1 BASIN ANALYSIS OF THE ALBIAN-SANTONIAN XIGAZE FOREARC, LAZI REGION, SOUTH-
2 CENTRAL TIBET
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13 **ABSTRACT**
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15 The Xigaze forearc basin records the evolution of the southern Lhasa terrane convergent margin,
16 largely affected by Neo-Tethyan subduction processes, prior to the Paleocene Tethyan
17 Himalaya—Eurasia collision. New geologic mapping and U-Pb detrital-zircon geochronologic
18 data from the Lazi region, 340 km southeast of Lhasa, show that forearc basin sedimentation
19 began ca. 110 Ma conformably atop the Yarlung-Tsangpo ophiolitic mélangé. By this time, the
20 arc-trench system along the southern margin of the Lhasa terrane (Eurasia) consisted of an
21 accretionary complex, overlying ophiolitic mélangé, the Xigaze forearc basin, and the Gangdese
22 magmatic arc. There is no geological evidence in the Lazi region for more than one subduction
23 zone between the southern Lhasa terrane margin and India. Sedimentological facies analysis
24 from Albian to Santonian clastic and carbonate sedimentary rocks preserved in the Xigaze
25 forearc basin indicate deep-marine sedimentation characterized by hemipelagic carbonate and
26 volcanogenic sediment-gravity-flow deposits. Sandstone modal petrographic and U-Pb detrital-
27 zircon geochronologic data reveal Asian continental margin, Gangdese magmatic arc, and central
28 to northern Lhasa terrane provenance. During this time, basin fill was deposited in cycles of high
29 and low sediment flux characterized by alternating successions of clastic turbidite inner- to
30 outer-fan deposits and hemipelagic limestone and marlstone sequences. Along-strike differences

in the timing of initial forearc basin sedimentation are likely the result of intra-basin topography and/or diachronous development of ophiolitic forearc basement.

Key Words: Forearc basin, Xigaze, Tibet, provenance, subsidence

INTRODUCTION

Forearc basins, located between the trench and the magmatic arc, are prominent tectonic elements of ocean-continent convergent margins. Strata deposited within forearc basins record changes in facies and depositional environments in response to accommodation space that is influenced by the tectonic evolution of the upper plate during subduction and magmatic-arc activity (Dickinson, 1995). Forearc basins often develop atop extended continental or trapped oceanic crust; the contact between the basin and its basement provides insight into geodynamic processes active before and during the onset of forearc sedimentation (Noda, 2016; Clift and Vannucchi, 2004; Dickinson, 1995). However, the primary relationships between forearc basin strata and underlying basement rocks are often difficult to resolve, as the majority of forearc basins are submarine and there is a paucity of well-preserved ancient forearcs on land.

The initiation of a forearc basin is largely contingent upon whether the continental margin is accretionary or erosive (Hamilton, 1969; Ernst, 1970; Karig and Sharman, 1975). In accretionary margins, there is sufficient transfer of material from the subducting plate to the overriding plate to build an outer forearc high, either by frontal offscraping or by underplating of the forearc wedge, which results in sufficient relief to create a forearc basin (Clift and Vannucchi, 2004; Simpson, 2010). In addition, whether a forearc basin develops above oceanic or extended continental crust controls the early stages of subsidence and the availability of accommodation space behind the growing accretionary prism (Dickinson and Suczek, 1979). Thick accumulations of forearc-basin strata are most commonly observed in regions where

oceanic crust forms the basement of the forearc basin (Bachman et al. 1983; Speed et al., 1989; Dickinson, 1995). Thus, knowledge of the composition of the forearc basement is crucial for understanding how a forearc basin initiates and evolves. In this study, we investigate the Xigaze forearc basin in southern Tibet, one of the best-preserved ancient forearc basins on Earth. We document the development of an accretionary margin in southern Tibet prior to modification by continent-continent collision, thereby informing our understanding of ancient forearc-basin systems.

The Xigaze forearc basin developed along the southern margin of Asia during the Early Cretaceous to Eocene as northward subduction of Neo-Tethyan oceanic lithosphere accommodated the northward motion and subsequent collision of the Indian plate during the Paleocene (Garzanti et al., 1987; DeCelles et al., 2014; Hu et al., 2015). The forearc basin is presently exposed as a 5-8-km-thick sequence of marine and nonmarine strata that extends laterally across 550 km of the Indus-Yarlung Suture Zone (IYSZ) (Figure 1A). A series of ophiolitic mélanges, termed the Yarlung-Tsangpo ophiolite (also referred to as the Xigaze ophiolite between Dazhuqu and Sangsang) (Girardeau et al. 1985), crop out along the southern margin of the Xigaze forearc basin. Sedimentological studies of the Xigaze forearc have focused mainly on outcrops exposed along its eastern limit, near Xigaze (Einsele et al., 1994; Dürr, 1996; Wang et al., 2012; An et al., 2014) and its western limit, near Saga (Ding et al. 2005; Orme et al. 2014; Hu et al., 2015) (Figure 1); however, no detailed sedimentology or stratigraphy has been documented for the central part of the basin. In addition, direct observations of the relationship between the forearc basin and Yarlung-Tsangpo ophiolite are limited to a few localities (An et al. 2014; Huang et al. 2015) and remain enigmatic, in that they are in many places overprinted by Cenozoic faulting along the southern limit of the forearc basin. Thus, significant portions of the

Xigaze forearc basin remain unstudied, limiting our understanding of the timing and mechanisms of forearc development.

This study focuses on the oldest sedimentary strata preserved in the Xigaze forearc basin to (1) document the relationship between the Yarlung-Tsangpo ophiolite and Xigaze forearc basin, (2) reconstruct the sedimentary environments of the southern margin of Asia during the initial stages of forearc deposition, and (3) reconstruct the basin history to evaluate possible mechanisms controlling forearc subsidence. We report new regional geologic mapping and sedimentologic, modal petrographic, and U-Pb detrital-zircon geochronologic data from Albian through Santonian clastic and carbonate sedimentary rocks preserved northeast of Lazi (also referred to as Latzi, Latze, or Lhatse) (Figure 1). The new data support hypotheses that forearc-basin sedimentation began conformably atop the Yarlung-Tsangpo ophiolite (Wu, 1984; Girardeau et al., 1985; Huang et al., 2015) and detail the paleogeography of a previously unexplored segment of the Xigaze forearc basin. These results help reconstruct the evolution of the southern margin of Eurasia (Lhasa terrane) prior to its collision with India in the Cenozoic.

GEOLOGIC SETTING

The Lhasa terrane, located between the Bangong suture to the north and the Indus-Yarlung suture to the south, is the southernmost of three terranes that together with remnants of the Indian continental margin make up the Tibetan Plateau (Burg and Chen, 1984; Allègre et al. 1984; Dewey et al. 1988). The Lhasa terrane consists primarily of Cretaceous-lower Cenozoic granitoid and volcanic rocks of the Gangdese magmatic arc, Cretaceous and Cenozoic volcanoclastic strata, and minor exposures of Jurassic, Paleozoic, and Cenozoic sedimentary and metasedimentary rocks (Schärer et al. 1984; Debon et al. 1986; Yin and Harrison, 2000; Kapp et al. 2003, 2007; Leier et al. 2007; He et al. 2007; Lee et al., 2009; Zhu et al. 2011).

Extending for > 2000 km across southern Tibet, the IYSZ formed during the collision between India and Asia during the early Cenozoic (Garzanti et al. 1987; DeCelles et al., 2014). The IYSZ consists of four tectonic elements: the Gangdese magmatic arc, the Xigaze forearc basin, the Yarlung-Tsangpo ophiolite, and an accretionary complex (Figure 1). The Gangdese magmatic arc and its extrusive equivalents were emplaced along the southern margin of the Lhasa terrane primarily between ca. 150 and ca. 40 Ma (Lee et al. 2009). The Yarlung-Tsangpo ophiolite consists of fragments of serpentinized mantle peridotite, gabbro, sheeted dikes, pillow basalt, and radiolarian chert (Göpel et al. 1984; Ziabrev et al. 2003; Hébert et al. 2012; Guilmette et al., 2012; Huang et al. 2015). Zircon U-Pb and $^{40}\text{Ar}/^{39}\text{Ar}$ dating from ophiolitic igneous rocks and biostratigraphy from radiolarian chert constrain its formation age to 128-120 Ma (Göpel et al. 1984; Ziabrev et al., 2003; Guilmette et al., 2009; Hébert et al. 2012; Dai et al., 2013). Whether the ophiolite formed above a subduction zone in a forearc position (Hébert et al. 2012) or was obducted onto India's northern margin prior to its collision with Asia (Burg and Chen, 1984; Girardeau et al. 1985; Searle et al. 1987; Ding et al. 2005; Guilmette et al. 2009) is debated. Recent paleomagnetic and geochronologic studies favor a model in which the ophiolite formed by forearc spreading along the southern margin of the Lhasa terrane during the early Cretaceous and therefore represents the basement to the Xigaze forearc basin (An et al. 2014; Huang et al. 2015). South of the Yarlung-Tsangpo ophiolite, exposures of sedimentary-matrix *mélange* comprise the Cretaceous-Eocene subduction-accretion complex (Cai et al. 2012).

The Xigaze forearc basin is preserved as an east-west-trending syncline for 550 km along the IYSZ (Figure 1). Along its southern margin, it is juxtaposed against fragments of the Yarlung-Tsangpo ophiolite, ophiolitic *mélange*, or sedimentary-matrix *mélange* by segments of a north-vergent thrust system. The Great Counter Thrust (GCT), which was active during the

Miocene (Yin and Harrison, 2000), juxtaposes the northern limit of the Xigaze forearc against Oligocene-Miocene conglomerates, which were deposited in buttress unconformity atop Gangdese arc rocks (Einsele et al. 1994; DeCelles et al. 2011).

STRATIGRAPHY

The stratigraphy of the Xigaze forearc basin is divided into two lithostratigraphic groups, the lower to upper Cretaceous Xigaze Group (Wen, 1974; Wu et al., 1977; Einsele et al. 1994; Wang et al. 2012; An et al. 2014) and the Paleocene to lower Eocene Tso-Jiangding Group (Ding et al., 2005; Orme et al., 2014; Hu et al. 2015). The Tso-Jiangding Group, deposited during the middle to final stages of forearc sedimentation, is not exposed in our study area and is not discussed herein.

The Xigaze Group comprises the Sangzugang, Chongdui, and Ngamring Formations (Figure 2). The Sangzugang Formation is 60 to 230 m thick and consists of gray-blue thick-bedded limestones rich in coral and bivalve fossils. It crops out in fault contact with the Kailas Formation and the Ngamring Formation, near Sangzugang village, east of Xigaze (Wu et al. 1977). Its original stratigraphic position is unknown but is interpreted by Liu et al. (1988) to have been deposited in a reef environment along the northern part of the forearc basin during the Aptian-Albian (Cherchi and Schroeder, 1980; Bassoullet et al. 1980).

The Chongdui Formation traditionally consists of two members, a lower chert and siliceous shale member and an overlying sandstone member with minor interbedded shale and limestone (Figure 2; Wu, 1984; Ziabrev et al. 2003). Recently, An et al. (2014) revised the stratigraphy to include only the chert member and interpreted the sandstone member as the base of the Ngamring Formation. Following this refinement, the Chongdui Formation is 85-100 m thick and comprises purple-green radiolarian chert interbedded with green siliceous shale and

reflects deep-marine deposition atop the Yarlung-Tsangpo ophiolite (Huang et al. 2015; An et al. 2014). Age constraints from biostratigraphy and detrital-zircon U-Pb geochronology date the formation to be Barremian to early Albian in age (128-107 Ma; Ziabrev et al. 2003; An et al., 2014; Huang et al. 2015).

The 1000-4100-m-thick Ngamring Formation comprises a series of fining-upward deep-marine turbidite sequences interbedded with thick channelized conglomerates, calcareous mudstones, and minor limestone facies (Figure 2; Einsele et al. 1994; Dürr, 1996; Wang et al., 1999; An et al. 2014). The Ngamring Formation conformably overlies the Chongdui Formation west of Ngamring County, but it is observed in fault contact with the Yarlung-Tsangpo ophiolite in the south and with the Kailas Formation or the Sangzugang Formation in the north at other localities (Wang et al. 2012). The maximum depositional age of the oldest strata in the Ngamring Formation, as constrained by the youngest detrital U-Pb zircon population, is between 84 and 107 Ma (Wu et al. 2010; An et al. 2014), in agreement with foraminiferal assemblages (Wan et al. 1998).

METHODS

Facies Analysis

Depositional environments of the basal Xigaze forearc basin were reconstructed from facies analysis of eight measured stratigraphic sections, totaling 4176 m in thickness (Figs. 3-7). Sections were measured at the centimeter scale using a Jacob's staff. Samples for sedimentary petrography, biostratigraphy, and geochronology were collected in the context of the measured sections. We use a lithofacies scheme modified after Miall (1978), Lowe (1982), Bouma (1962),

and Mutti (1992) for analysis of marine clastic facies (Table 1). The Dunham (1962) classification scheme was used for carbonate facies.

U-Pb Zircon Geochronology

Twenty-two medium- to coarse-grained sandstone samples were collected for detrital-zircon U-Pb geochronology to constrain the stratigraphic age of the investigated samples (e.g., Gehrels, 2014; Dickinson and Gehrels, 2009; Barbeau et al. 2009) and determine the provenance of basal forearc strata (Gehrels et al. 2008a, 2008b). Biostratigraphic age constraints were not used in this study owing to the poor preservation of microfossils. Following the methods of Gehrels et al., 2008 and Gehrels, 2014, zircon crystals were extracted by standard mineral-separation techniques, mounted in epoxy with fragments of the Sri Lanka (563.5 ± 3.2 Ma), FC-1 (1099.5 Ma), and R33 (419.3 Ma) standards, polished, and analyzed by a laser-ablation multicollector inductively coupled plasma mass spectrometer (LA-ICP-MS) at the University of Arizona LaserChron Center. Zircons were selected at random, and at least 100 grains per sample were analyzed (Appendix I).

Individual analyses carry analytical uncertainties of 1-2 % for $^{206}\text{Pb}/^{238}\text{U}$, $^{206}\text{Pb}/^{207}\text{Pb}$, and $^{206}\text{Pb}/^{204}\text{Pb}$ ages (2-sigma level). The $^{206}\text{Pb}/^{238}\text{U}$ and $^{206}\text{Pb}/^{207}\text{Pb}$ ages were used for grains less than and greater than 1000 Ma, respectively. Uncertainties are at the 1-sigma level and include only measurement errors. Individual zircons with > 20% discordance or > 5% reverse discordance were not considered. Detrital U-Pb ages are plotted on normalized age-probability diagrams, created using an Arizona LaserChron Center (www.alc.org) program developed using the routines of *IsoPlot* (Ludwig, 2008).

We also collected and separated three tuffaceous layers from the forearc stratigraphy and two igneous samples from the underlying ophiolite sequence for U-Pb geochronology. Pristine

195 euhedral zircons were picked, mounted in epoxy with the same standards used for detrital
196 analyses, and analyzed by LA-ICP-MS at the University of Arizona LaserChron Center. The
197 resulting mean ages are shown on Table 3 using the routines in *Isoplot* (Ludwig,
198 2008). Appendix I reports U-Pb analytical data that yielded isotopic data with acceptable
199 discordance, in-run fractionation, and precision.

200 **Sandstone Modal Analyses**

201 Forty-five samples of medium-grained sandstone were collected in the field and cut for
202 standard petrographic thin sections. Each thin section was stained for K-feldspar and Ca-
203 plagioclase, and 450 framework grains were counted according to the Gazzi-Dickinson point-
204 counting method (Gazzi, 1966; Ingersoll et al. 1984; Dickinson, 1985). Grain types and modal
205 parameters are listed and defined in Table 2, and recalculated data are available in Appendix II.
206 Standard ternary diagrams (Qm-F-Lt, Qt-F-L, Q-P-K, and Lm-Lv-Ls) and descriptive
207 petrographic classification after Garzanti (2016) are used for interpretation (Figure 10).

208 **Conglomerate Clast Counts**

209 Stratigraphic sections B (Figure 4) and H (Figure 7) provided the only opportunity for
210 provenance analysis of clast compositions. A minimum of 100 clasts were counted within several
211 100 cm x 100 cm grids following the method described by Graham et al. (1986). The results
212 from five clast counts are shown as pie charts in stratigraphic sections B and H (Figures 4 and 7).

214 **RESULTS**

215 **Sedimentology and Stratigraphy**

216 Eight stratigraphic sections were measured along a 65 km east-west-trending portion of
217 the Xigaze forearc basin, northeast of Lazi (Figures 1, 4-7). Lower Cretaceous sections A, B, C,

and D are in depositional or fault contact with the Yarlung-Tsangpo ophiolite and total 1623 m in thickness. Lower Cretaceous sections E, F, G, and H total 2553 m in thickness. Sections are described below from oldest to youngest. Sections young towards the axis of the syncline that consistently folds the Xigaze forearc-basin strata along strike (Figure 1).

Lower Cretaceous sections (~ 113-100 Ma)

Section A (29.199359°, 87.909876°)

Facies Description. This 153-m-thick section is in sharp contact with an underlying 3-m-thick sequence of thin-bedded red chert and siltstone, which in turn is underlain by a 1-m-thick felsic hypabyssal igneous rock and underlying ophiolitic mélange comprising serpentinite and gabbro clasts (Figures 3A, 4). Above the contact with the red chert is a 20-m-thick unit of alternating sandy limestone and marlstone beds with a tuffaceous interval at 10 m (A_10, 110.05 ± 1.3 Ma). Sandy limestone beds are 30-100 cm thick. Locally, in the sandy limestone interval, there are a series of 5-10-m-thick blue-gray wackestone and packstone blocks which extend laterally for ~ 250 m parallel to basin stratigraphy. Macrofossils include *Nerinea* gastropods, brain sponges, *scleractinean* corals, and rudist bivalves (Figure 3B).

At the 25 m level of the section, there is a facies change from predominantly sandy limestone to alternating coarse- to very coarse-grained, poorly sorted sandstone and clast-supported gravel-to-boulder conglomerate beds. Beds are 1-4 m thick. Conglomerate clasts include limestone, volcanic rocks (rhyolite), red chert and occasional shell material. Dominant lithofacies include massive sandstone (Sm), rare plane-parallel-laminated sandstones (Sh), clast-supported pebble-to-boulder conglomerate beds (Gcm), and pebble lenses in coarse-grained sandstone intervals. Beginning at the 85 m level, sandstone beds alternate with sandy limestone, siltstone, and carbonaceous marlstone for the remaining 68 m of the section. Sandstone beds are

mainly medium- to very coarse-grained and moderately to poorly sorted. Beds are 30-50 cm thick, decrease in grain size upward, and include massive to plane-parallel-laminated sandstone (Sm/Sh) and plane-parallel-laminated to rippled sandstone (Sh/Sr). Dewatering structures and shell fragments are also abundant in massive sandstone and sandy limestone beds.

Facies Interpretation. The sharp contact between sandy limestone lithofacies and red chert is depositional (Figure 3C). Based upon the lack of fossils, thickness of sandy limestone beds, and presence of marlstone, we interpret these facies to reflect pelagic sedimentation during sediment fallout (Mutti and Ricci Lucchi, 1972). In contrast, the association of macrofossils in local outcrops of blue-gray wackestone and packstone beds suggests deposition in a back-reef environment (Reading, 1986). The observed fossil assemblage is identical to that of the Aptian-Albian Sangzugang Formation, which represents a carbonate platform that formed along the southern margin of the Lhasa terrane during the Early Cretaceous (Liu et al. 1988). Therefore, we interpret these limestones as allochthonous blocks of Sangzugang Formation transported to the basin by debris flows or slumping during pelagic sedimentation (Flores, 1955; Neuendorf, et al., 2005). The overlying 60-m-thick sequence of massive sandstone and conglomerate supports this interpretation and is interpreted as the result of cohesive debris flows (Lowe, 1982; Shultz, 1984). Following Shultz (1984), lithofacies Gcm may reflect clast-rich or pseudoplastic debris flows derived from the continental slope. The association of lithofacies Sm, Sm/Sh, and Sh/Sr in the final 68 m are characteristic of deposition by low-density turbidity currents (Mutti, 1992). Lithofacies Sm, Sh, Sr, and Fsm are equivalent to Ta, Tb, Tc, and Td in the turbidite classification scheme of Bouma (1962) or F8-FD9 using Mutti (1992). The presence of shell material at the bases of fining-upward sandy-limestone beds also indicates a turbidity-current origin (Scholle, 1971; Eberli, 1988). Based on their grain size and bed thickness, we interpret

these low-density turbidite deposits to reflect deposition in the middle to inner parts of a turbidite fan (Mutti, 1977).

Section B (29.210252°, 87.949403°)

Facies Description. This 500-m-thick section comprises three distinct packages of beds that transition up section from interbedded fine-grained to coarse- to very coarse-grained sandstone facies to pebble-boulder conglomerates (Figure 4). The lower 312 m consists of beige to brown, fine-grained, moderately well-sorted sandstone and siltstone. Dominant lithofacies include massive siltstone (Fsl) and very fine-grained, massive to plane-parallel laminated sandstone (Sm/Sh). Beds are mostly 5-20 cm thick, but some fining-upward beds are 30-50 cm thick. Dewatering “flame” structures and convolute bedding are present at the bottoms of several fine-grained plane-parallel-laminated sandstone beds (Figure 3D).

The overlying 103 m is dominated by medium- to very-coarse grained, poorly sorted sandstone interbedded with siltstone. Dominant lithofacies include massive sandstone (Sm) and massive to plane-parallel-laminated sandstone (Sm/Sh), the latter of which often fines upward. Beds are 50-100 cm thick, tabular, and pinch out laterally. Flute clasts are present on the bases of a few massive sandstone beds (Figure 3e). The upper 85 m is dominated by poorly sorted, gravel to boulder matrix- (Gmm) and clast-supported conglomerates (Gcm). Clast-supported conglomerate beds are 0.5-2 m thick, lenticular, and often have erosional bases. Secondary lithofacies include 40-80-cm-thick massive sandstone (Sm) and plane-parallel-laminated sandstone (Sh) beds.

Facies Interpretation. The association of lithofacies Sm, Sh, Sm/Sh, and Fsl in the lower 312 m reflect deposition by sandy low- and high-density turbidity currents (Lowe, 1982; Mutti, 1992). Lithofacies Sm, with abundant dewatering structures, and lithofacies Sh, with intraclasts, are

interpreted to have formed by suspension settling and traction-carpet sedimentation respectively during flow of a high-density turbidity current and can be considered equivalent to S2 and S3, following the classification scheme of Lowe (1982). Following Bouma (1962), fine-grained lithofacies Sh and Fsl are Tc and Td, and reflect deposition during low-density turbidity flow (Talling et al., 2001; Mulder and Alexander, 2001; Talling et al., 2012) under lower-flow-regime conditions (Walker, 1965; Harms and Fahnestock, 1965; Ricci Lucchi, 1967; Allen, 1982) or waning dilute flow (Lowe, 1982). The thinness of sandstone beds is common on channel levees (Normark et al., 1980; Walker, 1975; Migeon et al. 2001; Talling et al. 2012) or distal basin plains (Walker, 1967; Mutti, 1977). Here, we interpret these as channel-wall to channel-levee deposits owing to their tabular geometry and relationship with overlying conglomerate beds (Figure 3F).

Up-section, from 312 m to 500 m, the increase in the abundance of massive sandstone (Sm) reflects a change in deposition to principally high-density turbidity flows where sediment concentration damps turbulence and prevents formation of bedforms (Lowe, 1988; Talling et al. 2012). Conglomerate lithofacies, namely Gcm, support this interpretation and are interpreted as a series of channel deposits in the middle- to inner-turbidite fan. Lithofacies Gmm is typical of deposition by plastic debris flows or hybrid flows (Talling et al. 2012). The increase in grain size and bed thickness indicates a more proximal setting compared to the basal 312 m (Sadler, 1982). Overall, this section documents the progradation of a submarine turbidite fan system southward.

Section C (29.206964°, 87.931374°)

Facies Description. This 680-m-thick section consists of an overall fining-upward sequence of siltstone and mudstone intercalated with thin sandstone and sandy limestone beds (Figure 5).

Dominant lithofacies include marlstone (Cm), massive, ungraded siltstone (Fsm), laminated siltstone (Fsl), massive sandstone (Sm), and massive to horizontally laminated sandstone (Sm/Sh). Sandstone beds in the first 486 m vary from fine- to coarse-grained and are 15-40 cm thick. Mudstone and siltstone beds range in thickness from 20 cm to 5 m. Several massive very coarse-grained sandstone beds pinch out laterally into marlstone. The upper 294 m is dominated by 5-20-m-thick marlstone interbedded with massive coarse- to very coarse-grained sandstone. Sandstone beds are 40 cm to 150 cm thick and generally fine upward. Subrounded to rounded, cobble-size limestone clasts are present in marlstone beds. Carbonate mounds and columnar carbonate deposits, 1-2 m tall and 0.5 m wide, are found within marlstone sequences (Figure 3G, inset).

Facies Interpretation. The association of lithofacies in the lower 396 m of Section C is characteristic of deposition from low-density turbidity currents (Mutti, 1992). Fine- to medium-grained sandstones with plane-parallel laminations likely represent dilute and expanded turbidity flows, indicative of channel levees (Mutti, 1977) or distal basin-plain deposits (Walker, 1967). Here, we interpret them as levee and interchannel deposits owing to lateral changes in sandstone beds which suggest a crevasse-splay origin (Mutti, 1977). Rounded limestone clasts, which lack macrofossils, may have been recycled from onshore fluvial deposits rather than being derived from debris flows sourced from the neighboring Sangzugang carbonate platform (Dürr, 1996). Marlstones likely reflect hemipelagic to pelagic sedimentation during periods of low sediment flux. The carbonate mounds and columnar structures are interpreted as gas hydrate and/or authigenic carbonate seeps, which are common features in many deep-marine basin-plain regions (Stakes et al. 1999; Conti and Fontana, 2011) and support the interpretation that these limestones were deposited in a basin-plain setting.

Section D (29.228768°, 88.117962°)

Facies Description. This 290-m-thick section comprises marlstone and siltstone interbedded with thin sandstone beds. Sandstone lithofacies are dominated by massive sandstone (Sm), with minor plane-parallel-laminated sandstone (Sh) and climbing-rippled sandstone (Sr) beds (Figure 5). Sandstone beds are medium- to coarse-grained and are normally graded or nongraded. Beds are 10-30 cm thick, with the exception of several 80-150-cm-thick massive sandstones with normal grading. Marlstone and siltstone beds are 2-10 m thick and have occasional lenses of massive sandstone that pinch out laterally.

Facies Interpretation. Fine- to medium-grained lithofacies Sm, Sh, and climbing Sr are characteristic of deposition by low-density turbidity currents (Mutti, 1992). Similarly to the upper part of Section C, these lithofacies are commonly found in channel levee to interchannel regions. The presence of climbing ripples suggests that these are proximal to the channel, rather than distal levee deposits. In contrast, coarse-grained massive sandstones can be thought of as Bouma A divisions, using the classification scheme of Bouma (1962) and are diagnostic of proximal channel-levee deposits whereby rapid suspension deposition by high-density turbidity currents causes a lack of bedload transport and development of structures (Lowe, 1982; Middleton, 1993).

Upper Cretaceous sections (~ 100-85 Ma)

Section E (29.226304°, 87.818874°)

Facies Description. This 253-m-thick section consists primarily of sandy limestone beds interbedded with siltstone and marlstone (Figure 6). In the first 60 m are several medium- to coarse-grained, poorly sorted, volcanoclastic sandstones, 30 cm to 2 m thick, that lack internal structure or grading. Sandy limestone beds are 3-40 cm thick. Marlstone and siltstone intervals

range from 1 m to 15 m thick and are often disrupted by a single sandy limestone bed. Several rounded, cobble-size clasts of limestone are present in marlstone intervals. Up-section, limestone beds become less abundant and thinner relative to marlstone intervals.

Facies Interpretation. Alternating sequences of thin, sheet-like limestone beds, which lack structure or grading, with thick marlstone intervals suggests deposition in a basin-plain setting (Mutti and Ricci Lucchi, 1972). Marlstones can be interpreted as Bouma (1962) division Te or Mutti and Ricci Lucchi (1972) Facies D₂ and likely reflect hemipelagic deposition by suspension settling. The massive medium- to coarse-grained volcanoclastic sandstone beds are likely the result of deposition by high-density turbidity flows (Lowe, 1982).

Section F (29.228399°, 87.756147°)

Facies Description. This 400-m-thick section consists of sandy limestone and sandstone beds intercalated with marlstone and siltstone (Figure 6). Sandstone lithofacies include very fine- to medium-grained massive sandstone (Sm) and plane-parallel-laminated sandstone (Sh) beds. Beds are mostly 5-15 cm thick, with the exception of a fining-upward package of coarse- to very-coarse-grained 20-30-cm-thick massive sandstone beds in the uppermost 43 m (Figure 6). Both sandy limestone and sandstone beds are tabular and laterally continuous until the outcrop becomes folded, at 400 m, owing to postdepositional deformation.

Facies Interpretation. Similarly to Section E, these sheet-like, thin-bedded, massive and nongraded sandstones, siltstones, and marlstones reflect deposition in a basin-plain setting (Mutti and Ricci Lucchi, 1972). Marlstones are likely hemipelagic and reflect deposition by suspension settling (Bouma, 1962). The fining-upward package of coarse-grained Sm lithofacies may reflect the appearance of an outer turbidite lobe or lobe fringe impinging on the basin plain (Mutti, 1977).

Section G (29.234266°, 87.854658°)

Facies Description. This 200-m-thick section comprises alternating sequences of sandstone and marlstone intervals (Figure 6). Dominant lithofacies include fine- to coarse-grained massive sandstone (Sm) and fine- to medium-grained plane-parallel-laminated sandstone (Sh). Sandstone beds are 10-60 cm thick, and several distinct fining- and thickening-upward packages are present. Mudstone intraclasts and subhorizontal burrows are present at the bases of several massive sandstone and marlstone beds, respectively.

Facies Interpretation. The association of the aforementioned lithofacies reflects deposition by high-density turbidity currents (Lowe, 1982). Sm, Sh, and marlstone lithofacies can be thought of as Ta and Tc-Te in the turbidite classification scheme of Bouma (1962) or F7-F8 using the scheme of Mutti (1992). The association of cyclic fining- and thickening-upward sandstone lithofacies is characteristic of sand lobes of an outer-turbidite-fan (Mutti and Ricci Lucchi, 1972; Walker, 1965; Mutti, 1977). Marlstone intervals reflect deposition by suspension settling, with bioturbated intervals reflecting periods of depositional hiatus.

Section H (29.251986°, 87.747083°)

Facies Description. This 1700-m-thick section comprises a succession of alternating fine- to medium-grained, moderately well-sorted sandstone, siltstone, and marlstone (Figure 7). In the first 745 m, dominant lithofacies include massive and laminated siltstone (Fsm and Fsl), fine- to medium-grained massive sandstone (Sm), and fine-grained rippled sandstone (Sr). Sandstone beds are 10-80 cm thick. In this sequence, there are several 3-5-m-thick beds of clast-supported, moderately sorted, pebble to boulder conglomerates (Gcm). From 745 to 1220 m, lithofacies consist primarily of medium-grained massive sandstone (Sm) and composite beds which fine

upward from massive sandstone to plane-parallel-laminated sandstone (Sm/Sh) (Figure 3H). Beds are mostly 20-100 cm thick. The upper 480 m consist primarily of 10-20-m-thick massive and laminated (Fsm and Fsl) siltstone intervals with sparse fine-grained, massive or rippled sandstone, and 5-10-cm-thick sandstone beds.

Facies Interpretation. Lithofacies in the lower 745 m are characteristic of deposition by low-density turbidity currents (Mutti, 1992). Lithofacies Sm, Sh, Sr, and Fsm are Ta, Tb, Tc, and Td within the turbidite classification scheme of Bouma (1962) or F8-F9 using the scheme of Mutti (1992). The fine to medium grain size, non-erosional bases, bed thickness, and amount of siltstone suggest deposition in the middle to outer regions of a turbidite fan (Mutti and Ricci Lucchi, 1972). Conglomerate lithofacies, namely Gcm, support this interpretation and are interpreted as a series of turbidite channels in middle-fan deposits. The lithofacies in the interval between 745 and 1220 m reflect a shift to deposition by high-density turbidity currents, likely in the middle to inner turbidite fan regions (Mutti, 1977; Talling et al. 2012). Marlstone and siltstone intervals in the upper 480 m reflect deposition by suspension settling. The lack of sandstone intervals in the upper 480 m may reflect an avulsion of the turbidite fan system or a period of depositional cessation by turbidity currents or retrogradation due to an increase in relative sea level, the latter of which is observed globally at ~ 84 Ma (Miller et al. 2005; Kominz et al. 2008).

Chronostratigraphy

At the locality of Section A, we interpreted a depositional contact between the ophiolite sequence and the overlying forearc stratigraphy based on a bedding-parallel contact and a lack of cataclasite, shear fabrics, and any other evidence of faulting (Figures 1C, 3A, 3B and 4). U-Pb

ages of zircons extracted from igneous rocks from the Xigaze ophiolite and a tuffaceous bed at the base of the forearc stratigraphy reveal the timing of initial forearc-basin sedimentation. Using the unmixing routine of *Isoplot*, ages are 110.5 ± 3.2 (A_7.6.14_3; n = 7) Ma and 111.7 ± 1.3 Ma (A_7.6.14_2; n = 26) for zircons derived from a gabbro in the ophiolitic mélange and overlying felsic hypabyssal intrusive, respectively (Table 3). The gabbro sample contained two older zircon grains that are not part of the youngest age population and are therefore excluded from the age calculation. A tuffaceous sample at the base of Section A yields an age of 110.1 ± 1.3 Ma (A_10; n = 44); three older and presumably inherited detrital ages were excluded from the age calculation.

Precise depositional ages of Cretaceous stratigraphic units in the Xigaze forearc are known from fossil assemblages (Liu et al. 1988), radioisotopic ages from tuffaceous layers (Ding et al., 2005; Orme et al. 2014; Dai et al., 2015), and maximum depositional ages deduced from detrital-zircon U-Pb age populations (Wu et al. 2010; Aitchison et al. 2011; Orme et al., 2014; An et al. 2014; Hu et al. 2015; Dai et al., 2015). Owing to the poor preservation of microfossils in our field area, we use the youngest distinct age population ($n \geq 3$) from U-Pb detrital geochronology to constrain a maximum depositional age (MDA) of specific stratigraphic horizons. The true depositional age (TDA) is constrained as younger than or equal to the calculated MDA. Thus, we use macrofossils and stratigraphic relationships to determine if the MDA of a sample approaches the TDA. This method of age control has proven to be robust in a forearc setting that is ideally located to capture first-cycle magmatic zircons with minimal lag time between magmatic crystallization and deposition (Orme et al. 2014; Sharman et al. 2014; Hu et al. 2015). In addition, there are no documented lulls in Gangdese magmatic-arc activity

during the time period of interest, suggesting that syndepositional magmatic zircons likely compose the youngest population of ages (Lee et al., 2009).

Table 3 presents five candidate MDAs for each sample following the methods outlined in Dickinson and Gehrels (2009) and Barbeau et al. (2009) (Appendix I). Our MDA calculations isolate the youngest distinct age population ($n \geq 5$) and perform a weighted mean age determination using *Isoplot*'s unmixing algorithm. Sections A-D yield Early Cretaceous maximum depositional ages of ca. 110-100 Ma, whereas Section C yielded a Late Cretaceous MDA of 89.5 ± 1.2 Ma. Sections E-H yielded Late Cretaceous maximum depositional ages of ca. 100-85 Ma (Table 3). Maximum depositional ages from 22 medium-grained sandstones collected from these sections yielded ages consistent with MDAs similar to those previously reported from the lower Xigaze Group (Wu et al. 2010; Aitchison et al. 2011; An et al. 2014). MDAs decrease up-section (Table 3) and are consistent with the three intervening tuffaceous beds (Figure 11; Appendix I; A_10, D_288, H_0). Therefore, we interpret these ages to be close to the true depositional age of the stratigraphic intervals sampled.

Provenance

Sandstone Petrography

Xigaze forearc-basin sandstones are composed of monocrystalline quartz (Qm), polycrystalline quartz (Qp), foliated polycrystalline quartz (Qpt), plagioclase feldspar (P), and volcanic lithic grains. No K-feldspar grains were found. Volcanic lithic grains, identified on the basis of variations in texture, include lathwork (Lvl), microlitic (Lvx), felsic (Lvfv), vitric (Lv), and mafic (Lv). Lathwork and microlitic types are the most abundant and consist of elongate (lath-shaped) plagioclase crystals set in a glassy, isotropic matrix. Vitric grains are glassy and have a pseudo-isotropic texture. Felsic and mafic grains comprise highly altered quartz and

plagioclase aggregates, with mafic grains rich in epidote. Sedimentary lithic grains include chert (C), carbonate (Lc), mudstone-shale (Lsh), and siltstone (S). Metamorphic lithic grains are limited to phyllite (Lph) and serpentine schist (Ss). Accessory minerals include epidote, chlorite, muscovite, biotite, and zircon.

Modal petrographic data are plotted on ternary diagrams (Figure 10). The sandstone compositions fall primarily into the feldspatho-lithic, quartzo-lithic, feldspatho-quartzose, and litho-quartzose descriptive petrographic classification fields (Garzanti, 2016). Samples from Early Cretaceous sections A, C, and D comprise feldspatho-lithic volcanicalstic sandstones with variable quartz and feldspar components. In contrast, upper Cretaceous sections E-H document a progressive increase in quartz relative to lithic-grain abundance. Lower Cretaceous Section B is anomalous relative to its neighboring sections. Samples from Section B comprise > 50% quartz grains and contain an abundance of sedimentary lithic grains, namely carbonate fragments.

Conglomerate Clast Counts

Clasts from the upper 83 m of Section B are dominated by volcanic (58-64%) and limestone (14-27%) clasts, with minor amounts of red chert, granite, and sandstone (Figure 4). Clasts are subrounded to rounded, suggesting fluvial transport prior to submarine debris-flow transport and deposition. Conglomerates from Section H are also dominated by volcanic clasts (74-84%) but show an increase in granite abundance (18%) and little or no chert or limestone clasts (Figure 7). Clasts are subangular, suggesting a more proximal source than Section B.

Detrital-Zircon U-Pb Ages

A total of 2164 zircon grains yielded U-Pb ages with acceptable precision and concordance for interpretation. Detrital-zircon U-Pb analyses for each section are plotted as cumulative distribution functions (CDFs) (Figure 11) and probability distribution plots

(Supporting Information I). Samples are dominated by grains < 200 Ma (Figure 11). Where present, grains > 400 Ma constitute less than 25% of the total population. These ages are distributed between 500-580 Ma, 1000-1300 Ma, and 2500-3000 Ma populations (Supporting Information I). Sections G and H contain the majority of grains > 400 Ma. All 25 samples have a major peak centered around 100 Ma. Sections C, G, and H also have a significant number of grains between 120 and 140 Ma (Figure 11). The youngest age populations from each sample define maximum depositional ages that young up-section (Table 3).

Provenance Interpretation

Modal sandstone compositions of Lazi region Xigaze forearc-basin strata are typical of sandstones derived from a magmatic arc (Dickinson et al. 1983; Marsaglia and Ingersoll, 1992; Garzanti et al. 2007). Samples exhibit an unroofing trend up-section, passing from feldspatho-lithic volcanoclastic sandstone to feldspatho-quartzose plutoniclastic sandstone (Garzanti, 2016). Paleocene-Eocene Xigaze forearc strata in the Saga region (Orme et al., 2014; Hu et al. 2015), located 275 km to the west, and in the Indus forearc basin (Garzanti and Van Haver, 1988), located 1200 km to the northwest, display a similar trend. Conglomerate compositions are consistent with this interpretation, as they are dominated by volcanic clasts. As previously identified, the U-Pb detrital zircon age spectra from this study supports the interpretation that the Gangdese magmatic arc and Lhasa terrane were the principal sources of Xigaze forearc detritus (Appendix I; Wu et al. 2010; An et al. 2014; Orme et al. 2014; Hu et al. 2015; Dai et al., 2015).

Paleontologic, petrographic, and detrital-zircon analyses of sections A and B suggest the presence of additional sources of forearc detritus aside from the Gangdese magmatic arc. Fossil assemblages from packstone beds near the base of Section A are consistent with those found in the Sangzugang Formation and are interpreted herein as allochthonous blocks that were shed into

the basin during the initial stages of deposition. Previously, remnants of the Sangzugang carbonate platform have been documented only east of the region studied here (Wu et al. 1977; Bassoullet et al. 1980; Einsele et al., 1994; Wang et al. 2012; An et al. 2014). In Section B, the presence of quartz-rich sandstones with abundant carbonate sedimentary lithic grains and rounded, pebble to boulder limestone clasts suggests a more distal carbonate source than the limestone found in Section A. Based on similar observations, Einsele et al. (1994) and Dürr (1996) proposed that the northern Lhasa terrane may have been a significant source of Xigaze forearc detritus, including minerals such as rutile and chromite, as well as the fluvial and carbonate facies preserved in the Cretaceous foreland basin. Their hypothesis predicts a U-Pb detrital-zircon age spectrum with abundant Paleozoic-Proterozoic zircons and a significant age peak between 120 and 140 Ma, which also characterizes the northern Lhasa terrane and Gangdese retroarc foreland basin (Leier et al. 2007; Gehrels et al. 2011). Our detrital-zircon age spectra support this hypothesis and indicate that river systems brought detritus from the northern Lhasa terrane to the forearc as early as Albian time (Figure 11).

DISCUSSION

Paleo-Depositional Environments and Basin Architecture

Lithofacies in Sections A-H reveal the presence of six primary depositional environments, including (1) continental slope, which are deposited by debris flows, (2) turbidite channel facies, (3) channel levee facies, reflecting deposition on the inner parts of a turbidite fan system, (4) middle and (5) outer-turbidite-fan deposit facies, and (6) basin-plain facies. To examine how these environments varied spatially and temporally, we quantify their abundance as a function of stratigraphic thickness in a measured section (Figure 8A) and as a percentage

relative to each other from 110 to 85 Ma (Figure 8B). From the oldest to youngest stratigraphic sections, the relative thickness of basin-plain, outer-turbidite-fan and middle-turbidite-fan facies increases; whereas inner-fan channel, levee deposits, and continental-slope facies decrease (Figure 1; Figure 8A). Figure 8B plots the relative abundance of each depositional environment at 5 Ma time intervals from 110 to 85 Ma. Initial sedimentation from ~ 110 to 100 Ma was dominated by slope and inner-turbidite-fan facies, which was followed by basin-plain deposition from ~ 100 to 93 Ma. These data suggest an overall deepening of the basin or migration of the main depocenter as a response to physiographic changes within the forearc during this time interval. At ~ 93 Ma, the abundance of basin-plain deposits decreased and middle- and outer-turbidite-fan deposits began to increase again. This suggests a cyclical nature of deposition in the basin whereby periods of clastic turbidite deposition were followed by mainly carbonate pelagic sedimentation. Einsele et al. (1994) first suggested that these cycles may represent distinct periods of filling controlled by changes in sea level. However, autocyclic processes also exert a strong control on sedimentary architecture, and further investigation is required to differentiate between these mechanisms. Interestingly, there is no correlation between sedimentation rate and the relative abundance of the various depositional environments. Undeformed accumulation rates, based upon stratigraphic thickness and maximum depositional ages at the top and bottom of each stratigraphic section, are between 0.02 and 0.31 mm/yr, with no discernible difference between environments. These first-order estimates of sedimentation rate are of the same magnitude as other estimates for the Xigaze forearc basin (Orme et al., 2014).

Using our interpreted depositional environments and chronostratigraphic age control from each section, we present a tentative correlation diagram of the basin from 110 to 85 Ma (Figure 9). The relative horizontal distance observed in the field is preserved (Figure 1).

Correlations are made based on the appearance and disappearance of specific depositional environments in each section, assuming no break in sedimentation and no erosional unconformities. This diagram also specifies the distance of each section from the outcrop trace of the north-vergent thrust fault located to the south of the basin (Figures 1, 9). This fault is likely responsible for the lack of exposed forearc strata older than ~ 95 Ma in the western part of our study area. Forearc strata of age similar to those preserved in the western part of the basin (Sections A-D) were likely eroded from the hanging wall of this fault, which was active between ~ 25 and 17 Ma (Quidelleur et al., 1997; Yin et al., 1999). This limits our ability to reconstruct the basin between 110 and 95 Ma along the length of our study area. However, from what we can observe, the spatial pattern of depositional environments observed near Lazi differs from the Xigaze forearc basin directly to the east. At the type localities near Xigaze, basin sedimentation during the same time interval (~ 110 to 85 Ma) was dominated by multiple major turbidite channels and turbidite fan lobes spaced 10 to 25 km apart (Figure 1A; Einsele et al. 1994; Dürr, 1996). In the study area, debris-flow and proximal turbidite channel and fan deposits are limited spatially to the eastern parts of the basin and temporally to the initial stages of deposition. This observation highlights the variability of depositional environments along strike and suggests that the Lazi region either lacked the extensive and coeval submarine canyon systems that characterized the basin to the east or only the more distal parts of the sedimentary environments are preserved.

Early Cretaceous Paleogeography

Our reconstruction of the paleogeography of the Xigaze forearc basin in the Lazi region from Albian (~ 110 Ma) through Santonian (~ 85 Ma) time is divided into four frames (Figure 12) and incorporates the spatio-temporal record of the Gangdese magmatic arc at this latitude

(compilation from Orme et al. 2014) and paleogeographic information from the central and northern Lhasa terrane (Leier et al. 2007; Dai et al., 2015). By ~ 110 Ma, forearc sedimentation began in the Lazi region atop red chert and ophiolitic *mélange*. Initial sedimentation was dominated by pelagic limestone with influx of continental material by submarine debris flows (Figure 12A). Slumping of the continental slope related to debris flows brought allochthonous blocks of the carbonate platform into the basin.

By 105 Ma, at the location of Sections A-C, submarine turbidite fans comprised the principal depositional systems (Figure 12B). As highlighted in Section B, the fan systems contained channels, channel levees, and interchannel facies and were progradational. Abundance of rounded limestone clasts, quartz-rich sandstones, and the U-Pb detrital-zircon age spectra (Figure 11) suggest derivation from the central to northern Lhasa terrane (Leier et al., 2007; Rao et al., 2015; Dai et al., 2015), suggesting that rivers with headwaters in this region transported sediment across the Gangdese magmatic arc into the forearc basin during Albian time. This interpretation is consistent with negative epsilon hafnium (ϵHf) values (-4.5 to 1.1) of zircons deposited in basal Chongdui unit of the Xigaze forearc basin (Dai et al., 2015), which are interpreted to reflect derivation from evolved Cretaceous volcanic rocks of the central Lhasa terrane. Alternatively, An et al. (2014) interpret that the lower Ngamring Formation (Albian-Cenomanian), east of Lazi, was derived from Early Cretaceous Gangdese magmatic arc and sedimentary cover along the southern Lhasa terrane based on positive ϵHf values in detrital zircons. These authors also document the presence of negative ϵHf values in the upper Ngamring and Padana Formations and interpret this as evidence for additional sources of sediment from the central Lhasa terrane during the Turonian-Campanian. No Hf isotopic data exist for detrital zircons in the Xigaze forearc near Lazi; therefore, we cannot directly address these competing

hypotheses. We favor the interpretation that the central and northern Lhasa terrane may have sourced some of the detritus preserved in Section B in the Lazi region of the Xigaze forearc basin, recording trans-arc sediment transport during Aptian time. This process has also been documented in other locations, including the Late Cretaceous—middle Eocene Great Valley forearc, California (Dumitru et al., 2013; Sharman et al. 2014), Cenozoic Japan forearc (e.g., Clift et al., 2013), and the modern Columbia River in the northwest United States.

Late Cretaceous Paleogeography

Between 100 and 92 Ma, there was a shift in depositional environments resulting in dominance of hemipelagic limestone deposition under the influence of occasional debris flows and turbidity flows (Figure 12C). Petrographic data suggest derivation from the Gangdese magmatic arc, which by 100 Ma was located along the southern margin of the Lhasa terrane. By 92 Ma, turbidite fan systems transported sediments derived from the continental margin into the basin via debris flows, low-density turbidity flows, and high-density turbidity flows (Figure 12D). Although petrographic results from strata deposited during this interval (Sections H and G) plot primarily in the feldspatho-quartzose ternary field, suggesting that derivation may have been from a more distal or quartz-rich source, the lack of limestone and subangularity of pebble to cobble sized clasts suggests that the basin was receiving detritus from exhumed portions of the Gangdese magmatic arc and proximal volcanic rocks. Similar changes in sediment provenance throughout the duration of forearc sedimentation have been observed in other forearc basins, including the Paleozoic to Cenozoic forearc basins of Japan (Aoiki et al. 2013).

Regional Tectonic Implications

The original depositional contact between forearc strata and the Yarlung-Tsangpo ophiolite is obscured along most of the IYSZ by Cenozoic thrust faults (Figures 1B, 9; Burg and Chen, 1984; Yin and Harrison, 2000). Its exposure in the Lazi region is an important constraint on the pre-collisional geometry of the southern Lhasa terrane. Our observations are consistent with paleolatitude determinations for the Yarlung-Tsangpo ophiolite and the Gangdese magmatic arc, which are indistinguishable by Early Cretaceous time (Huang et al. 2015). These data indicate that the Yarlung-Tsangpo ophiolite formed above a subduction zone in a forearc position, constituting the basement of the Xigaze forearc (Figure 13A; Wu, 1984; Hébert et al. 2012; Huang et al. 2015). Therefore, the southern margin of the Lhasa terrane consolidated into an arc-trench system by 110 Ma. There is no geological evidence for multiple arc-trench systems that accommodated closure of the Neo-Tethyan Ocean in the Lazi region.

U-Pb ages from a tuffaceous sandstone bed at the base of the forearc in Section A constrain the onset of forearc basin deposition to ~ 110 Ma. This constraint is younger than previous constraints for the beginning of forearc sedimentation, which are 116 Ma and 113 Ma to the west and east of our study area, respectively (Einsele et al. 1994; Wu et al. 2010; An et al. 2014; Dai et al., 2015; Huang et al. 2015). Variability along strike may reflect intra-basin topography or differences in the timing of exhumation in the footwall of east-west-striking detachment faults that dismembered south Tibetan ophiolites (Maffione et al. 2015). An alternative possibility arises from the fact that initial forearc sedimentation is contingent upon the presence of an outer forearc high, which acts as a barricade and traps sediment in the inner wedge environment (Dickinson, 1995). The discrepancy between the age of the Yarlung-Tsangpo ophiolite at other localities (120-128 Ma) and age of forearc basin sedimentation near

Lazi (110 Ma) suggests that a period of subduction erosion might have prevented the formation of an outer forearc high capable of trapping sediment before ~ 110 Ma.

Detailed sedimentologic and structural analysis of serpentinite- and sedimentary-matrix mélanges south of Lazi indicate formation in a subduction-accretion complex along the southern margin of the Lhasa terrane during Cretaceous time (Burg and Chen, 1984; Searle et al., 1987; Ding et al., 2005; Cai et al., 2012; An et al., 2015). Based on the transition from oceanic to terrigenous material in the Tangga mélange, Cai et al. (2012) constrain an Aptian maximum accretion age for the structurally highest and presumably first-formed portion of the accretionary prism. Structurally lower nappes were emplaced during the Late Cretaceous (Cai et al., 2012; An et al., 2015). As these data are consistent with the formation of an outer forearc high capable of trapping sediment by Early Cretaceous time, prior to the onset of forearc-basin sedimentation in the Lazi region, we favor a topographic or tectonic exhumation control on variations in the initiation age of forearc sedimentation. Future studies of the spatial and temporal relationships between forearc-basin deposition and development of the subduction-accretionary complex are crucial to improve our understanding the evolution of the Transhimalaya arc-trench system prior to India-Asia collision.

CONCLUSIONS

1. Eight new stratigraphic sections measured in the Xigaze forearc basin northeast of Lazi reveal > 4 km of turbidite fan and hemipelagic limestone and marlstone deposits. Locally, sedimentation began atop the Yarlung-Tsangpo ophiolite at 110 Ma and was dominated by debris flows. Between 105 and 90 Ma, the basin experienced cycles of high and low sediment

flux which deposited alternating successions of clastic turbidite inner-fan to outer-fan deposits and hemipelagic limestone and marlstone deposits.

2. Sandstone modal analyses, conglomerate clast compositions, and zircon U-Pb geochronology indicate that the provenance of forearc strata was mainly the Gangdese magmatic arc to the north (Figure 1). However, during the initial stages of forearc deposition, it is plausible that continental detritus was also derived from the central to northern Lhasa terrane.

3. The depositional contact observed between the Yarlung-Tsangpo ophiolite and Xigaze forearc deposits indicates that the ophiolite formed in a forearc position, constituting the basement of the Xigaze forearc. The age of initial forearc-basin sedimentation varies along strike, suggesting that forearc basement and/or accommodation space may have developed diachronously.

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Appendices

Appendix I: U-Pb detrital-zircon geochronologic data

Appendix II: Tabulated modal petrographic data

Tables

Table 1: Lithofacies used in this study

Table 2: Modal point-counting parameters

Table 3: Maximum depositional ages

Supporting Information

Supporting Information I: U-Pb relative age-probability density diagrams

Figure Captions

Figure 1: (A) Simplified geologic map of southern Tibet ca. 84° E and 91° E. The Xigaze forearc basin (green) extends for ~ 550 km along the Indus-Yarlung suture zone. Ophiolitic (dark purple) and sedimentary-matrix mélange (light purple) define the southern margin of the forearc. (B) Geologic map of the study area north of Lazi. (C) Region north of Jiwa where the forearc stratigraphy is in depositional contact with ophiolitic mélange.

Figure 2: Composite chronostratigraphic and lithostratigraphic diagram of Xigaze forearc strata adapted from Orme et al. (2014). Strata younger than ca. 85 Ma are not observed in the region of interest and therefore are not shown here.

Figure 3: Photographs of the Xigaze forearc outcrops. (A) Contact between Xigaze ophiolite mélange, red chert, and overlying basal sandy limestones. Field assistant for scale is standing on a white igneous rock. (B) Scleractinian coral in a packstone olistolith. (C) Looking west: depositional sequence, from south to north of ophiolitic mélange, red chert, pelagic limestones and olistostromes. White pole (bottom right) is 3 m in height. (D) Dewatering “flame” structures and plane-parallel lamination. (E) Flute clasts on the base of a massive sandstone bed. Local village boy is pointing up-section. (F) Turbidite channel levee (thinner beds) and channel deposits. Up-section is to the left, and the inner channel is towards the photographer. (G) Sandy limestones (brown) and hemipelagic marlstone intervals (white) overlain by channel facies (brown). Up-section is to the left. Inset: carbonate seep pipe and mound. (H) Middle- to outer-turbidite fan deposits. Jacob’s staff is on a Bouma Ta interval. Up-section is to the left.

Figure 4: Measured stratigraphic sections (in meters) A and B in lower Cretaceous strata. See Figure 1 for locations and Table 1 for explanation of lithofacies codes. Legend for all stratigraphic columns in Figures 4-7 is included.

Figure 5: Measured stratigraphic sections (in meters) C and D in lower and upper Cretaceous strata. See Figure 1 for locations, Figure 4 for explanation of colors and symbols, and Table 1 for lithofacies codes.

Figure 6: Measured stratigraphic sections (in meters) E, F, and G in upper Cretaceous strata. See Figure 1 for locations, Figure 4 for explanation of colors and symbols and Table 1 for lithofacies codes.

Figure 7: Measured stratigraphic sections (in meters) H in upper Cretaceous strata. See Figure 1 for locations, Figure 4 for explanation of colors and symbols, and Table 1 for lithofacies codes.

Figure 8: (A) Cumulative facies thickness for the six depositional environments interpreted from facies analysis described in the text. (B) Percentage of depositional environments as a function of time from 110 to 85 Ma.

Figure 9: Tentative correlation diagram for measured sections in the Xigaze forearc basin, northeast of Lazi. See Figure 1 for locations of sections. Relative horizontal position is maintained from Figure 1.

Figure 10: Ternary diagrams showing modal-framework grain compositions of sandstones from stratigraphic section in the Cretaceous strata of the Xigaze forearc basin, northeast of Lazi. Descriptive petrographic classification fields are after Garzanti (2016).

Figure 11: Cumulative distribution plot (CDP) for detrital zircons from this study from 0-4000 Ma and 75-200 Ma (inset). Samples from stratigraphic sections C, G, and H show an abundance of > 115 Ma zircons relative to other samples from the forearc basin. Tuffaceous samples have a characteristic “vertical” trend reflecting the presence of one age population. Sections A-H are in stratigraphic order using the same color scale as Figure 10.

Figure 12: Paleogeographic reconstruction of the Lazi region of the Xigaze forearc basin shown in Figure 1. (A) Reconstruction of the basin ca. 110 Ma during the initial stages of forearc sedimentation based on the interpreted depositional environments of Section A. (B) Time-transgressive reconstruction of the basin from 105 to 100 Ma based on the turbidity-flow deposits observed in Sections B, C, and D and their provenance results. (C) Reconstruction ca. 95 Ma based on the sediment-gravity-flow deposits observed in Sections C, E, and F. (D) Reconstruction ca. 90 Ma based upon interpreted depositional environments in Sections C, G, and H and their provenance results.

Figure 13: Schematic north-south cross sections showing the hypothetical evolution of the Xigaze forearc basin from Albian to Santonian time within the broader framework of subduction of Neotethyan lithosphere (black). The relative growth of the accretionary prism is based on Cai et al. (2012). Arrows denote the relative rate of convergence between Indian lithosphere and stable Asia (van Hinsbergen et al. 2011).

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1105

Section A

N 29.19936 E 87.90880
4148

LEGEND

Fsm	M	Marlstone
Fsl		Limestone
Sr		Sandy limestone
Sh		Wackestone
Sr/Sh		Packstone
Sm	Gcm	Covered
Sm/Sh	Gmm	Partially covered
●		U-Pb geochronology sample
■		Petrographic Sample

Other Symbols

Flute Casts	Shell material
Burrow/bioturbation	Intraclasts
Dewatering Structure	Gravel Lense
Clast-counts	limestone
volcanic	sandstone
granite	chert

Section B

N 29.21042 E 87.94588
4308 m

Section E
N 29.23094 E 87.82452
4050 m

Section F
N 29.2284 E 87.75615
4046 m

Section G
N 29.23834 E 87.85623
4100 m

Sections

▲	▲	▲	▲	■	■	■	■
H	G	F	E	D	C	B	A

D. ca. 90 Ma

C. ca. 95 Ma

B. ca. 105-100 Ma

A. ca. 110 Ma

Radiolarian chert	Limestone and marlstone
Ophiolitic melange	Turbidite deposits (Xigaze Forearc)
Carbonate platform	Terrestrial deposits (Takena Formation)

Table 1: Lithofacies used in this study (after Miall (1978), DeCelles (1991))

Lithofacies Code	Description	Interpretation
Fsm	Massive gray, black or beige siltstone; bioturbated	Suspension settling; Bouma (1962) Te
Fsl	Laminated gray or beige siltstone	Suspension settling during waning flow; Bouma (1962) Td-Te
Sm	Medium to very coarse massive sandstone	Traction carpet during high-density flow. Bouma (1962) Ta; Lowe (1982) S1; Mutti (1992) F7-F8
Sh	Fine- to medium-grained sandstone with plane-parallel laminations	Planar flow bed (upper and lower flow regime); unidirectional flow. Bouma (1962) Tb/Td; Mutti (1992) F9
Sr	Fine- to medium-grained sandstone with small ripples	Lower-flow-regime ripples; unidirectional flow. Bouma (1962) Tc; Mutti (1992) F9
Sm/Sh	Sandstone which grades from massive to plane-parallel laminated	Found only in turbidite deposits: Represents the transition from traction carpet to suspension settling deposition (Lowe (1982) S-facies); Bouma (1962) Ta/Tb; Mutti (1992) F9
Sr/Sh	Sandstone which grades from ripples to plane-parallel laminated	Found only in turbidite deposits: Represents the transition from a lower to upper flow regime; Bouma (1962) Tc/Td and Mutti (1992) F9
Gcm	Pebble to boulder conglomerate poorly sorted, clast supported	Pseudoplastic-debris-flow deposits (Shultz, 1984)
Gmm	Pebble to cobble conglomerate, poorly sorted, matrix supported	Plastic or cohesive debris-flow deposits (Shultz, 1984)
M	Marlstone (carbonaceous mud)	Dilute suspension settling above calcium carbonate compensation depth

Table 2: Modal Point-Counting Parameters

Symbol	Description
Grains > 60 um	
Qm	Monocrystalline quartz
Qp	Polycrystalline quartz
C	Chert
S	Siltstone
Qt	Total quartzose grains (Qm + Qp)
K	Potassium feldspar (perthite, microcline)
P	Plagioclase feldspar (Na and Ca varieties)
F	Total feldspar grains (K + P)
Fine grain lithics	
Lvv	Vitric volcanic grains
Lvx	Microlitic volcanic grains
Lvl	Lathwork volcanic grains
Lvf	Felsic volcanic grains (sericite + qtz + feldspar)
Lvm	Mafic volcanic grains (epidote + pyx + plag)
Lv	Total volcanic lithic grains (Lvv + Lvx + Lvl + Lvf + Lvm)
Lsh	Mudstone
Lph	Phyllite
Lsm	Mica schist
Lc	Carbonate lithic grains
Lm	Total metamorphic lithic grains (Lph + Lsm)
Ls	Total sedimentary lithic grains (Lsh + Lc + C + S)
Lt	Total lithic grains (Lv + Ls + Lm + Qp)
L	Total nonquartzose lithic grains (Lv + Ls + Lph + Lsm + Lc)
Accessory Minerals: epidote, chlorite, white mica, biotite, serpentine, zircon	

Table 3: Sandstone Maximum Depositional Ages (MDAs) calculated from zircon U-Pb geochronology

Sample	Method 1 Peak Age	Method 2 weighted mean	Method 3 unmixing algorithm	Method 4 youngest single grain	Method 5 TuffZirc (+/-)	Regional stratigraphic correlation
H_1700	85.92	85.5 ± 1.6	85.6 ± 1.7	83.7 ± 3.7	85.87 5.28/2.14	Ngamring
H_942	85.66	86.4 ± 1.8	86.37 ± 0.91	84.7 ± 2.6	85.49 2.62/0.76	Ngamring
H_626	90.36	86.0 ± 4.5	86.03 ± 1.3	84.6 ± 1.8	93.16 1.41/2.82	Ngamring
H_321	89.5	89.2 ± 1.1	88.5 ± 2.4	85.2 ± 3.8	89.46 2.38/1.59	Ngamring
H_223	90.36	89.0 ± 2.5	89.1 ± 2.6	86.6 ± 2.6	90.11 2.06/3.54	Ngamring
H_22	89.64	87.6 ± 3.0	91.1 ± 2.3	86.6 ± 2.3	88.04 5.0/1.43	Ngamring
H_0*	93.02	93.69 ± 0.32	92.49 ± 0.26	89.8 ± 1.3	92.89 0.22/0.34	Ngamring
G_197	86.24	85.6 ± 1.1	85.99 ± 1.7	83.9 ± 2.3	85.70 1.0/1.81	Ngamring
G_7	92.33	91.75 ± 0.83	91.87 ± 0.51	85.8 ± 3.6	91.66 0.89/1.94	Ngamring
F_38	96.41	95.8 ± 2.8	95.8 ± 3.4	94.5 ± 9.2	95.99 5.67/1.46	Ngamring
F_17	93.29	93.4 ± 3.7	95.4 ± 3.0	84.7 ± 3.1	94.56 1.67/5.78	Ngamring
E_200	93.4	92.4 ± 1.1	92.4 ± 3.6	90.6 ± 3.3	91.87 1.77/0.87	Ngamring
E_8	100.6	99.8 ± 1.2	99.8 ± 1.6	98.1 ± 2.5	99.59 0.57/0.33	Ngamring
D_288*	99.58	99.6 ± 2.3	99.6 ± 3.9	95.3 ± 8.9	100.37 1.49/3.05	Ngamring
D_35	102.2	104.8 ± 3.1	101.01 ± 9.5	97.2 ± 4.6	104.32 6.91/2.60	Ngamring
C_656	89	91.8 ± 2.9	89.5 ± 1.2	87.1 ± 1.7	89.84 2.82/2.69	Ngamring
C_207	99	99.78 ± 0.84	99.86 ± 0.72	95.1 ± 3.7	99.64 1.90/0.58	Ngamring
C_55	114	112.5 ± 1.9	112.5 ± 6.1	106.2 ± 5.7	112.41 1.66/1.6	Chongdui
B_422	109.2	105.6 ± 1.3	101.92 ± 7.1	93.4 ± 4.0	101.73 0.85/1.08	Ngamring
B_78	102.5	113.5 ± 1.9	102.04 ± 1.0	95.0 ± 1.7	115.20 1.99/0.53	Chongdui
B_5	102.75	109.18 ± 0.8	110.3 ± 0.78	88.3 ± 6.9	108.0 1.79/0.94	Chongdui
A_129	103.52	104.4 ± 1.1	102.78 ± 0.88	97.3 ± 2.0	105.39 1.43/1.62	Ngamring
A_96	106.27	106.82 ± 0.83	105.49 ± 0.79	98.1 ± 3.9	105.82 0.72/0.77	Ngamring
A_26	108.52	109.11 ± 0.84	107.26 ± 0.81	100.7 ± 2.5	108.75 0.79/1.04	Chongdui
A_10*	110.44	112.2 ± 1.3	110.05 ± 1.3	105.3 ± 1.9	112.44 0.88/1.27	Chongdui

Granitoid (2) and gabbro from ophiolitic mélange (3)

A_7.6.14_2	111.63	110.0 ± 1.2	111.75 ± 1.3	105.1 ± 2.0	110.87 1.39/0.61	Chongdui
A_7.6.14_3	110.68	110.0 ± 2.7	110.5 ± 3.2	100.6 ± 1.4	110.40 1.10/0.50	Ophiolitic mélange