History of subduction erosion and accretion recorded in the Yarlung Suture Zone, southern Tibet

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Abstract: The history of pre-Cretaceous subduction accretion and erosion along the Yarlung Suture Zone remains poorly constrained. We present new geological mapping along c. 200 km of the suture zone, 4881 detrital zircon U-Pb ages, and sandstone petrography for the subduction complex and Tethyan Himalayan strata. We provide the first documentation of the c. 158 Ma marine Xiazhia Formation, which contains volcanic clasts of intermediate to felsic volcanic rocks and ooids with both calcareous and volcanic cores. Based on our new data and synthesis of published data, we present a model in which the Zedong arc represents the southwards migration of the Gangdese arc onto a forearc ophiolite that was generated proximal to the southern Asian margin during Neotethyan slab rollback at 160–150 Ma. This contrasts with previous suggestions that the Zedong arc, Yarlung ophiolites and subduction complex rocks developed above an intra-oceanic subduction zone thousands of kilometres south of Asia. Although Gangdese arc magmatism began in the Middle Triassic, the only forearc units preserved are 160 Ma until collision between the Xigaze forearc basin and Tethyan Himalaya at c. 59 Ma. This suggests that almost all pre-Cretaceous forearc assemblages have been removed by subduction erosion at the trench.

Supplementary material: Zircon U-Pb geochronologic analyses and detail views of maps with stereonets are available at https://doi.org/10.6084/m9.figshare.c.4371404

The Yarlung Suture Zone (YSZ) separates the Mesozoic–Paleogene Asian (Gangdese) continental margin in the north from the Indian passive margin strata to the south (Fig. 1). Exposed within the suture zone, from north to south, are the Cretaceous–Eocene Xigaze forearc basin, Jurassic and Cretaceous ophiolites, and Cretaceous–Paleocene subduction complex rocks, all of which are fault bounded. The suture zone units are commonly interpreted to have formed along the active Asian continental margin during the northwards subduction of the Neotethys oceanic plate (e.g. Shackleton 1981; Tapponnier et al. 1981; Searle et al. 1987; Ensele et al. 1994; Dürr 1996; Dupuis et al. 2005b, 2006; Wu et al. 2010, 2014b; Cai et al. 2012; Dai et al. 2013, 2015; An et al. 2014, 2017; Zhang et al. 2014; Liu et al. 2015; Maffione et al. 2015a; W. Huang et al. 2015; Li et al. 2015b; Orme & Laskowski 2016; Xiong et al. 2016, 2017; Metcalf & Kapp 2017; H.-Q. Wang et al. 2017). An alternative interpretation suggests that the ophiolites and subduction complex rocks formed thousands of kilometres south of Asia in an intra-oceanic subduction zone, tectonically divorced from the Xigaze forearc basin and the Gangdese arc on the southern margin of Asia (e.g. Cao 1981; Aitchison et al. 2000, 2003, 2007a, b, 2011; Huot et al. 2002; McDermid et al. 2002; Malpas et al. 2003; Ziabrev et al. 2003, 2004; Abrajевичtch et al. 2005; Guilmette et al. 2009, 2012; Xia et al. 2011; Hebert et al. 2012; L. Wang et al. 2012; Bao et al. 2014; Chan et al. 2015; Baxter et al. 2016).

The subduction complex of the YSZ formed by the accretion of the Neotethyan ocean plate stratigraphy (Shackleton 1981; Tapponnier et al. 1981; Searle et al. 1987; Liu & Einsele 1996; Aitchison et al. 2000, 2003, 2011; Liu 2001; Matsuoka et al. 2001, 2002; Liu & Aitchison 2002; Aitchison & Davis 2004; Ziabrev et al. 2004; Dupuis et al. 2005a, b, 2006; Guilmette et al. 2008; Cai et al. 2012; Li et al. 2015c; An et al. 2017, 2018; Metcalf & Kapp 2017; H.-Q. Wang et al. 2017). The youngest fossils and sandstone blocks within the subduction complex are Late Paleocene (Tapponnier et al. 1981; Burg & Chen 1984; Burg et al. 1985; Liu & Aitchison 2002; An et al. 2017; Metcalf & Kapp 2017) but it remains uncertain when subduction complex rocks began to accrete, and the relative importance of subduction erosion and non-accretion. The basal Xigaze forearc basin was deposited during the Barremian (Marcoux et al. 1982; Girardeau et al. 1984; Wu 1984, 1986; Li & Wu 1985; Einsele et al. 1994; Dürr 1996; Ziabrev et al. 2003; Wu et al. 2010; Aitchison et al. 2011; Dai et al. 2015; W. Huang et al. 2015; Orme & Laskowski 2016; Wang et al. 2017a) and an old forearc basin deposits are known. Gangdese arc magmatism initiated by Middle Triassic or Early Jurassic time (e.g. Chu et al. 2006; Ji et al. 2009b; Zhu et al. 2011; Guo et al. 2013; Kang et al. 2014; Meng et al.)
2016; Ma et al. 2017b; R. Wang et al. 2017), however, leaving open the possibility that the YSZ may host remnants of older Asian margin subduction complex and forearc assemblages.

We first review the literature on YSZ units, and disambiguate the variety of names, descriptions and interpretations of subduction complex units that have been presented. We then present new geological mapping, sandstone petrographical studies, and 4881 new U–Pb detrital zircon ages from the subduction complex and Tethyan Himalayan strata. For sediments deposited near an active arc, the detrital zircon U–Pb maximum depositional age (MDA) is indistinguishable from the biostratigraphic age (Dickinson & Gehrels 2009; Painter et al. 2014). Since the subduction complex is structurally complex, many analyses are necessary to characterize its age and provenance. Finally, we present evidence for Jurassic forearc strata, the Xiazha Formation, along the northern edge of the subduction complex. Based on structural relationships, the timing of arc magmatism and ophiolite generation, and detrital zircon U–Pb ages, we interpret that the Xiazha Formation was deposited in a forearc setting proximal to the Asian continental margin and that much of the pre-Cretaceous forearc was removed by subduction erosion.

Geological background

The YSZ separates the Mesozoic–Paleogene Gangdese magmatic arc and overlying Oligocene–Miocene Kailas Formation in the north from Tethyan Himalayan strata that were deposited on the northern continental margin of India to the south. Assemblages in the YSZ include, from north to south, the
Cretaceous–Eocene Xigaze forearc basin, the ophiolite belt, the Cenozoic Luqu Conglomerate and accreted ocean plate stratigraphy (OPS).

Arc magmatism and the Kailas Formation

The Gangdese magmatic arc developed during northwards subduction of the Neotethys oceanic lithosphere beneath the southern margin of Asia. Subduction-related magmatism is as old as Middle Triassic–Early Jurassic (e.g. Chu et al. 2006; Ji et al. 2009b; Zhu et al. 2011; Guo et al. 2013; Kang et al. 2014; Meng et al. 2016; Ma et al. 2017b; R. Wang et al. 2017) (Fig. 2). The Zedong arc, which outcrops only near Zedong (Fig. 1), is composed of an overturned section of 160–155 Ma volcanic rocks, volcanioclastic chert and breccia, with intrusions of granitoids and hornblende-rich cumulates (Aitchison et al. 2000, 2007b; McDermid et al. 2002; L. Wang et al. 2012; Zhang et al. 2014), and its basement is composed of pillow basalts overlain by Bathonian–Lower Callovian red ribbon-bedded radiolarian chert (McDermid et al. 2002; Aitchison et al. 2007b) (Fig. 2). Some Zedong auto- clastic basaltic volcanic breccias are depletes in high field strength elements (HFSEs) and large ion lithophile elements (LILs), and plot in the shoshonitic field (Aitchison et al. 2007b; L. Wang et al. 2012), leading to one interpretation that the Zedong arc developed above an intra-oceanic subduction zone in the Tethys Ocean (e.g. Cao 1981; Aitchison et al. 2000, 2003, 2007a, b, 2011; Huot et al. 2002; McDermid et al. 2002; Malpas et al. 2003; Zia brev et al. 2003, 2004; Abrajevitch et al. 2005; Guilmette et al. 2009, 2012; Xia et al. 2011; Hebert et al. 2012; L. Wang et al. 2012; Bao et al. 2014; Chan et al. 2015; Baxter et al. 2016). However, Zedong arc rocks are altered (Aitchison et al. 2007b; Zhang et al. 2014), and alteration-immobile elements show a calc-alkaline rather than a shosho nitic signature (Zhang et al. 2014). The geochemistry of Zedong and Yeba volcanic rocks within the Gangdese arc to the north – interpreted as continental arc or back-arc (Zhu et al. 2008a, b; F. Huang et al. 2015; Ma et al. 2017a; Wei et al. 2017) – are indistinguishable, leading to the alternate interpretation that the Zedong arc is part of the juvenile Jurassic Gangdese arc (Zhang et al. 2014; Hu et al. 2016). The Oligocene–Miocene Kailas Formation (Heim & Gansser 1939; Aitchison et al. 2002; DeCelles et al. 2011, 2016; R. Leary et al. 2016) lies in a buttress unconformity on the southern Gangdese arc and everywhere along strike is bound to the south by the northernmost strand of the north-verging, southdipping, c. 19–13 Ma Great Counter Thrust (GCT) system, also known as the Renbu–Zedong Thrust (RZT) east of Xigaze (Heim & Gansser 1939; Burg et al. 1987; Ratschbacher et al. 1994; Yin et al. 1994, 1999; Quidelleur et al. 1997; Yin & Harrison 2000; Murphy & Yin 2003).

Xigaze forearc basin

The Early Cretaceous–Eocene Xigaze forearc basin is exposed in an east–west-trending synclinorium that plunges to the west (i.e. exposures generally young to the west). The forearc consists of the Bar remian–Campanian Xigaze Group (the Chongdoi/ Chongdai and Ngamring formations) and the Campanian–Ypresian Tso-Jiangding Group (the Padana, Qubeiya, Quxia and Jialazi formations) (Fig. 2). The Chongdoi Formation, which is depositional on the ophiolite massifs and ophiolitic mélanges, is the subject of much debate. The lower member consists of Late Barremian–Early Cenomanian radiolarian chert and mudstone with tuffaceous layers, and the upper member is composed of turbidite strata (Mar coux et al. 1982; Girardeau et al. 1984; Wu 1984, 1986; Li & Wu 1985; Einsele et al. 1994; Dürr 1996; Ziabrev et al. 2003; Wu et al. 2010; Aitchison et al. 2011; Dai et al. 2015; W. Huang et al. 2015; Orme & Laskowski 2016; Wang et al. 2017a). The onset of deposition and transition to turbidite sedimentation was diachronous, and could be explained by the topography within the basin and the diachronous uplift of the source region (Orme & Laskowski 2016; Wang et al. 2017a, b). The contact between the upper member of the Chongdoi Formation and the overlying Ngamring Formation is only locally exposed. The relatively poor age control on the Ngamring Formation at the time (Dürr 1996) and a palaeomagnetic study suggesting that the ophiolite developed at equatorial latitudes (Abrajevitch et al. 2005) were used to argue that the Chongdoi Formation is the forearc of an intra-oceanic arc (Cao 1981; Aitchison et al. 2000, 2003, 2011; Ziabrev et al. 2003). However, other studies document an Early Cretaceous palaeoaltitude within error of the southern margin of Asia (W. Huang et al. 2015), chronological and sedimentological similarities between the Chongdoi Formation and the Ngamring Formation (Einsele et al. 1994; Dürr 1996; Wu et al. 2010; Aitchison et al. 2011; An et al. 2014; Orme & Laskowski 2016; Wang et al. 2017a), olistostromal blocks of the Sangzugang Formation platform carbonates from the Lhasa terrane within the Chongdoi Formation (Einsele et al. 1994; Dürr 1996; Orme & Laskowski 2016), pre-Mesozoic zircons (Wu et al. 2010; W. Huang et al. 2015; Orme & Laskowski 2016; Wang et al. 2017a), and Cretaceous and older zircons with negative εHf values (Wu et al. 2010; Dai et al. 2015; Wang et al. 2017a), all consistent with deposition proximal to the Lhasa terrane. Taken together, the available geological data suggest that the Chongdoi Formation represents the basal strata to a forearc that formed along the southern
Fig. 2. Compilation of ages of units and events in the central Yarlung Suture Zone. ¹Gangdese arc compilation (Orme et al. 2015), Zedong arc magmatism (McDermid et al. 2002; Aitchison et al. 2007b; L. Wang et al. 2012; Zhang et al. 2014) and Zedong arc radiolarian basement (Aitchison et al. 2007b). ²Xigaze forearc basin biostratigraphic ages (Marcoux et al. 1982; Girardeau et al. 1984; Wu 1984, 1986; Li & Wu 1985; Liu et al. 1988; Einsele et al. 1994; Dürr 1996; Ziaibrev et al. 2003; Ding et al. 2005; C. Wang et al. 2012; Hu et al. 2015c) and
margin of the Lhasa terrane (Marcoux et al. 1982; Girardeau et al. 1984; Einsele et al. 1994; Dürr 1996; Wu et al. 2010; An et al. 2014; Dai et al. 2015; W. Huang et al. 2015; Orme & Laskowski 2016; Wang et al. 2017a). The similarity between the turbidites of the upper member of the Chongdoi Formation and those of the Ngamring Formation led An et al. (2014) to reclassify them both as the Ngamring Formation, limiting the Chongdoi Formation to the lower chert and tuffaceous mudstone.

The Aptian–Albian Sangzunang Formation (Bassoulet et al. 1980; Cherchi & Schroeder 1980; Liu et al. 1988; Einsele et al. 1994; An et al. 2014) is a platform carbonate found in fault contact with the Ngamring Formation to the south and the Kailas Conglomerate to the north (C. Wang et al. 2012; An et al. 2014). Blocks of the Sangzunang Formation are found in the Chongdoi and Ngamring formations, recording mass wasting from the continental margin into the newly formed supra-ophiolite forearc basin (Einsele et al. 1994; Dürr 1996; Orme & Laskowski 2016). The lower Cretaceous–Santonian Ngamring Formation is a series of turbiditic megasequences (Liu et al. 1988; Einsele et al. 1994; Dürr 1996; Ding et al. 2005; C. Wang et al. 2012; An et al. 2014; Dai et al. 2015; Orme et al. 2015; Orme & Laskowski 2016) deposited in middle–outer turbidite–fan and basin-plain settings (C. Wang et al. 2012; Orme & Laskowski 2016) (Fig. 2). The sediment was dominantly derived from the Gangdese magmatic arc with some metamorphic clasts (Dürr 1996), pre-Mesozoic zircon grains (Wu et al. 2010; Aitchison et al. 2011; An et al. 2014; Orme & Laskowski 2016), and Cretaceous and older zircons with negative εHf values (Wu et al. 2010; An et al. 2014) indicative of additional sources in the Lhasa terrane north of the Gangdese arc.

The Santonian–Early Campanian Padana Formation of the Tso-Jiangding Group is composed of sandstone and shale (Wu et al. 2010; An et al. 2014; Hu et al. 2015c) that were deposited in a shelfal–deltaic environment in the west and a fluvial environment in the upper portion in the east (An et al. 2014; Hu et al. 2015c; Orme et al. 2015) (Fig. 2). The Padana Formation represents filling of the forearc basin to near sea level (C. Wang et al. 2012; An et al. 2014; Hu et al. 2015c). The Late Campanian–Late Maastrichtian Quebeya Formation is composed of shallow-marine limestone with interbeds of sandstone, shale and mudstone (Liu et al. 1988; Ding et al. 2005; Wu et al. 2010; C. Wang et al. 2012; Hu et al. 2015c) (Fig. 2). The Quebeya Formation is separated from the overlying Quxia Formation by an angular unconformity (Ding et al. 2005), paraconformity (C. Wang et al. 2012), or a conformable, gradational contact (Orme et al. 2015). Biostratigraphic age control in the conglomeratic Quxia Formation is poor (Ding et al. 2005; C. Wang et al. 2012; Hu et al. 2015c) but, based on the ages of the underlying Quebeya Formation and the overlying Jialazi Formation, and detrital zircon ages within the Quxia Formation (Hu et al. 2015c), the Quxia Formation is constrained to be Danian–Selandian (Fig. 2). The Selandian–Ypresian Jialazi Formation is composed of shallow-marine sandstone and conglomerate interbeds in the lower part, and massive, rippled, planar cross-stratified and trough cross-stratified sandstone with palaeosol interbeds in the upper part (Liu et al. 1988; Wan et al. 2001; Ding et al. 2005; Aitchison et al. 2011; Hu et al. 2015c; Orme et al. 2015) (Fig. 2). Deposition of the shallow-marine to fluvial Jialazi Formation is interpreted to be synCollisional with the Tethyan Himalaya (Ding et al. 2005; Hu et al. 2015c; Orme et al. 2015), and Orme et al. (2015) suggested that the forearc basin began uplifting and eroding at c. 53 Ma as collision continued (Fig. 2). Detrital zircon ages in the Tso-Jiangding Group indicate derivation from the Gangdese arc and farther north in the Lhasa terrane, although some samples record an exclusive arc provenance (Wu et al. 2010; Aitchison et al. 2011; An et al. 2014; W. Huang et al. 2015; Orme & Laskowski 2016).

Fig. 2. Continued. U–Pb detrital zircon ages (Wu et al. 2010; Aitchison et al. 2011; An et al. 2014; Dai et al. 2015; Hu et al. 2015c; W. Huang et al. 2015; Orme et al. 2015; Orme & Laskowski 2016; Wang et al. 2017a), see the text for depositional environment interpretation references. Ophiolite zircon U–Pb ages (Göpel et al. 1984; McDermid et al. 2002; Malpas et al. 2003; Miller et al. 2003; Geng et al. 2006; Wang et al. 2006; Wei et al. 2006; Chan et al. 2007, 2015; Guilmette et al. 2007, 2009; Li et al. 2008; Xia et al. 2008, 2011; Zhu et al. 2009; Liu et al. 2011, 2016; Xiong et al. 2011; Dai et al. 2012, 2013; Bao et al. 2013) and amphibolite metamorphic sole ages (Guilmette et al. 2009, 2012). Xiazhua Formation, this study. Sample locations are shown in Figure 3. Baimang terrane radiolarian biostratigraphy: northern tract or Xialu chert (Girardeau et al. 1984; Wu 1993; Matsuoka et al. 2001, 2002; Ziabrev et al. 2004), southern tract (Ziabrev et al. 2004). Rongmawa Formation and Luogangcuo Formation U–Pb zircon MDA. (Left column) Rongmawa Formation this study; and (right column) Rongmawa Formation (Cai et al. 2012; Wang et al. 2018) and Luogangcuo Formation (An et al. 2018). Sample locations are shown in Figure 3. Siliciclastic-matrix mélange Asian-affinity sandstone blocks, U–Pb zircon MDA. (Left column) This study and (right column) Cai et al. (2012), An et al. (2017), H.-Q. Wang et al. (2017, 2018). Youngest ocean plate stratigraphy fossils from blocks and matrix (Tapponnier et al. 1981; Burg & Chen 1984; Burg et al. 1985; Liu & Aitchison 2002). Sample locations are shown in Figure 3. India–Eurasia convergence rate at the eastern syntax (dashed) and the western syntax (solid) from van Hinsbergen et al. (2011) in red and Gibbons et al. (2015) in black. Tethyan Himalayan strata ages and events.
The sandstone petrography of the forearc basin succession records progressive growth and unroofing of the Gangdese magmatic arc, and more minor but variable input from sources north of the arc (Einsle et al. 1994; Dürr 1996; An et al. 2014; Hu et al. 2015c; Orme et al. 2015; Orme & Laskowski 2016). Thermochronology of the forearc basin near Xigaze (Fig. 1) shows a northwards-younging trend in cooling ages, suggesting that the depocentre migrated northwards over time as the southern forearc was uplifted, beginning at 89–80 Ma (Li et al. 2017). However, thermochronology data from the forearc basin near Lazi and the Lopu Range indicate onset of cooling after c. 20 Ma (Orme 2017), so the Cretaceous uplift near Xigaze may be a local effect.

**Ophiolite belt and the Liuqu Conglomerate**

Ophiolite massifs and ophiolitic serpentine-matrix mélange (the Dazhuqu terrane of Aitchison et al. 2000) are present along much of the YSZ. Crystallization ages of the ophiolite rocks cluster around 160–150 and 132–120 Ma (Göpel et al. 1984; McDermid et al. 2002; Malpas et al. 2003; Miller et al. 2003; Geng et al. 2006; Wang et al. 2006; Wei et al. 2006; Chan et al. 2007, 2015; Guilmette et al. 2007, 2009; Li et al. 2008; Xia et al. 2008, 2011; Zhu et al. 2009; Liu et al. 2011, 2016; Xiong et al. 2011; Dai et al. 2012, 2013; Bao et al. 2013; Zhang et al. 2016) (Fig. 2). Mixed ages and geochemical signatures of ophiolitic units (sometimes within the same massif) have led to many different models of ophiolite generation, evolution and obduction/subduction. It has long been recognized that ophiolitic mantle rocks show a mid-ocean ridge (MOR) signature (often harzburgite) overprinted by a suprasubduction zone (SSZ) signature, with younger cross-cutting and subcreted units (herzolite, dolerite dikes, basalt, etc.) showing more of a SSZ signature (Huot et al. 2002; Dupuis et al. 2005a; Dai et al. 2011, 2012, 2013; Bao et al. 2013, 2014; Chan et al. 2015; Liu et al. 2015; Xiong et al. 2016, 2017). The ophiolitic rocks may have formed in the forearc or back-arc of an intra-oceanic arc (Aitchison et al. 2000, 2003; Huot et al. 2002; Malpas et al. 2003; Ziaibrev et al. 2003; Abrajevitch et al. 2005; Guilmette et al. 2009, 2012; Xia et al. 2011; Hebert et al. 2012; Bao et al. 2014; Chan et al. 2015; Baxter et al. 2016) or the forearc of the southern Lhasa terrane (Dai et al. 2013; An et al. 2014; Wu et al. 2014b; W. Huang et al. 2015; Liu et al. 2015; Maffione et al. 2015a; Xiong et al. 2016, 2017). The stratigraphic relationships and Lhasa terrane provenance of the Chongdoi Formation discussed above suggest that at least the Early Cretaceous ophiolite was generated in the forearc of the southern Lhasa terrane. In places, the Chongdoi Formation and extrusive basalts are depositional on mantle rocks, suggesting the development of detachment faulting during forearc hyperextension (Liu et al. 2014, 2015; Maffione et al. 2015b). Some studies suggest that the ophiolite entrained Asian sublithospheric mantle (Wu et al. 2014b; Liu et al. 2015; Maffione et al. 2015b; Griffin et al. 2016) and high-pressure phases from as deep as the mantle transition zone (e.g. Moores et al. 2000; McGowan et al. 2015). Numerical modelling by Butler & Beaumont (2017) shows that increased decoupling at the trench or slab breakoff and reinitiation of subduction could result in hyperextension of the Lhasa terrane forearc and exhumation of Asian sublithospheric mantle from the mantle transition zone. Hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages of dispersed amphibolite blocks within the ophiolitic serpentinite-matrix mélange are 132–127 Ma near Saga (Guilmette et al. 2012), 128–123 Ma near Xigaze (Guilmette et al. 2009) and 90–80 Ma near Dazhuqu (Malpas et al. 2003) (Figs 1 & 2). These blocks are interpreted to represent remnants of a dynamothermal or metamorphic sole which records inversion of the spreading ridge (Guilmette et al. 2008, 2009, 2012; Zhang et al. 2016) or obduction onto the Indian margin (Malpas et al. 2003).

The Liuqu Conglomerate is synkinematic with the GCT and is variably exposed in the suture zone along the ophiolite belt. The age of the Liuqu Conglomerate is poorly constrained, and published estimates are Paleocene–Eocene (Aitchison et al. 2000, 2011; Davis et al. 2002; Fang et al. 2006; Wang et al. 2010; Ding et al. 2017; Xu et al. 2018) and Oligo-Miocene age (Wei et al. 2011; Li et al. 2015a; R.J. Leary et al. 2016, 2017, 2018). Liuqu clasts are predominately derived from the Xigaze forearc basin strata, ophiolite, subduction complex and Tethyan Himalayan strata to the south (Davis et al. 2002; Wang et al. 2010; Li et al. 2015a; R.J. Leary et al. 2016), and paleocurrent indicators show northwards sediment transport (R.J. Leary et al. 2016).

**Subduction complex**

Accreted ocean plate stratigraphy and deformed Tethyan Himalayan strata are exposed south of the ophiolite, in the footwall of the north-dipping Yarlung Zangbo Mantle Thrust (YZMT). A variety of classifications, names and interpretations have been used for the units in this region (Table 1). Most workers in the subduction complex are aware of this variety, but the literature is replete with apparent contradictions in nomenclature and rock descriptions.

**Bainang terrane.** A chert-rich unit exposed locally south of the ophiolite is known variously as Jurassic–Cretaceous radiolarites, infra-ophiolitic thrust sheets of radiolarites, the Xialu chert, the Bainang
Siliciclastic-matrix mélange. South of the locally exposed Bainang terrane is a siliciclastic-matrix mélange of ocean plate stratigraphy and Tethyan Himalayan stratigraphy, known variously as sedimentary matrix mélange, flysch with exotic blocks, the Yamdrok mélange, tectonic mélange, Cretaceous chaotic sediments, the Pomunong mélange, the Renbu mélange, the Xiukang mélange or siliciclastic-matrix mélange (Table 1). Descriptions of the mélange vary because some interpretations include deformed Tethyan Himalayan strata to the south or the chert-rich unit to the north. Most workers interpret the mélange as a subduction complex (Table 1). Aitchison et al. (2000) introduced the Bainang terrane and interpreted it as a subduction complex, at the same time reinterpreting the Yamdrok mélange as deformed distal Indian passive margin and ‘more distal sediments’. This reinterpretation of the Yamdrok mélange was apparently based on the work of Liu (2001) south of Renbu and west of Gyangze (Fig. 1), well within an area that has been mapped as Jurassic–Eocene Tethyan Himalayan strata (Girardeau et al. 1984; Wu 1993; Matsuoka et al. 2001, 2002). Radiolarian ages range from Aalenian to Aptian (Girardeau et al. 1984; Wu 1993; Matsuoka et al. 2001, 2002). Radiolarian ages from Rhaetian until Aptian time when locally tuffaceous siliceous mudstone capped the section (Ziabrev et al. 2004) mapped the most extensive exposure of the chert-rich unit near Bainang, c. 40 km SE of Xigaze (Fig. 1). The northern tract is very similar to the Xialu section, with radiolarian chert dominating from Rhaetian until Aptian time when locally tuffaceous siliceous mudstone capped the section (Ziabrev et al. 2004) (Fig. 2). The southern tract is structurally lower than the northern tract, and to the north is composed of locally calcareous mudstone, calciturbidites with northwards palaeocurrent directions, tuffaceous chert, and mudstone, capped by red radiolarian chert (Ziabrev et al. 2004). Southwards, this stratigraphy becomes more deformed until it is dismembered within yellowish-grey calcareous shale (Ziabrev et al. 2004). Radiolarians extracted from the northern part of the southern tract record hemipelagic conditions from upper Norian/Rhaetian to Callovian time followed by pelagic conditions until the top of the preserved section in the Tithonian (Ziabrev et al. 2004) (Fig. 2). The northern tract of the Bainang terrane and Xialu chert are interpreted as ocean plate stratigraphy which entered a trench environment during Aptian time and was accreted to the Dazhuku terrane or Asia shortly after (Girardeau et al. 1984; Wu 1993; Matsuoka et al. 2001, 2002; Ziabrev et al. 2004; Chu et al. 2006; Ji et al. 2009b; Zhu et al. 2011; Guo et al. 2013; Kang et al. 2014; Meng et al. 2016). The southern tract of the Bainang terrane, exposed only near Bainang, is interpreted as ocean plate stratigraphy which developed close to the northern margin of India, underwent thermal subsidence related to Triassic rifting and was accreted at the trench shortly before India–arc collision during the Paleocene (Ziabrev et al. 2004).
| Authors | Year | Unit Name | Interpretation | Stage 1 (Ma) | Stage 2 (Ma) | Location |
|---------|------|-----------|----------------|-------------|-------------|----------|
| This study | | Xiazha Formation | Accreted Jurassic forearc | Siliciclastic-matrix mélange Subduction complex | Rongmawa and Luogangcuo formations Trench fill | Tethyan Himalaya Indian passive margin |
| Wang et al. (2018) | | NM Bainang terrane Subduction complex | Stage 1 (≤149–134 Ma) Subduction complex | Rongmawa Formation Trench fill | Tethyan Himalaya Indian passive margin |
| An et al. (2018) | | Xiukang mélange Subduction complex | | | | Tethyan Himalaya Indian passive margin |
| Metcalf & Kapp (2017) | | Bainang NI | Siliciclastic-matrix mélange Subduction complex | NM | Tethyan Himalaya Indian passive margin |
| An et al. (2017) | | Xiukang mélange Subduction complex | | NM | Tethyan Himalaya Indian passive margin |
| H.-Q. Wang et al. (2017) | | Sedimentary-matrix mélange Subduction complex | | NM | Tethyan Himalayan sequences Indian passive margin |
| Li et al. (2015b) | | NM NI | Renbu mélange Subduction complex | NM | Tethyan Himalaya Indian passive margin |
| Cai et al. (2012) | | Tangga mélange Subduction complex | Pomunong mélange Subduction complex | Upper Rongmawa Formation Trench fill | Northern Tethyan Himalaya Indian passive margin |
| Aitchison et al. (2011) | | Bainang terrane Subduction complex | Distal Tethyan Himalayan belt | | Indian passive margin |
| Guilmette et al. (2008) | | Yamrock mélange Subduction complex | | | Indian plate sediments Indian passive margin |
| Dupuis et al. (2005a, b, 2006) | | Yamrock mélange Subduction complex | | | Triassic flysch Indian passive margin Northern Tethyan sediments Indian passive margin |
| Aitchison et al. (2003); Aitchison & Davis (2004); Ziabrev et al. (2004) | | Bainang terrane Subduction complex | Yamdrok mélange | | Indian terrane Indian passive margin |
| Source                  | Geologic Unit                                    | Stratigraphy Details                                                                 | Location Notes                        | Notes                        |
|------------------------|--------------------------------------------------|--------------------------------------------------------------------------------------|---------------------------------------|------------------------------|
| Matsuoka et al. (2001, 2002) | Xialu chert                                      | Accreted ocean plate stratigraphy                                                   | Tethys Himalaya Indian passive margin | Tethys Himalaya Indian passive margin |
| Liu (2001); Liu & Aitchison (2002) | Bainang terrane                                 | Subduction complex                                                                  | Indian terrane Indian passive margin | Indian terrane Indian passive margin |
| Aitchison et al. (2000) | Bainang terrane                                 | Subduction complex                                                                  | Indian terrane Indian passive margin | Indian terrane Indian passive margin |
| Liu & Einsele (1996)    | Jurassic–Cretaceous oceanic sediments in Figure 2| Tectonic mélange/Cretaceous chaotic sediments Figure 2                                | Triassic rift clastics Indian passive margin | Tethyan shelf Indian passive margin |
| Wu (1993)               | Xialu chert                                      | Deposited in deep ocean                                                             | Flysch-like sediments Indian passive margin | NM Flysch-like sediments Indian passive margin |
| Searle et al. (1987)    | NM                                               | Yamdrok mélange                                                                     | Mesozoic flysch Indian passive margin | North Tethyan Himalaya Indian passive margin |
| Burg & Chen (1984); Burg et al. (1987) | Infra-ophiolitic thrust sheets of radiolarites/infra-ophiolitic sole | Wildflysch with exotic blocks Indian passive margin                                | Triassic–Liassic turbiditic flysch Indian passive margin | Himalayan sediments Indian passive margin |
| Girardeau et al. (1984) | Upper Jurassic–Lower Cretaceous red radiolarites | Wildflysch Metamorphic sole                                                        | Triassic flysch Indian passive margin | Himalayan sediments Indian passive margin |
| Tapponnier et al. (1981) | Deformed cherts and pelagic slates               | Upper Cretaceous wildflysch or mélange Possible subduction complex                 | Triassic flysch Indian passive margin | Indian shelf sediments Indian passive margin |
| Shackleton (1981)       | Jurassic (?)–Cretaceous radiolarite slices       | Sedimentary mélange Possible subduction complex                                     | Tethys Himalayan belt Indian passive margin | Tethys Himalayan belt Indian passive margin |
| Bally et al. (1980)     | Xialu section                                    | Mélange formations Possible subduction complex, dismembered thrust complex or olistoliths | Himalayan Tethys sediments Indian passive margin | Himalayan Tethys sediments Indian passive margin |

Names in bold include the siliciclastic-matrix mélange as defined in this study.
NM, no mention; NI, no interpretation.
therein), no amphibolite and only two local outcrops of blueschist have been documented in the siliciclastic-matrix mélangé, both in matrix shear zones near the ophiolite belt to the north (Ding et al. 2005; Li et al. 2007; H.-Q. Wang et al. 2017). Shackleton (1981) interpreted the southernmost part of the mélange as sedimentary in origin, although this may be in reference to the deformed Tethyan strata to the south that are included in the ‘sedimentary mélange’, while Searle et al. (1987) argued for a clearly tectonic origin. c. 100 km west of Xigaze, consistent with documentation of foliation, lineation and S-C fabrics in the siliciclastic matrix (Burg & Chen 1984; Cai et al. 2012). A local outcrop of sedimentary mélange recycled from the subduction complex is documented in the Lopu Range area (Metcalf & Kapp 2017), but the extent of sedimentary mélange within the subduction complex is unknown.

Blocks of ocean plate stratigraphy ranging in age from Permian to Paleocene could have been deposited well before accretion at the trench. Sandstone sourced from the upper plate and deposited in the trench would have been incorporated into the subduction complex within a few million years or less, their depositional age providing a robust constraint on accretion ages for these blocks. In the YSZ, detrital zircon U–Pb maximum depositional ages (MDAs) closely match biostratigraphic ages in both the Xigaze forearc basin (Fig. 2) and the early foreland basin in the Tethyan Himalaya (Hu et al. 2015b). By analysing sandstone blocks within the siliciclastic-matrix mélange, it is possible to determine the provenance, including continent of origin (India or Asia), and the MDA. Detrital geochronology of Cretaceous subduction complex units does not always provide a robust MDA because of the large number of age populations (Cai et al. 2012; An et al. 2017). Figure 2 distinguishes between samples with robust and poorly constrained (less than three zircon ages overlapping within 2σ uncertainty and/or the presence of significantly younger grains that do not overlap). The detrital zircon geochronology of sandstone blocks and matrix from the siliciclastic-matrix mélange shows both arc affinity and Tethyan Himalaya affinity. In the Sangsang area (Figs 1 & 3), Wang et al. (2018) divided the mélange into the northern mélange and the southern mélange, divided by a belt of basalt or basaltic breccias. Although the contact between the two mélanges was not observed, the northern mélange matrix contains more chert than the southern mélange matrix (Wang et al. 2018). Sandstone blocks and matrix in the northern mélange have MDAs of 149–134 Ma and sandstone petrography falling into the transitional to undissected arc field, while those in the southern mélange have MDAs of 87–78 Ma and sandstone petrography falling into the recycled orogen field (Fig. 2) (H.-Q. Wang et al. 2017, 2018). All samples from the southern mélange and some from the northern mélange contain Mesozoic and older zircon populations interpreted to be derived from the Lhasa terrane and Gangdese magmatic arc (H.-Q. Wang et al. 2017, 2018). In the Ngamring area (Figs 1 & 3), the siliciclastic-matrix mélange contains sandstone blocks with Paleozoic and older detrital zircon age populations, as well as Mesozoic populations that share the same age peaks and lulls with the Gangdese magmatic arc and Xigaze forearc basin (Cai et al. 2012). The MDAs of the Ngamring mélange sandstone blocks range from 85 to 200 Ma with individual grains as young as 71 Ma (Fig. 2), and the sandstone modal petrographical data plot in the recycled orogen field of Dickinson (1985), suggesting that the sandstone blocks were derived from the forearc basin during the Late Cretaceous (Cai et al. 2012). In the Renbu area (Fig. 1), the siliciclastic-matrix mélange contains sandstone blocks with Paleozoic and older detrital zircon age populations, as well as Mesozoic populations that share age peaks and lulls with the Gangdese magmatic arc and Xigaze forearc basin (Li et al. 2015b). Sandstone petrography on these blocks falls within the recycled orogen field, and zircon Hf isotopic signatures are consistent with the juvenile Gangdese magmatic arc and those of more evolved igneous rocks and detrital zircons in the central and northern Lhasa terrane (Li et al. 2015b). Thus, we interpret all the sandstone blocks and matrix in the mélange at Sangsang, Ngamring and Renbu to have been derived from Asia and incorporated into the subduction complex during oceanic subduction along the southern margin of Asia.

In the Lopu Range region, c. 50 km NW of Saga (Fig. 1), the siliciclastic-matrix mélange contains three types of sandstone blocks in terms of detrital zircon ages (Metcalf & Kapp 2017). The first contains no significant detrital zircon populations younger than 400 Ma, and the second has an additional Early Cretaceous peak age; both of these age spectra are consistent with Tethyan Himalayan strata (Metcalf & Kapp 2017). One sandstone block contains Mesozoic and older zircons, with the youngest population at c. 59 Ma; the zircon ages are similar to those within the Late Cretaceous–Paleogene Xigaze forearc basin strata and Paleocene strata of the oldest foreland basin deposits in the Sandanlin area (Metcalf & Kapp 2017). About 30 km east of Lazi (Figs 1 & 3), three types of sandstone blocks have been identified (An et al. 2017). The first is interpreted to be derived from Tethyan Himalayan strata, the second to represent trench fill that bypassed the filled forearc basin at c. 94 Ma and the third to be syncollisional foreland basin strata deposited at c. 54 Ma (An et al. 2017). Thus, there is a mixture of Tethyan and syncollisional sandstone blocks in the siliciclastic-matrix mélange near the Lopu
Fig. 3. Geological map of the central Yarlung Suture Zone. After Cai et al. (2012), Orme & Laskowski (2016), Orme (2017), Wang et al. (2018) and our own field observations. GCT, Great Counter Thrust; YZMT, Yarlung Zangbo Mantle Thrust. Note the scale bar for each local map. The dashed red line represents normal faulting. Larger versions of the maps with stereonets are included in the Supplementary material.
Range and Lazi. Additionally, Cretaceous sandstone blocks, absent near the Lopu Range, are mixed with the other sandstone blocks at Lazi (Shackleton 1981; Tapponnier et al. 1981; Searle et al. 1987; Dupuis et al. 2005b, 2006; Cai et al. 2012; Li et al. 2015b; An et al. 2017; Metcalf & Kapp 2017; H.-Q. Wang et al. 2017).

Trench fill. Cai et al. (2012) identified a turbiditic unit, the upper member of the Rongmawa Formation, which is in fault contact with the siliciclastic-matrix mélange to the north near Ngamring (Figs 1 & 3). The lower and middle members of the Rongmawa Formation are composed of chert, basalt, and red and green mudstone, typical lithologies in the siliciclastic-matrix mélange (Cai et al. 2012). We observed these exposures in the field and, because they have the same composition and style of deformation as the siliciclastic-matrix mélange, they should be considered part of that unit rather than part of the Rongmawa Formation. The upper Rongmawa Formation is similar to the Cretaceous sandstone blocks in the mélange in detrital zircon geochronology and sandstone petrography, and is interpreted as Asian-derived Late Cretaceous trench fill (Cai et al. 2012) (Fig. 2). Wang et al. (2018) documented Rongmawa Formation outcrops south of Sangsang (Figs 1 & 3) with a MDA of c. 85 Ma. The regional extent of this Cretaceous turbiditic unit is unknown. Near Saga, c. 600 km west of Lhasa (Fig. 1), An et al. (2018) identified a conglomeratic and turbidite unit, the Luogangcuo Formation, which overlies the siliciclastic-matrix mélange along a fault contact. The detrital zircon U–Pb ages and Hf isotope ratios are similar to those of Xigaze forearc basin strata, the conglomerates are poorly sorted with angular to subangular clasts dominated by chert, and petrographical data from the turbiditic sandstones plot in the transitional arc to recycled orogen provenance fields (Dickinson 1985; An et al. 2018). Olistostromal sandstone blocks have a main detrital zircon U–Pb peak age of c. 100 Ma (An et al. 2018). All of these attributes are consistent with derivation from the siliciclastic-matrix mélange and deposition into the trench south of the accretory prism (An et al. 2018). Although the Luogangcuo Formation lacks fossils, robust detrital zircon U–Pb MDAs suggest that it was deposited during the Late Cretaceous (Fig. 2) (An et al. 2018).

Tethyan Himalayan strata

The southern boundary of the siliciclastic-matrix mélange is the Zedong–Yangze Thrust (ZGT), which originally thrust the mélange southwards over Tethyan Himalayan strata. These deformed strata in the hanging wall have been identified as part of (1) the siliciclastic-matrix mélange or (2) the Indian passive margin, and are known variously as ‘Triassic flysch’, ‘flysch-like sediments’, ‘Triassic rift clastics’ or simply as part of the Tethyan Himalayan sequence (Table 1). The Tethyan Himalayan strata are divided into northern and southern subzones by the northern Himalayan ngeos domes (Liu & Einsele 1994). The northern subzone consists mostly of strongly deformed Triassic–Paleocene outer shelf, continental slope and abyssal strata, whereas the southern subzone consists mostly of Paleozoic–Eocene continental shelf strata (Liu & Einsele 1994, 1996, 1999). Shallow-marine carbonates were deposited on much of NE Gondwana during the Permian (Liu & Einsele 1994; Jin et al. 2015 and references therein). The Tethyan Himalayan strata were uplifted during the latest Permian initial stage of rifting, and large blocks of Permian limestone were broken up by faulting and mass wasting into olistostromal deposits during Triassic rifting (Tapponnier et al. 1981; Liu & Einsele 1994; Jin et al. 2015) (Fig. 2). Associated basalts show a within-plate geochemical signature with some continental crust input, distinct from the ocean island basalt (OIB) blocks in the siliciclastic-matrix mélange (Dupuis et al. 2005b). Synrift Triassic strata consist of thick (>5 km) turbidite sequences with minor thin limestone beds (Liu & Einsele 1994; Jin et al. 2015 and references therein). Post-rift subsidence during the Jurassic and Cretaceous led to deepening passive-margin facies and more olistostromal deposits during the Late Cretaceous (Liu & Einsele 1994, 1996, 1999).

Tethyan Himalayan strata generally contain no detrital zircon ages younger than c. 400 Ma (e.g. Gehrels et al. 2011) and have a highly quartzose, continental-interior sandstone petrographical composition (e.g. Décelles et al. 2014; Cai et al. 2016; J.-G. Wang et al. 2016; An et al. 2017; Metcalf & Kapp 2017), although signatures of Permo-Triassic and Early Cretaceous rifting events, as well as Paleocene India–arc or India–Asia collision, are locally recorded. In the eastern Tethyan Himalayan strata, some Upper Triassic strata are less compositionally mature (Li et al. 2010; Cai et al. 2016; J.-G. Wang et al. 2016), and contain detrital zircon signatures consistent with NW Australia and other Gondwana fragments, suggestive of sediment transport from eastern Gondwana rather than India (Cai et al. 2016; J.-G. Wang et al. 2016). The Comei and Rajmahal–Sylhet igneous provinces developed in NE India during the Early Cretaceous break-up of India and Australia (Fig. 2), and river systems distributed volcanic clasts along the northern margin of Greater India from east to west (Hu et al. 2015a and references therein). Lower Cretaceous volcanioclastic rocks, including the Wòlong Formation, are locally interbedded with 145–115 Ma intra-plate volcanic rocks (Hu et al. 2015a and references therein). Early Cretaceous zircon peak ages vary in Tethyan
strata. Those in the northern Tethyan Himalaya are similar to the Comei igneous province (c. 132 Ma), those in the Lesser Himalaya are similar to the Rajmahal–Sylhet igneous province (c. 117 Ma), and those in the southern Tethyan Himalaya fall in-between (Hu et al. 2015a). Arc-affinity Paleocene foreland basin deposits have been documented at Sangdalan (Ding et al. 2005; Aitchison et al. 2007a; J. Wang et al. 2011; DeCelles et al. 2014; Wu et al. 2014a; Hu et al. 2015b; Baxter et al. 2016), Gyangze (Cai et al. 2011; Wu et al. 2014a), south of Sangsang (H.-Q. Wang et al. 2017) and c. 30 km east of Lazi (An et al. 2017) (Fig. 1). For the purposes of this study, at c. 59 Ma, at least the region between the northern Tethyan Himalayan strata and the Xigaze forearc basin was sutured along the central portion of the YSZ.

Geology of the study area

Map patterns and faults

We mapped the YSZ units from Liuqu to Sangsang at a scale of 1:100 000. Geological mapping and sampling were conducted during traverses on foot. We present our new stratigraphic and structural data in Figure 3. More detailed maps with stereonets are available in the Supplementary material. The Cretaceous Xigaze forearc basin is exposed in the northern part of the study area, and is overthrust by ophiolite massifs and ophiolithic mélanges along the GCT (Fig. 3). The Liuqu Conglomerate is variably in the hanging wall and/or footwall of the GCT north of, south of or within the ophiolite (R.J. Leary et al. 2016). The Bainang terrane is variably exposed in the suture zone, and is either overthrust by the ophiolite along the north-dipping Yarlung Zangbo Mantle Thrust (YZM) or is itself over the ophiolite along a strand of the GCT (Fig. 3e). The exposure just west of Lang Xiazha village (Fig. 3e), the mélange matrix is locally chert-rich for c. 1.5 km to the south, and shale matrix dominates throughout the rest of the mélange. The mélanges fabric is predominately east–west-striking and south-dipping, but significant variations exist as well (Fig. 3). The distribution of block sizes is poorly sorted, ranging from decimetre- to kilometre-scale (Fig. 4a, b).

Siliciclastic-matrix mélange

The siliciclastic-matrix mélange is a block-in-matrix mélangé of ocean plate stratigraphy (Fig. 4a, b) in a matrix of purple and green shale and chert with occasional sandstone (Fig. 4c, d). Blocks of chert, metabasite, shale and limestone are most common, although blocks of sandstone are locally exposed (Fig. 4e–h). The farthest west siliciclastic-matrix mélange outcrop (c. 250 m long) mapped south of the Yarlung River (Fig. 3b) consists of basalt and purplish crystalline calcite, deformed together with mélange matrix (Fig. 4h). The largest blocks are limestones tens to hundreds of metres in stratigraphic thickness and extending for kilometres. Limestone blocks contain recrystallized fossils (mostly crinoid stems) and are locally pink. Blocks are variably deformed from intact pillow basalt (Fig. 4f), to folded chert, to pencil-cleaved and brecciated chert (Fig. 4e). The recessive matrix and its contacts with blocks are rarely exposed (Fig. 4c, g). South of Xiazha village (Fig. 3e), the mélange matrix is locally chert-rich for c. 1.5 km to the south, and shale matrix dominates throughout the rest of the mélange. The mélangé fabric is predominately east–west-striking and south-dipping, but significant variations exist as well (Fig. 3). The distribution of block sizes is poorly sorted, ranging from decimetre- to kilometre-scale (Fig. 4a, b).

Rongmawa Formation

The Rongmawa Formation is composed of fine- to medium-grained dark-grey litharenite sandstone beds interbedded with millimetre- to centimetre-scale dark-grey to black shale that is locally folded (Fig. 5a–e). In the valley south of Ngamring (Fig. 3c), the deformation and bed thicknesses increase to the north from centimetre-scale (Fig. 5b) up to tens of metres (Fig. 5c). The exposure just west of Lang Co (Fig. 3c) consists of millimetre- to centimetre-scale beds that are strongly deformed (Fig. 5d). In the next valley at 87° E (Fig. 3), the Rongmawa Formation is in contact with the siliciclastic-matrix mélange along a south-dipping thrust fault (Fig. 5f). Near the fault, Rongmawa Formation beds and mélange blocks are broken up and mixed in a fault zone a few tens of metres thick (Fig. 5g). Since the normal faulting in the valley north of Lazi (Orme 2017) offsets some contacts across the valley (Fig. 3e). All of the mapped suture zone units exhibit folding about an axis trending c. 105° (Fig. 3). Much of the map area is overlain by loess and aeolian soil deposited since the Last Glacial Maximum (Sun et al. 2007; Klinge & Lehmkühl 2015; Stauh 2015; Dong et al. 2017), complicating the projection of contacts along strike.
Fig. 4. Siliciclastic-matrix mélange field photographs. (a) & (b) Typical block-in-matrix mélange outcrops with good exposure of matrix. The matrix in (b) locally contains serpentinite. (c) & (d) Exposures of characteristic green and purple shale matrix with occasional sandstone or chert. (e) Block of brecciated chert. (f) Block of pillow basalt. (g) Block of sandstone (7.11.14.2KM) showing deformation, but poor exposure of the matrix surrounding the block. (h) Basalt and purplish crystalline calcite deformed together with the mélange matrix in the western structural window (Fig. 3).
Fig. 5. Rongmawa Formation field photographs. (a) & (b) Rongmawa Formation turbiditic sandstone with and without quartz veins. (c) Bed thickness increasing upsection to north. (d) Strongly deformed beds west of Lang Co (Fig. 3). (e) Westernmost exposure (Fig. 3). (f) Fault contact with siliciclastic-matrix mélange. (g) Road cut of the fault-zone breccia with siliciclastic-matrix mélange. Inset shows turbiditic bedding within the brecciated Rongmawa Formation.
Rongmawa Formation is primarily exposed in narrow valleys, and the surrounding hillsides are covered by loess, contacts outside of mapped areas are uncertain, but documented outcrops are exposed between 86.7 and 87.4° E (Fig. 3). Where exposed, the north–south map width of the Rongmawa Formation varies along strike from c. 1 to c. 10 km (Fig. 3).

**Tethyan Himalayan strata**

Deformed Triassic–Cretaceous Tethyan Himalayan strata are exposed south of the siliciclastic-matrix mélange (Fig. 3). The majority of the succession is composed of thick sequences of medium- to coarse-grained quartzarenite sandstone with only minor shale interbeds (Fig. 6a, c, d). Intervening units of dark shale contain some dismembered medium- to coarse-grained sandstone bodies (Fig. 6b, e). Minor centimetre- to decimetre-scale limestone units are interbedded with grey or dark shale, siltstone and marl. Large olistostromal limestone blocks with associated metabasite are locally present within the Tethyan strata (Fig. 3a). Tethyan strata are folded, and sandstone beds within thicker shale intervals are dismembered (Fig. 6a, b, e). About 6 km NW of Lazi (Fig. 3e), deformation gradually increases from folded and cleaved (Fig. 6f) to transposed (Fig. 6g). In the eastern part of the study area (Fig. 3a, g), Tethyan Himalayan strata are deformed to block-in-matrix mélange (Fig. 6h, i). All blocks in the Tethyan Himalayan mélange are composed of Tethyan Himalayan sandstone, shale, limestone, olistostromal Permian limestone and metabasite associated with olistostromes within a grey shale matrix. Blocks of pillow basalt and chert are absent, and sandstone is dominant. At first glance, the mélange of Tethyan Himalayan strata is difficult to distinguish from the siliciclastic-matrix mélange. However, the only exotic blocks in the mélange of Tethyan Himalayan strata are the olistostromal limestone and basalt. In addition, the matrix of the northern Tethyan Himalayan units is dark-grey shale compared to the purple and green siliciclastic or chert matrix of the siliciclastic-matrix mélange.

**Fault contact between siliciclastic-matrix mélange and Tethyan Himalayan strata**

The Tethyan Himalayan strata and siliciclastic-matrix mélange are in fault contact along a strand of the GCT. Generally, the Tethyan Himalayan strata are thrust northwards over the siliciclastic-matrix mélange; however, the fault geometry is complex, particularly NE of Lazi (Fig. 3e). West of Lazi (Fig. 3a–c), the fault is nearly planar and gently south-dipping (Fig. 7a, b); but immediately east of the Yarlung River north of Lazi (Fig. 3e), the fault is folded to the point of being overturned and changes from gently south dipping to vertical north–south striking to horizontal over a distance of a few hundred metres (Fig. 7c, d). A few kilometres east (Fig. 3e), the fault is offset by tens to a few hundred metres along a series of vertical, NNW–SSE-striking strike-slip faults (Fig. 7c). Farther eastwards, the Tethyan strata are in fault contact with the Liuqu Conglomerate or ophiolite, having been thrust northwards over the siliciclastic-matrix mélange (Fig. 3a, e, g). Farther south near the main highway (Fig. 3a, e, f), the siliciclastic-matrix mélange is exposed in a folded window structurally beneath the Tethyan Himalayan strata, >25 km from the northern extent of the Tethyan Himalayan strata, documenting major shortening along this strand of the GCT. West of Lang Co (Fig. 3), the fault contact steps several kilometres south near the Yarlung River before stepping north again at the western end of the study area where it is in contact with the Paleocene Sangdanlin Formation (H.-Q. Wang et al. 2017) (Fig. 3b). Another window of siliciclastic-matrix mélange is exposed along the Yarlung River to the south (Fig. 3a, b). From 87.5 to 87.75° E along the northern strand of the GCT (Fig. 3a, e), the mélange matrix also contains serpentinite within a c. 1 km-wide zone (Fig. 7a–d), but no serpentinite was observed farther west or along the margins of the structural window east of Lazi (Fig. 7e, f). Serpentinite in the mélange matrix could be the result of metasomatism of mafic blocks in a shear zone during slip along the fault or the remnants of a serpentinite diapir within the subduction complex which acted as a plane of weakness for fault propagation and slip.

**Xiazha Formation**

Two sections of intact stratigraphy are exposed between the ophiolite and siliciclastic-matrix mélange. We name these previously undescribed rocks the Xiazha Formation. The eastern section is located c. 16 km north of Lazi, adjacent to the village of Xiazha (Fig. 3a, e). On the east bank of the

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**Fig. 6.** Tethyan Himalayan strata field photographs. (a) & (b) Sandstone beds folded and dismembered within weaker shale. (c) Thick quartzarenite sandstone beds showing relatively little deformation. (d) Centimetre-scale sandstone beds with minor interbedded shale showing boudinage. (e) Sandstone bed dismembered within weaker shale. (f) Beds folded with axial-planar cleavage. (g) North of (f), the axial-planar cleavage is also folded. (h) Less deformed block of strata within a weaker shale matrix exposed in the east of the study area (Fig. 3). (i) Dark shale matrix of the block-in-matrix mélange of Tethyan Himalayan strata.
Yarlung River, the section is exposed between the ophiolite to the north and the siliciclastic-matrix mélangé to the south (Fig. 8a). Although the fault contact is locally non-planar, shear fabrics in the hanging wall are consistent with a gently (c. 20°) south-dipping fault (Fig. 8b). The section has an apparent stratigraphic thickness of c. 300 m, but dismembered beds (Fig. 8c) suggest that the section may be structurally thickened by tight folding and/or faulting. The section is largely composed of grey and black chert interbedded with shale and occasional coarser beds of sandstone and intraclastic chert (Fig. 8c). Distinctive fissile, green intraclastic beds contain very thin (<mm) fine-grained clasts (Fig. 8d). Just SW across the Yarlung River, the Bainang terrane is exposed in the same structural position as the Xiazha Formation (Fig. 3e). However, the lack of red radiolarian chert within the Xiazha Formation suggests that they are not equivalent.

The western section of the Xiazha Formation is located c. 2 km SW of Ngamring Lake (Fig. 3a, c), dips 20–40° to the south and has a stratigraphic thickness of c. 400 m (Fig. 8e). To the north, serpentined mafic rocks intruded by diorite dykes are deformed along with red ribbon chert in a gently folded broken formation at the southern edge of the ophiolitic mélangé (Fig. 8f). To the south, the Xiazha Formation is exposed beneath the siliciclastic-matrix mélangé in the footwall of a south-dipping fault (Fig. 3c). The western section is coarser than the eastern section (Fig. 8g), containing thicker and more numerous beds of sandstone with pebble- to cobble-sized volcanic and limestone clasts (Fig. 8h), intraclastic limestone (Fig. 8i), and green intraclastic beds (Fig. 8j). Intervening fine-grained and more recessive units are covered. Based on the presence of chert and limestone, the Xiazha Formation was deposited in a marine environment, and intraclastic beds suggest higher-energy turbiditic flows. Pebble to cobble, fine-grained intermediate to felsic volcanic clasts indicate their proximity to an active arc. The thicker, coarser, intraclastic beds in the western section were likely to have been deposited more proximal to the arc than the mostly shale and chert beds in the eastern section.

**Methods**

We sampled sandstones from across the siliciclastic-matrix mélangé (n = 8), Rongmawa Formation (n = 6), Xiazha Formation (n = 2), Tethyan strata (n = 7) and Tethyan Himalayan mélangé (n = 3) for U–Pb detrital zircon geochronology to determine their provenance and maximum depositional ages (MDAs). Zircons were recovered from samples using standard crushing and separation methods, mounted in epoxy, polished, imaged by back-scattered electrons using a scanning electron microscope, and dated using a laser-ablation multicollector inductively-coupled plasma mass spectrometer (LA-MC-ICPMS) at the Arizona LaserChron Center (methods described in Gehrels et al. 2008). Samples were analysed using either a 20 or 35 µm beam. We dated a total of 5420 grains, 539 of which were excluded because of discordance (>20%), reverse discordance (>5%), high common lead (>400 counts/s) or high uncertainties in ages (>10%). Most of the excluded grains had high common lead. Here we present 4881 new detrital zircon U–Pb ages. Because of the paucity of post-Cambrian zircons in Tethyan Himalayan strata, the MDA determined by zircon populations does not approximate the depositional age of most Tethyan Himalayan strata, so we report the youngest peak age (Fig. 3) rather than calculating MDAs for these samples. The youngest population of zircon ages in arc-derived sandstones may approximate their true depositional age (DeCelles et al. 2007; Dickinson & Gehrels 2009; Painter et al. 2014); we report MDAs for these samples. Previous detrital zircon studies from Cretaceous subduction complex units were not always able to determine a robust MDA for some samples because of the large number of age populations (Cai et al. 2012; An et al. 2017). We analysed 315 grains for each sample of subduction complex units in order to better resolve the youngest population of ages within the samples.

To calculate the MDAs, we first screened the youngest group of three or more ages that overlapped within 2σ uncertainty. Grains with anomalously high U concentrations that might be affected by lead loss and those with anomalously high U/Th ratios which might be affected by fluids post-crystallization were excluded in each case. We then calculated the unmixed age, which uses the Sambridge–Compston algorithm to fit Gaussian curves to the data, allowing a determination of age uncertainties but requiring a choice of the number of peaks to calculate (Ludwig 2012). Finally, we calculated the weighted mean age of the youngest unmixed age population to examine the mean square weighted deviation (MSWD) of that population. MSWD values ≤1 indicate that the age...
uncertainty is sufficient to explain the scatter in ages and gives us confidence that the grains are cogenetic, whereas MSWD values >1 indicate that the age uncertainty is insufficient to explain the scatter and that the grains may not be cogenetic.

We prepared thin sections for sandstone petrography from the 16 Jurassic and Cretaceous sandstone samples analysed for detrital zircon dating, and an additional Cretaceous sandstone. The thin sections were stained for K-feldspar and Ca-plagioclase. We counted 450 grains for each sample using the Gazzi–Dickinson method (Gazzi 1966; Dickinson 1970; Ingersoll et al. 1984).

Results

Collectively, the samples yielded five distinct detrital zircon age patterns (Fig. 9). The Triassic strata and mélange sandstones contain two age patterns: samples with no significant populations <400 Ma; and samples with Triassic peak ages between 240 and 214 Ma. The Xiazha Formation sandstones have single peak ages at 159.2 ± 1.9 and 157.1 ± 0.7 Ma. The siliciclastic-matrix mélange sandstones contain three age patterns: one block has a single peak age at 158.5 ± 0.5 Ma, consistent with the Xiazha Formation; Early Cretaceous MDAs of 108.8 ± 4 and 105.4 ± 0.4 Ma with 83 and 94% of the grains dated being <200 Ma; and Cretaceous MDAs of between 99 and 90 Ma with 14–22% of the grains dated being <200 Ma (Figs 2, 3 & 9). In the western part of the study area, the MDAs of the samples generally decrease from north to south in the siliciclastic-matrix mélange (Fig. 3b). The Rongmawa Formation sandstones have Cretaceous MDAs of between 90 and 85 Ma with 22–35% of the grains dated being <200 Ma (Figs 2, 3 & 9).

We conducted petrographical analyses on sandstone samples from the Xiazha Formation, siliciclastic-matrix mélange and Rongmawa Formation that yielded age populations of <200 Ma, and investigated other beds within the Xiazha Formation (Fig. 10). The Xiazha Formation sandstones are predominately oolitic, and the composition of ooid cores was recorded during point counting. In the eastern section, sample 7.8.12.3KM contains ooids with both calcareous and volcanic lithic cores (Fig. 10a). The cortices are dominantly calcareous but may also include silica, especially towards the edge of the ooid, and the matrix is composed of phosphate (Fig. 10a). There is local evidence of pressure dissolution along ooid grain boundaries. The intraclastic chert bed, sample 7.8.12.4KM, contains mostly organic-rich chert with some plagioclase and felsic volcanic clasts in a phosphatic and calcareous matrix (Fig. 10b). Sample 7.8.12.5KM, the distinctive green intraclastic layer cut parallel to bedding due to the friable layering, is also composed of ooids similar to the oolitic sandstone (Fig. 10c). The matrix contains less phosphate than the oolitic sandstone, but the phosphate is associated with chert, dolomite and pyrite (Fig. 10c).

In the western section, sample 7.19.14.2KM of the distinctive intraclastic green layer (also cut parallel to bedding due to the friable layering) is dominated by clasts of volcanic glass with minor quartz and plagioclase grains (Fig. 10d). Despite having similar hand-sample appearance, the thin-section compositions for samples 7.8.12.5KM and 7.19.14.2KM are different (Fig. 10c, d). We attribute this compositional variation to the layered nature of the beds and the necessity to cut thin sections parallel to bedding. Sample 7.19.14.3KM, an intraclastic limestone bed, is composed of millimetre-scale oncoids and smaller ooids similar to those in the eastern section (Fig. 10e). Sample 7.19.14.4KM is an oolitic sandstone similar to 7.8.12.3KM (Fig. 10a) with calcareous and volcanic lithic cores, and evidence of pressure dissolution along ooid grain boundaries, but phosphate is a minor component of the matrix (Fig. 10f). Ooid cortices are composed of calcite and interlayered silica (Fig. 10f). The silica is often elongated in a preferential direction that is the same throughout the sample, and in some places cross-cuts the outer calcareous cortex (Fig. 10f), consistent with secondary mineralization. Point-count analyses of clasts and ooid cores plot in the undissected arc field (Fig. 11).

Sample 7.25.14.1KM is a sandstone block from the siliciclastic-matrix mélange, and has a detrital zircon signature and sandstone petrography similar to the Xiazha oolitic sandstones, but contains no ooids (Figs 10g & 11). Samples 7.25.14.3KM and 7.25.14.4KM are coarse, contain subangular grains (Fig. 10h) and plot in the transitional arc field.

Fig. 8. Xiazha Formation field photographs: (a)–(d) eastern section (Fig. 3) and (e)–(j) western section (Fig. 3). (a) Eastern section exposed between the ophiolite and siliciclastic-matrix mélange. (b) Fault contact between the Xiazha Formation and the ophiolite. (c) Typical exposure of centimetre-scale dismembered beds of grey and black chert interbedded with shale and a coarser bed of sandstone (7.8.12.3KM). (d) View of a bedding plane of green intraclastic beds containing very thin (<mm) fine-grained clasts (7.8.12.5KM). (e) Western section exposed between the ophiolite and siliciclastic-matrix mélange. (f) Broken formation of chert within the ophiolite. (g) Typical exposure of the coarser western section. (h) Sandstone with pebble- to cobble-sized clasts of volcanic rock (LV) and limestone (LS) (7.19.14.4KM). (i) Intraclastic limestone with centimetre-scale and finer clasts (7.19.14.3KM). (j) View of bedding plane of green intraclastic beds containing very thin (<mm) fine-grained clasts (7.19.14.2KM).
Fig. 9. Detrital zircon U–Pb ages from sandstones in the Tethyan Himalayan strata, siliciclastic-matrix mélange, Xiazha Formation and Rongmawa Formation in the central suture zone. Ages are shown as histograms, probability density functions in coloured curves (the same colour scheme as Figs 2, 3 & 11) and kernel density estimations in black curves; \( n \) is the number of grains in the format \( n = \text{number in the subplot}/\text{number in the full plot} \). Inset plots show ages <200 Ma with a 5 myr histogram bin width, and full plots show ages 3500 Ma with a 50 myr histogram bin width. Note that the y-axis scale varies.
(Fig. 11), and the other Cretaceous mélangé sandstone blocks contain fewer lithic grains in a fine volcanic matrix with quartz fibres forming in pressure shadows (Fig. 10i) and plot in the continental interior field (Fig. 11). The Rongmawa Formation sandstones exhibit matrix and deformation (Fig. 10j) similar to the Late Cretaceous sandstone blocks in the siliciclastic-matrix mélangé, and plot in the recycled orogen to continental interior fields (Fig. 11).

Discussion

Provenance

The Xiazha Formation records a single zircon peak age (159–157 Ma), undissected arc provenance, high-energy ooids and a dysoxic, sulphate-reducing environment. Although the ooids were formed in a shallow-marine environment, their association with turbiditic deposits, shale and chert suggest that they were retransported into a deep-marine environment. The dominance of volcanic lithic clasts, ooids and oncoids derived above wave base, and interbedded chert and turbidites could be consistent with any forearc environment from the forearc basin to the trench. Deep-marine deposition in the forearc basin during ophiolite generation would be consistent with the basal Xigaze forearc basin (Marcoux et al. 1982; Girardeau et al. 1984; Wu 1984, 1986; Li & Wu 1985; Einsele et al. 1994; Dürr 1996; Ziabrev et al. 2003; Wu et al. 2010; Aitchison et al. 2011; Dai et al. 2015; W. Huang et al. 2015; Orme & Laskowski 2016; Wang et al. 2017a) but does not exclude deposition in a trench basin. Sample 7.19.14.4KM contains three grains in the 1293–214 Ma range, and sample 7.8.12.3KM contains six grains in the 3046–480 Ma range (Fig. 9). There are no significant >200 Ma populations (three or more grains overlapping within 2σ uncertainty), and a young 38 Ma grain in sample 7.8.12.3KM (Fig. 9) may be explained by contamination with a late-stage Gangdese arc grain from loessic soil which blankets the region (Sun et al. 2007; Klinge & Lehmkohl 2015; Staub 2015; Dong et al. 2017). The presence of >200 Ma grains suggests that the Xiazha Formation developed proximal to the southern Lhasa terrane, but could be explained by contamination from loessic soil or inheritance from minor continental fragments in an intra-oceanic subduction zone. Based only on detrital zircon ages, we cannot exclude the possibility that the Xiazha Formation may have formed in the forearc of an intra-oceanic subduction zone.

Jurassic mélangé sandstone sample 7.25.14.1KM has the same detrital zircon age (Fig. 9) and similar sandstone petrography (Fig. 11; Table 2) as the Xiazha Formation but is composed of volcanic and plagioclase grains without ooids (Fig. 10g), possibly deposited more distally from the Asian margin. This is further supported by the absence of limestone clasts in the mélangé sandstone block compared to the oolitic sandstones from the Xiazha Formation (Fig. 11). The three Jurassic samples are the oldest and most compositionally immature forearc deposits preserved in the YSZ (Figs 2 & 11). Early Cretaceous mélangé sandstone samples 7.25.14.3KM and 7.25.14.4KM plot in the transitional arc field (Fig. 11), and display prominent single peak ages with 6–17% of grains scattered between 3271 and 216 Ma (Fig. 9). In this case, there is good evidence for minor older continental input, so we interpret that these samples were derived from the Lhasa terrane. Late Cretaceous mélangé samples have significant other detrital zircon age populations (78–86% >200 Ma) (Fig. 9) and plot in the recycled orogen field (Fig. 11). The Late Cretaceous sandstone blocks from this study are more quartz-rich than the sandstone blocks from previous studies (Cai et al. 2012; An et al. 2017) (Fig. 11), but all are more quartz-rich than sandstones in forearc basin units that are similar in age and detrital zircon age populations (Einsele et al. 1994; Dürr 1996; An et al. 2014; Hu et al. 2015c; Orme et al. 2015; Orme & Laskowski 2016). To be incorporated into the mélangé, the sandstones must have been deposited in or near the trench, so the additional transport distance may have made the sandstones in the trench more compositionally mature than those in the forearc basin. Rongmawa Formation sandstones plot in the recycled orogen to continental interior fields (Fig. 11) and have significant other detrital zircon populations (65–78% >200 Ma) (Fig. 9), so we interpret that they were derived from the Lhasa terrane. Although the Rongmawa Formation sandstones are younger than the sandstone blocks from this study in the mélange (Fig. 3), they are compositionally less mature (Fig. 11).

The Rongmawa Formation sandstones are similar in detrital zircon geochronology (Fig. 12) and sandstone petrography (Fig. 11) to those reported in previous work by Cai et al. (2012), but analysing more grains per sample enabled determination of more robust MDAs (Figs 2, 3 & 9). Late Cretaceous mélange blocks similarly match previous detrital zircon geochronology (Fig. 12) while providing more robust MDAs, but their sandstone petrography is significantly more mature than that reported in previous studies (Fig. 11). Both the Rongmawa Formation and Late Cretaceous sandstone blocks have similar U–Pb zircon ages to those of the Upper Cretaceous Xigaze forearc basin strata. Early Cretaceous transitional arc (Fig. 11) samples with single dominant peak ages match well with the olistostromal sandstone blocks in the Luogangcuo Formation trench fill and with the Lower Cretaceous Xigaze forearc basin (Fig. 12) but were not previously documented in the
siliciclastic-matrix mélange. Latest Jurassic–earliest Cretaceous (149–134 Ma) undissected arc sandstones were documented near Sangsang (Wang et al. 2018), but the Jurassic samples presented here are the oldest (c. 159 Ma) and most compositionally immature within the suture zone (Figs 11 & 12).

Fig. 10. Photomicrographs of subduction complex units: (a)–(f) Xiazha Formation and (g)–(j) siliciclastic-matrix mélange. All images are in plane-polarized light. (a) Oolitic sandstone composed of volcanic lithic clasts and ooids with phosphate matrix. Ooids contain both calcareous and volcanic cores. (b) Clasts of organic-rich chert with minor plagioclase and felsic volcanic clasts in a phosphatic and calcareous matrix. (c) Oolitic sandstone with a phosphatic and calcareous matrix. Ooids contain more calcareous than volcanic lithic cores. (d) Similar hand-sample appearance to (c) but composed of clasts of volcanic glass with minor quartz and plagioclase grains in a phosphate matrix. (e) Oncoids with calcareous cores, felsic volcanic lithics, and ooids with both calcareous and volcanic lithic cores, all in a calcareous matrix. (f) Oolitic sandstone of volcanic lithic clasts and ooids with a phosphate matrix. Ooids contain both calcareous and volcanic cores, and some rims are siliceous. Many ooids show evidence of pressure dissolution.
Fig. 10. Continued. (g) Jurassic sandstone block in the siliciclastic-matrix mélangé dominated by angular volcanic lithic clasts. (h) Early Cretaceous sandstone block in the siliciclastic-matrix mélangé. (i) Late Cretaceous sandstone block in the siliciclastic-matrix mélangé. Quartz fibres formed in pressure shadows. (j) Deformed Rongmawa Formation. Quartz fibres formed in pressure shadows.

Fig. 11. Sandstone petrography ternary plots of subduction complex and Tethyan Himalayan units from this study, Cai et al. (2012), An et al. (2017, 2018) and Wang et al. (2018). Provenance fields after Dickinson et al. (1983). CI, continental interior; Lm, metamorphic lithics; Ls, sedimentary lithics; Lv, volcanic lithics; RO, recycled orogen; UDA, undissected arc. Recalculated petrographical data are displayed in Table 2.
Table 2. Recalculated petrographical data

| Unit          | Sample | Qm (%) | F (%) | Lt (%) | Qt (%) | F (%) | L (%) | Qm (%) | P (%) | K (%) | Lm (%) | Lv (%) | Ls (%) |
|---------------|--------|--------|-------|--------|--------|-------|-------|--------|-------|-------|--------|--------|--------|
| Xiazha Fm     | 7.8.12.3KM | 1.7    | 16.5  | 81.8   | 6.6    | 16.5  | 76.8  | 9.1    | 90.9  | 0     | 0      | 66.5   | 33.5   |
| Mélangé block | 7.11.14.2KM | 69.7   | 2.9   | 27.4   | 97.1   | 2.9   | 0.0   | 96.0   | 4.0   | 0     | 0      | 100    |        |
| Mélangé block | 7.11.14.3KM | 76.2   | 2.7   | 21.1   | 97.3   | 2.7   | 0.0   | 96.6   | 3.4   | 0     | 0      | 100    |        |
| Mélangé block | 7.11.14.4KM | 74.1   | 2.0   | 23.9   | 98.0   | 2.0   | 0.0   | 97.4   | 2.6   | 0     | 0      | 100    |        |
| Rongmawa Fm   | 7.18.14.3KM | 68.7   | 12.1  | 19.2   | 86.8   | 12.1  | 1.1   | 85.0   | 15.0  | 0     | 0      | 5.8    | 94.2   |
| Rongmawa Fm   | 7.18.14.4KM | 62.7   | 23.4  | 13.8   | 69.5   | 23.4  | 6.9   | 72.8   | 27.2  | 0.8   | 43.5   | 51.6   |        |
| Rongmawa Fm   | 7.18.14.5KM | 32.9   | 24.2  | 43.0   | 42.5   | 24.2  | 33.3  | 57.6   | 42.4  | 1.0   | 68.8   | 30.2   |        |
| Rongmawa Fm   | 7.18.14.6KM | 54.5   | 20.0  | 25.5   | 65.5   | 20.0  | 14.4  | 73.1   | 26.9  | 0.6   | 45.1   | 44.2   |        |
| Rongmawa Fm   | 7.18.14.7KM | 40.4   | 26.3  | 33.3   | 52.5   | 26.3  | 21.2  | 60.5   | 39.5  | 0     | 0      | 62.4   | 37.6   |
| Rongmawa Fm   | 7.18.14.8KM | 31.8   | 22.9  | 45.2   | 41.6   | 22.9  | 35.4  | 58.1   | 41.9  | 1.5   | 74.9   | 23.6   |        |
| Xiazha Fm     | 7.19.14.4KM | 0.5    | 8.7   | 90.8   | 3.6    | 8.7   | 87.7  | 5.3    | 94.7  | 0     | 0      | 59.7   | 40.3   |
| Mélangé block | 7.22.14.4KM | 71.7   | 3.1   | 25.2   | 95.6   | 3.1   | 1.3   | 95.8   | 4.2   | 0     | 4.4    | 0.9    | 94.7   |
| Mélangé block | 7.22.14.5KM | 72.5   | 10.0  | 17.4   | 86.5   | 10.0  | 3.1   | 87.8   | 12.2  | 0.1   | 15.4   | 83.3   |        |
| Mélangé block | 7.28.14.1KM | 2.9    | 7.8   | 90.0   | 4.1    | 7.8   | 88.9  | 29.5   | 70.5  | 0     | 0      | 98.2   | 1.8    |
| Mélangé block | 7.25.14.3KM | 12.1   | 42.2  | 45.7   | 16.2   | 42.2  | 41.6  | 22.3   | 77.7  | 0     | 0.5    | 75.5   | 24.0   |
| Mélangé block | 7.25.14.4KM | 8.7    | 58.3  | 33.0   | 8.9    | 58.3  | 32.8  | 13.0   | 87.0  | 0     | 0.7    | 97.2   | 2.1    |

Fm: Formation.

Subduction complex accretion ages

In the study area, the youngest fossils are Paleocene (Tapponnier et al. 1981; Burg & Chen 1984; Burg et al. 1985; Liu & Aitchison 2002), but the youngest sandstone blocks in the mélange are c. 85 Ma, although there is one matrix sandstone as young as 78 Ma (Fig. 2). The Rongmawa Formation is as young as c. 78 Ma (Fig. 2). The Rongmawa Formation is as young as c. 85 Ma (Wang et al. 2018; this study) or possibly but not demonstrably c. 70 Ma (Cai et al. 2012), and the Luogangcuo Formation is as young as c. 80 Ma (An et al. 2018). According to the sample locations from An et al. (2017), there may be blocks of the Sangdanlin Formation east of Lazi. However, the results of An et al. (2017) also place an Asian-affinity Cretaceous sandstone block in what we observed as bedded Triassic Tethyan Himalayan strata (Fig. 3), so the sample locations with respect to contacts need to be carefully examined. In any case, there are no documented Asian-affinity sandstones with robust MDAs between c. 78 Ma and collision at c. 80 Ma (Cai et al. 2018). To the west are shallow marine to fluviodeltaic, and record final basin filling (Wan et al. 2001; Ding et al. 2005; C. Wang et al. 2012; An et al. 2014; Hu et al. 2015; Orme & Laskowski 2016). If sediment supply to the forearc remained constant, much of the sediment would have bypassed the Xigaze forearc basin and been deposited in the trench. The sediment supply to the trench must have been cut off by disruption of sediment pathways (growth of an outer forearc high) or by decreasing sediment supply to the Xigaze forearc basin. Li et al. (2017) argued that growth of the subduction complex as a result of increased convergence created an outer forearc high and cut off the sediment pathway from the forearc basin to the trench. However, modern trenches with high convergence rates are more likely to have less sediment in the trench and be erosive or non-accretionary (Clift & Vannucchi 2004). While frontal erosion and basal accretion could cause forearc uplift, such as in the Coastal Cordillera of northern Chile (Delouis et al. 1998; Adam & Reuther 2000; Marquardt et al. 2004; Clift & Hartley 2007; Regard et al. 2010), we have already ruled out subduction erosion during Late...
Cretaceous–Paleocene time. An unconformity may have developed within the Xigaze forearc basin at c. 65 Ma (Ding et al. 2005; C. Wang et al. 2012), which in turn suggests that there was decreased sedimentation in the whole forearc region and decreased sediment supply from the Lhasa terrane.
and Gangdese arc. Thermochronological evidence of uplift of the southern forearc basin near Xigaze (Fig. 1) beginning at 89–80 Ma and subsequent northwards migration of the forearc basin depocentre (Li et al. 2017) is broadly contemporaneous with an increase in convergence rate (van Hinsbergen et al. 2011; Gibbons et al. 2015) (Fig. 2), but is not evident near Lazi or the Lopu Range where cooling began after c. 20 Ma (Orme 2017).

Although blocks of ocean plate stratigraphy in the siliciclastic-matrix mélange range from Permian to Late Paleocene, only the youngest blocks and sandstone deposited in the trench provide any constraints on age of accretion. Since the Xiazha Formation was possibly deposited on the upper plate, we cannot use it to constrain accretion age. The mélange block with similar age and sandstone petrography could have been deposited in the trench or been an olistostromal block derived from the upper plate. The oldest accretion documented in the subduction complex is recorded by 149 Ma sandstone blocks and matrix at the northern edge of the mélange near Sangsang (Wang et al. 2018). Wang et al. (2018) argued for a gap in accretion ages from 134 to 87 Ma near Sangsang, but we present blocks with 109–91 Ma MDAs in the same region which partially fill this gap. The remainder of the gap can probably be attributed to a low sediment supply to the trench during Early Cretaceous forearc extension and formation of the basal Xigaze forearc basin. There is some evidence of younging towards the trench near Sangsang, but nowhere else in the siliciclastic-matrix mélange. In fact, the unknown age distribution of blocks other than sandstone within the mélange, the unknown extent of recycling into the trench by mass transport, the abundance of GCT deformation and the exposure of blocks of Tethyan Himalayan strata at the northern edge of the mélange in the Lopu Range region suggest caution in interpreting the spatial distribution of subduction complex ages. Regardless, there are no units in the subduction complex which clearly document accretion prior to the latest Jurassic.

Tectonic model

Pre-Cretaceous southern margin of Asia. Subduction is suggested to have initiated along the southern margin of the Lhasa terrane by Middle Triassic time (e.g. Chu et al. 2006; Ji et al. 2009a; Zhu et al. 2011; Guo et al. 2013; Kang et al. 2014; Meng et al. 2016; J.-G. Wang et al. 2016; Ma et al. 2017b; R. Wang et al. 2017). The oldest strata in the Xigaze forearc basin are Early Cretaceous (Marcoux et al. 1982; Girardeau et al. 1984; Wu 1984, 1986; Li & Wu 1985; Einsele et al. 1994; Dürr 1996; Ziabrev et al. 2003; Wu et al. 2010; Aitchison et al. 2011; Dai et al. 2015; W. Huang et al. 2015; Orme & Laskowski 2016; Wang et al. 2017a), concurrent with the most recent ophiolite generation event (Göpel et al. 1984; Malpas et al. 2003; Miller et al. 2003; Geng et al. 2006; Wang et al. 2006; Wei et al. 2006; Chan et al. 2007, 2015; Guilmette et al. 2007, 2009; Li et al. 2008; Xia et al. 2008, 2011; Zhu et al. 2009; Liu et al. 2011, 2016; Xiong et al. 2011; Dai et al. 2012, 2013; Bao et al. 2013). The only pre-Cretaceous units exposed south of the Gangdese arc are the 160–155 Ma Zedong arc (McDermid et al. 2002; Aitchison et al. 2007b; L. Wang et al. 2012; Zhang et al. 2014), its Bathonian–Lower Callovian chert and basalt basement (McDermid et al. 2002; Aitchison et al. 2007b), the 160–150 Ma ophiolitic rocks (McDermid et al. 2002; Miller et al. 2003; Chan et al. 2007, 2015), the 159–146 Ma Jurassic sandstone blocks and matrix in the siliciclastic-matrix mélange (Wang et al. 2018; this study). Modern arc trench gaps suggest that the pre-Cretaceous forearc was >200 km wide (Fig. 13a). The absence of any pre-Cretaceous forearc basin and subduction complex sandstones (with well-constrained MDA) (Fig. 2), with the exception of the Jurassic Xiazha Formation and mélange sandstones, and the creation of an Early Cretaceous ophiolite that formed the basement of the c. 8 km-deep Xigaze forearc basin without major changes in the location of Gangdese arc magmatism suggest that the pre-Cretaceous forearc of the Lhasa terrane was dismembered and subducted, possibly along with an intra-oceanic arc.

Two models have been presented for the tectonic setting of the 160–155 Ma Zedong arc and associated Middle–Late Jurassic units: (1) the Zedong arc represents an intra-oceanic island arc that formed thousands of kilometres south of the Lhasa terrane in the Neotethys ocean (Aitchison et al. 2000, 2007a, b; McDermid et al. 2002; L. Wang et al. 2012); or (2) the Zedong arc represents the southern extent of the Gangdese arc (Zhang et al. 2014; Hu et al. 2016). All Zedong units formed during a pull in Gangdese arc magmatism (Orme et al. 2015 and references therein) and overlapped in time with the formation of the 160–150 Ma ophiolite. The geochemistry of mafic units from both the Jurassic and Cretaceous ophiolites is consistent with a supra-subduction zone spreading in a forearc, backarc or intra-arc setting (Aitchison et al. 2000, 2003; Huot et al. 2002; Malpas et al. 2003; Ziabrev et al. 2003; Abrajevitch et al. 2005; Guilmette et al. 2009, 2012; Xia et al. 2011; Hebert et al. 2012; Dai et al. 2013; An et al. 2014; Bao et al. 2014; Wu et al. 2014b; Chan et al. 2015; W. Huang et al. 2015; Liu et al. 2015; Maffione et al. 2015a; Baxter et al. 2016; Xiong et al. 2016, 2017).
Although the Zedong arc, Zedong basement and Xiazha Formation display many characteristics of an intra-oceanic arc setting, they can be explained without invoking another subduction zone. Trench retreat may have caused forearc extension along the southern margin of the Lhasa terrane, generating a marginal ocean basin in the upper plate and leading to southwards arc migration onto the newly created oceanic crust (Fig. 13b). As the arc swept south and then returned north to the Gangdese arc, ophiolitic rocks may have formed in a forearc, backarc or intra-arc setting. The Xiazha Formation may have
formed proximal to the Zedong arc in the forearc of a marginal basin rather than at a subduction zone thousands of kilometres from the Lhasa terrane. Forearc extension would have rifted most, if not all, of the older forearc region southwards (Fig. 13b). The magnitude of extension was sufficient to form the Zedong arc on newly formed oceanic crust in the upper plate either north or south of the spreading ridge (Fig. 13b). Invoking a separate subduction zone at this time is unnecessary to explain the geological setting of Late Jurassic units, does not explain the limited time span of these units (<15 myr) and requires removal of another forearc region in addition to that along the southern margin of Asia.

After c. 5 myr, magmatism terminated in the Zedong arc and migrated northwards back to the southern Lhasa terrane (Fig. 13c, d). Currently, the Zedong arc is exposed north of the ophiolite belt adjacent to the Gangdese arc, implying some forearc shortening that could have been accommodated along this contact by closure of the oceanic rift basin (Fig. 13d), removal of the pre-extension forearc at the trench by subduction erosion (Fig. 13c) or a combination of both these processes. Subduction complex rocks ranging in age from 159 to 134 Ma are exposed near Sangsang and, although the Zedong arc is only exposed near Zedong, the mélange sandstones contain abundant zircons of Zedong age found nowhere else in the Gangdese arc. No forearc basin strata of demonstrable 155–130 Ma age are exposed.

Cretaceous southern margin of Asia. Currently, the Xiazha Formation is exposed south of the ophiolite belt at the northern edge of the subduction complex, while the Zedong arc rocks are exposed between the ophiolite belt and the Gangdese arc, and thus the Xigaze forearc basin is most likely to have formed between them. Beginning at c. 130 Ma, another forearc extension event initiated between the Zedong arc and the Xiazha Formation, and extended enough to produce oceanic crust in the forearc (Fig. 13e). The Gangdese arc experienced another lull during forearc extension (Dai et al. 2013) but, unlike the Jurassic forearc extension event, no other arc units are documented. Xigaze forearc basin deposition on top of the ophiolite began shortly afterwards, as early as 129 Ma (based on detrital zircon ages) or late Barremian (based on radiolarians). Prior to accretion of the Bainang terrane during the Aptian (Matsuoka et al. 2002; Ziabrev et al. 2004), the pre-Cretaceous forearc region from arc to trench is suggested to have been consumed along a subduction interface, leaving only remnants of the Xiazha Formation. The amphibolite metamorphic sole of the ophiolite may represent basin closure, possibly by inversion of the spreading ridge in the upper plate (Guilmette et al. 2008, 2009). The remaining pre-Cretaceous forearc may have been subducted along the short-lived intra-basin subduction zone. Preservation of Jurassic sandstone in the northern mélange precludes subduction erosion along the Neotethyan trench. In the last few years, three new units (the Rongmawa, Luogangcuo and Xiazha formations) ranging in age from 159 to 80 Ma have been documented in the subduction complex, so it is quite possible that other dismembered remnants of the pre-Cretaceous margin are present within the subduction complex.

Forearc extension ceased at c. 120 Ma. The Xigaze forearc basin initially filled slowly with pelagic deposits (Chongdoi Formation) and then more rapidly with turbiditic deposits (Ngamring Formation) (Fig. 13f). Most of the robust MDAs from sandstones in the subduction complex are contemporaneous with this rapid filling phase (c. 100–80 Ma). Trench-fill sequences c. 88–80 Ma indicate high rates of sediment transport to the trench during relatively low convergence rates (Fig. 2). Beginning at c. 90–80 Ma, India–Asia convergence increased (van Hinsbergen et al. 2011; Gibbons et al. 2015) (Fig. 2). At c. 80 Ma, sediment supply to the forearc and trench decreased, although other ocean plate stratigraphy continued to accrete (Fig. 13g).

During initial subduction of Tethyan Himalayan strata in the trench at c. 59 Ma, Asian-affinity deposits entered the trench for the first time in about 25 myr (Fig. 13h). Deformation propagated southwards into the Tethyan Himalayan strata, forming the Tethyan thrust belt (Fig. 13i). During Miocene time, northwards thrusting along multiple strands of the GCT overprinted most of the original contacts in the suture zone.

Conclusions

Our new findings are summarized below:

- We identify four distinct units within the subduction complex: the bainang terrane, the Xiazha Formation, the siliciclastic-matrix mélange and the trench fill (Rongmawa Formation and Luogangcuo Formation). The Tethyan Himalayan strata are in places deformed to block-in-matrix mélange, but the lithologies are distinct from the siliciclastic-matrix mélange.
- The suture zone units are folded about east–west-trending fold axes and shortened across the north-directed Great Counter Thrust (GCT) system; one strand of this thrust locally transported Tethyan Himalayan strata >25 km northwards over the subduction complex (Fig. 3).
- We introduce the Xiazha Formation, which has a single U–Pb detrital zircon peak age at c. 158 Ma, has an undissected arc provenance, and contains clasts of intermediate to felsic volcanic rocks and ooids with both calcareous and volcanic
corres. Based on its age, sandstone petrographical signature and depositional environment, we interpret the Xiazha Formation to represent remnants of forearc strata that were deposited proximal to the Zedong arc and marginal to the Lhasa terrane.

- Latest Jurassic siliciclastic-matrix mélange blocks and matrix represent the oldest accretion of ocean plate stratigraphy exposed in the subduction complex. Any other forearc assemblages that developed during the previous c. 100 myr of Gangdese arc magmatism are not exposed – except for the Xiazha Formation, first documented in this study.

- The present geometry of suture zone units and our reconstructions (Figs 3 & 13) suggest that most of the pre-Cretaceous forearc of the Lhasa terrane was subducted during closure of the Late Jurassic or Early Cretaceous marginal basins.

- Sandstone blocks with dissected arc provenance and the Xiazha Formation have been identified in the northern margin of the Lhasa terrane.

- The majority of Asian-affinity sediment in the subduction complex is c. 100–80 Ma, and the youngest is 78 Ma, contemporaneous with shallow-marine deposition in the Xigaze forearc basin and increasing plate convergence rates. In contrast, ocean plate stratigraphy continued to accrete through the Paleocene. This suggests that the sediment supply to the trench may have abruptly decreased at c. 80 Ma.

The Rongmawa Formation, Luogangcuo Formation and the Xiazha Formation have been identified within the last decade, so it is plausible that more subduction complex units may be documented in the future and shed additional light on the subduction history along the southern margin of the Lhasa terrane.

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