

**Late Triassic orogenic assembly of the Tibetan Plateau: constraints from
magmatism and metamorphism in the east Lhasa terrane**

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22 Running title: Collisional orogeny of Paleo-Tethys ocean in the east Lhasa terrane

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24 Submitted to: *Journal of Petrology*

25 [July 12, 2022](#)~~[March 5, 2021](#)~~~~[February 20, 2021](#)~~

ABSTRACT

The early Mesozoic evolution of the Lhasa terrane, which represents a major component of the Himalayan-Tibetan orogen, remains highly controversial. In particular, geological units and events documented either side of the eastern Himalayan syntaxis (EHS) are poorly correlated. Here, we report new petrological, geochemical and geochronological data for co-genetic peraluminous S-type granites and metamorphic rocks (gneiss and schist) from the Motuo–Bomi–Chayu region of the eastern Lhasa terrane, located on the eastern flank of the EHS. Zircon U–Pb dating indicates that these units record both Late Triassic magmatic (216–206 Ma) and metamorphic (209–198 Ma) episodes. The granites were derived from a Paleoproterozoic crustal source with negative zircon $\varepsilon_{\text{Hf}}(t)$ values (–5.5 to –16.6) and T_{DM2} model ages of 1.51–1.99 Ga, and are interpreted to have formed by crustal anatexis of nearby metasediments during collisional orogeny and crustal thickening. The gneisses and schists experienced similar upper amphibolite-facies peak metamorphism and associated partial melting, followed by decompressional cooling and retrograde metamorphism. These rocks were buried to lower-crustal depths and then exhumated to the surface in a collisional orogenic setting during plate convergence. From comparison of these data ~~These data are then compared~~ to other metamorphic belts with similar grades and ages, and association of associated ~~associated~~ coeval granitic magmatism widespread in the central-east Lhasa terrane. ~~As a result,~~ we propose that the studied co-genetic magmatism and metamorphism in the Motuo–Bomi–Chayu region records Late Triassic accretion of the North Lhasa and South Lhasa terranes, which represents the first evidence of the Paleo-Tethys ocean (PTO) closure

in this part of Asia. These data provide new constraints on the spatial and temporal evolution of the Paleo-Tethyan Wilson Cycle and provide a ‘missing link’ to correlate the geology and tectonic history of the Lhasa terrane continental crust on either side of the EHS.

Keywords: collisional-orogenesis; east Lhasa terrane; magmatism and metamorphism; Paleo-Tethys Ocean; Zircon U–Pb geochronology

INTRODUCTION

The Himalayan Range and Tibetan Plateau formed due to collision between the Indian and Asian plates and closure of the Neo-Tethys Ocean at c. 50 Ma (Tapponnier *et al.*, 1986; O’Brien *et al.*, 2001; Najman *et al.*, 2010; Zhang *et al.*, 2012a; St-Onge *et al.*, 2013; Ding *et al.*, 2016), although this orogeny merely represents the youngest of multiple terrane-accretion events that have occurred along the southern margin of Eurasia since the Early Paleozoic (e.g. Kapp *et al.*, 2007; Yin & Harrison, 2000; Zhang *et al.*, 2014a). From north to south, the Tibetan Plateau includes the Songpan-Ganzi, Qiangtang, Lhasa, and Himalaya terranes, which are separated by the Jinsha, Bangong-Nujiang and Indus-Yarlung Tsangpo suture zones, respectively (Fig. 1a; Burg and Chen, 1984; Xu *et al.*, 1985, 2006, 2015; Searle *et al.*, 1987; Dewey *et al.*, 1988; Murphy *et al.*, 1997; Mo *et al.*, 2003, 2005, 2006, 2007, 2008; Pan *et al.*, 2004, 2006, 2012; Zhang *et al.*, 2012b, c, 2013, 2014a, b, c; Zhu *et al.*, 2013, 2015, 2016; Palin *et al.*, 2014, 2015; Ding *et al.*, 2015, 2016). The Lhasa

terrane records geological evidence of Paleoproterozoic magmatism (Zhang & Santosh, 2012; Lin *et al.*, 2013a), Mesoproterozoic metamorphism (Lin *et al.*, 2013a), opening of the Mozambique Ocean during the Neoproterozoic (Dong *et al.*, 2011a; Zhang *et al.*, 2012b, 2014a), assembly of Gondwana and subduction of the Proto-Tethyan ocean during the Early Paleozoic (Li *et al.*, 2008; Dong *et al.*, 2009; Ji *et al.*, 2009; Zhu *et al.*, 2012, 2013; Hu *et al.*, 2013; Ding *et al.*, 2015), subduction and closure of the Paleo-Tethys ocean (PTO) from the Permian to the Triassic (Yang *et al.*, 2006, 2007, 2009; Li *et al.*, 2009a, b, 2012; Zeng *et al.*, 2009; Dong *et al.*, 2011b, 2015; Lin *et al.*, 2013b; Cheng *et al.*, 2012, 2015; Weller *et al.*, 2015, 2016a, b; Chen *et al.*, 2017), and finally the formation and destruction of the Bangong-Nujiang Tethyan and Neo-Tethyan oceans during the Mesozoic and Cenozoic, respectively (Allègre *et al.*, 1984; Ding *et al.*, 2003; Mo *et al.*, 2003, 2005, 2006, 2007, 2008; Hou *et al.*, 2004, 2006; Zhu *et al.*, 2009, 2011, 2012, 2013, 2015, 2016; Zhang *et al.*, 2010, 2013, 2014c; Zhang & Santosh, 2012; Guan *et al.*, 2012; Palin *et al.*, 2014, 2015; Shui *et al.*, 2017).

Understanding the evolution of the PTO is significant for deciphering the tectonothermal history of the Lhasa terrane – and thus the Tibetan Plateau as a whole – as the PTO suture zone divides its central-east portions into northern and southern blocks with distinct geological histories (Fig. 1a; Yang *et al.*, 2006, 2007, 2009; Zhang *et al.*, 2014a; Cheng *et al.*, 2015; Weller *et al.*, 2015, 2016a; Chen *et al.*, 2017). Previous studies of Permian to Early Jurassic magmatic and high-grade metamorphic rocks located in central-east Lhasa terrane (westward of the eastern Himalayan syntaxis) suggest PTO

subduction and ocean-closure in this region (Kapp *et al.*, 2005; Yang *et al.*, 2006, 2007, 2009; Li *et al.*, 2009a, b, 2012; Zeng *et al.*, 2009; Zhang *et al.*, 2014a; Cheng *et al.*, 2015; Weller *et al.*, 2015, 2016a, b; Chen *et al.*, 2017). However, it is still unclear (1) if these events continued into the east Lhasa terrane, eastward of the eastern Himalayan syntaxis (EHS); (2) where the actual location of the PTO orogenic belt is in this region; and (3) the relative geochronology-age of metamorphic and magmatic events in this area. Together, this these represents an important gap in our understanding of the early evolution of this major orogenic system.

In this paper, we present new petrological, geochemical, and geochronological data for co-genetic amphibolite-facies metasediments and S-type granites from the Motuo–Bomi–Chayu region of the east Lhasa terrane, eastern flank of the EHS, which place new constraints on the timing and style of collisional orogeny and PTO closure during the Early Mesozoic. These integrated data represent the first robust constraints on the location and timing of North-South Lhasa microblock accretion east of the EHS and provide new constraints for tectonic reconstructions of terrane evolution in southeast Asia prior to India-Asia collision.

GEOLOGICAL BACKGROUND AND SAMPLES

The east-west oriented Lhasa terrane, southern Tibet, is 100–300 km wide and over 2000 km long (Fig. 1a). It is composed dominantly of Precambrian crystalline metamorphic basement overlain by Paleozoic to Mesozoic marine strata, volcanic rocks and

metasediments, and is intruded by Mesozoic and Cenozoic plutons (e.g. Yin & Harrison, 2000; Pan *et al.*, 2004, 2006, 2012; Metcalfe, 2006; Zhu *et al.*, 2009, 2011, 2012, 2016; Zhang *et al.*, 2010, 2012b, 2013, 2014a). The presence of Precambrian basement is proved shown by the discovery of Neoproterozoic granitic and mafic rocks and amphibolite- to granulite-facies metamorphic rocks in northern block, and Paleo- to Meso-Proterozoic granitic rocks in southern block, although these rocks were also further metamorphosed during the Neoproterozoic (Dong *et al.*, 2011a, 2020; Zhang *et al.*, 2012b, 2014a; Lin *et al.*, 2013a; Hu *et al.*, 2019 and references therein).

The study region discussed here is located in the easternmost segment of the Lhasa terrane, east/southeast of Namche Barwa, near to the towns of Motuo, Bomi, and Chayu (Fig. 1b). This area is characterized by regionally metamorphosed greenschist- to amphibolite-facies lithologies, Late Paleozoic to Cenozoic sediments, Early Mesozoic intrusions, and the exposed roots of the Gangdese Batholith (Fig. 1b). These metamorphic rocks consist mainly of orthogneiss, schist, marble, migmatite and amphibolite, which together are referred to as the Bomi Group and Demala Group (Peng *et al.*, 1999; Xie *et al.*, 2007; Dong *et al.*, 2011c, 2015). The radiogenic ages of 2264–2145 Ma, 1330–900 Ma and 600–520 Ma obtained by traditional dating methods show that remnants of Precambrian metamorphic basement are locally preserved (Peng *et al.*, 1999; Xie *et al.*, 2007; Dong *et al.*, 2011c, 2015), although recent metamorphic zircon U–Pb ages of ca. 217 Ma and ca. 22–16 Ma suggest that these rocks also experienced Mesozoic and Cenozoic thermal overprinting (Dong *et al.*, 2011c, 2015). Late Paleozoic sediments, Mesozoic marine sediments, and

minor Cenozoic sedimentary rocks mainly occur in the northeast, although Late Paleozoic sediments do not contact the Mesozoic strata and the Cenozoic sedimentary rocks unconformably overlie the Mesozoic sediments, indicating punctuated tectonics in this region. The Late Paleozoic sedimentary strata comprise Devonian marine sediments and Carboniferous-Permian volcanic rocks interbedding in marine clastic rocks, running NW-SE orientation in this region (Pan *et al.*, 2004; Wang *et al.*, 2008). Carboniferous, Triassic and Jurassic granites were recently recognized in this area, although these rocks have been partly transformed into orthogneiss during Mesozoic and Cenozoic thermal metamorphic overprinting (Pan *et al.*, 2004; Li *et al.*, 2013a; Dong *et al.*, 2015). The Gangdese batholith is predominantly composed of Cretaceous to Neogene granitoids, which formed as continental arc magmas during subduction and closure of the Neo-Tethyan ocean (Chiu *et al.*, 2009; Dong *et al.*, 2013; Li *et al.*, 2013b).

The samples documented here comprise magmatic and metamorphic rocks collected from two regions of the east Lhasa terrane (east of the EHS), but also located ~150 km apart: granites and metapelitic schists were collected from ~35 km southwest of Chayu, and metapelitic gneisses were collected from ~25 km southwest of Bomi (Fig. 1b). This region has been relatively understudied in comparison with outcrops west of the EHS and so the degree to which geological units and tectonic events correlate either side of the EHS is uncertain. Outcrop and petrological information for each sample is shown in Table 1. Mineral abbreviations are after Whitney and Evans (2010).

PETROLOGY AND MINERAL CHEMISTRY ANALYTICAL METHODS

Mineral composition

Mineral compositions were acquired using a JEOL JXA 8900 electron microprobe (EMP) housed at the Institute of Geology, Chinese Academy of Geological Science (CAGS), Beijing. Operating conditions comprised a 15-kV accelerating voltage, 5-nA beam current, 5-~~μmm~~ probe diameter, and count time of 10 s for peak and background. Natural garnet, biotite, plagioclase, and K-feldspar and synthetic silica standards were used as standards for calibrations and a ZAF correction was carried out. Analytical uncertainties for SiO₂, TiO₂, Al₂O₃, FeO, MnO, MgO, CaO, Na₂O, K₂O, and total are <1% at abundances >1 wt. % and <8% at abundances <1 wt. %. Compositional data collected for garnet, biotite, plagioclase, K-feldspar, and cordierite in all rock types are given in Supplementary Tables 1–5.

Whole-rock composition

All magmatic samples collected from the Chayu area were analysed for major and trace element contents, which are shown in Supplementary table 6. Whole-rock compositions were obtained at the National Research Centre for Geoanalysis, CAGS, Beijing. Standards GBW07103, GBW07121, and GBW07122 were used to monitor analytical quality control. Major-elements oxides, including loss on ignition (LOI), were analysed by X-ray fluorescence (XRF) on a Rigaku-3080 analyser, which has an analytical uncertainty of <0.5%. Concentrations of trace elements Zr, Nb, Cr, Sr, Ba, Ni, Rb and Y were determined

173 using a Rigaku-2100 XRF analyser, which has an analytical uncertainty of <3–5%. Other
174 trace elements and rare earth elements (REEs) were determined by inductively coupled
175 plasma mass spectrometry (ICP-MS) using a TJA-PQ-ExCell. Detailed description of the
176 ICP-MS method has been reported by Liang *et al.* (2000). Analytical uncertainties for
177 ICP-MS are 1–5% at abundances >1 ppm and 5–10% at abundances <1 ppm.

179 **Zircon U–Pb and Hf isotopes and trace element analysis**

180 Radiogenic isotope geochronology was performed on four of granite samples and all
181 metasediment samples to determine the timing of key tectonothermal events in the
182 Motuo–Bomi–Chayu region of the east Lhasa terrane. Zircon grains were separated from each
183 sample by magnetic and conventional heavy-liquid techniques at the Hebei Institute of
184 Regional Geology and Mineral Investigation. Cathodoluminescence (CL) images were taken
185 on a HITACHI S2250-N scanning electron microscope at the SHRIMP Unit at the Institute of
186 Geology, CAGS. Zircon U–Pb isotope and trace element analysis were performed
187 simultaneously on an Agilent 7500 ICP-MS equipped with a 193 nm ArF-excimer laser at the
188 State Key Laboratory of Geological Processes and Mineral Resources, China University of
189 Geosciences (Wuhan). Detailed operating conditions for the ICP-MS instrument and the laser
190 ablation system are as reported by Liu *et al.* (2010). Zircon 91500 was used as external
191 standard for U–Pb dating, which was analysed twice for every 5 analyses of the samples.
192 SRM610 were used as external standard for the trace element analysis. Time-dependent drifts
193 of U–Th–Pb isotopic ratios were corrected using a linear interpolation (with time) for every

five analyses according to the variations of zircon 91500. Zircon GJ-1 was used as standard to monitor the stability and accuracy of the U–Pb data acquired. Silicon (^{29}Si) was used as an internal standard. The U–Pb and trace element data were processed by ICPMSDataCal (Liu *et al.*, 2010) and Isoplot (Lugwig, 2003) was used to calculate isotopic ages and construct concordia diagrams. All LA-ICP-MS U–Pb data and the trace element compositions of the magmatic and metamorphic zircons from the granites and metasediments are presented in Supplementary table 7.

In situ Hf isotope compositions of zircon were obtained by a Neptune MC-ICP-MS at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences (Wuhan). The laser had a beam diameter of 44 μm , a frequency of 8 Hz, energy of 60 mJ, and a fluence of 5.3 J/cm². Analytical spots were chosen on the same domains with LA-ICP-MS spots. The zircon standards GJ-1 (Elhlou *et al.*, 2006) and 91500 (Blichert, 2008) were analysed as reference materials, and yielded weighted mean $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of 0.282012 ± 11 (1 σ , n = 4) and 0.282305 ± 10 (1 σ , n = 10), respectively. These values match the recommended values (0.282015 ± 19 for GJ-1, Elhlou *et al.*, 2006; 0.282308 ± 6 for 91500, Blichert, 2008). The initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios were calculated using the ^{176}Lu decay constant of $1.867 \times 10^{-11}/\text{yr}$ (Soderlund *et al.*, 2004). The $\epsilon_{\text{Hf}}(t)$ values were calculated using a chondritic $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282785 and $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.0336 (Bouvier *et al.*, 2008). The depleted-mantle-Hf model ages (T_{DM}) were calculated with respect to a depleted present-day mantle $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.28325 and $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.0384 (Griffin *et al.*, 2000). The Hf crustal model ages (T_{DM2}) of each zircon were

calculated by assuming its parental magma to have been derived from an average continental crust with $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$ (Griffin *et al.*, 2002). *In-situ* Lu–Hf isotopic ratios from two granites are presented in Supplementary table 8.

RESULTS

Magmatic rocks**Petrology**

Magmatic rocks

Magmatic rocks collected from the Chayu area include undeformed granite (samples T15–46–7, T15–46–8 and T15–46–9) and weakly deformed granite (samples T15–46–1, T15–46–3, T15–46–4, T15–46–5, T15–47–2 and T15–47–4). The undeformed granite displays a phaneritic texture and contains quartz (40–50%), K-feldspar (20–30%), plagioclase (20–25%), biotite (2–3%) and muscovite (2–3%), with accessory zircon, apatite, and sphene (Fig. 2a and c). The weakly deformed granites display a slight foliation in outcrop and consist of quartz (40–45%), K-feldspar (20–25%), plagioclase (25–30%), biotite (3–5%) and variable muscovite, garnet, and accessory minerals zircon and apatite (Fig. 2b and d). The finely crystallised K-feldspar grains are wrapped by quartz, feldspar and biotite grains, which together define a weak foliation (Fig. 2d).

Metamorphic rocks

Metamorphic rocks

Two high-grade gneisses (T15–32–3 and T15–33–7) were collected from southwest of

Bomi. They show prominent foliation in outcrop and comprise pale plagioclase-rich (Pl-rich) quartzofeldspathic leucosomes, which are generally aligned subparallel to the foliation, and dark melanosomes. The (Pl-rich) leucosomes range from millimeters to centimeters in thickness and often occur as interlayers or small- to large-scale lenses (Fig. 3a). This is diagnostic of partial melting during metamorphism and locally migmatization. Sample T15–32–3 is a metapelitic gneiss containing biotite, muscovite, quartz, plagioclase, and garnet, with minor sillimanite, ilmenite and chlorite (Fig. 4a and b). Large porphyroblasts of garnet are set in a matrix with a foliation defined by oriented mica and quartzofeldspathic domains (Fig. 4a). Chlorite partially pseudomorphs biotite and garnet (Fig. 4a), suggesting a retrograde origin. Sample T15–33–7 has a similar mineralogy to sample T15–32–3, but contains additional K-feldspar (Fig. 4c–f). In sample T15–33–7, muscovite occurs as both large lath-like porphyroblasts and randomly orientated fine-grained mattes (Fig. 4d), which suggest a retrograde origin (Ashworth, 1975, 1979; Tyler & Ashworth, 1982). Cuspate quartz also exhibits very small dihedral angles against plagioclase (Fig. 4e), and K-feldspar grains are surrounded by plagioclase rims (Fig. 4f). These microstructural features suggest the presence of partial melt during metamorphism (e.g. Holness & Clemens, 1999; Holness & Sawyer, 2008; Feisel *et al.*, 2018).

Two metapelitic schists (T15–43–1 and T15–43–3) were collected from southwest of Chayu. They have the appearance of stromatic migmatitic rocks and consist of millimeter to centimeter thick white-beige felsic domains (leucosomes), which occur as concordant layers or small-scale rootless folds, alternating with grey domains of fine-grained

257 quartzo-feldspathic material with interstitial biotite (Fig. 3b). These outcrop features imply
258 that these rocks experienced partial melting and the melt did not migrate over long distances.
259 Sample T15-43-1 is a garnet-two-mica schist containing garnet porphyroblasts with biotite,
260 muscovite, plagioclase, quartz and minor ilmenite as matrix minerals (Fig. 5a and b).
261 Muscovite occurs both as flakes that are aligned within the foliation and as larger,
262 unfoliated and subhedral to euhedral grains against garnet rims (Fig. 5a and b). These
263 microstructural relationships indicate two generations of muscovite growth (cf. Ashworth,
264 1975, 1979; Tyler & Ashworth, 1982). Garnet rims and internal fractures are occasionally
265 replaced by aggregates of biotite, muscovite, and plagioclase (Fig. 5a and b), indicating
266 minor retrogression. Sample T15-43-3 is a garnet-sillimanite-cordierite schist containing
267 biotite, plagioclase, K-feldspar, quartz, sillimanite, garnet and cordierite (Fig. 5c-f). The
268 foliation is defined by aligned fibrous/prismatic sillimanite and biotite flakes that wrap
269 around coarse K-feldspar, plagioclase and quartz in the matrix. Garnet porphyroblasts have
270 inclusion-rich cores, containing biotite, plagioclase and quartz, whereas rim domains are
271 mostly inclusion-free (Fig. 5d and f). Some garnet porphyroblasts are partly pseudomorphed
272 at their rims by biotite and cordierite (Fig. 5d-f). Minor perthite occurs in the matrix,
273 exhibiting micro-exsolved lamellae of plagioclase hosted by K-feldspar (Fig. 5e), and
274 polymineralic inclusions of biotite + plagioclase + quartz are occasionally preserved in
275 garnet cores (Fig. 5f), suggesting the studied rocks underwent partial melting (e.g., Holness
276 & Clemens, 1999; Waters, 2001; Holness & Sawyer, 2008; Zhang *et al.*, 2017; Feisel *et al.*,
277 2018). In all samples, garnet rims are partly replaced by biotite + plagioclase \pm muscovite

aggregates (Figs. 4a–c and 5a,b,d), recording back-reaction involving melt crystallization (e.g., Waters, 2001; Kriegsman & Alvarez-Valero, 2010).

Mineral chemistry

Bomi gneiss

Bomi gneiss samples T15–32–3 and T15–33–7 contain garnet porphyroblasts of composition $X_{\text{Fe}76-78}X_{\text{Mg}10-12}X_{\text{Ca}3-5}X_{\text{Mn}8-9}$ ($X_{\text{Fe}} = \text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mn} + \text{Mg} + \text{Ca})$, with X_{Ca} , X_{Mg} and X_{Mn} defined according to X_{Fe}) and $X_{\text{Fe}71-73}X_{\text{Mg}14-16}X_{\text{Ca}4-5}X_{\text{Mn}8-9}$, respectively (Supplementary table 1). Porphyroblasts in both samples thus lack systematic compositional zonation from core to rim. Matrix biotite has an X_{Mg} [= $\text{Mg}/(\text{Mg} + \text{Fe})$] of 0.35–0.42 in sample T15–32–3 and 0.45–0.49 in T15–33–7. The Ti contents of each are 0.15–0.18 and 0.21–0.23 cations per formula unit (cpfu) based on an 11-oxygen calculation, respectively (Supplementary table 2). Plagioclase in sample T15–32–3 is oligoclase with X_{Ca} [= $\text{Ca}/(\text{Ca} + \text{K} + \text{Na})$] of 0.18–0.23, and also oligoclase with $X_{\text{Ca}} = 0.26$ –0.29 in sample T15–33–7, whereas those surrounding K-feldspar are albite with $X_{\text{Ca}} = 0.02$ –0.03 (Supplementary table 3), and K-feldspar has X_{Na} [= $\text{Na}/(\text{Ca} + \text{K} + \text{Na})$] of 0.11–0.12 (Supplementary table 4).

These petrographic observations and patterns in mineral chemistry suggest that the Bomi gneisses contained a peak metamorphic mineral assemblage of Grt + Sil + Bt + Ms + Pl + Qz \pm Ilm + melt for sample T15–32–3 and Grt + Sil + Bt + Kfs + Pl + Qz + melt for sample T15–33–7. Retrograde cooling and exhumation likely initiated growth of secondary Ms, Chl, and Ab surrounding some K-feldspar grains.

300 *Chayu schist*

301 Garnet porphyroblasts from Chayu schist samples contain similar compositional
 302 characteristics, with homogenous broad cores and weakly zoned rims. Cores in sample
 303 T15-43-1 have composition $X_{Fe72-74}X_{Mg15-16}X_{Ca3-4}X_{Mn8-9}$, which change to outer-rim
 304 compositions $X_{Fe73-74}X_{Mg12-14}X_{Ca3-4}X_{Mn9-12}$ (Supplementary table 1; Fig. 6a). By contrast,
 305 garnet porphyroblasts in sample T15-43-3 have core compositions of
 306 $X_{Fe73-74}X_{Mg11-12}X_{Ca2-3}X_{Mn11-13}$, and rim compositions of $X_{Fe70-72}X_{Mg10-11}X_{Ca2-3}X_{Mn15-19}$
 307 (Supplementary table 1; Fig. 6b). These patterns are diagnostic of diffusion-driven
 308 homogenization at peak metamorphic conditions, followed by diffusion-controlled
 309 retrograde resorption during exhumation, which led to compositional inflections in outer
 310 rim domains (e.g. Florence & Spear, 1991; Spear, 1991, 1993; Kohn & Spear, 2000;
 311 Caddick *et al.*, 2010).

312 In sample T15-43-1, biotite grains in the matrix and adjacent to garnet rims have X_{Mg}
 313 values of 0.50–0.51 and 0.48–0.52, respectively, although the former have higher Ti
 314 contents (~0.08) than the latter (0.01–0.04) (Supplementary table 2; Fig. 6c). These
 315 compositional patterns confirm minor breakdown of garnet during cooling from peak
 316 conditions, leading to biotite formation during retrogression. Plagioclase grains in the
 317 matrix and adjacent to garnet rims have similar compositions $X_{Ca} = 0.23–0.30$,
 318 (Supplementary table 3). In sample T15-43-3, biotite inclusions within garnet cores have
 319 relatively high Ti contents of 0.16–0.18 cpdf and low X_{Mg} values of 0.46–0.49, whereas

biotites in the matrix have a relatively low Ti content (0.12–0.15 cpfu) and high X_{Mg} values of 0.49–0.51 (Supplementary table 2; Fig. 6c). Plagioclase in the matrix exhibits $X_{Ca} = 0.17–0.24$ (Supplementary table 3) and K-feldspar in the matrix has $X_{Na} = 0.14–0.20$ (Supplementary table 4). Cordierite grains adjacent to garnet rims have X_{Mg} values of 0.59–0.60 (Supplementary table 5).

These petrographic observations and patterns in mineral chemistry suggest that the Chayu schists preserve petrographic evidence of two stages of metamorphism. In sample T15–43–1, the peak metamorphic stage is defined by garnet and foliation-forming matrix minerals (Grt + Bt + Ms + Pl + Qz \pm Ilm) and retrograde metamorphism induced compositional inflections in garnet outer rims (e.g. Mn spikes; cf. Kohn & Spear, 2000) and peripheral breakdown to form secondary muscovite, biotite and plagioclase. In sample T15–43–3, peak metamorphism is characterized by garnet cores and their associated inclusions and the matrix minerals Sil + Kfs + Bt + Pl + Qz. Akin to sample T15–43–1, retrograde cooling induced garnet rims resorption and recrystallization to biotite and plagioclase, although here additional cordierite formed during decompression. These mineral assemblages and reaction textures indicate that all samples experienced a similar tectonothermal evolution, with changing P – T conditions quantified using thermobarometry (see below).

WHOLE-ROCK GEOCHEMISTRY

~~All magmatic samples collected from the Chayu area were analysed for major and trace~~

341 ~~element contents, which are shown in Supplementary table 6. Whole rock compositions~~
342 ~~were obtained at the National Research Centre for Geoanalysis, CAGS, Beijing.~~
343 ~~Major elements oxides, including loss on ignition (LOI), were analysed by X ray~~
344 ~~fluorescence (XRF) on a Rigaku 3080 analyser, which has an analytical uncertainty of~~
345 ~~<0.5%. Concentrations of trace elements Zr, Nb, Cr, Sr, Ba, Ni, Rb and Y were determined~~
346 ~~using a Rigaku 2100 XRF analyser, which has an analytical uncertainty of <3–5%. Other~~
347 ~~trace elements and rare earth elements (REEs) were determined by inductively coupled~~
348 ~~plasma mass spectrometry (ICP-MS) using a TJA PQ ExCell. Analytical uncertainties for~~
349 ~~ICP-MS are 1–5% at abundances >1 ppm and 5–10% at abundances <1 ppm.~~

350 **Whole-rock geochemistry**

351 All granites have high SiO₂ (69.55–76.12 wt. %), Al₂O₃ (12.33–14.58 wt. %) and Na₂O
352 + K₂O (4.36–8.59 wt. %) contents, and low Fe₂O₃^T (0.64–4.24 wt. %), MgO (0.11–2.96
353 wt. %), CaO (0.73–2.50 wt. %) and MnO (0.02–0.07 wt. %) contents (Supplementary table
354 6). Most are weakly peraluminous with A/CNK values of ~1.08, whereas sample T15–46–1
355 and T15–46–4 has an atypically higher value of 1.21 and 1.20 respectively (Supplementary
356 table 6).

357 On a primitive mantle-normalized spider diagram (Fig. 7a), all samples are enriched in
358 large ion lithophile elements (LILE), such as Rb, Th, U and K, and depleted in some high
359 field strength elements (HFSE), such as Nb, Ta, Sr and Ti. In addition, Ba is strongly
360 depleted compared to Rb and Th. When normalized to chondritic values, all granites are
361 strongly enriched and show similar fractionated REE patterns (LREE/HREE > 4.64,

(La/Yb)_N = 5.45–46.36), with LREE-enrichment, HREE-depletion, and negative Eu anomalies (δEu = 0.32–0.63), although sample T15–46–7 is only moderately enriched (Supplementary table 6; Fig. 7b).

Zircon U–Pb ages and Hf isotope

Zircons from granites T15–46–3, T15–46–5, T15–46–7, and T15–47–2 show similar internal structures, are colourless to pale brown, euhedral to subhedral prismatic in form (70–170 μm long), and have regular oscillatory zoning with inherited detrital cores in CL images (Fig. 9a–d). Most of the analysed zircon spots have relatively high Th/U ratios (>0.2) and REE contents, and are characterized by fractionated REE patterns with LREE depletion, HREE enrichment, and negative Eu anomalies (Supplementary table 7; Fig. 10a). The internal structure and compositional features suggest a magmatic origin (Hoskin and Schaltegger, 2003). Zircon data from these four samples yielded near-consistent $^{206}\text{Pb}/^{238}\text{U}$ ages, with weighted mean ages ranging from 216 Ma to 206 Ma (Fig. 11a–d). These ages are thus interpreted to represent magma emplacement/crystallization age.

Zircons from the gneiss and schist samples T15–32–3, T15–33–7, T15–43–1, and T15–43–3 are mostly colourless, oval or sub-rounded in shape, and show well-preserved core-rim zonation (Fig. 9e–f). Inherited cores have a variable size, irregular form and random zoning, implying that zircon cores are detrital/xenocrystic and the protoliths of the studied samples are sedimentary rocks. In contrast, metamorphic rims show weak patchy zoning (or have no zoning at all), and have relatively low Th/U ratios, low REE contents

and fractionated REE patterns with depleted LREE, flat HREE and slightly negative Eu anomalies (Supplementary table 7; Fig. 10b), typical of a metamorphic origin (Hoskin & Schaltegger, 2003). The analysed spots of the metamorphic rims yielded near-concordant $^{206}\text{Pb}/^{238}\text{U}$ ages, with weighted means of 198 ± 3.3 Ma, 202 ± 3.4 Ma, 209 ± 2.8 Ma and 203 ± 3.1 Ma for T15-32-3, T15-33-7, T15-43-1, and T15-43-3, respectively (Fig. 11e-h).

Together, these data are interpreted to record Late Triassic crustal magmatism at 216–206 Ma and regional metamorphism at 209–198 Ma in the east Lhasa terrane.

Zircons from two Late Triassic granites of the east Lhasa terrane (samples T15-46-3 and T15-46-5) have initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios ranging from 0.282308 to 0.282525 and variable negative $\epsilon_{\text{Hf}}(t)$ values (–8.6 to –16.2). These values correspond to T_{DM2} model ages of 1.51–1.99 Ga (Supplementary table 8; Fig. 12), indicating a Paleoproterozoic crustal magma source.

THERMOBAROMETRY DISCUSSION

Metamorphic P – T path of gneiss and schist

In order to constrain the tectonothermal evolution of metamorphic samples, both phase diagram-based thermobarometry and conventional techniques were employed to quantify P – T conditions for various stages of their prograde, peak and retrograde metamorphic evolutions.

In general, bulk-rock-specific phase diagrams (pseudosections) offer notably more precise constraints on P or T conditions of metamorphism than conventional techniques (cf.

404 Powell & Holland, 2008), thus allowing discrete changes in crustal depth and thermal state
405 during the burial and exhumation cycle to be defined (e.g. Weller *et al.*, 2015; Palin *et al.*,
406 2018). Typical uncertainties associated with cation exchange thermometers and net transfer
407 barometers are at least ± 50 °C and ± 1 kbar at 1. S.D., which is primarily a function of
408 uncertainty on thermodynamic end-member data and inaccuracies in the description of
409 activity-composition (a - x) relations describing elemental mixing in solid solutions (e.g.
410 Green *et al.*, 2016; Waters, 2019). While uncertainty on the absolute positions of
411 assemblage field boundaries on any individual pseudosection are of a similar magnitude to
412 those for conventional techniques (Powell & Holland, 2008; Palin *et al.*, 2016), as all phase
413 diagrams in this study were calculated using the same thermodynamic dataset and a - x
414 relations, similar absolute errors cancel, and the calculated phase equilibria shown below
415 are relatively precise to within ± 0.2 kbar and ± 10 – 15 °C (Worley & Powell, 2000). Such
416 values allow distinct differences between P - T paths to be determined, allowing
417 discrimination of depths of burial and calculated geotherms between localities to be
418 performed with confidence (e.g. Hernández-Uribe *et al.*, 2018; Hernández-Uribe & Palin,
419 2019; Li *et al.*, 2018). Nonetheless, conventional techniques must be employed in scenarios
420 where the bulk-rock composition of a rock is not representative of reaching ‘equilibration’
421 volume, such as for retrograde breakdown within porphyroblast strain shadows (Palin *et al.*,
422 2013; Waters, 2019). As such, the conditions of late-stage retrograde metamorphism were
423 identified using conventional techniques explicitly considering the minerals that formed
424 during these events: not the surrounding peak metamorphic assemblage.

Phase equilibria modelling

Phase equilibria modelling

All pseudosections were calculated using Perple_X (Connolly, 2005; version 6.8.4) and internally consistent thermodynamic dataset ds-55 of Holland and Powell (1998) in the system $\text{MnO-Na}_2\text{O-CaO-K}_2\text{O-FeO-MgO-Al}_2\text{O}_3\text{-SiO}_2\text{-H}_2\text{O-TiO}_2$ (MnNCKFMASHT). The following $a-x$ relationships were employed: biotite (Tajčmanová *et al.*, 2009), plagioclase and K-feldspar (Holland & Powell, 2003), garnet, muscovite, cordierite, staurolite, chlorite, ilmenite and silicate-melt (White *et al.*, 2014). Kyanite, sillimanite, quartz, and rutile are treated as pure phases. Pseudosections were calculated using XRF-derived bulk-rock compositions (Supplementary table 6) and fluid was considered as pure H_2O . All iron was considered as ferrous due to the lack of magnetite and other Fe^{3+} -rich minerals in the schists and gneisses.

Quantitative P - T pseudosections calculated for gneisses T15-32-3 between 600–800 °C and 4–9 kbar and T15-33-7 between 600–850 °C and 3–9 kbar are shown in Fig. 8a and b, respectively. Quartz and plagioclase are present throughout this P - T range and garnet is stable in both except for low- P conditions, where cordierite is stable instead. The measured H_2O contents for each sample saturate the solidus in each case, which is located at ~660–690 °C (Fig. 8a and b).

For sample T15-32-3, the observed peak mineral assemblage of Bt + Ms + Pl + Grt + Sil + Qz + melt is stable in a relatively broad field spanning 4.9–8.7 kbar and 670–745 °C

(Fig. 8a), although isopleths for garnet inner-rim composition ($X_{Mg} = 0.10\text{--}0.12$) constrain peak conditions within this range to a minimum pressure of ~6 kbar (pink field in Fig. 8a). Retrograde effects where garnet compositions have been modified by diffusional cation exchange with matrix phases are not considered here when determining peak metamorphic conditions. As such, sample T15-32-3 likely equilibrated at 5.7–7.5 kbar and 675–725 °C. For sample T15-33-7, the observed peak mineral assemblage of Grt + Sil + Kfs + Bt + Pl + Qz + melt is calculated to be stable between 3.2–9.6 kbar and 675–820 °C (Fig. 8b). This range was reduced by matching observed and calculated garnet compositions X_{Mg} (0.14–0.16), X_{Ca} (0.04–0.05) and biotite Ti contents (0.20–0.23 cpfu), as Ca in garnet and Ti in biotite have slow diffusivities; thus, their measured concentrations should represent those obtained at peak metamorphism. Isopleths delineating these ranges constrain peak P – T conditions to 5.3–6.7 kbar and 750–765 °C (pink field in Fig. 8b).

The calculated P – T pseudosection for garnet-biotite schist sample T15-43-1 shows that quartz, plagioclase and biotite are stable throughout the calculated P – T range, and garnet is stable everywhere except for <3.1 kbar and 675–685 °C. The fluid-saturated solidus is located at 670–690 °C, and the muscovite-out/sillimanite-in reaction has a positive slope between 630 °C and 750 °C. The observed peak-metamorphic assemblage (Grt + Bt + Ms + Pl + Qz + melt) is stable at 5.4–10 kbar and 675–730 °C, and garnet core X_{Mg} and plagioclase X_{Ca} isopleths constrain peak P – T conditions to be >7.9 kbar and 690–720 °C (pink field in Fig. 8c). Retrograde alteration at garnet rims, forming coarse aggregates of biotite, plagioclase, and muscovite, but no sillimanite, constrains immediate

post-peak retrogression to have a steep angle in P – T space. If heating had continued, or decompression had occurred isothermally, sillimanite would be expected to form within strain shadows, although this is not observed. Sample T15–43–3 contains sillimanite in the matrix, but is otherwise mineralogically similar to sample T15–43–1. In a calculated P – T pseudosection for T15–43–3, the interpreted peak assemblage of Grt + Sil + Kfs + Bt + Pl + Qz + melt is stable at 4.6–9 kbar and 710–800 °C, whereas the retrograde-metamorphic assemblage of Grt + Sil + Kfs + Crd + Bt + Pl + Qz yields a P – T range of 3.6–5.4 kbar and 705–800 °C (Fig. 8d). The highest Ti content of biotite isopleths was used to provide the maximum temperature and constrain the peak metamorphic P – T conditions of 4.6–8.7 kbar and 710–765 °C (pink field in Fig. 8d) and retrograde-metamorphic P – T conditions of 3.6–4.9 kbar and 727–760 °C (blue field in Fig. 8d).

These calculated phase relations infer a common clockwise prograde-to-peak P – T path for metamorphic rocks from the Bomi and Chayu regions, which formed along similar geothermal gradients and reached similar absolute P – T conditions (Fig. 8e and f), despite being located ~150 km apart along strike.

~~Conventional geothermobarometry~~

Geothermobarometry

The garnet–biotite (GB) thermometer (Holdaway, 2000) and garnet–biotite–plagioclase–quartz (GBPQ) barometer (Wu *et al.*, 2004) were applied to calculate the peak- and retrograde-metamorphic P – T conditions of the studied gneisses and

schists. These results were used both as an independent check on the conditions calculated via phase diagram analysis and due to local disequilibrium that may have developed during cooling and retrograde recrystallization, which obviates pseudosection modeling from interpreting the P – T conditions of retrograde exhumation far below the solidus.

For gneiss samples, flat garnet core compositions are interpreted to be the result of homogenization of cations at the thermal peak of metamorphism (e.g. Caddick et al., 2010). Garnet core compositions alongside those of matrix biotite and plagioclase produced peak-metamorphic P – T conditions of 6.1–6.7 kbar and 673–717 °C (sample T15–32–3; yellow-filled circle in Fig. 8a) and 5.3–6.4 kbar and 703–727 °C (sample T15–33–7; yellow-filled circle in Fig. 8b) using these calibrations. For schist samples, the garnet porphyroblasts show compositional zoning, and their cores and rims are interpreted to be formed during the peak and retrograde metamorphism respectively. Therefore, compositions of garnet core and matrix biotite and plagioclase that are not adjacent to garnet grains were selected to calculate the peak-metamorphic P – T conditions. These calibrations produced 5.5–6.1 kbar and 673–688 °C for sample T15–43–1 (yellow-filled circle in Fig. 8c), and garnet rim compositions were combined with compositions of neocrystalline biotite and plagioclase to constrain retrograde metamorphic P – T conditions of 3.9–4.5 kbar and 627–645 °C, shown by a blue-filled circle in Fig. 8c. No results could be determined for sample T15–43–3.

These peak metamorphic P – T conditions calculated via thermobarometry are similar to those constrained by phase equilibria modelling, although the calculated pressures of schist

sample T15–43–1 are significantly lower than those estimated via pseudosection analysis. Given that full bulk-rock compositions are not appropriate to forward-model mineralogical changes occurring in discrete domains, such as garnet strain shadows, the retrograde metamorphic P – T conditions for sample T15–43–1 determined by thermobarometry are considered most reliable in this case. These conditions lie below the fluid-saturated solidus and are interpreted to correlate with the retrograde closure temperature for effective cation exchange between garnet and the adjacent matrix (Ehlers *et al.*, 1994).

~~ZIRCON U–Pb GEOCHRONOLOGY AND LU–HF ISOTOPE DATA~~

~~Zircon U–Pb ages~~

~~Radiogenic isotope geochronology was performed on four of granite samples and all metasediment samples to determine the timing of key tectonothermal events in the Motuo–Bomi–Chayu region of the east Lhasa terrane. Zircon grains were separated from each sample by magnetic and conventional heavy-liquid techniques at the Hebei Institute of Regional Geology and Mineral Investigation. Cathodoluminescence (CL) images were taken on a HITACHI S2250-N scanning electron microscope at the SHRIMP Unit at the Institute of Geology, CAGS. Zircon U–Pb dating and trace element analysis were performed simultaneously on an Agilent 7500 ICP–MS equipped with a 193-nm ArF-excimer laser at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences (Wuhan). Detailed operating conditions for the ICP–MS instrument and the laser ablation system are as reported by Liu *et al.* (2010). Zircon 91500 was used as external~~

standard for U–Pb dating, which was analysed twice for every 5 analyses of the samples. SRM610 were used as external standard for the trace element analysis. Time dependent drifts of U–Th–Pb isotopic ratios were corrected using a linear interpolation (with time) for every five analyses according to the variations of zircon 91500. Zircon GJ-1 was used as standard to monitor the stability and accuracy of the U–Pb data acquired. Silicon (^{29}Si) was used as an internal standard. The U–Pb and trace element data were processed by ICPMSDataCal (Liu *et al.*, 2010) and Isoplot (Lugwig, 2003) was used to calculate isotopic ages and construct concordia diagrams. All LA-ICP-MS U–Pb data and the trace element compositions of the magmatic and metamorphic zircons from the granites and metasediments are presented in Supplementary table 7.

Zircons from granites T15-46-3, T15-46-5, T15-46-7, and T15-47-2 show similar internal structures, are colourless to pale brown, euhedral to subhedral prismatic in form (70–170 μm long), and have regular oscillatory zoning with inherited detrital cores in CL images (Fig. 9a–d). Most of the analysed zircon spots have relatively high Th/U ratios (>0.2) and REE contents, and are characterized by fractionated REE patterns with LREE depletion, HREE enrichment, and negative Eu anomalies (Supplementary table 7; Fig. 10a). The internal structure and compositional features suggest a magmatic origin (Hoskin and Schaltegger, 2003). Zircon data from these four samples yielded near-consistent $^{206}\text{Pb}/^{238}\text{U}$ ages, with weighted mean ages ranging from 216 Ma to 206 Ma (Fig. 11a–d). These ages are thus interpreted to represent magma emplacement/crystallization age.

Zircons from the gneiss and schist samples T15-32-3, T15-33-7, T15-43-1, and

T15-43-3 are mostly colourless, oval or sub rounded in shape, and show well preserved core-rim zonation (Fig. 9e-f). Inherited cores have a variable size, irregular form and random zoning, implying that zircon cores are detrital/xenocrystic and the protoliths of the studied samples are sedimentary rocks. In contrast, metamorphic rims show weak patchy zoning (or have no zoning at all), and have relatively low Th/U ratios, low REE contents and fractionated REE patterns with depleted LREE, flat HREE and slightly negative Eu anomalies (Supplementary table 7; Fig. 10b), typical of a metamorphic origin (Hoskin & Schaltegger, 2003). The analysed spots of the metamorphic rims yielded near concordant $^{206}\text{Pb}/^{238}\text{U}$ ages, with weighted means of 198 ± 3.3 Ma, 202 ± 3.4 Ma, 209 ± 2.8 Ma and 203 ± 3.1 Ma for T15-32-3, T15-33-7, T15-43-1, and T15-43-3, respectively (Fig. 11e-h).

Together, these data are interpreted to record Late Triassic crustal magmatism at 216–206 Ma and regional metamorphism at 209–198 Ma in the east Lhasa terrane.

Hf isotopes model ages

In situ Hf isotope compositions of zircon were obtained by a Neptune MC-ICP-MS at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences (Wuhan). The laser had a beam diameter of 44 μm , a frequency of 8 Hz, energy of 60 mJ, and a fluence of 5.3 J/cm². Analytical spots were chosen on the same domains with LA-ICP-MS spots. The zircon standards GJ-1 (Elhlou *et al.*, 2006) and 91500 (Blichert, 2008) were analysed as reference materials, and yielded weighted mean $^{176}\text{Hf}/^{177}\text{Hf}$ ratios of 0.282012 ± 11 (1 σ , n = 4) and 0.282305 ± 10 (1 σ , n = 10), respectively.

These values match the recommended values (0.282015 ± 19 for GJ-1, Elhlou *et al.*, 2006; 0.282308 ± 6 for 91500, Blichert, 2008). The initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios were calculated using the ^{176}Lu decay constant of $1.867 \times 10^{-11}/\text{yr}$ (Soderlund *et al.*, 2004). The $\epsilon_{\text{Hf}}(t)$ values were calculated using a chondritic $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282785 and $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.0336 (Bouvier *et al.*, 2008). The depleted mantle Hf model ages (T_{DM}) were calculated with respect to a depleted present-day mantle $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.28325 and $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of 0.0384 (Griffin *et al.*, 2000). The Hf crustal model ages (T_{DM2}) of each zircon were calculated by assuming its parental magma to have been derived from an average continental crust with $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$ (Griffin *et al.*, 2002). *In-situ* Lu-Hf isotopic ratios from two granites are presented in Supplementary table 8.

Zircons from two Late Triassic granites of the east Lhasa terrane (samples T15-46-3 and T15-46-5) have initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios ranging from 0.282308 to 0.282525 and variable negative $\epsilon_{\text{Hf}}(t)$ values (-8.6 to -16.2). These values correspond to T_{DM2} model ages of 1.51–1.99 Ga (Supplementary table 8; Fig. 12), indicating a Paleoproterozoic crustal magma source.

DISCUSSION

Late Triassic magmatism in the east Lhasa terrane

Four samples of granite collected from the Motuo-Bomi-Chayu region of the east Lhasa terrane yielded similar zircon U-Pb crystallization ages of 216–206 Ma, which correlate with U-Pb ages of magmatic zircons in biotite-hornblende schist (c. 217 Ma) from the same

area (Fig. 1; Dong *et al.*, 2011c). The granitic rocks studied here have negative $\epsilon_{\text{Hf}}(t)$ values of -8.6 to -16.2 and old crustal model ages (T_{DMC}) of ca. 1.51 to 1.99 Ga (Fig. 12), indicating that they were derived from partial melting of Paleoproterozoic crustal materials. Diorites in the same area with a slightly younger age (c. 194 Ma) have negative $\epsilon_{\text{Hf}}(t)$ values of -0.1 to -6.5 (Dong *et al.*, 2015), and so also likely formed from these older precursor rocks. Therefore, the east Lhasa terrane must have experienced widespread crustal-derived magmatism during the Late Triassic.

The Late Triassic granites studied here from the Chayu region plot in the S-type granite field on an A-C-F diagram (Fig. 13a) and have relatively high Al_2O_3 contents (12.33–14.58 wt. %), high A/CNK values (1.04–1.21), negative $\epsilon_{\text{Hf}}(t)$ values (-8.6 to -16.2) and are mixed magmas sourced from greywacke and pelite-derived melts (Fig. 13b). This parentage is supported by most zircons analysed from these granites containing inherited detrital cores (Fig. 9a–d) with metasedimentary protolith characteristics. These features suggest that the studied S-type granites likely formed by crustal anatexis of the metasediments during collisional orogeny (e.g. Zhang *et al.*, 2013; Chappell & White, 1992; Barbarin, 1998; Liegeois, 1998), as S-type granites are uncommon in non-convergent plate boundary or intraplate environments.

Late Triassic metamorphism in the east Lhasa terrane

Petrological observations, zircon geochronology, and phase equilibria modelling show that a major metamorphic episode affected the Motuo–Bomi–Chayu region of the eastern Lhasa

614 terrane during the Late Triassic. Gneiss sample T15–32–3 records similar P – T conditions
615 (5.7–7.5 kbar and 675–725 °C) to gneiss sample T15–33–7 (5.3–6.7 kbar and 750–765 °C),
616 and these are considered as peak metamorphism in the Bomi area. Given their proximity
617 with no evident structural discontinuities between outcrops, both samples must have
618 experienced the same tectonothermal evolution, even if the former preserves lower
619 temperature than the latter. The occurrence of fine-grained and large lath-like
620 porphyroblasts of retrograde muscovite formed in sample T15–33–7, alongside the absence
621 of cordierite, constrains the post-peak exhumation history to involve decompressional
622 cooling, as shown in Fig. 8e. By contrast, schists from the Chayu area ~150 km to the
623 southeast underwent a retrograde P – T path characterized by near-isothermal decompression
624 at high temperatures, causing the growth of cordierite and the consumption of garnet, and
625 then near-isobaric cooling towards to the inferred early prograde field (Fig. 8f). The studied
626 gneisses and schists thus experienced similar peak-stage upper amphibolite-facies
627 metamorphism and were buried to lower-crustal levels during orogenesis. These conditions
628 are consistent with Barrovian-type metamorphism, peaking at sillimanite-grade, which is
629 thought to be characteristic of collisional orogeny (e.g. England & Thompson, 1984; Palin
630 *et al.*, 2020); especially when followed by decompression along a cooling path associated
631 with tectonic exhumation (England & Richardson, 1977; England & Thompson, 1984;
632 Thompson & England, 1984; Harley, 1989). Slight differences in the character of the
633 exhumation path can be attributed to local thermal gradients, such as the presence of
634 magmatic intrusions in the Chayu region that allow the metamorphic rocks to remain hotter

during retrograde exhumation than those in the Bomi area. Variation in the style of exhumation along strike is not uncommon in collisional orogens at this length scale, and importantly has been well-documented in the Andes (Lease *et al.* 2016), which is a modern-day accretionary orogen with many tectonic parallels with the south Asian margin during closure of the Neo-Tethys ocean.

The metamorphic zircon rims from the gneisses and schists have low HREE concentrations and fractionated REE patterns with flat or even depleted HREE (Fig. 10b), indicating the metamorphic zircon grew coevally with garnet and plagioclase during medium- to high-grade metamorphism (Rubatto, 2002; Rubatto & Hermann, 2007; Rubatto *et al.*, 2013). Therefore, the ages of 209–198 Ma obtained from these grains are taken to represent the timing of peak metamorphism, which was characterized by P – T conditions just above the fluid-saturated solidus for metapelitic rocks (Palin & Dyck, 2020). These thermobarometric results are consistent with outcrop and microstructural observations of incipient partial melt development and the generation of zircon at or soon after the onset of cooling and melt crystallization (Bea & Montero, 1999; Kunz *et al.*, 2018).

Late Triassic magmatism in the east Lhasa terrane

Four samples of granite collected from the Motuo–Bomi–Chayu region of the east Lhasa terrane yielded similar zircon U–Pb crystallization ages of 216–206 Ma, which correlate with U–Pb ages of magmatic zircons in biotite-hornblende schist (c. 217 Ma) from the same area (Fig. 1; Dong *et al.*, 2011c). The granitic rocks studied here have negative $\epsilon_{\text{Hf}}(t)$ values

of -8.6 to -16.2 and old crustal model ages (T_{DMC}) of ca. 1.51 to 1.99 Ga (Fig. 12), indicating that they were derived from partial melting of Paleoproterozoic crustal materials. Diorites in the same area with a slightly younger age (c. 194 Ma) have negative $\epsilon_{Hf}(t)$ values of -0.1 to -6.5 (Dong *et al.*, 2015), and so also likely formed from these older precursor rocks. Therefore, the east Lhasa terrane must have experienced widespread crustal-derived magmatism during the Late Triassic.

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Tectonic implications

The Lhasa terrane represents the southern margin of Tibet and so is a key tectonic unit for documenting the spatio-temporal evolution of the India-Asia collision during the Cenozoic.

The terrane is bordered by the Bangong-Nujiang suture zone to the north and by the

677 Indus-Yarlung Tsangpo suture zone to the south, was initially divided into the northern,
678 central, and southern sub-terrane, separated by the Shiquan River-Nam Tso Mélange fault
679 and Luobadui-Milashan fault, respectively (e.g., Pan *et al.*, 2004, 2006; Zhu *et al.*, 2011,
680 2013). The southern part of Lhasa terrane preserves a semi-continuous record of the
681 Mesozoic northward subduction of the Neo-Tethyan ocean and the Cenozoic collision
682 between India and Asia (Allègre *et al.*, 1984; Acharyya, 2000; Yin & Harrison, 2000; Ding
683 *et al.*, 2001; Kapp *et al.*, 2003, 2007; Hou *et al.*, 2004, 2006; Chung *et al.*, 2005; Mo *et al.*,
684 2005; Zhang *et al.*, 2010; Xia *et al.*, 2011; Pan, *et al.*, 2012; Zhang & Santosh, 2012; Zhu *et*
685 *al.*, 2013; Dong *et al.*, 2018). However, the recent discovery of Sumdo high-pressure (HP)
686 metamorphic belt argued that Lhasa terrane also records the subduction and closure of
687 Paleo-Tethys ocean from the Late Paleozoic to the Early Mesozoic (Yang *et al.*, 2006, 2007,
688 2009; Li *et al.*, 2009b, 2012; Zeng *et al.*, 2009; Cheng *et al.*, 2015; Weller *et al.*, 2015,
689 2016a; Chen *et al.*, 2017). Together with the contemporaneous island volcanic rocks to the
690 north, dismembered ophiolite units and the regional angular unconformity between the
691 Middle and the Upper Permian, the belt is considered as a suture zone, which represents the
692 relics of the northward subduction of the Paleo-Tethys (Yang *et al.*, 2009; Zeng *et al.*, 2009;
693 Li *et al.*, 2009a, b; Cheng *et al.*, 2012, 2015; Weller *et al.*, 2015, 2016a). Separated by this
694 suture zone, the Lhasa terrane is now considered to consist of two discrete crustal fragments:
695 the North and South Lhasa terranes, with no central block (Fig. 15; Yang *et al.*, 2006, 2007,
696 2009; Li *et al.*, 2009a, b, 2012; Zeng *et al.*, 2009; Zhang *et al.*, 2014a; Cheng *et al.*, 2015;
697 Weller *et al.*, 2015, 2016a; Chen *et al.*, 2017).

698 The Sumdo Belt, central-east Lhasa terrane, contains high-pressure eclogite with a
699 Late Permian metamorphic age of 274–240 Ma (Yang *et al.*, 2006, 2007, 2009; Xu *et al.*,
700 2007; Chen *et al.*, 2008; Li *et al.*, 2009b; Zeng *et al.*, 2009; Cheng *et al.*, 2012, 2015; Weller
701 *et al.*, 2015, 2016a). These rocks must have formed prior to the onset of collisional orogeny.
702 Studies conducted on these rocks indicate that they experienced an amphibolite-facies to
703 epidote-amphibolite-facies retrograde metamorphism at 230–200 Ma, which matches the
704 evolution of many other metamorphic rocks with the similar P – T conditions and clockwise
705 P – T path (MP amphibolite-facies, typical of the Barrovian-type metamorphism) and ages
706 (Late Triassic to Early Jurassic, 225–192 Ma) along the central Lhasa terrane (Figs. 14 and
707 15), although they have slight variation in the character of the P – T paths due to different
708 local thermal gradients in this length scale collisional orogens. As such, all of these
709 lithologies constitute a large-scale Late Triassic to Early Jurassic metamorphic belt striking
710 east–west for at least 500 km, from the Nyainqentanglha in the west, through the Sumdo in
711 the central, to the Dongjiu region adjacent to Namche Barwa in the east (Fig. 15). This
712 linear belt is now considered as the primary record of the collisional orogeny between North
713 and South Lhasa terranes during the Early Mesozoic, and resulted from closure of the PTO
714 (Li *et al.*, 2008, 2009a, 2011, 2012; Dong *et al.*, 2011b; Lin *et al.*, 2013b; Cheng *et al.*, 2015;
715 Weller *et al.*, 2015, 2016a; Chen *et al.*, 2017). Additionally, the Early Mesozoic
716 metamorphism along the central Lhasa terrane is associated with widespread coeval granitic
717 magmatism, which is also interpreted to be the products of the collision between North and
718 South terranes (Fig. 15; Kapp *et al.*, 2005; Liu *et al.*, 2006; Zhang *et al.*, 2007; Li *et al.*,

2008, 2009a; Zhu *et al.*, 2011; Dong *et al.*, 2015; Weller *et al.*, 2016b). Therefore, we suggest that the Late Triassic magmatic and metamorphic rocks from the Motuo–Bomi–Chayu region of the eastern Lhasa terrane formed in a same tectonic setting of collisional orogeny between North and South Lhasa terranes, resulted from the closure of the PTO (Fig. 16). These data firstly indicate that the east Lhasa terrane east of the EHS, like the central-east Lhasa terrane to the west of the syntaxis, also witnessed closure of the Paleo-Tethys oceanic basin, and that the Sumdo metamorphic/orogenic belt documented in the east-central Lhasa terrane as recording the demise of the PTO should be extended eastward past the EHS into the east Lhasa terrane (Fig. 15).

There are key implications for this proposed extension of the Sumdo orogenic belt east of the EHS, where no HP or UHP eclogite-facies rocks have yet been discovered. The absence of canonical indicators of paleo-subduction in this region of the Lhasa terrane east of Namche Barwa, such as lithofacies that form only at convergent plate boundaries (e.g. mélangé, blueschist, MORB-type eclogite, jadeitite; Stern *et al.*, 2013; Palin & White, 2016), has historically hindered tracing the paleo-closure of the North and South Lhasa blocks. Our new data provide evidence that regional scale metamorphism and crustal thickening were occurring in this region simultaneously with units in Basong Tso and Sumdo, central Tibet (Weller *et al.*, 2015, 2016a), where HP mafic eclogite is well exposed. Further, Carboniferous-Permian volcanic rocks are documented on the southern margin of the North Lhasa plate in both the central Lhasa block and in the Motuo–Bomi–Chayu region (Fig. 15), indicating coeval along-strike subduction of the PTO (Wang *et al.*, 2008; Yang *et*

al., 2009). Why, then, are there not equivalent HP eclogite exposures east of the EHS? The exhumation potential of subducted oceanic lithosphere varies depending on a wide range of petrophysical factors (e.g. Guillot *et al.*, 2001; Warren, 2013). Further, a wide range of mechanisms has been proposed for allowing exposure at the Earth's surface following an initial period of rapid exhumation based on positive buoyancy (e.g. St-Onge *et al.*, 2013).

Firstly, it may be considered that the subducted PTO lithosphere experienced along-strike variation in dip angle, as noted today in the Andes (Wortel, 1984; Chen *et al.*, 2001) and proposed for the lack of Cenozoic (U)HP eclogite in the central and east Himalaya (O'Brien *et al.*, 2001; Leech *et al.*, 2005). Eclogite from the Sumdo complex near to Basong Tso (Fig. 15) reached peak P – T conditions of 27 kbar and 670 °C, equivalent to transport to a depth of ~95 km before exhumation (Weller *et al.*, 2016a). If slab subduction beneath the eastern Lhasa terrane during the Mesozoic occurred at a much steeper angle, it is possible that the subducted oceanic root achieved negative buoyancy at an equivalent time to the Basong Tso eclogites (Agard *et al.*, 2009), and so upon slab fragmentation descended terminally into the lower mantle. Alternatively, if the slab angle in this region was shallower than that interpreted for Basong Tso, relatively low-pressure eclogite may have formed (e.g. Hernandez-Urbe & Palin, 2019), although such low-angle subduction is often associated with formation of slab-derived magmas (adakites; Drummond *et al.*, 1996), which are absent from the Motuo–Bomi–Chayu region. Thus, this latter hypothesis appears unlikely based on the current understanding of the geology of this part of the Lhasa terrane.

If (U)HP eclogite formed during closure of the PTO east of the EHS, and was exhumed

and incorporated into the overlying crust, it may thus be present in the present-day subsurface, as the metamorphic pressures calculated from the rocks in this region are slightly lower than the eclogitic host gneisses in the Sumdo and Basong Tso region (cf. 9 kbar in Basong Tso; Weller *et al.*, 2015). Thus, we interpret that the level of exposure of the Lhasa terrane in Motuo–Bomi–Chayu region is slightly shallower than the temporal equivalent along-strike to the west. Such an assessment of orogen-parallel variation in exhumation rate should be considered in large-scale reconstructions of the evolution of the Tibetan region prior to the onset of uplift during Cenozoic collision with India.

CONCLUSIONS

The east Lhasa terrane witnessed Late Triassic felsic magmatism (216–206 Ma) and regional metamorphism (209–198 Ma). Metasedimentary gneisses and schists studied from the Motuo–Bomi–Chayu region, eastern flank of the EHS, experienced medium-pressure amphibolite-facies metamorphism and partial melting, followed by a decompressional cooling retrograde process. The Late Triassic granites are peraluminous S-type granites and derived from the partial melting of nearby metasediments, indicating localized melt transport. The coeval Late Triassic magmatism and metamorphism in the east Lhasa terrane are related to the collision between North and South Lhasa, which resulted from closure of the Paleo-Tethys Ocean. Finally, these new data show that the recently discovered Sumdo metamorphic/orogenic belt that formed during closure of the PTO should be extended eastward to at least the east Lhasa terrane.

782

783 **ACKNOWLEDGMENTS**

784 We thank Editor Adam Kent and an anonymous reviewer for meticulous and insightful
785 reviews, and Georg Zellmer for editorial handing. We also thank Dr. Changlei Fu, Yuelei
786 Yuan and David Hernández-Urbe for valuable discussions and suggestions, and Ph.D.
787 student Dongyan Kang, YuanYuan Jiang and Hongchen Mu for the field work. This study is
788 co-supported by the National Natural Science Foundation of China (91855210, 41941016,
789 41872069 and 41872070), the National Key Research and Development Project of China
790 (2016YFC0600310 and 2018YFC0603700), the Fundamental Research Funds for the
791 Central Universities of China (649911026) ~~the National Natural Science Foundation of~~
792 ~~China (41872029, 41230205, 41472056, 41202035 and 41602062)~~ and the China
793 Geological Survey (~~DD20160201~~ DD20190059 and DD20190011). ~~We thank Dr. Changlei~~
794 ~~Fu and Yuelei Yuan, and Ph.D. student David Hernández-Urbe for valuable discussions and~~
795 ~~suggestions. We also thank Ph.D. student Dongyan Kang, YuanYuan Jiang and Hongchen~~
796 ~~Mu for the field work.~~

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1284 **FIGURE CAPTIONS**

1285

Fig. 1. (a) Simplified geological map of the Lhasa terrane, showing the main suture zones and terranes. JSSZ, Jinsha suture zone; LSSZ, Longmu Tso-Shuanghu suture zone; BNSZ, Bangong-Nujiang suture zone; SDSZ, Sumdo Paleo-Tethys suture zone; ITSZ, Indus-Yarlung Tsangpo suture zone; ATF, Altyn Tagh Fault; KJF, Karakorum-Jiali Fault; RRR, Red River Fault; EHS, eastern Himalayan syntaxis; NQ, North Qiangtang terrane; SQ, South Qiangtang terrane; NL, North Lhasa terrane; SL, South Lhasa terrane. (b) Geological map of the east Lhasa terrane, showing the sample locations and magmatic and metamorphic ages reported in this work. The literature data are after Dong *et al.*, (2011c, 2015).

Fig. 2. Field photographs and photomicrographs of the magmatic rocks from the east Lhasa terrane, showing the texture and mineral components of granite. Ms, muscovite; Bt, biotite; Pl, plagioclase; Kfs, K-feldspar; Q, quartz.

Fig. 3. Outcrops of metapelitic gneiss (a) and schist (b) in the east Lhasa terrane.

Fig. 4. Photomicrographs of mineral assemblages and microstructures of the gneisses. (a) and (b) Sample T15–32–3 with large garnet porphyroblasts surrounded by an aligned matrix defined by biotite, muscovite, plagioclase, quartz and minor sillimanite. Garnet and biotite are partly replaced by chlorites. (c) and (d) Sample T15–33–7 showing an isotropic sillimanite–biotite–plagioclase–quartz matrix with garnet and K-feldspar porphyroblasts,

1307 and large lath-like or randomly orientated fine-grained muscovite. (e) A very small cusped
1308 quartz dihedral angle against plagioclase in sample T15-33-7. (f) K-feldspar grains
1309 surrounded by plagioclase rim in sample T15-33-7. Ms, muscovite; Bt, biotite; Pl,
1310 plagioclase; Kfs, K-feldspar; Q, quartz; Sil, sillimanite; Grt, garnet; Chl, Chlorite; Ilm,
1311 ilmenite.

1312

1313 **Fig. 5.** Photomicrographs of mineral assemblages and microstructures of the schists. (a) and
1314 (b) Sample T15-43-1 exhibiting large garnet porphyroblasts, that rims replaced by
1315 biotite-muscovite-plagioclase aggregates, wrapped by a matrix foliation comprised of
1316 biotite, muscovite, plagioclase, quartz and minor ilmenite. Muscovite occurs both as aligned
1317 flakes within foliation and as larger, subhedral to euhedral, unfoliated grains against garnet
1318 rims. (c) and (d) Sample T15-43-3 containing garnet porphyroblasts surrounded by a
1319 matrix foliation defined by biotite, sillimanite, plagioclase, K-feldspar, and quartz. Garnet
1320 rims are partly pseudomorphed by biotite and cordierite. (e) Micro-exsolved lamellae of
1321 plagioclase hosted by K-feldspar in sample T15-43-3. (f) Garnet grains have inclusion-rich
1322 cores and inclusion-absent rims that show textural equilibration with matrix phases. Ms,
1323 muscovite; Bt, biotite; Pl, plagioclase; Kfs, K-feldspar; Q, quartz; Sil, sillimanite; Grt,
1324 garnet; Crd, Cordierite; Ilm, ilmenite.

1325

1326 **Fig. 6.** Compositional profiles of garnet porphyroblasts from the schists (a, sample
1327 T15-43-1; b, sample T15-43-3) and X_{Mg} vs. Ti (cpfu) diagram for biotite (c).

1328

1329 **Fig. 7.** (a) Primitive mantle normalized trace element diagrams and (b) chondrite
1330 normalized rare earth element (REE) diagrams of granites. The trace element data for
1331 primitive mantle and REE data for chondrites are after Sun and McDonough (1989).

1332

1333 **Fig. 8.** Pressure–temperature (P – T) pseudosections for samples (a) T15–32–3, (b)
1334 T15–33–7, (c) T15–43–1 and (d) T15–43–3, calculated using the bulk-rock compositions
1335 given in Supplementary table 6. (e) and (f) Summary plot showing the interpreted P – T
1336 evolution of studied gneisses and schists respectively. The red bold fonts refer to the
1337 observed mineral assemblage. The bold brown lines mark the positions of the solidus, bold
1338 yellow lines mark the stability of muscovite. The blue, purple, green and brown lines
1339 represent plagioclase (X_{Ca}), biotite (Ti) and garnet (X_{Mg}) and (X_{Ca}) isopleths, respectively.
1340 The pink- and blue-filled polygons represent the peak and retrograde metamorphic
1341 conditions. The yellow-filled circles represent the peak-metamorphic conditions by
1342 thermobarometry. The bold black lines and dashed lines with arrow refer to the inferred
1343 prograde and retrograde P – T paths. Peak apparent thermal gradients were calculated
1344 assuming a crustal density of 3000 kg/m³ and a linear gradient.

1345

1346 **Fig. 9.** Representative cathodoluminescence (CL) images of zircon grains from studied
1347 rocks showing the analysed spot locations and related ages (in Ma).

1348

Fig. 10. Chondrite-normalized REE patterns of zircons from granites (a) and metamorphic rocks (b). Chondrite values are after Sun and McDonough (1989).

Fig. 11. Zircon U–Pb concordia diagrams for studied rocks.

Fig. 12. Zircon $\varepsilon_{\text{Hf}}(t)$ values vs. U–Pb ages diagram of the Late Triassic granites.

Fig. 13. (a) A–C–F discrimination diagram for I-type and S-type magmas (after Chappell and White, 1992) and (b) Rb/Sr vs. Rb/Ba discrimination diagram for source of the granites (after Sylvester, 1998).

Fig. 14. Summary of the inferred P – T – t paths for the studied east Lhasa terrane gneisses and schists, and comparison with those reconstructed for the coeval amphibolite-facies metamorphic rocks from central-east Lhasa terrane. The arrow lines represent the P – T paths. The geothermal gradients of 20, 27 and 45°C/km are shown. Aluminosilicate phase relations are after Pattison (1992). The literature data are after Dong *et al.* (2011b), Lin *et al.* (2013b), Weller *et al.* (2015) and Chen *et al.* (2017).

Fig. 15. The distribution characteristics of the Paleo-Tethys Ocean orogenic belt and the related magmatic and metamorphic rocks. The abbreviations are the same as in Fig. 1.

Fig. 16. Schematic plate tectonic evolution model of the east Lhasa terrane during the Early Mesozoic. The abbreviations are the same as in Fig. 1.

TABLE CAPTIONS

Table 1. The major features of the studied rocks from the east Lhasa terrane.

Supplementary table 1. The compositions of representative garnet from the east Lhasa terrane metamorphic rocks.

Supplementary table 2. The compositions of representative biotite from the east Lhasa terrane metamorphic rocks.

Supplementary table 3. The compositions of representative plagioclase from the east Lhasa terrane metamorphic rocks.

Supplementary table 4. The compositions of representative K-feldspar from the east Lhasa terrane metamorphic rocks.

Supplementary table 5. The compositions of representative cordierite from the east Lhasa terrane metamorphic rocks.

1391 **Supplementary table 6.** Major (wt. %) and trace (ppm) element data of the studied rocks
1392 from the east Lhasa terrane.

1393

1394 **Supplementary table 7.** LA–ICP–MS U–Pb dating and rare earth element results of the
1395 magmatic and metamorphic zircons.

1396

1397 **Supplementary table 8.** Hf isotopic data of zircons for the granites from the east Lhasa
1398 terrane.