Late Cretaceous metamorphism and anatexis of the Gangdese magmatic arc, South Tibet: implications for thickening and differentiation of juvenile crust

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ABSTRACT

Magmatic arcs are the primary sites of growth of post-Archean continental crust; however, the mechanisms and processes for transforming primary arc crust into mature continental crust are subject to disagreement. We conducted a detailed petrologic and geochronological study on mafic and felsic migmatites from the eastern Gangdese magmatic arc, typical of continental arcs worldwide. The studied mafic migmatites contain amphibole, garnet, plagioclase, epidote, muscovite, quartz, rutile and ilmenite in melanosomes, and plagioclase, garnet, epidote, muscovite and quartz in leucosomes. The leucosomes occur as diffusive patches, concordant bands, or concordant and discordant networks and veins in the melanosomes. The migmatites have protolith ages of ~157 Ma and ~90–100 Ma, and metamorphic ages of ~83–87 Ma, and underwent high-pressure granulite-facies metamorphism under *P*–*T* conditions of ~850–880 °C and 15–17 kbar, with a counterclockwise *P*–*T* path characterized by early heating and burial, associated partial melting, and late near-isobaric cooling with residual melt crystallization. Significant melt (>16 wt. %) generated during heating and loading has a granitic composition. This study indicates that the eastern Gangdese magmatic arc experienced crustal thickening during the Late Cretaceous late-stage evolution of the arc due to accretion and loading of mantle-derived magma, and tectonic shortening and thrusting of the arc crust. The thickened lower arc crust underwent significant partial melting, and the resulting granitic melt formed a potential source for the Late Cretaceous granitoids of the upper arc crust. These results provide robust evidence for crustal thickening and chemical differentiation of the Gangdese arc during the growth of juvenile crust prior to the India–Asia collision, as well as new insights into formation and evolution of continental magmatic arcs in general.

Key words: thickening; partial melting; chemical differentiation; juvenile crust; Gangdese magmatic arc

INTRODUCTION

Magmatic arcs are widely considered to be the primary sites for the growth of post-Archean continental crust (Davidson & Arculus, 2006; Miller & Snoke, 2009; Brown & Ryan, 2011; Jagoutz & Schmidt, 2012; Ducea *et al*., 2015; Jagoutz & Kelemen, 2015). Bulk continental crust has an andesitic composition that cannot have been in equilibrium with the upper mantle (e.g., Rudnick, 1995; Jagoutz & Kelemen, 2015; Lee & Anderson, 2015), so most models for the generation of the continental crust involve at least two stages of differentiation—the extraction of basaltic magma from the mantle and the intracrustal differentiation of that basalt (Taylor & McLennan, 1985; Rudnick, 1995; Rudnick & Gao, 2003; Hawkesworth

& Kemp, 2006). Proposed intracrustal differentiation mechanisms include fractional crystallization (Taylor, 1967; Davidson & Arculus, 2006; Jagoutz, 2014; Keller *et al*., 2015; Jagoutz & Kelein, 2018), crustal melting (Brown & Rushmer, 2006; Garrido *et al*., 2006; Brown, 2010; Yakymchuk *et al*., 2016), magma hybridization and assimilation (Hildreth & Moorbath, 1988; Otamendi *et al*., 2009), melt-rock interaction (Cooper *et al*., 2016; Daczko *et al*., 2016), relamination of buoyant, subducted rocks (Hacker *et al*., 2011, 2015; Kelemen & Behn, 2016), and a combination of some or all of these mechanisms (Hildreth & Moorbath, 1988; Chapman *et al*., 2014).

Although fractional crystallization of basaltic magma is considered key for intracrustal differentiation, partial melting within magmatic arc crust is likely also necessary for transforming this juvenile material into mature continental crust (Jagoutz & Klein, 2018). This is because buoyancy-driven extraction of partial melt from the lower arc crust and its emplacement in the upper crust can lead to vertical differentiation into a mafic lower portion and a felsic upper portion (Brown & Rushmer, 2006; Garrido *et al*., 2006; Brown, 2007, 2010). Furthermore, dense residues produced by anatexis in a thickened lower crust may delaminate or founder into the less-dense underlying mantle, which is likely key for generating an average bulk continental crust of andesitic composition (Kay & Kay, 1993; Stern & Scholl, 2010; Hacker *et al*., 2011; Jagoutz & Behn, 2013; Ducea *et al*., 2015). In particular, in magmatic arcs with crustal thicknesses greater than 50–60 km, such as the Andes, South America, Gangdese, southern Tibet, Kohistan, western Himalayas, and Fiordland, New Zealand, the thickened juvenile lower crust has probably experienced significant partial melting, and in turn caused intense granitic magmatism and widespread crustal differentiation.

Disagreements related to differentiation mechanisms and their relative contributions highlight two key issues that must be addressed. The first issue is whether intracrustal differentiation occurs during arc construction, during arc–continent collision, or both for mature magmatic arcs with a complex evolution history from subduction to collision. The second issue is whether different differentiation mechanisms occur at different stages in the evolution of a magmatic arc and what mechanism predominates during different stages; for example, fractional crystallization is probably a dominant mechanism for crustal differentiation during the subduction period, whereas partial melting is more likely to be a main process during the collision period (DeBari & Greene, 2011).

The Mesozoic to Early Cenozoic magmatic rocks in the Gangdese arc, southern Tibet, formed during subduction of the Neo-Tethyan oceanic lithosphere beneath the southern margin of Asia, and so represents juvenile continental crust (Harris *et al*., 1988; Chu *et al*., 2006; Mo *et al*., 2007, 2008; Wu *et al*., 2007, 2010; Wen *et al*., 2008a, 2008b; Ji *et al*., 2009; Zhang *et al*., 2010a; Zhu *et al*., 2011; Niu *et al*., 2013; Hou *et al*., 2015b). The modern sections of this arc represent some of the thickest continental crust on Earth today (up to 60–80 km; Hirn *et al*., 1984; Molnar, 1988; Zhao *et al*., 1993; Yakovlev & Clark, 2014). Therefore, it is widely accepted that the Gangdese arc underwent intense crustal thickening, and that the Gangdese granitoids, representing major components of the Gangdese arc, were mostly derived from partial melting of the thickened lower arc crust during the Mesozoic–Cenozoic (Yin & Harrison, 2000; Ding & Lai, 2003; Kapp *et al*., 2005, 2007a, 2007b; Mo *et al*., 2007; Chung *et al*., 2009; Guan *et al*., 2012; Ji *et al*., 2012; Yakovlev & Clark, 2014; Ding *et al*., 2014; Zhu *et al*., 2017; Tang *et al*., 2019). However, the timing and mechanisms of crustal thickening and differentiation remain unclear because the petrological and thermal evolution of metamorphosed and partially melted lower crustal materials of the Gangdese arc have only been studied in brief (Zhang *et al*., 2010b, 2013, 2014a; Guo *et al*., 2012, 2013; Palin *et al*., 2014).

A series of medium- to high-grade metamorphic rocks occur widely in the eastern segment of the Gangdese arc (Fig. 1). These rocks, mostly formed from Mesozoic to Early Cenozoic arc-related magmatic rocks, form the middle and lower crustal components of the Gangdese arc (Zhang *et al*., 2010b, 2013, 2014a, 2014b, 2015, 2019a; Searle *et al*., 2011; Guo *et al*., 2012, 2013). The well-exposed lower crustal sections provide a rare opportunity to directly study the processes by which juvenile continental crust forms in convergent plate margins (Searle *et al*., 2011; Zhang *et al*., 2020; Guo *et al*., 2020).

In this paper, we present a petrological, geochemical and geochronological study of Late Cretaceous migmatites from the exhumed lower crustal sections of the eastern Gangdese arc. The aims are to reveal the building process of Gangdese juvenile crust, to probe the evolution of magmatic arcs prior to deformation, metamorphism and magmatism associated with collisional orogenesis, and to explore the complete evolutionary history of composite arcs with both subduction- and collision-related magmatic processes. Our results show that the Gangdese arc underwent crustal thickening during the Late Mesozoic, and the juvenile arc crust experienced significant chemical differentiation prior to the collision of India and Asia. This study provides new insights into the formation and evolution of the Gangdese arc, and the tectonothermal processes that have driven the growth and differentiation of the continental crust since the onset of plate tectonics.

GEOLOGICAL SETTING

The Gangdese magmatic arc, South Tibet, represents the central-eastern segment of the Trans-Himalayan magmatic arc that strikes E–W for ~2400 km (Fig. 1a; Debon *et al*., 1986; Ding *et al*., 2003; Chu *et al*., 2006; Wen *et al*., 2008b; Pan *et al*., 2012; Hou *et al*., 2015a, b; Zhu *et al*., 2015, 2018). The arc consists mainly of the Cretaceous–Early Tertiary Gangdese batholith, and Paleogene Linzizong volcanic succession, with some Triassic–Cretaceous intrusive and volcanic-sedimentary rocks (Mo *et al*., 2007, 2008; Wen *et al*., 2008a, 2008b; Guo *et al*., 2011; Zhu *et al*., 2011, 2018; Lee *et al*., 2012; Pan *et al*., 2012; Wang *et al*., 2016; Li *et al*., 2018). The Gangdese batholith is mainly composed of granodiorite and granite, with a small amount of gabbro and diorite (Chung *et al*., 2005; Mo *et al*., 2007, 2008; Zhang *et al*., 2010a, 2010b, 2013, 2019a; Zhu *et al*., 2011, 2018; Zheng *et al*., 2012, 2014; Ma *et al*., 2013a, 2013b, 2013c, 2014). The batholith documents a period of long-lasting magmatism (ca. 180–50 Ma), with two main magmatic pulses in the Late Cretaceous and the Paleocene to Early Eocene, respectively (Chu *et al*., 2006, 2011; Wu *et al*., 2007, 2010; Ji *et al*., 2009, 2014; Guo *et al*., 2011; Zhu *et al*., 2011, 2017, 2018; Liu *et al*., 2014; Zhang *et al*., 2020). The Late Cretaceous magmatic pulse was likely caused by Neo-Tethyan mid-ocean ridge subduction (Zhang *et al*., 2010a, 2019b; Guo *et al*., 2011, 2013; Zheng *et al*., 2014; Zhu *et al*., 2015, 2018) or the roll-back of subducted Neo-Tethyan lithosphere (Ma *et al*., 2013a, 2013b), whereas the Paleocene to Early Eocene magmatic pulse may relate to the roll-back and subsequent breakoff of the deeply subducted Neo-Tethyan oceanic plate (Lee *et al*., 2009; Guo *et al*., 2011; Zhu *et al*., 2015, 2018; Ji *et al*., 2016) or to the partial melting of underthrusting Tethyan oceanic crust (Niu *et al*., 2013) after the initial Indo-Asian collision. The collision likely occurred at the Early Cenozoic at ~65–50 Ma (Rowley, 1996; Mo *et al*., 2007; Najman *et al*., 2010; Wu *et al*., 2014; Zhu *et al*., 2015; Ding *et al*., 2016a, 2016b; Hu *et al*., 2016).

The Eastern Himalayan Syntaxis, southeastern Tibet, exposes three tectonic units: the Indian-plate Himalayan belt in the south, the Indus–Yarlung Tsangpo suture zone in the middle, and the Asian-plate eastern Gangdese magmatic arc (eastern Lhasa terrane) in the north (Fig. 1; Booth *et al*., 2004; Geng *et al*., 2006; Zhang *et al*., 2010a, 2010b, 2012; Xu *et al*., 2012). The Himalayan belt includes the Tethyan Himalayan Sequences and Greater Himalayan Sequences (Fig. 1b). The Indus–Yarlung Tsangpo suture zone consists of strongly deformed Early Cretaceous mélange with remnants of the Neo-Tethyan oceanic crust (Geng *et al*., 2006). The eastern Gangdese arc is dominated by Paleozoic sedimentary rocks, Jurassic volcanic rocks, Jurassic to Cretaceous diorite–granite, Late Cretaceous (~95–100 Ma) gabbro–granodiorite (Lilong batholith), Late Cretaceous (~71–87 Ma) granite (Wolong batholith), and Paleocene to Eocene gabbro and granite. All these rocks underwent Mesozoic and Cenozoic metamorphism, and were intruded by Oligocene granite (Fig. 1b). In the Zhaxi–Milin–Dongdui–Bujiu area, the meta-plutonic and meta-sedimentary rocks record upper-amphibolite to high-pressure granulite-facies metamorphism (Fig. 1b), and therefore are interpreted to represent lower crustal components of the arc (Zhang *et al*., 2020). Geological mapping and dating show that the protoliths of these high-grade metamorphic and anatectic rocks include Late Cretaceous pyroxenite, gabbro, diorite and granodiorite, Late Jurassic diorite, and Paleocene to Eocene gabbro and granite, with small amounts of Mesoproterozoic diorite and Late Cambrian granite, Early Permian gabbro, and Late Carboniferous sedimentary rocks (Fig. 2). The Late Cretaceous rocks of the Lilong batholith are high-grade metamorphosed pyroxenite, gabbro and diorite in its northeastern part, and low-grade metamorphosed to unmetamorphosed gabbro, diorite and granodiorite in its southwestern part (Fig. 1b). The meta-sedimentary rocks include paragneiss (felsic migmatite), schist (pelitic migmatite), marble and calc-silicate rock, and occur locally as lenses with variable sizes ranging from several meters to several thousands of meters within the metamorphosed plutonic rocks (Fig. 2). The high-grade metamorphosed and anatectic rocks consist mainly of mafic and felsic migmatites (Figs. 2 and 3). These migmatites mostly underwent strong ductile deformation, and show sub-vertical and pervasive foliation (Figs. 2, 3a and 3i). The strongly deformed migmatites are widely distributed in the exposed lower arc crust. In contrast, a few migmatites have a relatively homogeneous massive structure (Fig. 3b and c), and are distributed in regions of weak deformation that locally occur in the lower arc crust. In places, garnet-rich leucosomes occur as concordant and discordant veins within the migmatites (Fig. 3d–f). The large and euhedral garnet crystals (up to 2 cm in diameter) are set in the leucosome (Fig. 3f). Apparent retrogression and melt-mineral back reactions caused the garnet of the mafic migmatites to partly or completely transform into amphibole and epidote (Fig. 3d–f).

SAMPLES

The present study focuses on garnet-bearing mafic and felsic migmatites. These rocks are widely distributed in the Zhaxi–Milin–Dongdui–Milin Airport area (Fig. 2), so are interpreted to represent the main components of the Late Cenozoic lower crust of the Gangdese arc (Zhang *et al*., 2020). The mafic migmatites were collected from a metamorphic and anatectic Late Cretaceous gabbro of the Lilong batholith root in the Zhaxi–Milin area (Fig. 2). Most migmatites are stromatic, although some resemble diatexite migmatite, based on the definition of Sawyer (2008). The stromatic migmatites contain laterally continuous leucosomes that parallel the foliation, and therefore show banded structure defined by alternating bands of melanosome and leucosome (Fig. 3a). The diatexite migmatites lack paleosome, and have melts that are pervasively distributed throughout (Fig. 3b and 3c). Both of the stromatic and diatexite migmatites are composed of darker-colored and lighter-colored parts with transitional contacts (Fig. 3a–c). The darker-colored parts consist mainly of amphibole, plagioclase and garnet, with small amounts of epidote, muscovite and quartz, whereas the lighter-colored parts consist mainly of plagioclase, epidote, muscovite, quartz and garnet, with small amounts of amphibole (Fig. 3a–c). In this paper, referring to the definition of Sawyer (2008), the darker-colored parts of the migmatite are called the melanosome, and the lighter-colored parts are called the leucosome.

The studied garnet-bearing felsic migmatite was collected from a metamorphosed Late Jurassic diorite in the Dongdui area (Fig. 2). It commonly shows a strong foliation and banded structure that is defined by alternating melanosome and leucosome (Fig. 3g–i). The melanosome consists of plagioclase, biotite, amphibole, quartz and garnet; the leucosome contains plagioclase, quartz, garnet and amphibole, and occurs not only as bands, but also as networks or concordant and discordant veins (Fig. 3g–i).

The low-grade meta-gabbros, collected from the southwestern part of the Lilong batholith, near Lilong village (Fig. 1b), were used for geochemical compositional comparison between the gabbro and mafic migmatite (anatectic gabbro) from the same batholith to evaluate the possible compositions of the extracted melts during the anatexis of gabbros. The Lilong village gabbros commonly show massive structure, and contain clinopyroxene, orthopyroxene, plagioclase, amphibole, epidote, biotite, quartz and magnetite (Supplementary Data Table S1). The primary (magmatic) clinopyroxene and orthopyroxene are partially or completely replaced by metamorphic amphibole, epidote and biotite.

Mineral and liquid abbreviations used in this paper are as follows: Alm = almandine, Amp = amphibole, Bt = biotite, Cpx = clinopyroxene, Ep = epidote, Grs = grossular, Grt = garnet, Ilm = ilmenite, Ky = kyanite, L = liquid, Ms = muscovite, Mt = magnetite, Opx = orthopyroxene, Pg = paragonite, Pl = plagioclase, Prp = pyrope, Qtz = quartz, Rt = rutile, Sph = sphene, and Spe = spessartine.

ANALYTICAL AND PHASE EQUILIBRIA MODELLING METHODS

Mineral compositions were determined using a JEOL JXA 8900 electron microprobe with a 15 kV accelerating voltage, 5 nA beam current, 5 μm beam diameter, and count time of 10 s for peak and background, at the Institute of Geology, Chinese Academy of Geological Sciences. Natural and synthetic mineral standards were used and ZAF corrections were carried out. Whole-rock major and trace element contents were measured at the National Geological Analysis Center of China, Beijing. Major element oxides including total Fe (FeO^{total}) as Fe_2O_3 were determined by X-ray fluorescence spectrometry (XRF) (Rigaku-3080) with an analytical uncertainty of <0.5%. The FeO content was obtained independently by titration and Fe_2O_3 content was calculated as $Fe_2O_3 = Fe_2O_3 - 1.1114FeO$. The trace elements Zr, Nb, V, Cr, Sr, Ba, Ni, Rb, and Y were analyzed using a different XRF instrument (Rigaku-2100) with an analytical uncertainty of <3%–5% relative. Other trace elements and rare earth elements (REEs) were analyzed by inductively coupled plasma mass spectrometry (ICP-MS; TJA-PQ-ExCell); relative analytical uncertainties are 1%–5% when the abundance is greater than 1 ppm, and 5%–10% when the abundance is less than 1 ppm.

In-situ U–Pb dating and trace element analysis of zircon were simultaneously conducted by LA–ICP–MS at the Wuhan Sample Solution Analytical Technology Co., Ltd., China. Detailed operating conditions for the laser ablation system and the ICP–MS instrument are described by Zong *et al.* (2017). Laser sampling was performed using a GeolasPro laser ablation system that consists of a COMPexPro 102 ArF excimer laser (wavelength of 193 nm and maximum energy of 200 mJ) and a MicroLas optical system. An Agilent 7700e quadrupole ICP–MS instrument was used to acquire ion-signal intensities. Helium was applied as a carrier gas. Argon was used as the make-up gas and mixed with the carrier gas via a T-connector before entering the ICP–MS. The spot size and frequency of the laser were set to 32 μm and 5 Hz, respectively. Each analysis incorporated a background acquisition of 20 s followed by 50 s of data acquisition from the sample. All traces were verified for flat signals to ensure that ablation did not inadvertently create mixed analyses by penetrating compositionally and chronologically different domains below the imaged surface. Zircon standards GJ-1 and Plešovice were used as secondary reference materials during LA–ICP–MS analysis. As listed in Supplementary Data Table S8, GJ-1 and Plešovice yield concordant or nearly concordant U–Pb ages of 585.9–616.5 Ma (weighted mean age = 602.8 ± 4.6 Ma (2σ, n = 16, MSWD = 1.2)) and 345.8–330.5 Ma (mean age = 336.5 ± 2.2 Ma (2σ, n = 16, MSWD = 1.07)), respectively, which are well within uncertainty of recommended values (GJ-1: 599.8 ± 1.7 Ma, Jackson *et al*., 2004, 602.1 ± 4.9 Ma, Liu *et al*., 2010; Plešovice: 337.1 ± 0.4 Ma; Sláma *et al*., 2008). NIST610 and Si were used to calibrate the trace element concentrations as external reference material and internal standard element, respectively. The software package Iolite was used for data reduction, and the exponential function was used to calibrate the downhole fractionation (Paton *et al*., 2010). Concordia diagrams and weighted mean calculations were made using Isoplot/Ex_version 4.15 (Ludwig, 2003).

The pressure–temperature (*P*–*T*) conditions of metamorphism and partial melting of the mafic migmatite were constrained by using phase equilibria modelling, which has numerous advantages over conventional cation exchange or net transfer thermobarometers (e.g., Powell & Holland, 2008). Moreover, recently parameterized activity–composition (a–x) relations for high-*T* amphibole, augitic clinopyroxene, and metabasic melt presented by Green *et al*. (2016) now permit quantitative investigation of high-grade metamorphism and partial melting of mafic rock types using a phase diagram-based approach (e.g., Green *et al*., 2016; Palin *et al*., 2016a, 2016b). The phase equilibria modelling was performed in the Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O–TiO_2–O_2 (NCKFMASHTO) system using GeoPS version 2.1 (http://www.geology.ren/) and THERMOCALC version 3.45 (Powell & Holland, 1988, update 2016). Both modeling methods yielded nearly identical *P*–*T* pseudosections. The simulations used the internally consistent thermodynamic dataset of Holland & Powell (2011; update ds62, 6 February, 2012), and the following a–x relations: metabasite melt, augite and hornblende (Green *et al*., 2016); garnet, orthopyroxene, biotite and chlorite (White *et al*., 2014a); olivine and epidote (Holland & Powell, 2011); magnetite–spinel (White *et al*., 2002); ilmenite–hematite (White *et al*., 2000); C1 plagioclase and K-feldspar (Holland & Powell, 2003); and muscovite–paragonite (White *et al*., 2014b). Pure phases included quartz, albite, rutile, sphene (titanite), and aqueous fluid (H_2O).

PETROLOGY AND GEOCHEMISTRY

Twenty-six mafic migmatite samples with banded or massive structures were studied. Their mineral assemblages, modes, and sampling locations are listed in Supplementary Data Table S1. These migmatites have similar assemblages of garnet, amphibole, plagioclase, epidote, muscovite, quartz, ilmenite, rutile and zircon (Figs. 3a–f and 4); however, they have highly variable mineral modes, with garnet of 2–60 vol.%, amphibole of 10–45 vol.%, plagioclase of 5–45 vol.%, epidote of 3–29 vol.%, muscovite of 2–15 vol.%, and quartz of 1–5 vol.%. These variations reflect different volume proportions of melanosome and leucosome, and variable degrees of retrogression. Garnet commonly occurs as coarse-grained porphyroblasts and contains inclusions of amphibole, plagioclase, quartz, ilmenite and rutile (Fig. 4). Some garnet grains are anhedral and are partially replaced by amphibole and epidote along their rims and within fractures (Fig. 4). Amphibole occurs as coarse-grained porphyroblasts and as a fine-grained matrix phase. The amphibole porphyroblasts have quartz and rutile inclusions, and exsolved ilmenite needles in their cores, and are partially replaced by epidote along their margins (Fig. 4a–d). Plagioclase is mostly a fine-grained matrix phase, and is partially replaced by scattered epidote and muscovite within their interior and along their rims (Fig. 4).

Mineral composition data were obtained from thirteen mafic migmatite samples, with analytical results given in Supplementary Data Tables S2–S6. Garnet porphyroblasts from the different samples have similar compositional ranges, with pyrope of 0.176–0.294, grossular of 0.150–0.198, almandine of 0.504–0.629, and spessartine of 0.014–0.048. Garnet porphyroblasts from five samples have similar and relatively homogenous compositions in their cores (Figs. 5 and 6 a–e; Supplementary Data Table S2). However, the garnet porphyroblast in sample 76-42 has an increasing grossular component and decreasing pyrope component from core to rim (Fig. 6f). Some garnet porphyroblasts have increasing almandine and spessartine components and decreasing pyrope component in their rims (Figs. 5 and 6a, c; Supplementary Data Table S2).

In general, all amphiboles from different textural domains can be classified as calcic amphibole, with Na_B = 0.252–0.503, Ca_B = 1.497–1.748 (23 oxygen's on an anhydrous basis) Supplementary Data Table S3), including pargasite and ferropargasite. Amphibole porphyroblasts show increasing Al and decreasing Mg from core to rim (Fig. 7). The TiO_2 and FeO contents of the cores of amphibole porphyroblasts may have decreased significantly from their initial (growth) compositions due to exsolution of abundant ilmenite needles (Fig. 4d) taking up high TiO_2 and FeO contents. Matrix amphibole has similar compositions to the rims of amphibole porphyroblasts (Supplementary Data Table S3), with lower Mg and higher Al than porphyroblast cores (Fig. 8).

Plagioclase from thirteen migmatite samples has similar compositions, with anorthite components of 0.090–0.210 and albite components of 0.788–0.908 (Supplementary Data Table S4). Epidote in thirteen samples has similar compositions with Fe^{3+} = 0.429–0.732 (pfu), X_{Mg} (=Mg/(Mg+Fe) × 100) of 0.29–2.9%, and pistacite (Ps = $Fe^{3+}/(Fe^{3+}+Al)$) content of 0.151–0.261 (Supplementary Data Table S5). Muscovites (white micas) include phengite, with Si of 3.082–3.223 (pfu; O = 11) and Na = 0.106–0.298, and paragonite with Si of 2.977–3.021 and Na = 0.738–0.788 (Supplementary Data Table S6).

These petrological and compositional features show that the mafic migmatite contains at least two generations of mineral assemblages. The early assemblage is represented by the cores of garnet and amphibole porphyroblasts, and their inclusions of plagioclase, quartz, rutile and ilmenite. The late assemblage is represented by the rims of garnet and amphibole porphyroblasts, and matrix amphibole, plagioclase, epidote, muscovite, quartz, ilmenite and rutile. If recrystallization occurs continuously during prograde metamorphism (Guiraud *et al*., 2001), but texturally late assemblages reflect final melt crystallization, the early mineral assemblage corresponds with maximum temperatures, while the late assemblage reflects crystallization of the last remaining melt upon crossing the solidus, down-grade.

Whole-rock geochemical analyses of 26 samples show mafic migmatite from the Zhaxi–Milin area varies in composition, with total ranges as follows: SiO_2 of 42.98–54.27 wt.%, Al_2O_3 of 16.48–23.57 wt.%, Fe_2O_3 of 0.92–7.34 wt.%, FeO of 1.06–11.51 wt.%, MgO of 2.48–5.30 wt.%, TiO_2 of 0.39–2.41 wt.%, MnO of 0.12–0.33 wt.%, CaO of 6.66–10.09 wt.%, Na_2O of 3.00–4.81 wt.%, and K_2O of 0.40–1.38 wt.% (Supplementary Data Table S7; Fig. 9). The garnet- and amphibole-rich rocks have lower SiO_2, and higher FeO^{total} than the plagioclase-rich rocks. The mafic migmatite and low-grade meta-gabbro mostly have fractionated REE patterns with LREE enrichment and HREE depletion relative to chondrites except for the garnet-rich sample 76-6 that has a HREE-rich pattern (Fig. 10a). These mafic rocks are also enriched in Rb, Ba, Sr and K, and depleted in Nb, Ta, Zr and Ti relative to primitive mantle (Fig. 10b). The migmatite samples mostly have higher Al_2O_3, TiO_2, FeO^{total}, MgO, CaO, Y, Co, V and HREE contents, and lower SiO_2, Na_2O, K_2O, Rb, Sr, Ba, and LREE contents compared with the low-grade meta-gabbros (Supplementary Data Table S7; Figs. 9, 10 and 11).

ZIRCON U–Pb AGES AND TRACE ELEMENT COMPOSITIONS

More than 1000 zircon grains were separated from each of five mafic migmatites and one felsic migmatite. Zircon occurs both as inclusions and as a matrix mineral. The zircon U–Pb age and trace element data are given in Supplementary Data Table S8. Cathodoluminescence images of representative zircon grains are shown in Fig. 12, and zircon U–Pb concordia diagrams and chondrite-normalized REE patterns are shown in Figs. 13 and 14.

Zircon from samples 76-12, 76-33, and 77-11 mostly has stubby prismatic shapes and clear core-rim textures (Fig. 12a–c). Core domains mostly have wide-banded zones, whereas the zircon rims have slight patchy zoning or no zoning. Fifteen analytical spots from the zircon core domains of sample 76-12 yielded similar and concordant $^{206}Pb/^{238}U$ ages of 91.9–84.1 Ma, with a weighted mean age of 87.2 ± 1.2 Ma (2σ, MSWD = 1.04) and average Th/U value of 0.451. Nine spots from the zircon rim domains of this sample yielded concordant and slightly younger $^{206}Pb/^{238}U$ ages of 86.0–79.3 Ma, with a mean age of 83.0 ± 1.6 Ma (2σ, MSWD = 0.94; Fig. 13a) and average Th/U value of 0.373. For sample 76-33, eleven analytical spots from the core domains yielded concordant $^{206}Pb/^{238}U$ ages of 92.6–81.5 Ma, with a mean age of 86.0 ± 1.8 Ma (2σ, MSWD = 1.7) and average Th/U value of 0.447, and 10 spots from the rim domains yielded concordant and slightly younger $^{206}Pb/^{238}U$ ages of 87.5–76.4 Ma, with a mean age of 82.8 ± 2.1 Ma (2σ, MSWD =0.99; Fig. 13b) and average Th/U values of 0.290. For sample 77-11, twelve analytical spots on the zircon cores yielded concordant $^{206}Pb/^{238}U$ ages of 160.0–153.8 Ma, with a mean age of 157.0 ± 1.4 Ma (2σ, MSWD = 0.49) and average Th/U value of 1.052, and fifteen spots on the rims yielded nearly concordant and young $^{206}Pb/^{238}U$ ages of 87.1–81.4 Ma, with a mean age of 85.1 ± 0.9 Ma (2σ, MSWD = 1.9; Fig. 13c) and average Th/U value of 0.353. The core domains of zircon have relatively high HREE contents (HREE = Er + Tm + Yb + Lu = 207–3102 ppm, av. 701 ppm) and markedly fractionated HREE patterns with significant negative Eu anomalies (δEu = 0.07–0.47), whereas the zircon rims have relatively low HREE contents (6.47–242 ppm, av. 72.6 ppm) and weakly fractionated HREE patterns, mostly with slight negative Eu anomalies (δEu = 0.18–0.86; Supplementary Data Table S8; Fig. 13d–f).

Zircon from the samples 76-6, 76-7 and 76-42 mostly shows subrounded forms and weakly patchy zones in cathodoluminescence images (Fig. 12d–f). Eleven analytical spots from the zircons of sample 76-6 yielded concordant $^{206}Pb/^{238}U$ ages of 90.0–85.6 Ma, with a mean age of 87.2 ± 2.1 Ma (2σ, MSWD = 0.19; Fig. 14a) and average Th/U of 0.314. Fourteen analytical spots from zircon in sample 76-7 yielded similar and concordant $^{206}Pb/^{238}U$ ages of 89.4–82.2 Ma, with a mean age of 86.6 ± 1.6 Ma (2σ, MSWD = 0.41; Fig. 14b) and average Th/U of 0.220. Ten analytical spots from zircon in sample 76-42 yielded sub-concordant $^{206}Pb/^{238}U$ ages of 89.9–83.5 Ma, with a mean age of 86.5 ± 1.8 Ma (2σ, MSWD = 0.52; Fig. 14c) and average Th/U value of 0.219. Zircon from these three samples has relatively low HREE contents (8.57–68.3 ppm, av. 27.7 ppm) and slightly fractionated or flat HREE patterns, mostly with slight negative Eu anomalies (δEu = 0.40–0.93; Supplementary Data Table S8; Fig. 14d–f).

METAMORPHIC AND ANATECTIC CONDITIONS

Metamorphic conditions of mafic migmatite were constrained by phase equilibria modelling for sample 76-12 based on the measured bulk rock composition (Supplementary Data Table S7). Modelled O_2 was calculated by the measured whole-rock FeO and Fe_2O_3 contents, with bulk-rock Fe^{3+}/Fe^{total} equal to $(2\times O)/FeO^{total}$. H_2O content was calculated from the volume proportions and water contents of hydrous minerals (amphibole, epidote and muscovite) determined via point counting, assuming stoichiometric OH. The modelled *P–T* pseudosection ranges from 3–20 kbar and 600–900 °C. The solidus is water-saturated and has a negative *P–T* slope between ~705 °C at ~3 kbar and ~625 °C at ~9.2 kbar and is water-undersaturated between ~9.2 kbar and 20 kbar (Fig. 15). Garnet is stable above ~8–12 kbar, and rutile is stable above ~11–13.5 kbar between 740 and 900 °C. Epidote is stable above 6–20 kbar between 600 and 845 °C. Muscovite is unstable above 600 °C at ~7.8 kbar and ~835 °C at 20 kbar. The observed early peak-metamorphic mineral assemblage of Grt + Amp + Pl + Qtz + Ilm + Rt is stable at *P–T* conditions of 11–17.6 kbar and 740–880 °C in the presence of melt, and the late retrograde mineral assemblage of Amp + Grt + Pl + Ep + Ms + Qtz + Rt is stable at 11.2–16.2 kbar and 600–730 °C in the absence of melt (Fig. 15a).

Metamorphic conditions of the migmatite are further constrained by mineral compositional isopleths. The calculated garnet X_{Mg} (=Mg/(Mg+Fe+Ca)) and plagioclase An (=Ca/(Ca+Na+K)) isopleths are shown in Fig. 15b. The cores of garnet porphyroblasts have higher X_{Mg} than their rims, consistent with garnet core formation at high-*T* peak metamorphic conditions. The highest X_{Mg} isopleths (0.26–0.28) of the garnet cores intersect the stability field of the peak-metamorphic mineral assemblage at ~850–880 °C between 15 kbar and 17 kbar (Fig. 15b). Therefore, the peak metamorphic conditions of the migmatite appear equivalent to the upper stability field of the peak mineral assemblage (the yellow-filled circle marked with P (= "Peak"); Fig. 15). At such peak-metamorphic conditions, the migmatite contains ~16–18 vol. % melt (Fig. 15b). The minimum An isopleth (0.16) of the matrix plagioclases intersects the stability field of retrograde mineral assemblage at 11.5–14.5 kbar between 600 °C and 720 °C (Fig. 15b), and intersects the system solidus at ~720 °C and ~14.5 kbar (the yellow-filled circle marked with R (="Retrograde"; Fig. 15b).

DISCUSSION

Metamorphic and anatectic *P–T* path

Potential melt segregation and loss complicates modelling the metamorphic evolution of partially melted rocks (Indares *et al*., 2008; Groppo *et al*., 2012; Guilmette *et al*., 2011). The studied migmatite likely underwent significant melt loss (see discussion below). Nonetheless, the pseudosection constructed using the measured bulk-rock composition (representing the present rock composition after partial melt loss), allows the exploration of the phase equilibria in the high-*T* part of the metamorphic history and the assessment of the near-to-the peak and the retrograde evolution (White *et al*., 2004; Indares *et al*., 2008; Groppo *et al*., 2012). Sample 76-12 contains pervasive leucosomes, and displays a massive structure (Figs. 3b and 4c). In this case, XRF-measured bulk-rock compositions from a large rock segment (~5 kg) are considered as the effective composition of the rock system. Therefore, the *P*–*T* conditions for the early and late mineral assemblages as constrained via phase equilibria modelling can be interpreted to be the peak-metamorphic and retrograde conditions of the mafic migmatite, respectively (Fig. 15). Combining the peak with retrograde metamorphic conditions, the migmatite records a segment of an anticlockwise *P*–*T* path, characterized by approximately isobaric cooling from high-*P* granulite facies to upper amphibolite facies, with final textural equilibration upon partial melt crystallization (White and Powell, 2010; Palin *et al*., 2018).

Recent studies (Niu *et al*., 2019; Qin *et al*., 2019) indicate that the garnet amphibolites and garnet-mica schists from this study area underwent Late Cretaceous high-*P* granulite-facies metamorphism and partial melting under *P*–*T* conditions of ~16–17 kbar and ~820 °C, similar to the present phase equilibria modelling for the mafic migmatites. Previous high-*T* and high-*P* experimental studies demonstrate that the mafic rocks with garnet, amphibole, plagioclase, rutile and melt, but without clinopyroxene can be formed at *P*–*T* conditions of ~820–880 °C and ~10–16 kbar (López & Castro, 2001), and ~800–850 °C and ~13–16 kbar (Qian & Hermann, 2013). These experiments support the interpretation that the mafic migmatites, consisting mainly of garnet and amphibole, can be stable under metamorphic conditions of the high-*P* granulite-facies.

The studied 26 mafic migmatite samples from different locations have variable mineral modes (Supplementary Table S1) and whole-rock compositions (Supplementary Table S7; Fig. 9). However, as described above, the major minerals garnet, amphibole, plagioclase, epidote and muscovite in all the analyzed 13 migmatite samples have similar compositions, indicating that these minerals have crystallized from the migmatites with similar whole-rock compositions before partial melt loss. Therefore, the average compositions of the 26 mafic migmatite samples we studied should be considered representative of the compositions of the migmatites. The whole-rock composition of sample 76-12 is the nearly same as the average composition of the 26 migmatites (Supplementary Table S7). Therefore, we consider that the sample 76-12 is representative of an effective whole-rock composition for the studied mafic migmatites, and the results calculated by this single sample can be applied generally to understand the petrogenetic evolution of the mafic migmatites from the Zhaxi–Milin area.

It is difficult to constrain the prograde metamorphic *P*–*T* path due to lack of petrographic information in these samples. However, several lines of evidence suggest that the rocks underwent prograde heating and burial. (1) The migmatites show structure consistent with high-*T* metamorphism and anatexis, such as the diffusive and banded leucosomes, concordant and discordant vein leucosomes, and equant garnet in the leucosomes (Fig. 3). Apparently, these migmatites are not the direct crystallization product of deep-emplacement gabbroic magma. (2) Late Jurassic diorites and Late Carboniferous sedimentary rocks, together with host Late Cretaceous gabbro, underwent Late Cretaceous high-*P* metamorphism and anatexis (Niu *et al*., 2019; Qin *et al*., 2019; this study). This indicates that the plutonic and sedimentary rocks at middle to upper crustal levels were buried in the thickened lower crust after the emplacement of Late Cretaceous gabbro. (3) Garnet in the pelitic migmatites, which occurs within mafic migmatites and has the same metamorphic ages as the host mafic migmatites, displays growth compositional zones that suggest prograde heating and loading (Qin *et al*., 2019). Thus, we consider that the mafic and pelitic migmatites have a common prograde metamorphic *P*–*T* path from the upper-middle crust to the thickened lower crust.

Gabbros in magmatic arcs typically crystallize at a relatively high temperature of >900 °C and low pressure of <1.0 GPa, and most then experience isobaric cooling to reach thermal equilibrium with the adjacent middle or lower crust before high-*T* metamorphism and remelting (Pattinson, 2003; Berger *et al*., 2009; Castro *et al*., 2013). Some studies demonstrate that meta-gabbros of arc roots undergo prograde metamorphism from amphibolite-facies to granulite-facies conditions (e.g., Berger *et al*., 2009). For instance, meta-gabbros in the lower crust of the exhumed Kohistan arc, western segment of Trans-Himalayan magmatic arc, record a prograde metamorphic *P*–*T* path from ca. 700 °C and 7 kbar to ca. 900 °C and 13 kbar (Yoshino *et al*., 1998; Yoshino & Okudaira, 2004).

Prograde metamorphism and associated partial melting cannot be reasonably constrained by the phase equilibria modelling using the measured whole-rock composition because the migmatite is an open system involving melt loss and/or melt gain (White *et al*., 2004; Indares *et al*., 2008; Groppo *et al*., 2012). However, phase equilibria modelling under closed and open systems by Palin *et al*. (2016b) shows that during prograde metamorphism between 700 °C/11 kbar and 900 °C/15 kbar water-saturated mafic rocks melt via amphibole-breakdown melting, garnet grows as a peritectic phase, and plagioclase remains stable. This result is consistent with our petrographic observations and modelling, which show that the garnet, plagioclase, amphibole, quartz and melt form as the peak-metamorphic phases of the mafic migmatite.

Mineral and melt mode variations along the retrograde *P*–*T* path of the mafic migmatite, calculated by the phase equilibria modelling for sample 76-12 using the measured whole-rock composition and the same approach as *P*–*T* pseudosection modelling described above, show that the modes of garnet and plagioclase decrease, whereas volumes of amphibole, muscovite, epidote and quartz increase. These predictions are consistent with petrographic interpretations, that plagioclase is partly replaced by epidote and muscovite, and garnet by amphibole and epidote during retrograde metamorphism and residual melt crystallization.

Protolith and metamorphic ages

High HREE contents, fractionated REE patterns and high Th/U values are typical of magmatic zircons, whereas low HREE contents, slightly fractionated to flat HREE patterns and low Th/U ratios are typical of zircon from high-grade metamorphic and anatectic rocks with garnet (Schaltegger *et al*., 1999; Vavra *et al*., 1999; Harley *et al*., 2007). In this context, the internal zoning and compositional features of zircon cores from our samples – pronounced CL banding, relatively high HREE contents, fractionated REE patterns, negative Eu anomalies, relatively high Th/U – indicate that they likely formed during magma crystallization. In contrast, the internal zoning and compositions features of zircon rims – weak patchy CL zoning, relatively low HREE contents, weakly fractionated HREE patterns, small or no Eu anomalies, and lower Th/U – indicate that they formed during metamorphism and anatexis of the migmatites. Because formation of peritectic garnet would explain relatively low HREE contents and slightly fractionated HREE patterns, metamorphic rims of zircons in our samples probably formed during or after garnet growth. Decomposition of plagioclase characterizes retrograde metamorphism of mafic migmatites and would cause any coevally formed metamorphic zircons to have small or absent Eu anomalies, as we observe (Fig. 13d–f). We propose that the ages of ~87 Ma, ~86 Ma and ~157 Ma obtained from the zircon cores represent protolith (intrusion) ages of the mafic and felsic migmatites, whereas the ~83 Ma and ~85 Ma ages obtained from the zircon rims represent the retrograde metamorphism and melt crystallization. That is, zircon ages suggest that the Late Jurassic (~157 Ma) diorite and the Late Cretaceous (~87–86 Ma) gabbro record Late Cretaceous (~85–83 Ma) metamorphism. If so, metamorphism of Late Cretaceous gabbros occurred soon after their emplacement. These ages are consistent with previous studies that indicated the meta-plutonic and meta-sedimentary rocks of the eastern Gangdese arc have Late Jurassic to Late Cretaceous protolith ages, and Late Cretaceous metamorphic ages (Guo *et al*., 2013; Qin *et al*., 2019; Zhang *et al*., 2020).

For mafic migmatite samples 76-6, 76-7 and 76-42, the patchy zoning, lack of a core–rim structure, low Th/U values (averages of 0.314, 0.220 and 0.219) and HREE contents, and slightly fractionated HREE patterns, mostly with small negative Eu anomalies (Supplementary Data Table S8; Figs. 12d–f and 14d–f) indicate that the zircon is of metamorphic origin, like the rims of zircon with core–rim structure. The zircon ages of ~87 Ma probably represent the metamorphic age of the migmatites. However, these zircon ages are older than the metamorphic zircon rims from the three other migmatite samples (~85–83 Ma). We consider that the relatively older ages represent prograde metamorphism and anatexis, and the younger ages represent retrogression and melt crystallization.

Low- to high-grade meta-gabbro of the Lilong batholith has been dated with zircon U–Pb methods, and ranges from 100 to 88 Ma (Fig. 1b; Zhang *et al*., 2010a, 2014a; Guo *et al*., 2013; Ma *et al*., 2013a, 2013b). Possibly, our relatively young "protolith" ages (~87–86 Ma) from the mafic migmatites were reset during metamorphism and anatexis, whereas the older ages (~90–100 Ma), mostly obtained by the magmatic zircons of low-grade meta-gabbro, reflect protolith ages of mafic migmatites.

In summary, the metamorphism and anatexis of the migmatites from the Late Cretaceous lower crust of eastern Gangdese arc occurred during the Late Cretaceous, ~83–87 Ma. This event relates to the late subduction of Neo-Tethyan oceanic lithosphere that initiated in the Late Triassic and ended in the Late Cretaceous at ~65 Ma, and was slightly later than the Late Cretaceous main pulse of the Gangdese arc magmatism at ~90 Ma (Ji *et al*., 2009; Zhu *et al*., 2018; Zhang *et al*., 2020).

Partial melting, melt composition and melt loss

The deeply exposed lower crust of magmatic arcs commonly consists of high-grade metamorphic rocks with characteristic features of intense partial melting and melt loss (Hacker *et al*., 2015; Miller & Snoke, 2009; DeBari & Greene, 2011; Jagoutz & Schmidt, 2012; Chapman *et al*., 2014; Ducea *et al*., 2015). Our field and petrological observations show that the migmatites from the eastern Gangdese lower arc crust contain abundant (up to ~15–20 vol.%) leucosome bands and networks (Fig. 3a, 3c and 3f–h), concordant and discordant vein leucosomes (Fig. 3d, 3e and 3i), high modes (~15–50 vol.%) of garnet, and equant large (up to 2 cm in diameter) garnet crystals in the leucosomes (Fig. 3d–f). Moreover, peak-metamorphic minerals experienced limited retrogression. These features all suggest that the studied migmatites underwent partial melting and associated melt segregation, migration and loss. Our phase equilibria modelling shows that around 16–18 wt. % melt was produced at the peak-metamorphic stage of the mafic migmatite (Fig. 15b). This estimate is a minimum because the sample could have lost melt previously. Such large volumes of partial melt and contemporary deformation of the migmatites (Fig. 3a) provide very favorable conditions for melt segregation, migration and extraction. Previous studies suggest that partial melt may be lost from a system when its proportion rises above a critical threshold of 7–10 vol. % (Rosenberg & Handy, 2005; Brown, 2007, 2010; Indares *et al*., 2008; Groppo *et al*., 2012), and that compaction and deformation play an important role in segregation and extraction of melts in lower crusts (e.g., Vigneresse *et al*., 1996; Brown, 2007).

Our modelling shows that the partial melt generated by anatexis of mafic migmatite under the peak-metamorphic condition of ~850 ˚C and ~16 kbar have a granitic composition with SiO_2 = 70.02, Al_2O_3 = 18.51, FeO^{total} = 0.11, MgO = 0.04, CaO = 2.55, Na_2O = 5.10 and K_2O = 3.66 (in wt. %; normalized to anhydrous). The melt composition seems to be inconsistent with the observed leucosomes containing abundant plagioclase and lacking K-feldspar. However, leucosomes often have distinct compositions from an anatectic melt due to fractional crystallization and separation of the fractionated melt (e.g., Sawyer, 2008).

As shown in Fig. 1b, gabbro of the Lilong batholith was metamorphosed under a range of metamorphic conditions, from granulite-facies in the northeast, through amphibolite-facies, to epidote-amphibolite and greenschist-facies in the southwest. The studied mafic migmatites are products of high-grade metamorphism, partial melting and melt loss of gabbros from the Zhaxi–Milin area, whereas gabbro near the Lilong village underwent only greenschist- to epidote amphibolite-facies metamorphism (Fig. 1b). We assume the low-grade meta-gabbros have similar whole-rock compositions to the expected protoliths of the mafic migmatites prior to melt loss. However, the mafic migmatites mostly have higher TiO_2, Al_2O_3, FeO^{total}, MgO, CaO, Y, Co, V, and HREE contents, and lower SiO_2, Na_2O, K_2O, Rb, Sr, Ba, Th, and LREE (Figs. 9, 10 and 11). These differences are consistent with loss of granitic melt during partial melting, consistent with predictions of phase equilibria modeling. The extracted melts have relatively high Sr and LREE, and low Y and HREE concentrations, and therefore show compositional affinity to adakitic rocks with high Sr/Y ratios and significantly fractionated REE patterns. This agrees with the widely accepted view that high Sr/Y adakitic granitoids can be generated by partial melting of high-*P* mafic granulites or eclogites in the thickened lower crust (e.g., Chung *et al*., 2003; Hou *et al*., 2004).

Late Cretaceous crustal thickening

In places, the Gangdese arc is twice as thick as average continental crust, but there is still much disagreement about the timing and mechanisms of the arc crustal thickening. Although some previous studies of structure and magmatism suggest that the Gangdese arc underwent enhanced crustal thickening during the Mesozoic prior to India–Asia collision (England & Searle, 1986; Murphy *et al*., 1997; Ding & Lai, 2003; Ding *et al*., 2003, 2014; Kapp *et al*., 2005, 2007a, 2007b; Ji *et al*., 2012), most workers propose that thickening occurred after c. 55 Ma as a direct consequence of continental convergence (Molnar *et al*., 1993; Yin & Harrison, 2000; Williams *et al*., 2001; Chung *et al*., 2003, 2005; Hou *et al*., 2004, 2015a; Yin, 2006; Guo *et al*., 2007; Mo *et al*., 2008; Wang *et al*., 2014a, 2014b; Yang *et al*., 2015, 2016a, 2016b). For example, Zhu *et al*. (2017) argued that the Gangdese arc had a normal crustal thickness of ~37 km before collision, thickened to ~50–58 km at 55–45 Ma, and finally attained its present thickness of ~68 km at ~20–10 Ma. The new results presented here show that both Late Jurassic (~157 Ma) and Late Cretaceous (~90–100 Ma) igneous rocks underwent Late Cretaceous (~83–87 Ma) high-*P* granulite-facies metamorphism at 15–17 kbar. This suggest that the eastern Gangdese arc crust was at least 50–55 km thick during late stage subduction of the Neo-Tethyan oceanic lithosphere (Fig. 16).

Common thickening mechanisms of continental magmatic arc crust include accretion and loading of mantle-derived magma, and tectonic shortening and thrusting. The anticlockwise *P*–*T* path found by our work, characterized by a heating and burial prograde path and a near-isobaric and cooling retrograde path, is consistent with a continental arc setting that is related to the accretion of mantle-derived magmas (Bohlen, 1987, 1991; Harley, 1989). Moreover, this study shows that the Late Cretaceous gabbro underwent high-*P* metamorphism soon after its emplacement, indicating that mantle-derived magma accretion and loading probably were the main mechanism for thickening of the Gangdese arc crust (Fig. 16). This interpretation is consistent with the pulse of Late Cretaceous mantle-derived igneous rocks throughout the Gangdese arc (Wen *et al*., 2008a, 2008b; Ji *et al*., 2009; Zhang *et al*., 2010a; Ma *et al*., 2013a, 2013b; Zhu *et al*., 2018). Previous studies also indicate that magmatic accretion and loading resulting in crustal thickening and granulite-facies metamorphism is one of the most important processes in subduction-related magmatic arc roots (Brown, 1996; Gibson *et al*., 1988; Yoshino *et al*., 1998; Müntener *et al*., 2000; Collins, 2002; Yoshino & Okudaira, 2004; Stowell *et al*., 2010). For the eastern Gangdese arc, pre-collisional crustal thickening could also have occurred via tectonic shortening and thrusting due to the Late Cretaceous compression during the shallow subduction of the young

Neo-Tethyan oceanic lithosphere soon after subduction of a spreading mid-oceanic ridge (Murphy *et al*., 1997; Ding & Lai, 2003; Kapp *et al*., 2005; Mo *et al*., 2007; We *et al*., 2008a; Zhang *et al*., 2010).

Chemical differentiation of juvenile crust

The continental crust is thought to be chemically stratified into a felsic upper portion and mafic lower portion (Rudnick & Gao, 2003). Such chemical differentiation occurs over a wide range of time scales and different geodynamic settings. The Gangdese arc underwent significant crustal thickening during the Indo-Asian collision, and the Cenozoic granitoids, especially the adakitic rocks, are considered to be the products of partial melting of thickened mafic lower crust of Gangdese arc (Chung *et al*., 2003, 2005; Hou *et al*., 2004; Chen *et al*., 2011; Guan *et al*., 2012; Zhang *et al*., 2015; Zhu *et al*., 2017; Ding & Zhang *et al*., 2018). This interpretation implies that the Gangdese arc crust experienced chemical differentiation during the arc-continent collision.

Our results suggest that the eastern Gangdese arc underwent crustal thickening, partial melting of thickened lower crust, and formation of granitic melts during Late Mesozoic subduction of oceanic lithosphere prior to collision (Fig. 16). The studied migmatites, derived from the Late Cretaceous and Late Jurassic gabbroic to dioritic intrusions, are widespread in the exposed lower arc crustal sections of the eastern Gangdese arc (Figs. 1b and 2). Moreover, these intrusions have chemical affinities to arc-type magmatic rocks, characterized by fractionated REE patterns, enrichment of Rb, Ba, Sr and K, and depletion of Nb, Ta and Ti (Fig. 10). This is consistent with that the Late Cretaceous was the main period of mantle-derived magmatism and juvenile crustal growth of the arc (Zhu *et al*., 2018; Zhang *et al*., 2019a). If so, the Late Cretaceous migmatites are probably widespread in the unexposed lower arc crust, and significant amounts of melts extracted from the migmatites provide the potential source for the granitic rocks in the upper arc crust. In fact, the Late Cretaceous (~85–70 Ma) is the main period of granitic magmatism, characterized by the widespread occurrence of granitoids (with SiO_2 contents of 65–75 wt. %) in the eastern Gangdese arc (Wen *et al*., 2008a, 2008b; Ji *et al*., 2009, 2014; Zhu *et al*., 2018; Zhang *et al*., 2019a). The Late Cretaceous granitoids in the Wolong batholith extend ca. 60–70 km and were emplaced in the upper crustal level of the eastern Gangdese arc (Fig. 1b), yet were derived from partial melting of juvenile lower crust (Wen *et al*., 2008a; Tang *et al*., 2019). Moreover, the Wolong granitoids show a geochemical affinity to adakite-like rocks, consistent with their derivation from the partial melting of thickened lower crust (Wen *et al*., 2008a; Tang *et al*., 2019). The presence of voluminous adakitic granitoids provides robust evidence for the Late Cretaceous thickening and remelting of the juvenile crust of the eastern Gangdese arc. In this case, we consider that the Late Cretaceous granitoids are likely products of partial melting of the thickened lower crust that is composed of mafic and felsic migmatites derived from the granulite-facies metamorphism of the Late Cretaceous and Late Jurassic basaltic to intermediate intrusions (Fig. 16). Therefore, the chemical differentiation of arc crust induced by partial melting of the thickened lower crust has occurred during building of the juvenile crust of the Gangdese arc, prior to Indo-Asian collision.

CONCLUSIONS

Widespread mafic and felsic migmatites from the eastern Gangdese lower arc crust underwent peak metamorphism at granulite facies *P*–*T* conditions of 850–880 ˚C and 15–17 kbar, and record an anticlockwise *P*–*T* path, characterized by quasi-isobaric cooling. The migmatites have protolith ages of Late Cretaceous (~90–100 Ma) and Late Jurassic (~157 Ma), and high-grade metamorphic and anatectic ages of Late Cretaceous (~83–87 Ma), indicating that metamorphism occurred soon after the emplacement of Late Cretaceous gabbroic pluton. A significant volume (>16 vol. %) of granitic melt was generated by the partial melting of the voluminous migmatites, and provided a potential source for Late Cretaceous granitoids. The eastern Gangdese magmatic arc experienced significant crustal thickening due to the accretion and loading of mantle-derived magma and coeval tectonic deformation. Simultaneous chemical differentiation occurred through intense partial melting of thickened lower crust and the migration and ascension of generated melt during the building of juvenile crust prior to the arc-continent collision.

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Figure Captions:

Fig. 1 (a) Simplified geological map of the Tibetan Plateau, showing the distribution of the magmatic rocks of the Gangdese arc along the southern part of the Lhasa terrane. ATF-Altyn Tagh fault, BNS-Bangong-Nujiang Suture zone, IYS-Indus-Yarlung Tsangpo Suture zone, JSS-Jinsha Suture zone, KKF-Karakorum fault, KLS-Kunlun Suture zone, LSF-Longmenshan fault, LSS-Longmu Co-Shuanghu Suture zone. (b) Geological map of the eastern Gangdese arc (the Eastern Himalayan Syntaxis; modified after Zhang *et al*., 2020). Published zircon U–Pb ages and sampling locations of the low-grade meta-gabbros of Lilong batholith are from Zhang *et al*. (2010a), Guo *et al*. (2013) and Ma *et al*. (2013a).

Fig. 2 Geological map of the eastern Gangdese lower arc crust (modified after Zhang *et al*., 2020). Stars with sample numbers show approximate locations of representative migmatite samples used in this study.

Fig. 3 Field photographs of mafic and felsic migmatites. The coins for scale are 2 cm in diameter. (a) Mafic migmatite with banded (stromatic) structure, defined by alternating bands of melanosome and leucosome. The melanosome consists mainly of amphibole, garnet and plagioclase, and the leucosome contains abundant plagioclase and garnet, with minor quartz and amphibole. The parallel bands form a sub-vertical foliation. (b) Mafic migmatite with massive structure (diatexite); the garnet- and plagioclase-rich leucosomes homogenously distribute in the amphibole- and plagioclase-rich melanosomes. (c) Mafic migmatite contains diffuse leucosomes that consist of plagioclase, garnet and minor quartz and amphibole. (d) Mafic migmatite contains garnet-rich leucosome veins, sub-parallel to the host rock foliation. A late amphibolebearing and garnet-free granitic vein crosscuts the garnet-rich veins. (e) Mafic migmatite contains garnet-rich leucosome vein. (f) Peritectic garnet and diffuse leucosomes of the mafic migmatite. Euhedral, equant, and large garnet crystals (up to 2 cm) are set in a small volume of leucosome. (g) Felsic migmatite contains abundant garnet-bearing leucosome networks sub-parallel to the migmatite foliation. (h) Felsic migmatite shows banded structure, defined by alternating bands of melanosome and leucosome. The melanosome consists of plagioclase, biotite, amphibole, quartz and garnet, and the leucosome contains plagioclase, quartz and garnet. Some leucosomes occur as networks. (i) Felsic migmatite contains garnet-bearing leucosomes that occur as the concordant and discordant veins.

Fig. 4 Photomicrographs of the mafic migmatites (all plane-polarized light). Yellow lines across the garnet and amphibole grains in (a), (c) and (e) are locations of the compositional profiles in Figs. 6 and 7. (a, b) Sample 76-6, consisting mainly of garnet, amphibole, plagioclase and epidote with minor quartz, muscovite, rutile, and ilmenite. The garnet has inclusions of plagioclase, quartz and rutile, is partially replaced by amphibole and epidote along grain margins and fractures, and occurs locally as skeletal remnants. Plagioclase is partially replaced by fine-grained and scattered epidote and muscovite. (c, d) Sample 76-12, consisting mainly of garnet, amphibole, plagioclase and epidote with minor quartz, muscovite, rutile and ilmenite. Garnet porphyroblasts have quartz inclusions, and are partially replaced by amphibole and epidote. Amphibole porphyroblasts have rutile and quartz inclusions, and exsolved ilmenite needles in their cores (see inset image), and are partially replaced by epidote along their rims. Matrix plagioclase is partially replaced by fine-grained and scattered epidote and muscovite. (e) Sample 76-7, consisting mainly of garnet, plagioclase, amphibole, epidote, and minor muscovite, quartz, ilmenite and rutile. Garnet is partially replaced by amphibole and epidote along their margins and fractures. Plagioclase is partially replaced by fine-grained and scattered epidote and muscovite. (f) Sample 76-42, garnet porphyroblasts contain amphibole, rutile, ilmenite and quartz inclusions, and are partially replaced by amphibole and epidote along their rims and fractures.

Fig. 5 Compositional maps of garnet porphyroblasts of the mafic migmatites. Yellow lines are locations of the compositional profiles in Fig. 6. “Hotter” colors (red, yellow) are higher concentration than “cooler” colors (blue, black).

Fig. 6 Compositional profiles of garnet porphyroblasts of the mafic migmatites. The profile locations of samples 76-6, 76-7 and 76-12 are shown in Figs. 4 and 5 by the yellow lines.

Fig. 7 Compositional profiles of amphibole porphyroblasts of the mafic migmatites. The profile location of sample 76-6 is shown in Fig. 4a by the yellow line.

Fig. 8 Compositional variation of the calcic amphibole porphyroblast core (early amphibole) and porphyroblast rim and matrix (late amphibole) of the mafic migmatites.

Fig. 9 Harker diagrams of major and trace elements of the mafic migmatites and low-grade meta-gabbros.

Fig. 10 (a) Chondrite-normalized REE patterns and (b) primitive mantle-normalized trace element patterns of the mafic migmatites.

Fig. 11 Low-grade metamorphosed gabbro-normalized major and trace element patterns of the mafic migmatites. The thin lines with different colors refer to the 26 mafic migmatite samples listed in Supplementary Table S7. The red thick line is an average composition of the 26 samples. The normalized value is an average composition of the 16 low-grade meta-gabbro samples listed in

Supplementary Table S7.

Fig. 12 Cathodoluminescence images of zircon from the mafic (a, b, d–f) and felsic (c) migmatites, showing locations of analyzed spots (circles) and ages (in Ma). All scale bars are 100 μm.

Fig. 13 U–Pb concordia diagrams (a–c) and chondrite-normalized REE patterns (d–f) of zircon cores and rims from the mafic (samples 76-12 and 76-33) and felsic (sample 77-11) migmatites.

Fig. 14 U–Pb concordia diagrams (a–c) and chondrite-normalized REE patterns (d–f) of zircon without core-rim texture from the mafic migmatites (samples 76-6, 76-7, and 76-42).

Fig. 15 (a) *P*–*T* pseudosection calculated for mafic migmatite sample 76-12, using the measured whole-rock composition. Stability limits of representative minerals are shown by thick lines with different colors. The observed peak and retrograde mineral assemblages are shown in red fonts. (b) *P*–*T* pseudosection superimposed with isopleths of melt mode (in vol. %, thin red lines), X_{Mg} [= Mg/(Mg+Fe+Ca)] of garnet (blue lines), and An [=Ca/(Ca+Na+K)] of plagioclase (black dotted lines). The yellow-filled circles marked with P and R represent the calculated peak and retrograde metamorphic conditions, respectively. The red thick lines with arrows refer to the retrograde *P*–*T* path of the migmatite.

Fig. 16 Late Cretaceous tectonic model of the eastern Gangdese magmatic arc. Accretion and loading of mantle-derived magma (Late Cretaceous gabbro-diorites) and coeval tectonic deformation resulted in crustal thickening. The thickened lower crust underwent high-*T* and high-*P* metamorphism and associated partial melting. The migration and ascension of generated granitic melts resulted in the formation of Late Cretaceous I-type granites, and the chemical differentiation of juvenile arc crust.

SUPPLEMENTARY DATA TABLES

Table S1 Mineral assemblage, mode and sampling location of the migmatites and low-grade meta-gabbros

Table S2 Chemical compositions of garnet in the mafic migmatites (wt.%)

Table S3 Chemical compositions of amphibole in the mafic migmatites (wt.%)

Table S4 Chemical compositions of plagioclase in the mafic migmatites (wt.%)

Table S5 Chemical compositions of epidote in the mafic migmatites (wt.%)

Table S6 Chemical compositions of muscovite in the mafic migmatites (wt.%)

Table S7 Whole-rock major (wt.%) and trace (ppm) element compositions of the mafic migmatites and low-grade meta-gabbros

Table S8 Zircon U–Pb dating and trace element analytical results of the mafic and felsic migmatites