Early Cenozoic partial melting of meta-sedimentary rocks of the eastern Gangdese arc, southern Tibet, and its contribution to syn-collisional magmatism

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ABSTRACT

Continental magmatic arcs are characterized by the accretion of voluminous mantle-derived magmatic rocks and the growth of juvenile crust. However, significant volumes of meta-sedimentary rocks occur in the middle and lower arc crust, and the contributions of these rocks to the evolution of arc crust remain unclear. In this paper, we conduct a systematic study of petrology, geochronology and geochemistry of migmatitic paragneisses from the eastern Gangdese magmatic arc, southern Tibet. The results show that the paragneisses were derived from Late Carboniferous greywacke, and underwent an Early Cenozoic (69–41 Ma) upper amphibolite-facies metamorphism and partial melting at *P*–*T* conditions of ~11 kbar and ~740 °C, and generated granitic melts with enriched Hf isotopic compositions (anatectic zircon $\varepsilon_{Hf\,(t)}$ = −10.57 to +0.78). Our study shows that the widely distributed meta-sedimentary rocks in the eastern Gangdese arc deep crust have the same protolith ages of Late Carboniferous, and record northwestward-decreasing metamorphic conditions. We consider that the deeply buried sedimentary rocks resulted in the compositional change of juvenile lower crust from mafic to felsic, and formation of syn-collisional S-type granitoids. The mixing of melts derived from mantle, juvenile lower crust and ancient crustal materials resulted in the isotopic enrichment of the syn-collisional arc-type magmatic rocks of the Gangdese arc. We suggest that crustal shortening and thrusting, and the accretion of mantle-derived magma during the Indo-Asian collision transported the supracrustal rocks to the deep crust of the Gangdese arc.

INTRODUCTION

Continental magmatic arcs form at convergent plate margins where subduction-related magmas intrude overriding continental lithosphere. Such arcs are tectonically complex, typically form areas of thick crust, and have overall compositions that are higher in silica than island arcs (Ducea et al., 2015). As subduction is a global and continuous process, arc magmatism is thought to have been the primary mechanism for growth of the continental crust since the onset of plate tectonics in the Late Archean (Jagoutz and Schmidt, 2012; Jagoutz and Kelemen, 2015; Palin et al., 2016a; Palin et al., 2020). An increasing number of studies demonstrated that early formed supracrustal sedimentary rocks within a growing magmatic arc occur in the deep crust of arcs, and underwent high-grade metamorphism and partial melting (Saleeby et al., 1990; Whitney et al., 1999; Babeyko et al., 2002; Miller, et al., 2009; Ducea et al., 2015). However, contributions of the meta-sedimentary rocks to arc magmatism and as components of juvenile lower crust are rarely studied (Chapman et al., 2013, 2014; Chin et al., 2013; Xu et al., 2013; Zhang et al., 2013, 2015; Guo et al., 2020), and the mechanism of transporting the sedimentary rocks into the deep arc crust remains highly controversial (e.g., Chin et al., 2013; Guo et al., 2020; Zhang et al., 2020b). In addition, protoliths and deposition ages of the widely distributed meta-sedimentary rocks in the eastern Gangdese arc, southern Tibet, and spatial changes of metamorphic conditions and timing of exposure of deep crustal rocks also needs further study in order to construct crustal sections and a tectonic evolution history of the Gangdese arc.

The Gangdese magmatic arc, southern Tibet, records a complete growth history from long-lasting subduction of the Neo-Tethyan oceanic lithosphere during the Mesozoic to subsequent Cenozoic collision between the Indian and Asian continents, and represents one of the archetypal composite continental volcanic arcs in the world (Fig. 1A; e.g. Coulon et al., 1986; Debon et al., 1986; Harris et al., 1988a, b; Pearce and Mei, 1988; Yin and Harrison, 2000; Ding et al., 2003; Hou et al., 2004; Mo et al., 2005a, b). It is therefore an ideal natural laboratory for studying subduction- and collision-related magmatism, and continental crustal growth. The eastern Gangdese arc exposes a series of middle- to high-grade metamorphic rocks, which were exhumed during Neogene rapid uplift and erosion. These rocks are considered to be representative of middle and lower crustal components of the Gangdese arc (Searle et al., 2011; Zhang et al., 2014b, 2020b; Guo et al., 2020).

Here, we report petrological, geochemical and geochronological data for migmatitic paragneisses from the eastern Gangdese arc. Our results indicate that the paragneisses and associated rocks were derived from Late Carboniferous sedimentary rocks of the Lhasa terrane, and buried to the lower arc crust during the Early Cenozoic collisional orogeny. The large volumes of supracrustal rocks significantly changed the component of the juvenile lower crust and the isotopic compositions of syn-collisional magmatic rocks. These results provide key insights into the formation and evolution of the Gangdese arc and other continental magmatic arcs worldwide.

GEOLOGICAL SETTING

From north to south, the Himalayan-Tibetan orogen consists of the Songpan-Ganze, North Qiangtang, South Qiangtang, Lhasa, and Himalayan terranes (Fig. 1A) (Yin and Harrison, 2000, and references therein). The Gangdese magmatic arc is located within the southern Lhasa terrane, which is dominated by voluminous Cretaceous–Tertiary plutons – the Gangdese batholith – and the Paleocene Linzizong volcanic succession, with minor Triassic–Cretaceous intrusive and volcanic-sedimentary rocks (Fig. 1A; Allégre et al., 1984; Coulon et al., 1986; Debon et al., 1986; Pan et al., 2004; Chung et al., 2005; Wen et al., 2008a, b; Zhu et al., 2015, 2018; Wang et al., 2016). The Gangdese magmatic arc preserves a protracted and episodic period of magmatism spanning the Late Triassic (ca. 210 Ma) to the Late Miocene (ca. 10 Ma) (Mo et al., 2005a; Ji et al., 2009; Zhu et al., 2018). Mesozoic magmatism occurred during subduction of the Neo-Tethyan oceanic lithosphere, whereas Cenozoic magmatism is related to the continental collision (e.g., Chung et al., 2005; Mo et al., 2005b; Zhu et al., 2018; Zhang et al., 2018, 2019). The Early Cenozoic (65–40 Ma) magmatic rocks are represented by the Gangdese batholith and Linzizong volcanic succession that spread throughout the southern Lhasa terrane, and form the main body of the Gangdese magmatic arc (Fig. 1A; Chung et al., 2005; Mo et al., 2005a, b; Ji et al., 2009; Zhu et al., 2011, 2018).

The study area considered here is located at the eastern end of the Gangdese arc, on the western flank of the eastern Himalayan syntaxis, and contains three tectonostratigraphic units: the Himalayan sequence, the Yarlung-Tsangpo suture zone, and the eastern Gangdese arc (Fig. 1A and B; Yin and Harrison, 2000). The Himalayan sequence includes the Tethyan Himalayan sequence and the Greater Himalayan sequence; the Yarlung-Tsangpo suture zone contains remnants of the Neo-Tethyan Ocean that existed between the Asian and Indian plates; and the eastern Gangdese arc includes Paleozoic sedimentary rocks (mainly Carboniferous strata), Jurassic volcanic rocks, Jurassic to Cretaceous granitoids, Late Cretaceous gabbro-granodiorite (Lilong batholith), Paleocene to Eocene granite, and Oligocene granite (Fig. 1B). The pre-Oligocene rocks underwent multi-stages of amphibolite- to granulite-facies metamorphism and partial melting during the Late Cretaceous and Eocene (Wang et al., 2008; Dong et al., 2010a, b, 2012; Zhang et al., 2010b, c, 2013, 2014b, 2015; Guo et al., 2011, 2012; Xu et al., 2013; Ding and Zhang., 2018; Palin et al., 2014; Kang et al., 2019; Niu et al., 2019; Qin et al., 2019). The protoliths of these metamorphic rocks include sedimentary and igneous rocks of various ages, and therefore termed the Nyingchi complex by Zhang et al. (2013, 2014c). The complex is characterized by a northwestward-decreasing trend in metamorphic grade from the granulite-facies belt in the southeast, through an amphibolite-facies belt, to an epidote amphibolite- and greenschist-facies belt in the northwest (Fig. 1B). These metamorphic belts form the lower, middle and upper crustal levels of the eastern Gangdese arc, respectively (Zhang et al., 2020b).

The meta-sedimentary rocks, including the paragneisses, pelitic schists, marbles and calc-silicate rocks, widely occur in the eastern Gangdese arc (Fig. 1B). Field observation shows the migmatitic sedimentary rocks occur as bodies with various sizes within the metamorphosed magmatic rocks (juvenile lower crust), and the migmatitic sedimentary rocks commonly contains small volumes of migmatitic and magmatic rocks. Moreover, the anatectic magmatic and sedimentary rocks commonly display transitional contacts. In addition, the Paleocene gabbros and granitoids contains abundant inclusions of the anatectic magmatic and sedimentary rocks. In this paper, we focused on paragneisses in the Zire area (Fig. 1B). The paragneisses and associated pelitic schists, marbles and calc-silicate rocks occur as a large block within the metamorphosed Paleocene granitoids in the Zire-Jiemai-Nyingchi area, and have the appearance of stromatic migmatites, and display intense folding (Fig. 2). At the outcrop scale, the gneisses display a banded structure defined by alternating bands of black melanosome and white leucosome (Figs. 2A and 2B). The melanosomes consist of coarse-grained garnet, plagioclase, K-feldspar, biotite, muscovite, and quartz. The leucosomes occur as ~1–10 cm thick bands or veins that run parallel to and cross the foliation of host gneisses (Fig. 2A). The leucosomes account for ~20 vol. % of the rock, and mainly comprise medium-grained plagioclase, K-feldspar, quartz and garnet (Fig. 2). Garnet is heterogeneously distributed and has a range of sizes. Smaller garnet grains mostly occur in the leucosome whereas larger garnets are commonly found in the melanosome (Figs. 2B and 2C). Discordant leucosome veins (Fig. 2A) and their local enrichment (Fig. 2C) in the migmatitic gneisses provide favorable evidence for the migration and loss of melt from the melanosome.

ANALYTICAL METHODS

Cathodoluminescence (CL) and back scattered electron (BSE) images were obtained using a CAMECA electron microprobe at Institute of Geology, Chinese Academy of Geological Sciences (CAGS), Beijing. They were used to check the internal texture of single zircon and monazite, then to select suitable positions for U–Th–Pb dating and zircon Hf isotope analysis.

Mineral chemical compositions were acquired using a JEOL JXA 8900 electron microprobe (EPM) with a wavelength-dispersive detection mode housed at the Institute of Geology, CAGS. Operation conditions were 15 kV accelerating voltage, 5 nA beam current, 5 μm probe diameter, and counting time of 10 s for peak and background. Natural or synthetic standards were used for EPM analysis, and oxide ZAF corrections were applied. Whole-rock chemical compositions were obtained at the National Geological Analysis Center of China, Beijing. Major element oxides were determined by X-ray fluorescence (XRF) (Rigaku-3080), with precision better than 0.5% including loss on ignition (LOI).

Zircon and monazite U–Th–Pb dating and trace element analysis were conducted simultaneously by laser ablation-inductively coupled plasma mass spectrometry (LA–ICPMS) at the Wuhan Sample Solution Analytical Technology Co., Ltd., China. Detailed operating conditions for the laser ablation system and the ICP–MS instrument and data reduction are described in Zong et al. (2017). Laser sampling for zircon and monazite 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 ICP–MS instrument was used to acquire ion-signal intensities. A common Pb correction was made utilizing the correction function (Andersen, 2002). The spot size and frequency of the laser were set to 32 μm and 5 Hz, respectively, for zircon dating. The spot size of the laser was set to 10 μm or 16 μm for monazite dating. Zircon 91500 and monazite 44069 were used as external standards for zircon U-Pb dating and monazite Th-Pb, respectively, and zircon GJ-1 and Plešovice and monazite TRE (Trebilcock) were used as internal standards. Trace element compositions of zircon and monazite were calibrated against NIST SRM610. The age and trace element data of zircon standards 91500, GJ-1 and Plešovice and monazite standards 44069 and TRE are listed in Table DR4[1] and Table DR5 (see footnote 1), respectively. Zircon standards 91500, GJ-1 and Plešovice yield concordant or near-concordant $^{206}Pb/^{238}U$ ages of 1052–1073 Ma [weighted mean age = 1062.4 ± 4.3 Ma (2σ, n = 58, MSWD = 0.02)], 596–605 Ma [weighted mean age = 601.2 ± 3.9 Ma (2σ, n = 16, MSWD = 0.11)] and 334–342 Ma [mean age = 338.3 ± 2.1 Ma (2σ, n = 16, MSWD = 0.26)], respectively, which are consistent with recommended values in the uncertainty

(91500: 1062.4 ± 0.4 Ma; Wiedenbeck et al., 1995. 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). The mean $^{208}Pb/^{232}Th$ age for monazite standards 44069 and TRE were 417–433 Ma [weighted mean age = 424.9 ± 2.0 Ma (2σ, n = 36, MSWD = 0.57)] and 265–277 Ma [mean age = 270.1 ± 2.0 Ma (2σ, n = 12, MSWD = 0.98)], respectively, consistent with the recommended value (44069: 424.9 ± 0.4 Ma; Aleinikoff et al., 2006. TRE: 272 ± 2 Ma; Tomascak et al., 1996). Each analysis incorporated a background acquisition of approximately 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. Off-line selection and integration of background and analyzed signals, and time-drift correction and quantitative calibration for trace element analysis and U–Th–Pb dating, were conducted by the Excel-based software ICPMSDataCal (Liu et al., 2010).

Zircon U–Pb geochronology was also conducted using the LA–ICPMS housed in the Mineral and Fluid Inclusion Microanalysis Lab, Institute of Geology, CAGS. The NWR 193^{UC} laser ablation system (Elemental Scientific Lasers USA) was equipped with Coherent Excistar 200 excimer laser and a Two Volume 2 ablation cell. The laser ablation system was coupled to an Agilent 7900 ICPMS (Agilent, USA). The detailed analytical methods are described in Yu et al. (2019). LA–ICPMS tuning was performed using a 50-micron diameter line scan at 3 μm/s on NIST 612 at ~3.5 J/cm^2 with repetition rate 10 Hz. Adjusting the gas flow to get the highest sensitivity (^{238}U ~ 5×10^5 cps) and the lowest oxide ratio (ThO/Th < 0.2%). P/A calibration was conducted on the NIST 610 using a 100-micron diameter line scan. Zircon 91500 and GJ-1 were used as primary and secondary reference materials respectively. The 91500 was analyzed twice and GJ-1 analyzed once every 10-12 analysis of the sample. As listed in Table DR4, 91500 and GJ-1 yield concordant or near-concordant $^{206}Pb/^{238}U$ ages of 1044–1082 Ma [weighted mean age = 1061.8 ± 2.5 Ma (2σ, n = 51, MSWD = 0.80)] and 594–617 Ma [weighted mean age = 601.9 ± 2.0 Ma (2σ, n = 27, MSWD = 1.90)], respectively, which are consistent with recommended values in the uncertainty. Multiple groups of 10 to 12 sample unknowns were bracketed by triplets of primary and secondary zircon standards. Typically, 35–40 s of the sample signals were acquired after 20 s gas background measurement. Using the exponential function to calibrate the downhole fractionation (Paton et al., 2010). NIST610 and ^{91}Zr were used to calibrate the trace element concentrations as external reference material and internal standard element respectively. The spot size of the laser was set to 25 μm or 30 μm for this zircon dating. The Iolite software package was used for data reduction (Paton et al., 2010).

In-situ Hf isotope ratio analysis of zircon was conducted using a Neptune Plus multicollector ICPMS (MC–ICPMS) (Thermo Fisher Scientific, Germany) in combination with a Geolas HD excimer ArF laser ablation system (Coherent, Gottingen, Germany) hosted at the Wuhan Sample Solution Analytical Technology Co., Ltd., China. The energy density of laser ablation that was used in this study was

~7.0 J/cm^2. Each measurement consisted of 20 s of acquisition of the background signal followed by 50 s of ablation signal acquisition. Detailed operating conditions for the laser ablation system and the MC–ICPMS instrument and analytical method are the same as description by Hu et al. (2012a, b).

In this study, age calculations were performed using the ISOPLOT/Ex_ver3 program (Ludwig, 2003). We reported zircon $^{207}Pb/^{206}Pb$ ages for >800 Ma, and the $^{206}Pb/^{238}U$ ages for younger zircons. Th- and Pb-rich monazites often lead to a large range of U–Pb age error, therefore, this paper reported monazite $^{208}Pb/^{232}Th$ ages. The chondrite normalized values were after Sun and McDonough (1989).

PETROLOGY

The studied samples include six garnet-bearing gneisses (91-5, 91-9, 91-16, 91-19, 91-30 and 91-41) and one leucosome (91-35). Three gneisses (91-9, 91-19 and 91-41), and one leucosome (91-35) were selected for detailed petrology. Mineral abbreviations used in this paper follow the guidelines of Whitney and Evans (2010).

Petrography and Whole-Rock Chemical Compositions

The gneisses exhibit well-defined banding, with small fractions of leucosome and large proportions of melanosome (Figs. 2B and 3A). The leucosomes consist mainly of plagioclase, K-feldspar and quartz with minor garnet, biotite, muscovite, rutile and ilmenite, and have a medium- to fine-grained size (Fig. 3B). Garnets generally lack inclusions and display a granoblastic texture in leucosomes (Fig. 3B). The melanosomes contain plagioclase, K-feldspar, quartz and biotite with minor garnet, muscovite, rutile and ilmenite, and display clear foliation defined by elongated and aligned biotite and quartz ribbons (Figs. 3A and 3C). Coarse-grained garnet and plagioclase occur as porphyroblasts, whereas fine- to medium-grained plagioclase, K-feldspar, biotite and quartz characterize the matrix (Figs. 3C, 3D, and 3E). Garnet porphyroblasts commonly have inclusions of quartz, biotite, feldspar, muscovite and ilmenite, which are generally preserved in cores (Figs. 3D and 3F). It is notable that polymineralic inclusions, consisting of biotite + plagioclase + quartz, plagioclase + quartz + ilmenite, and muscovite + biotite, are visible within garnet (Figs. 3D and 3F). Plagioclase inclusions in garnet are locally surrounded by a thin film of K-feldspar (Fig. 3D). The biotite in polymineralic inclusions is partially surrounded by plagioclase with a cuspate shape (Fig. 3F). Some garnets are partially replaced by symplectitic corona of biotite + plagioclase + quartz along their rims (Figs. 3D and 3G). The variably recrystallized quartz ribbons, defining the main foliation, are resorbed by fine-grained biotite, plagioclase and K-feldspar, forming embayments (Fig. 3C).

Whole-rock major element data for the three gneisses are presented in Table 1, which show that all are compositionally similar, having relatively high SiO_2 (74.42–

75.50 wt. %), Al_2O_3 (10.42–10.81 wt. %), FeO (5.02–5.04 wt. %), K_2O (2.01–3.62 wt. %) and Na_2O (1.39–2.53 wt. %), but low MgO (0.70–1.50 wt. %), CaO (0.72–1.31 wt. %) and MnO (0.29–0.91 wt. %).

Mineral Chemistry

Representative mineral compositions for the gneisses are listed in Data Repository Tables DR1–DR3, respectively, and characteristic features are described below.

Garnet in sample 91-19 contains end-member components almandine (X_{Fe} = Fe/(Ca + Mg + Fe + Mn), X_{Mg}, X_{Ca} and X_{Mn} defined similarly to X_{Fe}) = 0.66–0.75, pyrope (X_{Mg}) = 0.11–0.15, grossular (X_{Ca}) = 0.05–0.11 and spessartine (X_{Mn}) = 0.02–0.18 (Table DR1). These porphyroblasts display compositional zoning mostly characterized by an increasing X_{Mg} and homogeneous X_{Mn} in the core, except for some cores containing atypically Mn-rich (Fig. 4A), and then show an increase in X_{Mn} as well as decreases in X_{Mg}, X_{Fe} and X_{Ca} towards the rim (Fig. 4A). Thus, the rim composition is typical of diffusional resetting during retrograde metamorphism (e.g., Spear et al., 1990; Spear, 1991). Garnets within sample 9-9 and 9-19 have similar compositions, with X_{Fe} = 0.61–0.68, X_{Mg} = 0.10–0.13, X_{Ca} = 0.03–0.05 and X_{Mn} = 0.15–0.23 (Table DR1), which resemble the composition of garnet rims from sample 91-19 (Fig. 4B).

Biotite from the gneisses has a wide range of 0.08–0.18 Ti atoms per formula unit (apfu) (Table DR2). In general, matrix biotite is the most Ti-rich (0.12–0.18 apfu) and has relatively high X_{Mg} [= Mg/(Fe^{2+} + Mg); 0.82–0.93], compared with those in the symplectitic corona after garnet (Ti = 0.09–0.12 apfu; X_{Mg} = 0.79–0.93) and in inclusions in garnet (Ti = 0.08–0.10 apfu; X_{Mg} = 0.75–0.82) (Fig. 4C). Plagioclase has a composition that varies according to its textural position. Matrix plagioclase has homogeneous compositions, being classified as oligoclase with An [= Ca/(Ca + K + Na)] = 0.19–0.23 (Table DR3; Fig. 4D), whereas plagioclase porphyroblasts are compositionally zoned, with decreasing An contents from the core to rim (Fig. 4D). The rim compositions (An = 0.21) are like those of fine-grained plagioclase in the matrix, whereas the core has the highest An content (0.26). Plagioclase in corona has a relatively low CaO contents (An = 0.21–0.22) (Fig. 4D).

These petrological and mineral compositional features show that the gneisses record three stages of metamorphism (M1–M3). The earliest stage, prograde metamorphism (M1), is recorded by garnet and plagioclase porphyroblasts inner core compositions and associated inclusions hosted in these minerals; the mineral assemblage for this stage is Grt (inner core) + Bt + Ms + Pl (core) + Kfs + Qz + Ilm. Peak-metamorphism (M2) is characterized by the outer core (maximum X_{Mg}) of porphyroblastic garnet, the rim of porphyroblastic plagioclase, and matrix minerals; the mineral assemblage for this stage is Grt (outer core) + Bt + Kfs + Pl (rim) + Qz + Rt. The retrograde stage (M3) is characterized by the rims of porphyroblastic garnet and its corona minerals as well as those minerals resorbing the quartz ribbon; the mineral assemblage for this stage is Grt (rim) + Bt + Ms + Pl + Kfs + Qz + Ilm.

METAMORPHIC CONDITIONS

Phase Equilibrium Modeling

Paragneiss samples 91-9, 91-19, and 91-41 consist mainly of melanosomes with very fine bands of leucosome. Whole-rock XRF data were obtained from large blocks of each sample that contained similar melanosome–leucosome proportions in order to obtain a representative estimation of the metamorphic equilibration volume during anatexis (cf. Palin et al., 2016b). The metamorphic and anatectic *P*–*T* conditions were constrained via phase equilibrium modeling using an average bulk-rock composition of the three samples (Table 1). All equilibria were constructed using PERPLE_X (Connolly, 2005; version 6.7.4) and the internally consistent thermodynamic data set of Holland and Powell (1998) in the 10-component system MnO–Na_2O–CaO–K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O–TiO_2 (MnNCKFMASHT). Ferric iron was omitted from the model system due to the absence of Fe^{3+}-rich oxides and other Fe^{3+}-bearing silicate minerals (e.g., epidote). The fluid was considered as pure H_2O due to the absence of carbonate or graphite. The activity-composition relations used for modeling include biotite (White et al., 2014), garnet (White et al., 2014), cordierite and staurolite (White et al., 2014), feldspar (Holland et al., 2013), white mica (White et al., 2014), ilmenite (White et al., 2014), chlorite and chloritoid (White et al., 2014), amphibole (Diener et al., 2007) and silicate melt (White et al., 2014).

Field and microscopic evidences indicate that the studied rocks have experienced intense partial melting and melt loss. A *P*–*T* pseudosection constructed based on the measured bulk-rock composition is considered representative of the final stage of melt crystallization, and thus, reflects the metamorphic evolution after the loss of melt (White and Powell, 2002; Kelsey et al., 2003). In addition, melt loss is unlikely to have occurred during crystallization (i.e. at decreasing temperature), thus the measured bulk composition can be applied to explore the metamorphic evolution from peak-*T* to the final melt crystallization, but may not be valid for the prograde evolution (Indares et al., 2008; Groppo et al., 2010, 2012, 2013; Palin et al., 2018).

As shown in the pseudosection calculated for the *P*–*T* range of 3–13 kbar and 600–900 °C (Fig. 5A), garnet, plagioclase and quartz are stable at all given *P*–*T* conditions. Muscovite is unstable above ~630–750 °C, biotite disappears above ~820–850 °C, and rutile stabilizes above ~5–12 kbar. The solidus is located at ~660–690 °C, although is partly fluid-undersaturated for the measured bulk-rock H_2O content. These phase equilibria are consistent with those observed in Al-poor metasediments, such as (meta)greywacke (Palin and Dyck, 2021). Peak and retrograde *P*–*T* conditions were constrained by comparing the observed and modeled assemblages in the pseudosection. The observed peak M2 assemblage, Grt (outer core) + Bt + Kfs + Pl (rim) + Qz + Rt, is stable at ~7–13 kbar and ~720–830 °C in the presence of melt, and the retrograde M3 assemblage, Grt (rim) + Bt + Ms + Pl + Kfs + Qz + Ilm, is stable at a narrow field of ~5–8 kbar and ~660–710 °C (Fig. 5A).

We further used mineral compositional isopleths to put tighter constrains on the *P–T* conditions within the assemblage stability field. Calculated X_{Mg} [= Mg/(Ca + Mg + Fe + Mn)] and X_{Ca} [= Ca/(Ca + Mg + Fe + Mn)] isopleths of garnet, and Ti (= Ti apfu) of biotite are shown in Fig. 5A. Intersections between the X_{Mg} and X_{Ca} isopleths that represent the observed garnet compositions have been widely applied in estimating *P–T* paths (e.g., Tinkham and Ghent, 2005). For the outer core of garnet, when the X_{Mg} reached the maximum, and the X_{Mn} has increased (Fig. 4A), indicating the garnet occurred diffusion. Combining the maximum X_{Mg} (~0.15) and X_{Ca} (~0.10) of garnet outer core composition thus yields a *P–T* conditions lower than peak conditions. In contrast, due to Ca being the slowest-diffusing divalent cation in garnet (Guilmette et al., 2011), the X_{Ca} (~0.10) of the garnet outer core combined with maximum Ti value (~0.18) of the matrix biotite intersect at *P–T* conditions of ~11.3 kbar and ~745 °C (the yellow circle in Fig. 5A), falling on the boundary of the stability field of the observed M2 mineral assemblage (Fig. 5A). This is taken to represent a reasonable estimate of the conditions of peak-metamorphism.

Garnet core-to-rim compositional profile suggests that garnets were affected by retrograde diffusion (Fig. 4A). As a result, retrograde metamorphic conditions can be constrained by the isopleths of X_{Mg} and X_{Ca} for the garnet rim. The garnet rim composition with minimum X_{Mg} (~0.10) and X_{Ca} (~0.05) suggests a *P–T* condition of ~6.2 kbar and ~680 °C (the white circle in Fig. 5A). It is worth noting that the constrained *P–T* conditions for the latest stage of metamorphism are very close to the solidus, representing the conditions of melt crystallization. This is consistent with the textural evolution of minerals in anatectic rocks commonly ending with the solidification of melts (Forshaw et al., 2019), that is, the actual observed mineral assemblages (M3) should be those corresponding to the final crystallization of the melt at the solidus (e.g., White and Powell, 2002; Indares et al., 2008; Guilmette et al., 2011); not necessarily those that were present at peak metamorphism.

The calculated melt isomodes indicate ~8 wt. % of melt would have been produced at peak-metamorphic *P–T* conditions (Fig. 5B), representing the preserved melt in a closed-system environment. Figure 5B shows the mineral proportions at suprasolidus conditions, where garnet mode decreases whereas the plagioclase increases during the retrograde process. This is consistent with the petrographic observation that the garnet rims are replaced by biotite + plagioclase + quartz corona (Figs. 3D and 3G).

Conventional Thermobarometry

Due to the potential issue of partial rock-wide chemical disequilibrium in migmatites owing to differences in the length- and time-scales of major elements at crustal *P–T* conditions (Powell et al., 2019), M2 and M3 *P–T* conditions for the gneisses were also calculated using conventional thermobarometry, which only requires chemical equilibrium to have been reached in local mineralogical domains. This was achieved using a combination of the garnet-biotite (GB) thermometer (Holdaway, 2000) with the garnet-biotite-plagioclase-quartz (GBPQ) barometer (Wu et al., 2004). For the peak-metamorphic stage (M2), we use compositions of garnet outer core with the maximum X_{Mg}, matrix plagioclase with the lowest CaO content and matrix biotite with highest TiO_2 content. This assemblage yields a *P*–*T* condition of ~10.8 kbar and ~740 °C (the yellow rectangle in Fig. 5A), which are close to the results obtained with phase equilibrium modeling (the yellow circle in Fig. 5A). For the retrograde stage (M3), we use compositions of garnet rim with maximum X_{Mn}, and adjacent coronas of plagioclase-biotite, yielding *P*–*T* conditions of ~6.1–6.6 kbar and ~670–680 °C (the white rectangle in Fig. 5A), which are in agreement with the results from our modeling (the white circle in Fig. 5A).

GEOCHRONOLOGY

Zircon U–Pb Geochronology

Zircon grains from the migmatitic gneisses and leucosome display euhedral to subhedral prismatic shapes and range from 130 to 200 μm in size. CL images show that the zircons are characterized by well-preserved core-rim structure (Fig. 6). The inherited detrital cores have near-rounded shapes and oscillatory, sector, or no zoning patterns (Fig. 6), a common feature of zircons from sedimentary rocks (Corfu et al., 2003). Zircon rims are euhedral to subhedral edges and weak oscillatory zoning (Fig. 6). In some samples, zircon rims are too thin to conduct LA–ICPMS analyses. All zircon U–Pb dating and trace element analysis results are given in Table DR4.

Zircon U–Pb dating results of the gneiss samples 91-9, 91-19 and 91-30 were performed at the CAGS. Ninety-four, 94 and 86 analyses were collected from the samples above, respectively, from which 90, 91 and 85 analyses are from the inherited cores, and four, three, and one analyses are from the overgrowth rims, respectively (Fig. 7). The zircon cores of sample 91-9 yield concordant or near-concordant U–Pb ages ranging from 310 to 2706 Ma (Figs. 7A and 7B; Table DR4). Four analyses in zircon rims yield $^{206}Pb/^{238}U$ ages of 61–46 Ma with low Th/U ratios of 0.02–0.03 (Fig. 7B; Table DR4). For sample 91-19, analyses on the zircon cores have concordant or near-concordant ages ranging from 332 to 2713 Ma (Figs. 7C and 7D; Table DR4). Three analyses of zircon rims have a $^{206}Pb/^{238}U$ age of 72–55 Ma with low Th/U ratios of 0.01–0.04 (Fig. 7D; Table DR4). Zircon cores analyses from the sample 91-30 give concordant or near-concordant U–Pb ages of 2601–315 Ma (Figs. 7E and 7F; Table DR4). One zircon rim spot yields a $^{206}Pb/^{238}U$ age of 60 Ma (Fig. 7F; Table DR4).

Zircon rims from four gneiss samples (91-5, 91-9, 91-16 and 91-41) and one leucosome sample (91-35) were analyzed at the Wuhan Sample Solution Analytical Technology Co., Ltd., and yield $^{206}Pb/^{238}U$ ages of 57–48 Ma, 53–45 Ma, 56–49 Ma, 54–47 Ma and 54–49 Ma, respectively (Figs. 8A, 8C, 8E, 8G, and 8F; Table DR4).

The analyzed zircon rim domains from the five samples have relatively high and variable rare earth element (REE) contents of 628–2916 ppm, and low Th/U ratios of 0.01–0.09 (Table DR4). The REE patterns are characterized by steep heavy REE (HREE) slopes [$(Lu/Gd)_{CN}$ = 26.39–153.98] with variable negative Eu anomalies (Eu/Eu* = 0.01–0.11) (Table DR4; Figs. 8B, 8D, 8F, 8H, and 8J).

Monazite U–Th–Pb Geochronology

Monazite grains in four gneiss samples (91-5, 91-9, 91-16 and 91-41) and one leucosome sample (91-35) are commonly ~80–100 μm in diameter and contain dark cores surrounded by bright rim in BSE images (Fig. 9). Monazite dating and trace element analysis results are given in Table DR5. Grains from all five studied samples yield $^{208}Pb/^{232}Th$ ages of 69 Ma, 68–60 Ma, 65–61 Ma and 67–60 Ma for cores, whereas rims yield younger $^{208}Pb/^{232}Th$ ages of 57–41 Ma, 56–43 Ma, 57–45 Ma, 54–41 Ma and 58–41 Ma, respectively (Fig. 10). Note that partial monazite domains with relatively old age have relatively low HREE contents (362–4579 ppm) and slight Eu anomalies (Eu/Eu* = 0.03–0.39) (Figs. 10D and 8H; Table DR5), whereas the monazite domains with younger age have stronger negative Eu anomalies (Eu/Eu* = 0.02–0.26) with significantly fractionated REE patterns (HREE = 514–5697 ppm) (Figs. 10D, 10F, and 10H; Table DR5).

Summary

The age spectrum for all the detrital zircon cores (266 analyses) from the three migmatitic gneisses shows four age clusters at ca. 1800–1700 Ma, ca. 1100–900 Ma, ca. 600–500 Ma and ca. 400–300 Ma (Fig. 11A; Table DR4). Zircon rims yielded an age group of 57–45 Ma (Figs. 11A and 11B). Monazite U–Th–Pb analyses show that monazite grains from the gneisses and leucosome recorded two $^{208}Pb/^{232}Th$ age populations: 69–60 Ma for the cores and 58–41 Ma for the rims (Fig. 11B). Kelsey et al. (2008) ascribed discrepant zircon and monazite ages from the same rocks to differences in the temperatures at which each mineral grows during metamorphism. As temperature increases and an increasing amount of melt is produced, zircon is preferentially dissolved into these leucosomes; however, zircon often grows during cooling from high-grade conditions when melt crystallizes, and thus zircons ages most commonly reflect the cooling part of a *P*–*T* path (Kelsey and Powell, 2011). Conversely, monazite always has a complex internal structure with multiple periods of growth, which records the multistage evolution history of the rocks (Qiu et al., 2011) and more care must be taken to interpret the petrological significance of any ages obtained.

ZIRCON Hf ISOTOPIC COMPOSITION

Zircon rims from the five gneiss samples were also analyzed for their Hf isotopic composition. The results are listed in Table DR6 (see footnote 1) and are shown graphically in Fig. 12. The Hf isotopic analyses display $^{176}Lu/^{177}Hf$ ratios ranging from 0.000538 to 0.002004 and $^{176}Hf/^{177}Hf$ ratios ranging from 0.282436 to 0.282758 (Table DR6). These zircon rims yield similar and mostly negative $\varepsilon_{Hf\,(t)}$ values, range from to −10.57 to +0.78 with corresponding crustal model ages (T^{C}_{DM}) ranging from ca. 1082 to ca. 1802 Ma (Fig. 12; Table DR6). The $\varepsilon_{Hf\,(t)}$ values are calculated based on the monazite ages of 64–60 Ma because the ages are considered to represent the times of partial melting and melt loss of the studied gneisses, as discussed below.

DISCUSSION

Protolith and Deposition Age of Paragneiss

Internal structures of zircon from these gneisses show that inherited cores are large and have a diverse morphology, suggesting multiple origins (Fig. 6). The broad age range of 2713–310 Ma also suggests that the gneiss protoliths have a sedimentary origin, as magmatic protoliths are less likely to incorporate parent material of such variable heritage. The gneisses have a relatively high SiO_2 (>70 wt. %) and FeO (>5 wt. %; Table 1) contents, suggesting that the rocks were probably derived from a greywacke. The maximum depositional ages of the sedimentary protoliths can be constrained by the youngest concordant age of detrital zircon cores. The youngest $^{206}Pb/^{238}U$ ages are 310 ± 6 Ma for sample 91-9, 332 ± 6 Ma for sample 91-19, and 315 ± 4 Ma for sample 91-30 (Fig. 7; Table DR4), indicating that the gneiss protoliths were deposited no earlier than ~310 Ma.

The detrital zircon age spectrum indicates that all samples received detritus from Mesozoic to Proterozoic sources with four significant peaks at ca. 1800–1700 Ma, ca. 1100–900 Ma, ca. 600–500 Ma and ca. 400 Ma (Fig. 11A). These detrital zircons have similar age spectra to those of detrital zircons in Late Carboniferous strata of the Lhasa terrane (Leier et al., 2007; Pullen et al., 2008). As shown in Figure 1B, the Carboniferous strata widely exposed in the northwestern part of the study area. We therefore consider the protoliths of the studied paragneiss are most likely the Late Carboniferous strata from the Lhasa terrane. Such detrital zircon age spectrum are similar to those from meta-sedimentary rocks of the Nyingchi, Bujiu and Shejila areas (Dong et al., 2010a; Dong, 2011; Zhang et al., 2012, 2020a). This indicates that the widely distributed meta-sedimentary rocks in the deep crust of Gangdese arc have the same protolith ages, and were derived from a set of contemporaneous strata of the Lhasa terrane.

Metamorphic Conditions and Times, and Their Spatial Change

Previous studies have shown that the Nyingchi complex experienced amphibolite- to granulite-facies metamorphism during the Paleocene-Eocene (Wang et al., 2008; Guo et al., 2012; Palin et al., 2014; Kang et al., 2019). Dong (2011) and Dong et al. (2012) shown that the schists and paragneisses from the Nyingchi area underwent Early Cenozoic (70–50 Ma) lower amphibolite-facies metamorphism with *P*–*T* conditions of 4–6 kbar and 630–700 °C. Zhang et al. (2020a) showed that the schists from the Jiemai area experienced Early Cenozoic (70–50 Ma) upper amphibolite-facies metamorphism and partial melting with *P*–*T* conditions of ~8 kbar and ~740 °C. Zhang et al. (2013) revealed that the meta-plutonic and meta-sedimentary rocks in the Bujiu-Gama area underwent Early Cenozoic (67–52 Ma) granulite-facies metamorphism and partial melting at *P*–*T* conditions of ~9–11 kbar and ~800–830 °C. Our results from phase equilibrium modeling and conventional thermobarometry indicate that the paragneisses from the Zire area underwent Early Cenozoic (see discussion below) upper amphibolite-facies under conditions at ~11 kbar and ~740 °C (Figs. 5 and 13), indicating that the paragneisses equilibrated at middle-lower crustal depths of ~35 km. Therefore, the work and previous results clearly indicate that the Nyingchi complex has northwestward-decreasing *P*–*T* conditions from a granulite-facies belt in the Gama-Bujiu area, through an upper amphibolite-facies belt in the Zire-Jiemai area, to a lower amphibolite-facies belt in the Nyingchi area (Fig. 1B).

This study shows that monazite rim domains with relatively young $^{208}Pb/^{232}Th$ ages (58–41 Ma) have strongly negative Eu anomalies (Figs. 10D, 10F, and 10H; Table DR5), indicating that monazite rims grew at the same time as plagioclase (Rubatto, 2002; Rubatto et al., 2013). These ages constrain the time and duration of the cooling and decompression history of the paragneiss (Fig. 13), as plagioclase should stabilize during retrograde metamorphism (Fig. 5B). Our petrographic observations also show that the garnet was partly replaced by biotite + plagioclase + quartz corona during melt crystallization. In contrast, the monazite core domains with relatively old $^{208}Pb/^{232}Th$ ages (69–60 Ma) have relatively low HREE contents (Figs. 10D and 10H; Table DR5), indicating that the growth of the monazite cores was associated with the growth of garnet. These ages represent the time and duration of heating and burial metamorphism, and associated partial melting of the paragneiss because the peritectic garnet growth during melting. Hence, we infer that the age gap (60–58 Ma) between the two age groups represent the age of peak metamorphism.

The zircon metamorphic rims from the paragneisses and the leucosome exhibit subhedral and long prismatic shape, weak oscillatory zoning in CL images, very low Th/U values (0.003–0.1), negative Eu anomaly and steep fractionated HREE patterns (Fig. 8; Table DR4), suggesting that they crystallized from anatectic melt (Rubatto, 2002; Wu and Zheng, 2004; Rubatto et al., 2009). The $^{206}Pb/^{238}U$ ages of 57–45 Ma obtained from the rims are therefore interpreted as the age of melt crystallization.

Overall, this study obtains a long-lasting (69–41 Ma) high-*T* metamorphic process. Moreover, our results probably indicate that the upper amphibolite-facies metamorphism and associated partial melting of the paragneisses initiated at the Early Paleocene, and retrograde cooling lasted until the Eocene (Fig. 13). This conclusion is consistent with the previous studies, which demonstrated the metamorphic rocks from the Gama-Bujiu-Jiemai-Nyingchi area experiencing a prolonged Early Cenozoic (70–45 Ma) metamorphic process (e.g., Booth et al., 2009; Guo et al., 2011, 2012; Dong et al., 2012; Zhang et al., 2013, 2020a; Palin et al., 2014; Kang et al., 2019).

Recent studies have shown that the Early Cenozoic metamorphism of the Nyingchi Complex followed an anticlockwise *P*–*T* path (Fig. 13; e.g., Zhang et al., 2013; Mu, 2018; Kang et al., 2019). Therefore, we infer that the studied paragneiss might experience a similar anticlockwise *P*–*T* path, and the retrograde process was characterized by cooling and decompression to shallow crustal levels (Fig. 13).

Partial Melting and Melt Compositions

As shown in Figure 13, the metamorphic *P*–*T* paths of the deep crustal rocks from the Gama-Jiemai area mostly are in the high-*T* amphibolite-facies and granulite-facies field and in the presence of melt, indicating that these rocks underwent significant partial melting. This is consistent with the studied paragneisses containing abundant leucosomes (Fig. 2). Therefore, we conclude that the anatexis was widespread throughout the middle to lower crustal levels of the eastern Gangdese arc during the Early Cenozoic.

The studied gneisses show the following microstructural characteristics: (i) garnet porphyroblasts contain abundant monomineralic or polymineralic inclusions of biotite, plagioclase, K-feldspar and quartz (Figs. 3D and 3F), typical peritectic growth during melting (Guilmette et al., 2011; Rubatto et al., 2013). (ii) In the polymineralic inclusions, thin films of K-feldspar are locally present along the plagioclase rim (Fig. 3D) and plagioclase surround the biotite with cuspate shape (Fig. 3F), which are interpreted to the presence of former melt (e.g., Holness and Clemens, 1999; Holness and Sawyer, 2008). (iii) Biotite + plagioclase + quartz intergrowth (corona) replacing garnet (Fig. 3G), suggests reaction between garnet and melt during the melt crystallization (Waters, 2001; Cenki et al., 2002; Kriegsman and Álvarez-Valero, 2009; Groppo et al., 2010). (iv) Quartz ribbons have corroded rims against the matrix (Fig. 3C), suggesting that the quartz ribbons have been partially consumed during the melt crystallization (Indares and Dunning, 2001; Jordan et al., 2006). These microstructures observed at the thin section scale indicate that the gneiss records a complex partial melting and crystallization of melt.

Field observations show that the leucosome content of the studied gneisses reaches ~20 vol. % at the outcrop scale (Fig. 2). Such large volumes of melts indicate the partial melts may have escaped from their sources, given that this melt proportion is above the viscosity change threshold of ~7-10 vol.% (Rosenberg and Handy, 2005). Local enrichment of leucosomes in the migmatites (Fig. 2C) provides direct evidence that the rocks have lost a portion of melt. Moreover, preservation of peritectic garnet (Fig. 3D) suggests that part of the melt produced was removed from the gneisses;

otherwise garnet would have been consumed during melt crystallization (White and Powell, 2002).

In most natural migmatites, the exact melt loss history (including the total amount and composition of the removed melt, and the number of episodes of melt loss) is very complex. Therefore, we simply calculate the composition of the melt at the peak conditions (~11 kbar and ~750 °C). The generated melts have a weakly peraluminous granitic compositions with SiO_2 = 72.69, Al_2O_3 = 15.88, K_2O = 4.69, Na_2O = 5.65, CaO = 0.55 (in wt. %), and A/CNK = 1.03 (Table 1). The Lu–Hf isotopic analyses of anatectic rims of zircon indicate that the partial melts have generally negative $\varepsilon_{Hf\ (t)}$ values from −10.57 to +0.78 (Fig. 12; Table DR6). Because the studied paragneisses probably represent metamorphosed greywacke, the generated melt compositions are necessarily different from typical peraluminous granites derived from partial melting of pelitic rocks.

Contributions to the Juvenile Lower Crust and Syn-collisional Magmatism

The juvenile lower crusts of magmatic arcs form via accretion of mafic magma derived from the mantle. They consist mainly of gabbros and amphibolites that are typically metamorphosed under amphibolite to granulite facies and exhibit anatectic structures (Depine et al., 2008; Ducea et al., 2015). However, the lower crusts of continental arcs, such as the Cascades arc in the northwestern United States, and the Sierra Nevada arc in California, are dominated by intermediate and felsic granulites in addition to mafic granulites (Saleeby et al., 2003; Miller and Snoke, 2009; DeBari and Greene, 2011; Jagoutz and Schmidt, 2012; Hacker et al., 2008, 2015; Jagoutz and Kelemen, 2015). Detailed geological mapping showed that the Gangdese arc lower crust contains large volumes of intermediate to felsic migmatites, and have a wide compositional range from ultramafic to felsic, with an average andesitic composition (SiO_2 of 57.61 wt. %, Mg# of 0.49) (Zhang et al., 2020b). Accordingly, there must be additional mechanisms to convert juvenile lower crust from a primitive basaltic composition to a mean andesitic one. Our study shows that paragneisses are widely distributed in the deep crust of the Gangdese arc, and these rocks have high SiO_2 contents (74.42–75.50 wt. %), and so are distinct from the mafic composition of juvenile lower crust. Therefore, incorporation of voluminous paragneisses could result in felsic component increases of the Gangdese juvenile lower crust and geochemical changes from a former mafic to a later intermediate composition. As argued by Chin et al. (2013) and Zhang et al. (2020b), the sediments transporting into deep arc crust is the key process in transformation of juvenile lower arc crust from mafic to intermediate composition, as indicated by geophysical observations (Hacker et al., 2015).

Previous studies have shown that Early Cenozoic S-type granites formed in the eastern Gangdese arc (Guo et al., 2012; Zhang et al., 2013; Ma et al., 2017). Guo et al. (2012) and Zhang et al. (2013) attributed the formation of these S-type granites to anatexis of the meta-sedimentary rocks of the Nyingchi complex. Our study clearly demonstrates that the partial melting of paragneisses produced the granitic melts with enriched isotopic compositions (Fig. 12), and large volumes of partial melts have migrated and been lost from local systems. We propose, therefore, that the meta-sedimentary rocks provide a potential source for syn-collisional S-type granites of the eastern Gangdese arc. In fact, some S-type granites have lower $\varepsilon_{Hf\,(t)}$ values than those probably derived from the paragneisses (Fig. 12). We speculate that those granites were derived from partial melting of Precambrian basement of the Lhasa terrane, which occurs locally in the lower crust of the eastern Gangdese arc (Zhang et al., 2020b).

Magmatism in the Gangdese arc can be divided into two separate stages. The early stage of Mesozoic magmatism (ca. 210–65 Ma) was related to the northwards subduction of the Neo-Tethyan oceanic lithosphere beneath the Lhasa terrane (Rowley, 1996; Aitchison et al., 2007; Ji et al., 2009; Kang et al., 2014; Meng et al., 2016a, b; Wang et al., 2016). The late stage of Early Paleocene-Late Miocene (ca. 65–10 Ma) magmatism occurred during the Indo-Asia collision (Chung et al., 2005; Mo et al., 2007, 2008; Chu et al., 2011; Zhu et al., 2011, 2018; Zhang et al., 2018, 2019). These subduction-related igneous rocks are dominated by intermediate to gabbroic rocks, and mostly show positive and high $\varepsilon_{Hf(t)}$ values (up to +20; Fig. 12) and $\varepsilon_{Nd(t)}$ values (up to +5.5; Chu et al., 2006, 2011; Mo et al., 2007, 2008; Ji et al., 2009; Ma et al., 2013a, b, c; Wang et al., 2016; Zhu et al., 2018; Zhang et al., 2019, 2020b). Therefore, these mantle-derived rocks formed the juvenile crust of the Gangdese arc (e.g., Zhu et al., 2008; Ji et al., 2014; Zhang et al., 2014b; Zheng et al., 2014).

When compared with pre-collisional magmatic rocks, the syn-collisional (65–40 Ma) rocks of the Gangdese arc display some distinct differences: (1) the magmatism is characterized by notable compositional diversity from mafic to felsic rocks, being dominated by intermediate-acid magmatic rocks and small amounts of mafic rocks (Fig. 12); (2) the presence of adakitic granitoids (Ji et al., 2012; Jiang et al., 2014; Ma et al., 2014); (3) extensive mingling of the mantle- and crust-derived magmas (Mo et al., 2005a, 2007, 2008; Dong et al., 2006; Zhu et al., 2018); and (4) large variation in Sr–Nd–Hf isotopic composition and even show clearly negative of $\varepsilon_{Nd(t)}$ and $\varepsilon_{Hf(t)}$ values (Fig. 12; Chu et al. 2006, 2011; Wu et al., 2007, 2010, 2014; Ji et al., 2009, 2017; Zhao et al., 2011; Lee et., 2012; Ma et al., 2014; Hou et al., 2015; Pan et al., 2016; Zhang et al., 2020b). The wide occurrence of Early Cenozoic adakitic granitoids show that the Gangdese arc underwent significant crustal thickening, and the thickened juvenile lower crust experienced intense partial melting (e.g. Mo et al., 2007; Guo et al., 2012; Ji et al., 2012; Jiang et al., 2014; Ma et al., 2017; Zhu et al., 2015, 2017a). The enriched isotopic compositions of arc-type magmatic rocks show the incorporation of older crustal materials into the magmatic source. Previous researchers have proposed the following possible explanation: the assimilation of ancient basement of the Lhasa terrane (Zhu et al., 2017b; Guo et al., 2011, 2012),the incorporation of subducted India continental materials (Chu et al., 2011; Ji et al., 2012), and the relamination of crustal materials removed from the Gangdese arc by tectonic erosion (Guo et al., 2019).

Based on the widespread occurrence of Late Paleozoic sedimentary rocks in the deep crust of eastern Gangdese arc, the structural relationships and transitional contacts between the anatectic magmatic and sedimentary rocks, and the Early Cenozoic magmatic rocks with inclusions of the migmatitic magmatic and sedimentary rocks, we consider that the mixing and mingling of melts derived from the mantle, juvenile lower crust and supracrustal rocks, and the assimilation between the arc magmas and supracrustal rocks resulted in the Early Cenozoic magmatic rocks having significant lower zircon $\varepsilon_{Hf(t)}$ values than those of the pre-collisional magmatic rocks (Fig. 12), and the Late Paleozoic meta-sedimentary rocks and more ancient crustal materials provide an enriched isotopic source. As argued by Chapman et al. (2014), supracrustal assemblages are adopted as potential high Sr_i–low ε_{Nd} end-members for the Salinian arc crust in California. In addition, it is possible that some of the Early Cenozoic granites and gabbros with enriched isotopic compositions were derived from partial melting of relaminated crustal materials that were removed from the Gangdese belt by tectonic erosion, and partial melting of lithospheric mantle which was metasomatized by inputs from relaminated crustal materials, respectively (Guo et al., 2019).

Mechanism of Transporting Sedimentary Rocks into Lower Arc Crust

Although many studies suggested that sedimentary rocks could be transported into the roots of continental arcs, where they can undergo high-grade metamorphism and partial melting (e.g., Stern, 2011), the mechanisms responsible are sometimes unclear. Possibilities include: (1) crustal thickening and thrusting (Whitney et al., 1999; Valley et al., 2003; Chin et al., 2013); (2) viscous downflow of country rock in response to diapiric ascent of plutons (Saleeby et al., 2003); (3) underplating by buoyant slab-derived sedimentary diapirs (Behn et al., 2011; Hacker et al., 2011); or (4) subduction erosion and relamination of material to the base of the crust (Hacker et al., 2011, 2015; Guo et al., 2019).

In the first hypothesis, the Gangdese arc has a thickened crust at the Early Cenozoic (Zhu et al., 2017a and references therein). Crustal thickening is thus one possible way in which the sedimentary rocks could have been transported into the deep crust. The second hypothesis, involving the downward transport of upper crustal rocks by viscous return flow (Saleeby et al., 2003), is unlikely given that wall rocks are probably highly viscous owing to their felsic composition and cold ambient temperatures compared to ascending plutons. Thus, high viscosities are barrier to wall rock return-flow (Chin et al., 2013). With respect to the third scenario, the studied paragneisses derived from the Late Carboniferous strata of Lhasa terrane, strongly indicating that the rocks originally deposited at the Lhasa terrane, rather than the subducted Neo-Tethyan oceanic slab. In terms of subduction erosion, Xu et al. (2013) and Guo et al. (2019) proposed that the supracrustal materials relaminated to the lower arc crust were dominantly derived from the forearc basin. In this case, the relaminated crustal materials should contain abundant inherited detrital zircons from the Mesozoic magmatic rocks of the Gangdese arc. However, the studied paragneisses from the Zire region do not have inherited detrital zircons with ages of < 300 Ma (Fig. 11A), clearly indicating the protoliths of paragneisses were not deposited in the forearc basin. Based on the widely distributed meta-sedimentary rocks in the deep crust of Gangdese arc have the same protolith ages, and the meta-sedimentary rocks have a transitional contact with the low-grade metamorphosed Late Carboniferous strata in northeastern part of the studied area, we consider that the paragneisses and associated rocks were metamorphosed product of the deeply buried Late Carboniferous strata, rather than derived from the mélanges in subduction channel proposed by the tectonic erosion model.

During the continental collision, the Gangdese arc experienced significant tectonic thickening, in addition to the major contribution of mantle-derived magmatic accretion (England and Searle, 1986; Yin et al., 1999; Kapp et al., 2005, 2007; Mo et al., 2007; Ji et al., 2012; Zhu et al., 2017a). For the eastern Gangdese arc, Early Cenozoic gabbros only locally occur in the exposed lower crustal sections (Guo et al., 2020; Zhang et al., 2020b). Therefore, the tectonic process was probably the major cause of arc crustal thickening. In fact, the accretion and loading of voluminous mantle-derived magmas in the middle crust can also transport sedimentary rocks into the lower arc crust. In this case, we suggest that crustal shortening and thrusting, and mantle-derived magmatic accretion transported the Late Carboniferous sedimentary rocks into the middle and lower crust of the eastern Gangdese arc during the Indo-Asian collision (Fig. 14). With the influx of mantle-derived magma, the deeply buried sedimentary rocks underwent amphibolite- to granulite-facies metamorphism and partial melting, and produced S-type granites (Fig. 14). Moreover, the mixing of melts derived from the mantle, juvenile lower crust and supracrustal rocks resulted in the isotopic compositional enrichments of syn-collisional arc-type magmatic rocks. Thus, as convergent margin processes have operated on Earth since at least the Late Archean (c. 3 Ga; Palin and Santosh, 2021), these unusual signatures should be preserved elsewhere within the geological record and may represent a valuable signature of arc formation and evolution in terranes.

CONCLUSIONS

(1) Migmatitic paragneisses from the Zire area, eastern Gangdese arc, underwent a prolonged Early Cenozoic (69–41 Ma) high-grade metamorphism and partial melting at *P*–*T* conditions of ~11 kbar and ~740 °C.

(2) The widely distributed meta-sedimentary rocks in the eastern Gangdese arc deep crust have the same protolith ages of Late Carboniferous, and show northeastward-decreasing metamorphic *P*–*T* conditions.

(3) The melts derived from the partial melting of paragneisses have a granitic

composition, and enriched isotopic compositions, and form a potential source for the syn-collisional S-type granites of the Gangdese arc.

(4) The voluminous meta-sedimentary rocks in the deep crust altered the components and compositions of juvenile lower crust of the Gangdese arc. The mixing of melts derived from the depleted mantle, juvenile lower crust and meta-sedimentary rocks resulted in the isotopic compositional enrichments of syn-collisional arc-type magmatic rocks of the Gangdese arc.

(5) The crustal shortening and thrusting, and accretion of mantle-derived magma have transported the supracrustal rocks into the deep crust of the Gangdese arc during the Early Cenozoic continental collision.

SUPPLEMENTARY TABLES

Table DR1. Representative chemical compositions (wt. %) of garnet from the paragneisses.

Table DR2. Representative chemical compositions (wt. %) of biotite from the paragneisses.

Table DR3. Representative chemical compositions (wt. %) of plagioclase from paragneisses.

Table DR4. Zircon U–Pb dating and trace element data of the paragneisses and leucosome.

Table DR5. Monazite U–Th–Pb dating and trace element data of the paragneisses and leucosome.

Table DR6. Zircon Lu–Hf isotope data of the paragneisses and leucosome.

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Figure captions

Figure 1. Sketch geological map of the Tibetan Plateau and Gangdese magmatic arc (A) and geological map of the eastern Gangdese arc (B) (modified after Zhang et al., 2020b).

Figure 2. Field photographs of the studied migmatitic paragneisses of the eastern Gangdese arc. (A) Migmatite shows distinct banded structure, defined by alternating bands of melanosomes and leucosomes, and folding deformation. Noted that the leucosomes occur locally as concordant and discordant thick veins. (B) Migmatite displays a banded structure. The melanosome contains abundant biotite and garnet, with minor quartz and feldspar, while the leucosome consists mainly of feldspar and quartz with minor biotite and garnet. (C) The leucosomes are locally rich in the migmatite, and consist of feldspar, quartz and minor garnet and mica.

Figure 3. Photomicrographs (A), (B), (C), (D), (E) and BSE images (F), (G) of the studied migmatitic paragneisses and leucosome. (A) Migmatitic paragneiss (sample 91-19), showing distinct foliation, defined by aligned biotite, elongated feldspar and quartz, as well as alternative bands of melanosome and leucosome. (B) Migmatitic leucosome (sample 91-35) shows granoblastic texture, and consists of plagioclase, quartz, K-feldspar, garnet and biotite. (C) Migmatitic melanosomes (sample 91-41) shows distinct foliation. Noted that quartz ribbon displays an embayment margin, that is rimmed by biotite, feldspar and quartz. (D) In the migmatitic paragneiss (sample 91-19), garnet porphyroblast contains abundant mono- and poly-mineral inclusions. Noted that the plagioclase within the garnet is surrounded by thin film of K-feldspar. Black line crossing the garnet grain is the location of the compositional profile shown in Fig. 4A. (E) In the migmatitic paragneiss (sample 91-9), plagioclase occurs as porphyroblast. Black line crossing the plagioclase grain refers to compositional analysis shown in Fig. 4D. (f) In the migmatitic paragneiss (sample 91-19), garnet contains plagioclase, muscovite, and ilmenite inclusions and polymineralic inclusions of biotite + plagioclase + quartz, plagioclase + quartz + ilmenite, and muscovite + biotite. Notice that biotite is partially surrounded by plagioclase with cuspate shape. (G) In the migmatitic paragneiss (sample 91-19), garnet is replaced by symplectitic corona of biotite-plagioclase-quartz along its margin.

Figure 4. Compositional diagrams of minerals in the migmatitic paragneisses (91-9, -19, -41). (A) Compositional profile of garnet porphyroblasts of sample 91-19. X_{Mg} = Mg/(Ca + Mg + Fe + Mn), X_{Ca} = Ca/(Ca + Mg + Fe + Mn), X_{Fe} = Fe/(Ca + Mg + Fe + Mn), and X_{Mn} = Mn/(Ca + Mg + Fe + Mn). The profile location is shown by the black line in Fig. 3D. (B) Pyrope-almandine + spessartine-grossularite diagram of garnets. (C) X_{Mg} versus Ti (a.p.f.u.) diagram of biotite. X_{Mg} = Mg/(Fe^{2+} + Mg). (D) Ab–An–Or diagram of plagioclase. The profile location of plagioclase porphyroblast was shown by the black line in Fig. 3E.

Figure 5. *P*–*T* pseudosections for the migmatitic paragneisses calculated using the measured bulk composition. The solidus is shown by red thick line, the disappearance (out) lines of various minerals are shown by the thick line with different colors in the left panel. The dashed blue, green and orange lines represent isopleths of X_{Mg} and X_{Ca} in garnet, and Ti in biotite, respectively (the left panel). Isomodes of melt, plagioclase (Pl) and garnet (Grt) are shown in the right panel. The yellow and white circles refer to the peak and retrograde metamorphic *P*–*T* conditions constrained by the mineral composition isopleths, respectively. The yellow and white rectangles represent the peak and retrograde metamorphic *P*–*T* conditions constrained by the conventional thermobarometry,

respectively.

Figure 6. Cathodoluminescence (CL) images of zircons from the migmatitic paragneisses (A-F) and leucosome (G), showing the analysed spot locations and relevant ages (in Ma). The scale bars are 100 μm.

Figure 7. U–Pb concordia diagrams (A, C, E) and $^{207}Pb/^{206}Pb$ or $^{206}Pb/^{238}U$ age probability plots (B, D, F) for zircons of the migmatitic paragneisses.

Figure 8. U–Pb concordia diagrams (A, C, E, G, I) and chondrite-normalized REE patterns (B, D, F, H, J) of the zircon metamorphic rims in the migmatitic paragneisses and leucosome.

Figure 9. BSE images of monazite from the migmatitic paragneisses (A-D) and leucosome (E), showing the analyzed spot locations and relevant ages (in Ma). The scale bars are 100 μm.

Figure 10. U–Th–Pb concordia diagrams (A, C, E, G, I) and chondrite-normalized REE patterns (B, D, F, H, J) of monazites in the migmatitic paragneisses and leucosome.

Figure 11. (A) Age frequency diagram of zircons from the migmatitic paragneisses. (B) Age frequency diagram of monazites and anatectic rims of zircon from the migmatitic paragneisses and leucosome.

Figure 12. Diagram of ages versus $\varepsilon_{Hf\ (t)}$ values of magmatic zircons from the magmatic rocks and anatectic zircons from the studied paragneisses of the eastern Gangdese arc. Data sources for gabbroic and I- and S-type granitic rocks are: Zhang et al., 2010a, 2013, 2014a, b, 2020b; Guo et al., 2011, 2012, 2013; 2019; Guan et al., 2012; Zheng et al., 2012, 2014; Ma et al., 2013a, b; Ji et al., 2014, 2017; Ding and Zhang, 2018.

Figure 13. Inferred *P–T–t* path for the studied migmatitic paragneiss and comparison with those reconstructed for the metapelitic rocks from the eastern Gangdese arc. All numbers refer to metamorphic ages in Ma. The solidus is the same as in Fig. 5A. Metamorphic facies boundaries are after Vernon and Clarke (2008).

Figure 14. Early Cenozoic tectonic model for the eastern Gangdese arc. Crustal shortening and thrusting induced by the Indo-Asian collision, and underplating and accretion of mantle-derived magma resulted in distinct crustal thickening of the Gangdese arc, and burial of sedimentary rocks in lower arc crust. The partial melting of meta-sedimentary rocks and juvenile lower arc crust generated S-type and I-type granites, respectively. The mixing of melts derived from meta-sedimentary rocks, mantle and juvenile lower crust, and the assimilation between the arc magmas and supracrustal rocks occurred in MASH (melting, assimilation, storage, and homogenization) zone of the arc root.

Table caption

Table 1. Chemical compositions of the studied paragneisses and the calculated melt (wt. %)