The Liuqu Conglomerate, southern Tibet: Early Miocene basin development related to deformation within the Great Counter Thrust system

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ABSTRACT

The Liuqu basin formed during the early Miocene between ophiolitic mélanges (to the south) and uplifted Cretaceous forearc deposits (to the north) along a central, 150-km-long segment of the India-Asia suture zone in southern Tibet. Sedimentological analysis shows the Liuqu Conglomerate to be composed of mixed fluvial and sediment-gravity flow lithofacies assemblages locally interbedded with mature paleosols. We interpret the Liuqu Conglomerate as coarse-grained fluvial and alluvial-fan deposits. Structural analysis indicates that the Liuqu Conglomerate was deposited in a contractional setting. Paleocurrent and provenance data demonstrate that sediment was transported north-northwest from the hanging wall of a coeval thrust fault system along the southern limit of Liuqu outcrops. Detrital zircon U-Pb ages (and Hf [t] isotope ratios) cluster around 80–110 Ma (ɛHf[t] = −23.5–14.6), 120–135 Ma (ɛHf[t] = −12.6–13.1), 150–170 Ma (ɛHf[t] = −14.1–14.7), 500–600 Ma (ɛHf[t] = −26–3.4), and 1100–1200 Ma (ɛHf[t] = −2.76–2.9), requiring input from both Gangdese and mélange sources. Asian zircons were recycled northward after being incorporated into accretionary mélanges along the southern Asian margin prior to India-Asia collision. The age of the Liuqu Conglomerate is still somewhat uncertain, but new chronologic data, including biotite 40Ar/39Ar data, detrital zircon fission-track analyses, and δ13C compositions of soil carbonates, all converge on ca. 20–19 Ma as the most probable age. Together, these results indicate that part of the north-to-south sediment transport system that existed prior to India-Asia collision and into the Eocene was reversed by ca. 20 Ma. The Liuqu Conglomerate may represent deposits associated with the paleo–Yarlung River.

INTRODUCTION

The India-Asia suture zone records the closure of the Neotethys Ocean and the collision between India and Asia at ca. 60 Ma (Beck et al., 1996; Rowley, 1998; Zhu et al., 2005; Ding et al., 2005; Cai et al., 2012; Orme et al., 2014; DeCelles et al., 2014; Hu et al., 2015a, 2015b). In addition to the Asian Xigaze forearc basin, ophiolitic fragments, and remnants of an accretionary mélangé, this zone contains Cenozoic coarse-grained terrestrial sedimentary rocks, including the Liuqu Conglomerate (J. Yin et al., 1980). How this several-kilometer-thick conglomerate unit was deposited and preserved in an area that would otherwise be expected to be dominated by uplift and exhumation remains an intriguing problem. Previous workers have assumed that the Liuqu Conglomerate was deposited in a contractional tectonic setting (Davis et al., 2002; Aitchison et al., 2007; Wang et al., 2010), but this hypothesis has not been rigorously tested. Given that other mid-Cenozoic coarse-grained deposits within the India-Asia suture zone have been interpreted as extensional basin deposits (DeCelles et al., 2011; Wang et al., 2013), the a priori assumption that the Liuqu basin was contractional is unwarranted. Whether basins within the India-Asia suture zone formed as a result of contraction or extension is critical to understanding the underlying geodynamics of the Indian and Asian plates in the collision zone.

The Liuqu Conglomerate has been the subject of much debate, in large part because of uncertainties in its depositional age. Several workers have interpreted these rocks to have formed during the collision of an island arc with India prior to India-Asia collision, based primarily on a lack of petrologic evidence of Asian-derived sediment (Davis et al., 2002, 2004; Aitchison et al., 2011). Other workers have found evidence of both Indian- and Asian-derived sediment, requiring the Liuqu Conglomerate to have been deposited after India-Asia collision (Wang et al., 2010). Analysis of plant fossils has produced age estimates of Cretaceous–Eocene (Tao, 1988; Fang et al., 2006), and an Oligocene age was assigned based on palynologic analysis (Wei et al., 2011). The absence of tuffaceous rocks in the Liuqu Conglomerate has contributed to the difficulty in applying radiometric dating; however, recent work points to an early Miocene age based on detrital zircon U-Pb (Leary et al., 2012) and (U-Th)/He (G. Li et al., 2015) analyses.

In this study, we present an in-depth analysis of the age, depositional environment, provenance, and causes of basin formation for the Liuqu Conglomerate. The age of the Liuqu Conglomerate is refined through 40Ar/39Ar dating of a crosscutting dike, detrital zircon U-Pb and zircon fission-track dating, reanalysis of published pollen data, and carbon stable isotope analyses of soil carbonates. Depositional environment was
determined through documentation of facies in 12 measured sections totaling ~4.3 km of stratigraphic thickness. Samples for petrography and detrital thermochronology and geochronology were collected and documented in these sections. Provenance was assessed using conglomerate clast counts, petrographic point counting of sandstones, and detrital zircon U-Pb dating and Hf isotope analysis.

GEOLOGIC AND TECTONIC SETTING

The southern Tibetan Plateau formed during the northward subduction of Neotethyan oceanic lithosphere beneath the southern margin of Asia and the subsequent collision of the Indian continent with this margin (Argand, 1924; Molnar and Tapponnier, 1975; Powell and Conaghan, 1975; Allègre et al., 1984; Molnar and Stock, 2009; Wang et al., 2014). Many workers have concluded that collision occurred in central Tibet between 55 Ma and 50 Ma (Powell and Conaghan, 1975; Besse et al., 1984; Patriat and Achache, 1984). However, new data linking sedimentary rocks deposited on the north Indian shelf with Asian provenance suggest that collision was under way by as early as 60 Ma (Yin and Harrison, 2000; Orme et al., 2014; DeCelles et al., 2014; Hu et al., 2015a, 2015b). Southern Tibet encompasses four major tectonic domains (Fig. 1), including the Lhasa terrane, Xigaze forearc basin, India-Asia suture zone, and the Tethyan Himalayan fold-and-thrust belt. The Lhasa terrane is dominated by igneous rocks that formed initially in an Andean-type arc as a result of northward subduction of the Neotethyan oceanic slab; subduction related magmatism continued until ca. 40 Ma (Harrison et al., 2000; Wen et al., 2008; Ji et al., 2009; Zhu et al., 2011). The Linzizong volcanics represent the extrusive parts of the Paleocene–Eocene arc and have been interpreted as the result of initial India-Asia collision (Zhou et al., 2004; Mo et al., 2008), Neotethyan slab breakoff (Yin and Harrison, 2000; Lee et al., 2009), and lithospheric delamination (Ji et al., 2014). The Gangdese batholith and related plutons are interpreted as remnants of the arc’s deeper magmatic roots (Allègre et al., 1984; Wen et al., 2008; Mo et al., 2008; Ji et al., 2009; Zhu et al., 2011).

South of the Gangdese arc, the Xigaze forearc is the remnant of a once much wider forearc basin (Fig. 2; Ratschbacher et al., 1994). These rocks were deposited in submarine channels and fans, shallow marine platforms, and fluvial systems along the southern margin of Asia beginning in Aptian time and continuing through early Eocene time (Einsele et al., 1994; Dürr, 1996; Wu et al., 2010; Wang et al., 2012; Orme et al., 2014; An et al., 2014; Hu et al., 2015b). The Xigaze forearc is highly deformed and juxtaposed against the southern edge of the Gangdese Batholith by the Great Counter Thrust, a system of north-verging thrust faults associated with the latest stages of deformation in the Tethyan fold-and-thrust belt (Gansser, 1964; Yin et al., 1999; Murphy and Yin, 2003). To the south, portions of the Xigaze forearc are thrust southward over India-Asia suture zone rocks (Ratschbacher et al., 1994; Wang et al., 2000; Hébert et al., 2012; Cai et al., 2012), whereas in other places, this contact is depositional (Allègre et al., 1984; Wang et al., 2012; An et al., 2014; Orme et al., 2014; Huang et al., 2015).

Directly south of the Xigaze forearc basin, a structurally complex band of rocks is collectively referred to as the India-Asia suture zone (Searle
The Liuqu Conglomerate, southern Tibet | RESEARCH

Figure 2. Geologic map of the central India-Asia suture zone after Cai et al. (2012) and schematic cross sections of Liuqu Conglomerate exposures. Measured section numbers are given in parentheses.
Hébert, et al. (2012), and ophiolitic-, serpentinitic-, and sedimentary-matrix mélangé complexes (Cai et al., 2012; An et al., 2015). Ophiolitic rocks have been dated as Jurassic–Cretaceous (McDermid et al., 2002; Malpas et al., 2003; Guillet et al., 2009) and most likely formed originally in a subsupduction-zone environment (Guillet et al., 2009).

It has been suggested that these rocks were associated with the northern Indian margin prior to India-Asia collision (Burg and Chen, 1984; Girardeau et al., 1984; Searle et al., 1987; Malpas et al., 2003; Ding et al., 2005), but recent work in the Lazi (also spelled “Latze” in many publications) area (Fig. 2) has demonstrated that they were in fact accreted to the southern margin of Asia prior to the collision with India (Cai et al., 2012; Orme et al., 2014). South of the ophiolites, mélangé zones (the Bainang terrane of Aitchison et al., 2000) represent the remnants of the accretionary prism formed by off-scraping of material from the subducting Neotethyan slab (Cai et al., 2012; An et al., 2015).

South of the mélangé complex, there are the sedimentary rocks of the Tethyan Himalaya. These Paleozoic–Eocene rocks were deposited on the passive north Indian margin after the Lhasa terrane rifted away from Gondwanaland (Garzanti et al., 1987; Liu and Einsele, 1994; Jadoul et al., 1998). Beginning after India-Asia collision, they were deformed into a south-verging thin-skinned thrust belt (Burg and Chen, 1984; Ratschbacher et al., 1994; Murphy and Yin, 2003). Numerous small, north-verging thrust faults thought to be associated with the Great Counter Thrust cut Tethyan and mélangé rocks (Ratschbacher et al., 1994; Yin et al., 1999).

Marine sedimentation in basins within the Tethyan system was terminated during the Eocene, ~10 m.y. after the onset of India-Asia collision (Ding et al., 2005; Najman et al., 2010; Hu et al., 2012). However, localized nonmarine deposition is documented along the suture zone in southern Tibet long after this. A nearly continuous zone of conglomeratic rocks crops out over ~1300 km along strike of the suture zone between rocks of the Xigaze forearc basin and the Gangdese arc. Various formation names have been assigned to these rocks, including the Kailas Formation (Heim and Gansser, 1939; Gansser, 1964; DeCelles et al., 2011), the Quwu Conglomerate (Aitchison et al., 2002; Ding et al., 2005), the Dazuqu Formation (Aitchison et al., 2002), the Luobusa Formation, and the Gangrinboche Conglomerate (Aitchison et al., 2002, 2007). Here, we collectively refer to these rocks as the Kailas Formation in deference to Gansser’s (1964) original terminology. The Kailas Formation at its type area in southwestern Tibet (near Mount Kailas) was deposited between 26 and 21 Ma and is younger to the east (Leary et al., 2016); the youngest documented Kailas Formation deposition has been dated at 18 Ma southeast of Lhasa (Kong et al., 2015). Deposition of the Kailas Formation has been attributed to extension associated with Indian slab shearing (DeCelles et al., 2011; Leary et al., 2016), and initiation of the paleo–Yarlung River (Wang et al., 2013; S. Li et al., 2015; Wang et al., 2015).

By ca. 17 Ma, the entire central suture zone experienced a rapid pulse of exhumation (Carrapa et al., 2014). This would have involved removal of large volumes of elastic material and has also been attributed to incision by the paleo–Yarlung River. There is evidence that the paleo–Yarlung River may have been active and draining into Myanmar prior to its capture by the Brahmaputra between 21 and 18 Ma (Robinson et al., 2014; BracciaI et al., 2015; for a contrasting interpretation, see Lich et al. 2015). It has also been suggested that the paleo–Yarlung River flowed through a transverse paleo–Subansiri River prior to 4–3 Ma (Cina et al., 2009; Zhang et al., 2012). However, detrital zircon provenance data, low-temperature thermochronometric data, and knickpoint modeling suggest that the Yarlung-Brahmaputra system achieved its modern configuration by ca. 10 Ma (Lang and Hunting ton, 2014; Schmidt et al., 2015).

**STRUCTURAL CONTEXT OF THE LIUQU CONGLOMERATE**

Geologic mapping of the Liuqu Conglomerate was carried out over three field seasons from 2011 to 2013. Overall, the Liuqu Conglomerate is folded into a several-kilometer-wide syncline by movement of faults along both its northern and southern boundaries. The maximum north-south exposure width is <5 km, and it is impossible to reconstruct the original north-south extent of Liuqu deposition. However, Liuqu Conglomerate lithofacies are very coarse grained and proximal to their source terranes, suggesting that the present exposure limits are not much different from the original depositional extent. The mapping indicates that the Liuqu Conglomerate is in fault contact with adjacent rocks at nearly every exposure along its entire 150 km length of outcrop, except rare depositional contacts between the Liuqu Conglomerate and bedded red chert in section 4LQ (Fig. 3), and with ophiolitic basalt near the town of Bainang, southeast of Xigaze (Fig. 2; Davis et al., 2002). Along its northern outcrop boundary, the Liuqu Conglomerate is typically overthrust by ophiolitic rocks, whereas to the south, it is typically overthrust by ophiolitic sequences or Tethyan strata.

At its type section near the town of Liuxiang (Liuqu), the formation is folded into a 2-km-wavelength south-verging syncline. The stratigraphically lowest part of this section is structurally overturned, and it is overthrust by ophiolitic rocks along a fault dipping 60° toward 030°. Growth strata are present in this zone (Figs. 2 and 3), with bedding dip varying from 30° toward 000° (overturned) to 20° toward 220° (upright) over a stratigraphic distance of ~1 km. Several intraformational angular unconformities are preserved in this interval. The southern edge of this exposure is bounded by Tethyan and mélange rocks that are thrust northward over the Liuqu Conglomerate.

Near the town of Saguil (1LQ; Fig. 2), ophiolitic rocks are exposed to the south of, and thrust northward over, the Liuqu Conglomerate along a fault dipping 35° toward 170°; a 0.2-km-wavelength syncline at the top of the section records this deformation, but no growth strata are preserved. To the east, near the town of Xiaolu (Fig. 2), the Liuqu Conglomerate is overturned, dipping north, and is exposed in a 0.5-km-wide outcrop belt in thrust contact with red chert to the north and Tethyan rocks to the south.

**SEDIMENTOLOGY**

**Description**

The depositional environment of the Liuqu Conglomerate was reconstructed based on sedimentological data collected at 12 locations (Figs. 4–9). Sections were measured with a Jacob staff and tape measure at the centimeter scale. Paleoflow directions were reconstructed using clast imbrications measured at 43 locations (at least 10 measurements per station, retrodeformed using the freeware computer program StereoNet on R.W. Allmendinger). The lithofacies present in the Liuqu Conglomerate have already been well documented in the sedimentologic literature; here, we provide only brief descriptions and interpretations of these facies in Table 1.

The maximum preserved thickness of the Liuqu Conglomerate is ~2 km (Fig. 10) at the type section near the town of Liuxiang (Liuqu). Exposures along strike range from several hundred meters to over a kilometer thick. It is unknown exactly how much additional stratigraphic thickness has been eroded or truncated by faults, but unreset zircon U-Th/He ages in the Liuqu Conglomerate (S. Li et al., 2015) indicate that this must have been less than ~7 km (Carrapa et al., 2014).

Dominant lithofacies include clast-supported, massive pebble-cobble conglomerate (Gcm; Fig. 3G); clast-supported, horizontally stratified, and bedded red chert in section 4LQ (Fig. 3), and with ophiolitic basalt near the town of Bainang, southeast of Xigaze (Fig. 2; Davis et al., 2002). Along its northern outcrop boundary, the Liuqu Conglomerate is typically overthrust by ophiolitic rocks, whereas to the south, it is typically overthrust by ophiolitic sequences or Tethyan strata.
Figure 3. Continued on following page.
Figure 3 (continued). Photographs of Liuqu Conglomerate exposures. (A) Looking west over the section near Sagui. Ophiolitic sequences are thrust northward over the Liuqu Conglomerate. (B) Looking west at the section near Liuqu. Ophiolitic sequences (upper right) are thrust southward over Liuqu Conglomerate growth strata. (C) Looking west near Liuxiang (Liuqu) showing Liuqu growth strata. Arrows indicate stratigraphic up. (D) Depositional contact between the Liuqu Conglomerate and red chert at section 4LQ. (E) Clast-supported, poorly sorted, horizontally stratified conglomerate (Gch) typical of the Liuqu Conglomerate. (F) Clast-supported, moderately sorted, horizontally stratified conglomerate (Gch) at section 1LQ. (G) Channelized, massive clast-supported conglomerate incised into paleosol horizon at section 7LQ. (H) Thick paleosol intervals at section 1LQ. (I) Poorly organized, matrix-supported boulder conglomerate near section 8LQ.
Figure 4. Measured section 1LQ. DZ—detrital zircon, SS—sandstone, ZFT—zircon fission track. Vertical axes are in meters. See Table 1 for lithofacies codes.
are either not graded or show slight normal grading; rare inversely graded beds are present in some sections. Matrix-supported conglomerates are angular, are extremely poorly sorted, and have maximum clast sizes of up to 1.5 m. Conglomeratic intervals range from ~0.5 m to 10 m in thickness. Nearly all have scoured, irregular basal contacts. Conglomeratic beds are laterally continuous at outcrop scale (>100 m), although lenticular channelized beds are preserved in some locations (Fig. 11). Conglomerates and mudstones are intercalated such that most conglomerates are separated by at least one mudstone, and it is common for three or more stacked mudstone intervals to separate conglomerates. Individual mudstone intervals are typically 10–50 cm thick, mottled, mottled, red or green in color, and often contain root traces. Although sparse, carbonate nodules are present in some of these deposits. Mudstones are clay-rich and have slickensided surfaces; very little coarse-grained material is preserved in most fine-grained intervals.

Interpretation

We interpret the facies and facies architecture of the Liuqu Conglomerate to represent deposition in a stream-dominated alluvial-fan environment (Rust, 1977; Allen, 1981; Ridgway and DeCelles, 1993; Kumar et al., 2007). Thick, clast-supported conglomerate beds with sharp basal contacts are interpreted as high-concentration traction deposits (or sheetfloods), whereas similar beds with erosive, channelized bases are interpreted to represent filling of fan distributary channels by noncohesive flows (Gloppen and Steel, 1981; Todd, 1989; Lawton et al., 1999). Matrix-supported conglomerates are interpreted as the result of proximal, viscous debris flows (e.g., Shultz, 1984).

Massive, mottled, and slickensided mudstone intervals within the Liuqu Conglomerate are interpreted as paleosols. These are interpreted as having formed on inactive portions of the fan surface and likely represent considerable periods of nondeposition (Mack et al., 1993; Tabor et al., 2006). These intervals bear striking visual similarity to paleosols in the Siwalik Group of India and Pakistan (Tandon and Narayan, 1981; Quade and Cerling, 1995; Quade et al., 1995; Brozović and Burbank, 2000; Ojha et al., 2000), which were deposited in a warm, subhumid environment. This is supported by the presence of root casts and by paleobotanic (Tao, 1988; Fang et al., 2006) and palynologic (Wei et al., 2011) evidence. The abundant slickensided surfaces are interpreted as pedogenic slickensides, formed by physical alignment of clays during swelling and shrinking of soil by wetting and drying (Gray and Nickelsen, 1989). The presence of sparse carbonate nodules supports this and indicates that these soils experienced at least some seasonal drying. Together, these observations suggest that the Liuqu Conglomerate was deposited in a warm, seasonal, subhumid, and low-elevation setting.

ANALYTICAL METHODS

Detrital U-Pb Geochronology

Detrital zircon samples were collected from nine fine-, medium-, and coarse-grained sandstone beds (from sections 1LQ, 4LQ, 5LQ, 7LQ, and 9LQ; samples LQ-1, LQ-2, and LiuquTuff were collected separately) and six fine-grained intervals (from sections 1LQ, 8LQ, and 14LQ) within the Liuqu Conglomerate (see Figs. 2 and 4–9 for locations). These samples were processed according to standard crushing, sieving, water table, magnetic, and heavy liquid procedures. Zircon grains were then mounted in epoxy pucks, polished, and mapped with high-resolution backscattered electron (BSE) imaging at the University of Arizona LaserChron Center. U-Pb ages of zircon grains were determined using laser
Figure 6. Measured section 5LQ. Legend same as Figures 4–5. Vertical axes are in meters. See Table 1 for lithofacies codes.
Figure 7. Measured section 6LQ. DZ—detrital zircon, SS—sandstone, ZFT—zircon fission track. Vertical axes are in meters. See Table 1 for lithofacies codes.

Figure 8. Measured sections 7LQ, 8LQ, and 14LQ. DZ—detrital zircon, SS—sandstone, ZFT—zircon fission track. Vertical axes are in meters. See Table 1 for lithofacies codes.
The Liuqu Conglomerate, southern Tibet

**TABLE 1. LITHOFACIES AND INTERPRETATIONS USED IN THIS STUDY**

<table>
<thead>
<tr>
<th>Lithofacies code</th>
<th>Description Interpretation</th>
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<tbody>
<tr>
<td>Fal</td>
<td>Laminated red, green, or gray siltstone</td>
</tr>
<tr>
<td>Fp</td>
<td>Massive, bioturbated, mottled siltstone; usually red; occasional carbonate nodules</td>
</tr>
<tr>
<td>Sm</td>
<td>Massive medium- to fine-grained sandstone; bioturbated</td>
</tr>
<tr>
<td>St</td>
<td>Medium- to very coarse-grained sandstone with trough cross-stratification</td>
</tr>
<tr>
<td>Sh</td>
<td>Fine- to medium-grained sandstone with plane-parallel lamination</td>
</tr>
<tr>
<td>Gcm</td>
<td>Pebble to cobble conglomerate, moderately sorted, clast supported, unstratified, poorly organized</td>
</tr>
<tr>
<td>Gcm(i)</td>
<td>Pebble to cobble conglomerate, well sorted, clast supported, unstratified, imbricated (long axis transverse to paleoflow)</td>
</tr>
<tr>
<td>Gch</td>
<td>Pebble to cobble conglomerate, well sorted, clast supported, horizontally stratified</td>
</tr>
<tr>
<td>Gch(i)</td>
<td>Pebble to cobble conglomerate, well sorted, clast supported, horizontally stratified, imbricated (long axis transverse to flow)</td>
</tr>
<tr>
<td>M</td>
<td>Micritic massive gray and yellow marl</td>
</tr>
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**Note:** Modified after Miall (1978) and DeCelles et al. (2011).

Figure 9. Measured sections 10LQ, 11LQ, and 12LQ. DZ—detrital zircon, SS—sandstone, ZFT—zircon fission track. Vertical axes are in meters. See Table 1 for lithofacies codes.
Figure 10. Compiled and correlated (where applicable) measured section from the Liuqu Conglomerate. Arrows indicate paleocurrent directions calculated from clast imbrications. Horizontal arrows indicate modern distance between sections. Stereonet shows paleocurrent directions; light gray arrow is the average.
ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS; Gehrels et al., 2006, 2008; Gehrels and Pecha, 2014) at the University of Arizona LaserChron Center.

**Detrital Zircon Hf Analysis**

Hf isotope analyses were performed on three samples from section 1LQ (see Figs. 2 and 4 for locations) utilizing techniques described by Gehrels and Pecha (2014). Grains with a wide range of U–Pb ages were selected for analysis, and between 31 and 47 grains were analyzed in each sample; 120 grains were analyzed in total. See the Data Repository Item¹ for a description of analytical methods.

**¹GSA Data Repository Item 2016214,** which includes the zircon U-Pb, zircon Hf, and U-Pb age data for detrital zircons, is available at www.geosociety.org/pubs/ft2016.htm, or on request from editing@geosociety.org.

**Ar/Ar Analysis**

Samples for ⁴⁰Ar/³⁹Ar analysis were collected from lamprophyre dikes crosscutting the lower Liuqu Conglomerate near section 8LQ. Biotite was selected for dating because it was the most abundant mineral phase, and individual crystals were large enough to be picked by hand. Samples were analyzed at Arizona Noble Gas Laboratory at the University of Arizona. The closure temperature for argon in biotite ranges between 280 °C and 345 °C, depending on rate of cooling; lower closure temperatures correspond to slower cooling rates (Harrison et al., 1985). Biotite ⁴⁰Ar/³⁹Ar ages are often interpreted as cooling ages in metamorphic settings (e.g., Blanckenburg et al., 1989; Hacker and Wang, 1995), but they can also be used to determine crystallization ages of dikes and extrusive volcanic rocks (e.g., Fleming et al., 1997; Teixeira et al., 1999; Layer et al., 2001). See Data Repository for additional methods description.

**Carbonate Stable Isotopes**

Soil carbonate nodules were sampled from eight locations within sections 1LQ, 4LQ, 5LQ, 6LQ, and 7LQ (see Figs. 2–9 for locations), and detrital limestone clasts from three locations (5LQ, 6LQ, and 7LQ) were collected and analyzed for δ¹³C and δ¹⁸O. All carbonate samples were powdered using a dental drill, baked in vacuo at 150 °C for 3 h, and analyzed with a Finnegan MAT 252 mass spectrometer at the University of Arizona. Results were normalized to NBS-19. Precision for analysis of standards was ±0.1‰ for δ¹³C and ±0.06‰ for δ¹⁸O (1σ). The δ¹³C results are presented in the Data Repository.

Soil carbonate isotopic composition is influenced by the concentration of CO₂ in the atmosphere during formation, the isotopic composition of that CO₂, the respiration rate of the soil, and the type of plants present in that soil (C₃ or C₄; Cerling, 1984; Cerling et al., 1989; Ekart et al., 1999). Of these factors, only respiration rate is locally controlled. Comparison of Liuqu carbon isotopic values to global values through time removes that CO₂, the respiration rate of the soil, and the type of plants present in the Liuqu Conglomerate; (2) ⁴⁰Ar/³⁹Ar ages from a crosscutting dike for a minimum age constraint; (3) the stable isotope composition of paleosol carbonates; and (4) reinterpretation of previously reported palynological evidence.

In total, 1500 new zircon U-Pb analyses can be combined with 518 previously published ages (Wang et al., 2010; Aitchison et al., 2011) in an attempt to determine a maximum depositional age for the Liuqu Conglomerate. However, from a combined data set of 2018 individual analyses, no statistically robust maximum depositional age younger than 94.5 Ma can be calculated for any single sample. The youngest single detrital zircon grain yielded a U-Pb age of 13.7 ± 0.7 Ma. When all analyses are binned, minor populations occur at 18 Ma (two grains), 33 Ma (two grains), 53 Ma (two grains), and 70 Ma (three grains). Although neither of the populations at 18 Ma or 33 Ma is large enough to be statistically

**Zircon Fission Track**

Zircon fission-track analysis provides an estimate of the time at which zircon grains or crystals pass through the 210–240 °C closure temperature, depending on rate of cooling (Brandon et al., 1998; Yamada et al., 1995; Bernet, 2009). In addition to providing information about the timing of burial heating and exhumation of sedimentary deposits, detrital zircon fission-track analyses can also be used to establish maximum depositional ages and place first-order constraints on provenance if they retain their “detrital” ages and are not reset during burial heating (e.g., Garver et al., 1999; Bernet et al., 2001; Najman et al., 2005).

Five samples for detrital fission-track analysis were collected from sections 5LQ, 7LQ, 9LQ, 12LQ, and 14LQ. All samples were collected from sandstones, except for the 14LQ, which was collected from a fine-grained interval. Zircons were separated following exactly the same procedure as for U-Pb analysis as described already. Grains were mounted into Teflon and etched according to the methods outlined in Huford et al. (1991). These samples were irradiated at the Oregon State University Triga reactor in Corvallis, Oregon. Neutron fluence was determined through simultaneous irradiation of IRMM-541 uranium-dosed glasses. Spontaneous and induced tracks were counted using an Olympus BX51 microscope at 1250x magnification at the University of Arizona Fission Track Laboratory. Where possible, over 100 grains per sample were counted; however, in three samples, inclusions, fractures, and low zircon yield reduced the number of countable grains. Individual grain ages were calculated using the zeta calibration method of Huford and Green (1983). Grain populations were determined using Mark Brandon’s BinomFit program (http://markbrandon.commons.yale.edu/software/121-2/).

**Modal Sandstone Petrography and Clast Counts**

The modal-framework grain compositions of 14 medium- to coarse-grained sandstone samples collected from sandy intervals within the Liuqu Conglomerate were determined by point counting according to the Gazzi-Dickinson method (Ingersoll et al., 1984) modified to count 450 grains in each sample. Staining was applied to aid in the identification of Ca-plagioclase and K-feldspars. Point-counting parameters are listed in Table 2. In addition, at least 100 conglomerate clasts were counted at 21 locations throughout measured sections of the Liuqu Conglomerate. Clasts were counted on a grid scaled to the average clast size at each location.

**CHRONOSTRATIGRAPHIC CONSTRAINTS**

Here, we report on and synthesize several lines of evidence to arrive at a conservative age estimate for the Liuqu Conglomerate: (1) detrital zircon U–Pb and fission-track analyses that provide a maximum age for the Liuqu Conglomerate; (2) ⁴⁰Ar/³⁹Ar ages from a crosscutting dike for a minimum age constraint; (3) the stable isotope composition of paleosol carbonates; and (4) reinterpretation of previously reported palynological evidence.

In total, 1500 new zircon U-Pb analyses can be combined with 518 previously published ages (Wang et al., 2010; Aitchison et al., 2011) in an attempt to determine a maximum depositional age for the Liuqu Conglomerate. However, from a combined data set of 2018 individual analyses, no statistically robust maximum depositional age younger than 94.5 Ma can be calculated for any single sample. The youngest single detrital zircon grain yielded a U-Pb age of 13.7 ± 0.7 Ma. When all analyses are binned, minor populations occur at 18 Ma (two grains), 33 Ma (two grains), 53 Ma (two grains), and 70 Ma (three grains). Although neither of the populations at 18 Ma or 33 Ma is large enough to be statistically
rigorous, and could be the result of Pb loss or contamination, their presence does suggest that the Liuqu Conglomerate is Oligocene or younger.

Two zircon fission-track samples (Fig. 12; Table 3) yielded Oligocene–Miocene populations. A sample from near the base of section 5LQ yielded a youngest peak age of 25.1 ± 4.5/5.5 Ma, made up of six grains out of a total of 114 grains counted. A sample from section 14LQ (near the top of the Liuqu type section) yielded a youngest peak of 15.1 ± 3.0/3.7 Ma, made up of five grains out of a total of 38 grains counted. These ages suggest that the upper part of the Liuqu Conglomerate is no older than 18.8 Ma, including analytical uncertainty.

The 40Ar/39Ar ages from a lamprophyre dike crosscutting the Liuqu Conglomerate provide a minimum depositional age. Two biotite samples from a north-south–striking, ~45°E-dipping dike that crosscuts the upper Liuqu type section yielded 40Ar/39Ar plateau ages of 20.1 ± 1.0 Ma (Fig. 12) and 20 ± 3.0 Ma (Data Repository).

Palynological assemblages reported by Wei et al. (2011) led those authors to suggest that the Liuqu Conglomerate is Oligocene in age. They described a diverse assemblage consisting mostly of temperate deciduous angiosperms with conifers and a few evergreen, broad-leaved angiosperms. In addition, they showed rarer occurrences of several herbaceous species that are more typically found in the Miocene, although not totally unknown in the Oligocene. These include Chenopodipollis (chenopods and amaranths), Cyperaceaepollis (sedges), Echitricolporites (Compositae family), and the Graminidites (grass family). Wei et al. (2011) concluded that these herbaceous species were too rarely occurring to indicate an age as young as Miocene. However, we point out that while rare, the
TABLE 2. MODAL PETROGRAPHIC POINT-COUNTING PARAMETERS

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Qm</td>
<td>Monocrystalline quartz</td>
</tr>
<tr>
<td>Qp</td>
<td>Polycrystalline quartz</td>
</tr>
<tr>
<td>Qt</td>
<td>Foliated polycrystalline quartz</td>
</tr>
<tr>
<td>Qss</td>
<td>Monocrystalline quartz in sandstone or quartzite lithic grain</td>
</tr>
<tr>
<td>C</td>
<td>Chert</td>
</tr>
<tr>
<td>Cb</td>
<td>Black chert</td>
</tr>
<tr>
<td>Ch</td>
<td>Chaledony</td>
</tr>
<tr>
<td>Kt</td>
<td>Total quartzose grains (Qm + Qp + Qms + C +Cb + Ch)</td>
</tr>
<tr>
<td>K</td>
<td>Potassium feldspar (including perthite, myrmekite, microcline)</td>
</tr>
<tr>
<td>P</td>
<td>Plagioclase feldspar (including Na and Ca varieties)</td>
</tr>
<tr>
<td>F</td>
<td>Total feldspar grains (K + P)</td>
</tr>
<tr>
<td>Lv</td>
<td>Mafic volcanic grains</td>
</tr>
<tr>
<td>Lv</td>
<td>Felsic volcanic grains</td>
</tr>
<tr>
<td>Lv</td>
<td>Vitric volcanic grains</td>
</tr>
<tr>
<td>Lv</td>
<td>Microlitic volcanic grains</td>
</tr>
<tr>
<td>Lv</td>
<td>Total volcanic lithic grains (Lv_m + Lv_f + Lv_v + Lv_x)</td>
</tr>
<tr>
<td>Lsh</td>
<td>Mudstone</td>
</tr>
<tr>
<td>Lph</td>
<td>Phyllite</td>
</tr>
<tr>
<td>Lsm</td>
<td>Schist (mica schist)</td>
</tr>
<tr>
<td>Lc</td>
<td>Carbonate lithic grains</td>
</tr>
<tr>
<td>Musc</td>
<td>Muscovite</td>
</tr>
<tr>
<td>Lm</td>
<td>Total metamorphic lithic grains (Lph + Lsm + Qpt)</td>
</tr>
<tr>
<td>Ls</td>
<td>Total sedimentary lithic grains (Lsh + Lcc + C +Cb + Qss)</td>
</tr>
<tr>
<td>Lt</td>
<td>Total lithic grains (Ls + Lm + Lp + Qp)</td>
</tr>
<tr>
<td>L</td>
<td>Total nonquartzose lithic grains (Lv + Ls+ Lph + Lsm + Lc)</td>
</tr>
</tbody>
</table>

Note: Accessory minerals: tourmaline, kaolinite, zircon, pyroxene.

...herbaceous species found do occur consistently in most of the Wei et al. (2011) samples. Chenopods and sedges were found in 12 of the 16 samples analyzed by Wei et al. (2011), whereas composites and grasses were found in five and three samples, respectively. The more common occurrence of the herbaceous species in the Qiuwu is, in our opinion, sufficient evidence to indicate an early Miocene age for the pollen assemblage described by Wei et al. (2011).

A final age constraint comes from the carbon isotopic data from paleosol carbonates in the Liuqu Conglomerate, which can be used to broadly constrain its depositional age. The average δ13C value of Liuqu soil carbonates is ~9.4‰ ± 1.13‰. The values overlap with global averages (Fig. 12) only prior to 50 Ma or after 20 Ma and suggest that the Liuqu Conglomerate was not deposited between those times (Ekart et al., 1999).

Because no single geochronologic technique has been able to provide a definitive depositional age for the Liuqu Conglomerate, we integrated all available data sets with the goal of arriving at an age that is consistent with the largest number of available chronometers. When all of the available data are considered (Fig. 13), ca. 20–19 Ma emerges as the most viable depositional age for the Liuqu Conglomerate. If correct, this would mean that deposition of the Liuqu Conglomerate partly overlapped in time with the last stages of deposition of the Kailas Formation.

The only age data that fall outside this range (if analytical uncertainty is included) are the youngest single detrital zircon U-Pb age (13.7 Ma; this study) and the Eocene age interpreted by Fang et al. (2006) based on plant fossils. The youngest single detrital zircon U-Pb age is not a robust indicator of maximum depositional age (Dickinson and Gehrels, 2009), as there is no way to rule out Pb loss or contamination. Although Fang et al. (2006) did not find any plant species that would specifically indicate an age younger than Eocene, they noted that the species assemblage they documented overlaps strongly with that found in the Qiuwu Formation, which has been dated as latest Oligocene to mid-Miocene (Li, 2004; S. Li et al., 2015). It is important to note that plant fossils in the Qiuwu Formation incorrectly indicated a Late Cretaceous to early Paleogene age (Wang et al., 2013, and references therein) prior to revision by palynologic and radiometric methods (Li, 2004; Wang et al., 2013; S. Li et al., 2015), and the possibility of a similar discrepancy between apparent plant fossil age and true age should not be ruled out for the Liuqu Conglomerate.

We stress that establishing the age of the Liuqu Conglomerate is a work-in-progress because of the unusual geochronologic challenges posed by dating of nonfusceous conglomerates and paleosols. To our knowledge, paleomagnetism has not been attempted on the Liuqu Conglomerate because the paleosols and certainly the conglomerates are not ideal for such analysis. There are no reported tuffs in the Liuqu Conglomerate, which would be surprising if the Liuqu were older than 40 Ma, when arc-related volcanism was still very active, but is unsurprising if the true age is younger than 30 Ma, when arc volcanism had greatly diminished (Lee et al., 2009; Ji et al., 2009; Zhu et al., 2011). Reduced volcanic activity in the region also reduces the likelihood of obtaining detrital zircons with ages close to the age of deposition. The likelihood is further diminished by the small catchment area probably represented by Liuqu-age paleoires, as well as their north-directed paleoflow, toward rather than away from the Gangdese arc.

PROVENANCE

Modal Sandstone and Conglomerate Compositions

Quartzose grains were identified as monocrystalline quartz (Qm), polycrystalline quartz (Qp), chert (Qch), or tectonized polycrystalline quartz (Qpt). Feldspars were identified as potassium types (orthoclase, perthite, myrmekite, microcline) or plagioclase (P). Lithic grains are abundant in Liuqu sandstones and were identified as shale (Lsh), phyllite (Lph), limestone (Lc), vitric volcanic (Lv), or felsic volcanic (Lv). Accessory minerals include muscovite, serpentine, and rare tourmaline.

The composition of Liuqu Conglomerate sandstones is variable (Fig. 14); however, no clear up-section or along-strike trends are present. Sandstone provenance was dominated by sedimentary and low-grade metasedimentary rocks, such as those in the regions directly north and south of the Liuqu basin.

Dominant conglomerate clast lithologies are green lithic sandstone, red and green chert, and brown sandstone. Serpentinitic clasts, white sandstone, quartzite, limestone, and basalt are present in lower abundances. Consistent with the apparent absence of depositional age zircons, no granite or volcanic clasts were identified anywhere within the Liuqu Conglomerate. No strong up-section trends were observed anywhere except for section 1LQ, where sandstone replaces chert as the dominant clast lithology in the upper portion of the section. However, the clast composition of the Liuqu Conglomerate does vary somewhat along strike. Chert, serpentinite, basalt, and gabbro are much more abundant in western Liuqu exposures (Fig. 14), whereas green lithic sandstone is the dominant lithology to the east. The only exception to this pattern is section 1LQ, where chert and sandstone clasts occur in roughly equal proportions.

Detrital Zircon Ages and Hf Isotope Compositions

Most Liuqu Conglomerate zircon U-Pb age spectra are broadly similar and were combined for regional provenance interpretation. Samples collected from sandstones and fine-grained intervals within the same section showed no significant variation in their age spectra and little up-section variation. Although some along-strike heterogeneities in age spectra are...
Figure 12. (A) Detrital zircon fission-track results from the Liuqu Conglomerate. (B) $^{40}$Ar/$^{39}$Ar plateau and inverse isochron ages from biotite in lamprophyre dikes cutting the Liuqu Conglomerate. MSWD—mean square of weighted deviates. (C) Carbon isotopic composition of global soil carbonates (Ekart et al., 1999) compared to Liuqu values. VPDB—Vienna Peedee belemnite.

### TABLE 3. DETRITAL ZIRCON FISSION-TRACK DATA

<table>
<thead>
<tr>
<th>Sample</th>
<th>N</th>
<th>P1</th>
<th>P2</th>
<th>P3</th>
</tr>
</thead>
<tbody>
<tr>
<td>5LQ-196</td>
<td>114</td>
<td>25.1–4.5/+5.5</td>
<td>5.30%</td>
<td>45.80%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>72.8–7.5/+8.3</td>
<td>48.90%</td>
<td></td>
</tr>
<tr>
<td>9LQ-500</td>
<td>105</td>
<td>65.0–10.1/+11.9</td>
<td>45.80%</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>203–25.5/+29.0</td>
<td>45.80%</td>
<td></td>
</tr>
<tr>
<td>5LQ-50</td>
<td>38</td>
<td>15.1–3.0/+3.7</td>
<td>15.20%</td>
<td>84.80%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>32.6–8.7/+11.8</td>
<td>15.20%</td>
<td></td>
</tr>
<tr>
<td>9LQ-500</td>
<td>105</td>
<td>65.0–10.1/+11.9</td>
<td>15.20%</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>203–25.5/+29.0</td>
<td>15.20%</td>
<td></td>
</tr>
<tr>
<td>7LQ-500</td>
<td>105</td>
<td>65.0–10.1/+11.9</td>
<td>15.20%</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>203–25.5/+29.0</td>
<td>15.20%</td>
<td></td>
</tr>
<tr>
<td>12LQ-24</td>
<td>26</td>
<td>32.6–8.7/+11.8</td>
<td>15.20%</td>
<td>84.80%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>123.7–15.3/+17.4</td>
<td>15.20%</td>
<td></td>
</tr>
<tr>
<td>12LQ-24</td>
<td>26</td>
<td>32.6–8.7/+11.8</td>
<td>15.20%</td>
<td>84.80%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>123.7–15.3/+17.4</td>
<td>15.20%</td>
<td></td>
</tr>
</tbody>
</table>

Note: N—number of grains counted in each sample. Binomial fitted age peaks (P1, P2, P3; calculated using BinomFit) are given in Ma with errors representing 95% confidence. Relative size of each peak is given in percent. See Methods sections for sample preparation details.
Figure 13. Compilation of all available geochronologic data for the Liuqu Conglomerate. Dark bars represent time during which deposition of the Liuqu Conglomerate is permitted by each chronometer, with lighter bars showing uncertainty, when available. “Binned population” makes use of all available zircon data. References: carbon stable isotopes—Ekart et al. (1999); Great Counter Thrust (GCT) movement—Murphy et al. (2002) and Yin (2006); plant fossils—Tao et al. (1988) and Fang et al. (2006); palynology—Wei et al. (2011) and this study; zircon U-Th/He—Li et al. (2015). ZFT—zircon fission track.

Figure 14. (A) Results of sandstone point counts. Petrographic fields are from Dickinson (1985). (B) Results of conglomerate clast counts for the Liuqu Conglomerate plotted by along-strike outcrop location (Fig. 2). (C) Total clast counts for all sites. SS—sandstone.
present, these differences reflect small, nonintegrated catchment areas for Liuqu alluvial fans. The detrital zircon U-Pb ages from the Liuqu Conglomerate (Fig. 15) cluster at 95 Ma, 130 Ma, 155 Ma, 500–550 Ma, and 1100–1200 Ma. Grains younger than 80 Ma are rare.

The $\varepsilon_{\text{Hf}}(t)$ values (Fig. 16) range between -27.6 and 14.7. These values are scattered, but age/$\varepsilon_{\text{Hf}}(t)$ values cluster at 80–120 Ma/$\varepsilon_{\text{Hf}}(t) = 5.8–14.6$; 500–600 Ma/$\varepsilon_{\text{Hf}}(t) = -11$; and 500–580 Ma/$\varepsilon_{\text{Hf}}(t) = -26$ to -17.3. The most prominent age peak within the Liuqu Conglomerate age spectra is at ca. 95 Ma; the $\varepsilon_{\text{Hf}}(t)$ value of grains in this population ranges from 5.8 to 14.6. Ages also cluster around 126 Ma; these grains typically have $\varepsilon_{\text{Hf}}(t)$ values ranging from -13.1 to 12.6.

**Zircon Fission Track**

Most fission-track samples yielded a minor young peak between 15 Ma and 32 Ma and a major peak between 60 Ma and 100 Ma made up of grain ages ranging from 50 Ma to 150 Ma (Fig. 12; Table 3). Sample 5LQ-196 also yielded a minor peak at 200 Ma, and sample 9LQ-500 yielded a minor peak at 320 Ma.

**Provenance Interpretation and Discussion**

Consistent with north-northwest paleocurrent indicators, the sandstone petrographic and conglomerate clast-count data suggest that the Liuqu Conglomerate was derived from source terranes to the south. Chert grains (all varieties) were likely derived from radiolarian chert exposed in the suture zone (Guilmette et al., 2009; Bédard et al., 2009). Green, lithic sandstone was most likely derived from the mélangé complex south of the suture zone (Dupuis et al., 2006; Cai et al., 2011; An et al., 2015). White sandstone, quartzite, and limestone were also likely derived from the mélangé (Dupuis et al., 2006; An et al., 2015), with a probable ultimate origin in the Tethys Himalayan domain (Liu and Einsele, 1994; An et al., 2015). We interpret the along-strike changes in composition to reflect the relative proportion of sediment derived from ophiolitic sequences. Some ophiolitic rocks were present within the source area of westerly sections of the Liuqu Conglomerate, whereas the source areas in more easterly locations were dominated by sedimentary matrix mélanges rather than ophiolites and ophiolitic matrix.
mélanges. This along-strike variation suggests that the preserved Liuqu Conglomerate exposures were derived from separate catchments.

The absence of granite and volcanic clasts in the Liuqu Conglomerate precludes direct sediment contribution by the Gangdese arc or the Linzizong volcanics to the north. This contrasts with the Kailas Formation, which received abundant granitic and volcanic detritus from ca. 26 Ma to 18 Ma (Aitchison et al., 2009; DeCelles et al., 2011; Wang et al., 2013; Leary et al., 2016), indicating that the Gangdese arc must have been incised and shedding sediment by that time. The lack of Gangdese clasts suggests that a sedimentologic barrier blocked coarse-grained Gangdese material from transiting into the Liuqu basin.

Detrital zircon grains with U-Pb ages between 80 Ma and 110 Ma and juvenile (>0) \( \epsilon_{Hf} \) values must have come originally from the Gangdese arc. There are no identified igneous rocks to the south from which these zircons could have been derived (Fig. 16; Ji et al., 2009; Zhu et al., 2011; Ji et al., 2012, 2014). However, grains of the same age are abundant in sandstone blocks incorporated into the mélange to the south of the Liuqu Conglomerate (Cai et al., 2012), and this is likely the source of these grains.

Additional provenance insight is provided by qualitative comparison of probability density function (PDF) curves and binned age histograms (Fig. 15). A prominent age peak between 80 and 100 Ma overlaps with Xigaze Group zircons (Orme et al., 2014; An et al., 2014). This peak also overlaps with ages from the accretionary mélange, much of which was originally derived from the Xigaze Group (Cai et al., 2012; An et al., 2015). A second prominent age peak within the Liuqu spectra is centered at 128 Ma and closely matches a similar peak from the accretionary mélange to the south of the Liuqu Conglomerate (Cai et al., 2012), and this is likely the source of these grains.

Of 2018 individual zircon analyses considered here (published and new), only 16 grains are younger than 79 Ma. For comparison, 271 grains were dated between 79 and 118 Ma. The >60 m.y. gap between depositional age and the age of the youngest major zircon population indicates that the rocks from which the Liuqu Conglomerate was derived were, for the most part, no younger than 80 Ma, and grains younger than 80 Ma were likely derived from volcanic air-fall events. This is consistent with the inference that the Liuqu Conglomerate was derived from local catchments incising the accretionary mélange complex and possibly the basal part of Xigaze forearc basin succession, as these rocks contain zircon populations capable of producing the age spectra found in the Liuqu Conglomerate.

The wide spread in zircon fission-track ages (Fig. 12; Table 3) in each sample indicates that values were not reset during burial heating, and thus individual grains record their cooling history prior to their inclusion in the Liuqu Conglomerate. This is consistent with the findings of G. Li et al. (2015), which showed that detrital zircon (U-Th)/He ages from the Liuqu Conglomerate were not reset by burial. Because the zircon He system is reset at a lower temperature than the zircon fission-track system (Brandon et al., 1998; Yamada et al., 1995; Reiners et al., 2004; Bernet, 2009; Guenthner et al., 2013), it would have been impossible for burial heating to reset zircon fission-track ages for the Liuqu Conglomerate without also resetting zircon U-Th/He ages.

Zircons with fission-track ages younger than 50 Ma are far more abundant than zircons with U-Pb ages younger than 50 Ma. Thus, we argue that these young fission-track ages are unlikely to be igneous cooling ages, and we interpret them to represent cooling associated with exhumation along thrust faults in the mélange to the south. The most abundant zircon fission-track ages are in the 50–150 Ma range. These ages closely match the U-Pb ages of Liuqu zircons (Fig. 15) as well as the U-Pb age spectra within the mélange (Cai et al., 2012). We interpret these ages to represent primary cooling ages of Gangdese arc zircons. This implies that only small volumes of Liuqu Conglomerate source rocks were in areas that were heated above ~210 °C after the incorporation of Gangdese zircons into the mélange.

Liuqu Conglomerate provenance data are best explained by a two-stage transport history. Prior to the India-Asia collision, Asian sediment was shed southward across an overfilled forearc basin and into the subduction trench, and some of these sediments were accreted to the southern margin of Asia along with mélange rocks. This concept is consistent
with work in the Xigaze forearc basin (Orme et al., 2014), which shows a transition from deep to shallow marine facies at 81 Ma. This change likely marks the onset of a filled basin state and would have allowed sediment to bypass the forearc depocenter. Sediment bypassing the forearc would have been derived directly from the Gangdese arc and would have contained large numbers of 80–120 Ma zircons with juvenile Hf t values. Because the deposits incorporated into the accretionary mélangé would have been distal to the arc, the grain size of this sediment would have been sand sized or smaller. Coarse-grained sediment would have been trapped proximal to the arc in locations such as the northern forearc basin. This is consistent with the work of Cai et al. (2012), who mapped blocks of Asian-derived sandstone in the accretionary mélangé; no coarse-grained clastic material was found in that study.

By the time of Liuqu Conglomerate deposition around 20 Ma, the local topographic gradient in the suture zone must have been reversed, and sediment was transported to the north into what may have been the early Yarlung River. The fluvial and alluvial systems would have been draining small catchment areas containing rocks of mélangé and possibly Tethyan Himalayan affinity. Gangdese zircons from sandstone units within the mélangé would have been recycled northward and deposited in the Liuqu Conglomerate. Although this sediment would have contained Gangdese zircons, no volcanic or granitic clasts would have been available for northward transport out of the mélangé or Tethyan terranes, since Asian coarse-grained sediment did not reach those units prior to India-Asia collision.

Whether the Liuqu Conglomerate was partially derived from Tethyan Himalayan rocks is still uncertain. Wang et al. (2010) argued that detrital zircons from the Liuqu Conglomerate with ages between 113 Ma and 139 Ma and Hf t values between −7.9 and -1 must have been derived from the Wölong volcanics in the Tethyan Himalaya (Hu et al., 2010). Although this remains a possible source for these zircons, an increasingly large Hf data set from the Gangdese arc (Zhu et al., 2011) suggests that 113–139 Ma zircons with Hf t values between −7.9 and -1 could have been produced in the Gangdese arc; these new data suggest that the Wölong volcanics are not a required source for the Liuqu Conglomerate. Here, we propose that lithology and isotopic provenance data allow, but do not require, Tethyan Himalayan rocks as a source terrane for the Liuqu Conglomerate.

**CAUSES OF LIUQU BASIN FORMATION**

Extensional processes have been invoked to explain other coarse-grained deposits in the region (DeCelles et al., 2011; Wang et al., 2013; Leary et al., 2016); however, we interpret the Liuqu Conglomerate to have been deposited in a contractional, wedge-top environment (DeCelles and Giles, 1996) in which sediment was accommodated within thrust structures. This conclusion is primarily supported by growth strata in the Liuqu Conglomerate type section (Fig. 3). Here, bedding is 30° overturned immediately beneath a thrust contact with the Xigaze ophiolite and rapidly shallows to dips of ~20° south of the fault. This indicates that the fault, which we interpret as a back-thrust splay related to the Great Counter Thrust, was active during Liuqu deposition (e.g., Manayón et al., 1986; Lawton et al., 1999; Zapata and Allmendinger, 1996; Vergès et al., 2002). Northward paleocurrent indicators preclude simple thrust growth-structure geometry, where sediment is shed in the same direction as thrust vergence (Vergès et al., 2002). Rather, sediment transport in the Liuqu Conglomerate was opposite of thrust vergence, and thus the fault in the core of the growing antiform must have been blind. This implies that sediment accommodation was structurally produced in a triangle zone within the Great Counter Thrust wedge-top (Fig. 17). As such, sediment accommodation was produced by damming behind thrust-produced structural relief rather than flexural subsidence.

The facies and architecture of the Liuqu Conglomerate are consistent with this interpretation and with wedge-top depozones in orogens worldwide (DeCelles, 1994; Lawton et al., 1999; Heermance et al., 2007). Common architectural features of this depozone are laterally continuous conglomerate intervals with scoured and channelized bases separated by fine-grained intervals. Also, the Liuqu Conglomerate lacks any facies or architectural elements commonly associated with continental rift basins such as lacustrine, turbidite, or playa deposits (Friedmann and Burbank, 1995; Gawthorpe and Leeder, 2000), although this absence does not preclude their existence in more distal portions of the Liuqu basin.

**GEODYNAMIC IMPLICATIONS**

New age constraints for the Liuqu Conglomerate change our understanding of the tectonic framework in which these rocks were deposited. Although initially interpreted to be related to island-arc accretion onto India (Davis et al., 2002; Aitchison et al., 2007) or initial India-Asia...
Beginning around 26 Ma in western Tibet and ca. 24 Ma near the Liuqu Conglomerate exposures (Leary et al., 2016), the India-Asia suture zone experienced a major geographic and sedimentologic reorganization. Prior to India-Asia collision, sediment eroded from the Gangdese arc was transported across the forearc basin and shed into the subduction trench. Postcollisional deposits within the suture zone are sparse, and there is little record of sediment transport patterns between 50 Ma and 26 Ma. Beginning ca. 26 Ma, extension attributed to Indian slab rollback and breakoff (DeCelles et al., 2011; Leary et al., 2016) produced a narrow basin in which the Kailas Formation was deposited ~20 km north of present Liuqu exposures. The Kailas Formation contains organic-rich, deep-lacustrine facies and coal (DeCelles et al., 2011), suggesting that the Kailas basin formed at low elevation. Thermochronologic data indicate that the Kailas basin was subsequently exhumed by 17–16 Ma (Carrapa et al., 2014). The upper part of Kailas Formation contains sediment derived from the south and has been interpreted to be a result of uplift in response to initial Great Counter Thrust slip (DeCelles et al., 2011). Uplift may have been a response to Indian slab breakoff followed by rapid northward underthrusting of India (Fig. 17), and rapid exhumation during the early Miocene probably reflects integration of the regional Yarlung River drainage system (Carrapa et al., 2014).

The Liuqu Conglomerate was deposited several million years after the onset of Kailas basin extension in the central suture zone (Leary et al., 2016). Unlike the majority of the Kailas Formation, there is clear evidence that the Liuqu Conglomerate was deposited in a contractional environment. Thus, we suggest that the Liuqu Conglomerate was deposited during or immediately after Indian slab breakoff as contractional deformation resumed in the suture zone, but before any significant isostatic rebound elevated the suture zone (van der Meulen et al., 1999; Wortel and Spakman, 2000; Leary et al., 2016). In this scenario, the Liuqu Conglomerate may represent the southern, wedge-top extent of the upper Kailas Formation depositional system. In this scenario, the stark differences in clast composition and detrital zircon ages between the Liuqu Conglomerate and Kailas Formation would require that modern exposures of both units represent proximal portions of each system, and the point at which these systems joined is not preserved. However, upper portions of the Kailas Formation do contain chert and green lithic sandstone clasts (Leary et al., 2016), which may indicate that the final stages of deposition at those locations included some sediment derived from the same source areas as the Liuqu Conglomerate.

A pulse of exhumation within the suture zone at ca. 17 Ma (Carrapa et al. 2014) has been attributed to capture of the paleo–Yarlung River by the Brahmaputra system during the early Miocene (Bracciali et al., 2015), and there is evidence that the paleo–Yarlung River may have been active and draining into southeast Asia prior to that time (Robinson et al., 2014). Northward paleocurrent indicators in the Liuqu Conglomerate suggest that a drainage divide must have separated the suture zone from the Himalayan rivers to the south; this is consistent with the absence of evidence in detrital zircon and petrographic data sets for significant transport of detritus from the Gangdese arc and India-Asia suture zone southward into the Miocene Siwalik Group foreland basin deposits along the south flank of the Himalayan thrust belt (DeCelles et al., 1998; Szulc et al., 2006; Najman, 2006). The data presented here also run counter to the model presented by Tremblay et al. (2015) that argued that Gangdese arc sediment was transported directly southward across the suture zone and Himalayan fold-and-thrust belt until ca. 10 Ma. Here, we suggest that the Liuqu Conglomerate represents the southern portion of the early paleo–Yarlung River catchment shortly after establishment of an east-west drainage system. From the location of current exposures, Liuqu sediment would have been transported farther northward into the central suture zone before traveling axially out of the suture zone.

CONCLUSIONS

The Liuqu Conglomerate was deposited in stream-dominated alluvial fans within the India-Asia suture zone. Sediment was eroded from small catchments incising thrust sheets within the accretionary mélangé and shed northward. No first-cycle sediment from north of the suture zone was incorporated into the Liuqu Conglomerate, and a hydrologic divide must have existed between the Liuqu basin and the Gangdese arc. Second-cycle Gangdese zircons were incorporated into the Liuqu Conglomerate after being recycled from rocks of the accretionary mélangé south of the suture zone.

New age constraints on the Liuqu Conglomerate suggest that it was deposited at ca. 20 Ma, although further refinement is necessary. This would have coincided with the transition from slab rollback–driven extension in the central suture zone at 24 Ma (Leary et al., 2016) to contractional deformation within the Great Counter Thrust system. This also places Liuqu deposition just before, or coeval with, the initiation and/or capture of the paleo–Yarlung River at ca. 18 Ma (Robinson et al., 2014; Bracciali et al., 2015), and we suggest that the Liuqu Conglomerate represents the southern portion of the paleo–Yarlung River catchment.


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