnetization of specimens. Two sites, one with normal polarity and one with reversed polarity, are shown from both the north and south segment of the Tongde magnetostratigraphic section. A low coercivity component of magnetization was typically removed between AF0 and AF100 or TT150. The high coercivity component was typically removed between TT80° - 250° and TT570 or sometimes TT680°, suggesting that the ChRM is carried by magnetite and hematite (see Supplementary Information).
Supplementary Figure 5.4: Reversal test for Tongde magnetostratigraphic section. There are 20 normal polarity sites and 29 reversed polarity sites. Normal polarity sites have a mean, tilt-corrected declination of 11.0°, a mean, tilt-corrected inclination of 33.4°, and a 95% confidence limit on the mean normal polarity direction of 10.8°. Reversed polarity sites have a mean tilt-corrected declination of 183.2°, a mean tilt-corrected inclination of -43.5°, and a 95% confidence limit on the mean directions of 7.5%. The black circles define the 95% confidence limit of the mean inclination and declination directions. The sites pass a C-quality reversal test. Because the beds dip very shallowly in this section (between 0° and 2°), we present only the tilt-corrected reversal test. The results of the reversal test in geographic coordinates are essentially identical.
Supplementary Figure 5.5. Volumetric denudation (V) along the Yellow River in Guide, Gonghe, and Tongde basins. $V^*$ represents the volume of eroded material per unit stream length in each basin, and progressively decreases in the upstream direction.
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Supplementary Table 1. Summary of biostratigraphic constraints on the age of the uppermost fill in the Gonghe basin complex (Zheng et al., 1985).
Supplementary Table 5.2. Locations and measured cosmogenic inventories for cosmogenic burial age samples.

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Supplementary Table 5.3. Tongde magnetostratigraphic data.
Site names followed by ** indicate sites that were rejected because the VGP latitude was less than 45°.
Decl. = Declination, Incl. = Inclination, (g) = geographic, (t) = tilt corrected
<table>
<thead>
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<th>Sample</th>
<th>Field ID</th>
<th>Material</th>
<th>Latitude (N)</th>
<th>Longitude (E)</th>
<th>$\delta^{13}$C</th>
<th>$^{14}$C age BP (yr)</th>
<th>Uncertainty (yr)</th>
<th>Calibrated Max Years BP (2(\sigma))</th>
<th>Calibrated Min. Years BP (2(\sigma))</th>
<th>Height (m)</th>
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Supplementary Table 5.4. Description of $^{14}$C samples. Calibrated ages were determined using Calib 5.017, using the IntCal04 calibration curve (Stuiver, 2004). 2\(\sigma\) age range is reported for calibrated $^{14}$C ages.
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