Late Triassic intra-oceanic arc system within Neotethys: Evidence from cumulate appinite in the Gangdese belt, southern Tibet

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ABSTRACT

The evolution of the Neotethyan realm played an important role in shaping the Phanerozoic paleogeographic framework of Earth, as well as the formation of the Tibetan Plateau. However, there is considerable debate regarding the timing of the opening and initial phases of subduction of the Neotethys Ocean. The Gangdese magmatic belt is located along the southern margin of the Lhasa terrane in southern Tibet and was formed during the subduction of Neotethyan oceanic lithosphere. In this paper, we discuss a Late Triassic cumulate appinite suite along the southern margin of the Gangdese magmatic belt. The appinite suite exhibits a cumulate structure, with hornblende and plagioclase being the primary mineral phases. Isotopic data indicate a hydrous magma source derived from the mantle wedge that was modified by slab dehydration. Geochemical discriminators suggest that the appinite suite was formed in an intra-oceanic arc setting with crystallization ages of ca. 220–213 Ma. Hornblende, hornblende-plagioclase, and ilmenite geothermometers yielded crystallization temperatures of 750–900 °C for the appinite. Hornblende and hornblende-plagioclase geobarometers yielded emplacement depths between 14.5 and 19.5 km, which is consistent with arc-related cumulates. The occurrence of this appinite constitutes a line of evidence for intra-oceanic arc magmatism that was coeval with similar magmatism in Turkey; this suggests that there was a vast east-west intra-oceanic subduction system within the Neotethys. A pre–Middle Triassic opening of the Neotethys would be required to explain the vastness of this subduction system. Our research provides a robust constraint for evaluating the Mesozoic framework of the Neotethyan realm and the evolutionary history of the Gangdese magmatic belt in southern Tibet.

INTRODUCTION

The Indus-Yarlung Tsangpo ophiolitic mélangé zone (or suture zone), confined between the northern Lhasa terrane and southern Tethys Himalaya, is considered to be a remnant of Neotethyan oceanic lithosphere (Fig. 1). However, the onset of the opening and subduction of the Neotethys Ocean is still enigmatic (Cai et al., 2016; Torsvik et al., 2012; Zhu et al., 2011). The formation of the Tibetan Plateau and the Gangdese magmatic belt in southern Tibet are closely related to the evolution of the Neotethys. Thus, a full understanding of the precollisional magmatism and forearc accretionary system of the Gangdese magmatic belt is crucial not only to elucidate the subduction-related magmatic processes, but also to shed light on the evolutionary history of the Neotethys.

The temporal constraints on the opening of the Neotethys Ocean are controversial. Some models suggest that the Lhasa terrane, which is an integral part of the Cimmerian continent, rifted off from the Gondwanan landmass during the Early Permian (Ueno, 2003; Zanchi and Gaetani, 2011). Other paleogeographical models suggest that the Neotethyan Ocean opened between the Late Permian to Early Triassic and had developed considerable width by ca. 250 Ma (Kroner et al., 2016; Stampfli and Borel, 2002; Torsvik et al., 2012). A Permian–Triassic age for the opening is supported by the presence of abundant Middle–Late Triassic radiolarians within the Indus–Yarlung Tsangpo suture zone (Mo et al., 2005a; Wang et al., 2002; Yang et al., 2002). In addition, newly reported Early Triassic to Jurassic calc-alkaline igneous rocks occur in the Gangdese magmatic belt of southern Tibet, which would support the subduction of the Neotethyan oceanic lithosphere by that time (Ji et al., 2009; Wang et al., 2016; F. Liu, 2017, personal commun.). Despite these data, few direct lines of evidence for the Neotethyan oceanic system have been preserved or reported (especially ophiolites and intra-oceanic arcs), except for the Permian rift-related volcanic rocks (Mo et al., 2005a; Zhu et al., 2010). The Indus-Yarlung Tsangpo ophiolites were previously considered to be remnants of the Neotethys; however, their formation ages cluster around ca. 130–120 Ma (Liu et al., 2016; Wu et al., 2014). Additionally, their tectonic setting has been correlated with forearc hyperextension (Huang et al., 2015; Maffione et al., 2015). Therefore, the possible preservation of older remnants of the eastern Neotethys is still an open question.

The initiation of northward subduction of Neotethyan oceanic lithosphere beneath the southern Lhasa terrane is equally controversial. More specifically, contrasting geodynamic models have been proposed for subduction-related magmatism in the Gangdese magmatic belt. Some researchers have suggested that the initial subduction of Neotethyan...
Figure 1. Tectonic framework of the Tibetan Plateau (modified after Kapp and Guynn, 2004; G.W. Li et al., 2016; Xu et al., 2015) showing the location of the Gangdese magmatic belt.
oceanic lithosphere beneath the southern margin of the Lhasa terrane occurred during the Early Cretaceous and was completed by the Late Cretaceous to early Eocene (Wu et al., 2016; Zhu et al., 2011). According to those models, the Middle Triassic to Jurassic magmatic rocks enclosed within the Gangdese magmatic belt were related to the southward subduction of the Bangong-Nujiang oceanic slab beneath the Lhasa terrane (Zhu et al., 2016). Other studies have contended that the subduction of Neotethyan oceanic lithosphere beneath the southern Lhasa terrane began during the Early Jurassic or even Middle Triassic (Kang et al., 2014; Meng et al., 2016; Wang et al., 2016; Zhang et al., 2007). These new discoveries include the Early to Middle Jurassic volcanic rocks of the Yeba, Bima, and Xiongcun Formations, in combination with synchronon plutons (Guo et al., 2013; Tang et al., 2015), as well as Middle–Late Triassic volcanic rocks in the Changguo area of southern Lhasa (Wang et al., 2016). Considering the competing models, more work is required to unravel the evolutionary history of the Gangdese magmatic belt.

Many intra-oceanic subduction systems within the Neotethys have been described. In the western segment, a Middle–Late Triassic intra-oceanic subduction system was proposed based on the arc-type basalt intercalated with early Late Triassic pelagic limestone from the Lycian nappes of southwestern Anatolia (Sayit et al., 2015) and Middle Triassic back-arc-type basalt intercalated with radiolarian chert in the Mersin melange of southern Turkey (Tekin et al., 2016). In northern Pakistan, the Kohistan complex has been interpreted as an intra-oceanic island arc (Burg, 2011; Tahirkheli et al., 1979; Yoshida et al., 1996), as were the Xiongcun Formation volcanics, Zhonga ophiolitic massif, and Zedong terrane (Aitchison et al., 2000; Dai et al., 2011; Lang et al., 2014; Tafti et al., 2014; Tang et al., 2015). However, L.L. Zhang et al. (2014) challenged the traditional view of the Zedong terrane and argued that the magmatic rocks formed in an active continental margin setting. In general, intra-oceanic subduction systems occupy nearly 40% of modern convergent margins around the globe and are marked by chains of oceanic island arcs (Leat and Larter, 2003). In the Gangdese magmatic belt, voluminous Triassic–Jurassic magmatic rocks have been discovered (Meng et al., 2016; Wang et al., 2016; F. Liu, 2017, personal commun.). Is it possible that some of the rocks within the Gangdese magmatic belt were associated with intra-oceanic subduction systems of the Neotethys?

Based on the foregoing introduction, two key questions are raised: (1) When did the onset of northward subduction of the Neotethys occur (Late Triassic, Early Jurassic, or Early Cretaceous)? (2) Are there additional and older intra-oceanic remnants of the Neotethys within the Gangdese magmatic belt? In this work, we carried out detailed studies of a Late Triassic appinite within the southernmost part of the Gangdese magmatic belt with the aim to decipher the early stage framework of the Neotethys.

GEOPOLITICAL SETTING

The Cimmerides, which were composed of several ribbon-like continental fragments, rifted off Gondwana during the Phanerozoic and were conveyed northward to the southern margin of the Eurasian continent, triggering the opening of the Neotethys (Kroner et al., 2016; Metcalfe, 1999; Şengör, 1979). From west to east, the Cimmerides include terranes located in Turkey, Iran, and Tibet (Xu et al., 2015). The Lhasa terrane was the last to collide with the Eurasian continental margin (Ji et al., 2009; Mo et al., 2005a; Sun et al., 2016; Zhu et al., 2015).

Bounded by the northern Bangong-Nujiang suture zone and the southern Indus-Yarlung Tsangpo suture zone (Fig. 1), the Lhasa block is a composite terrane. Recent discoveries have revealed that the Lhasa terrane can be divided into northern and southern subterranes along the Sumdo eclogite and peridotite belt (Yang et al., 2007; Z.M. Zhang et al., 2014). Voluminous igneous rocks occupy most of the southern subterranes, which is known as the Gangdese magmatic belt (Fig. 2). The Gangdese
maggmatic belt is marked by multistage magmatism that resulted from the subduction of the Neotethyan oceanic lithosphere and the Indo-Asian collision. The Indo-Asian collision produced voluminous magmatic rocks, such as the Linzizong volcanics and the Quxu Batholith (Lee et al., 2012; Mo et al., 2005b; Wen et al., 2008; Zhu et al., 2015; Ma et al., 2017). Postcollisional (30–10 Ma) adakitic and ultrapotassic rocks attributable to postcollisional effects are also found in this region (Chung et al., 2009).

In addition, significant crustal shortening within the Lhasa terrane, as well as in the Himalayan terranes, occurred due to the continued Indo-Asian collision (Dupont-Nivet et al., 2010; Xu et al., 2013; Yi et al., 2015). Over the past 10 yr, numerous zircon Hf isotopic studies on igneous rocks have been published. The results show that most of the precollisional magmatic rocks are marked by highly positive epsilon Hf values, implying that significant pulses of episodic juvenile material addition occurred during the Mesozoic and early Cenozoic (Hou et al., 2015; Ji et al., 2009; Wen et al., 2008).

Located immediately to the south of the Gangdese magmatic belt, the Indus-Yarlung Tsango suture zone was thought to contain remnants of Neotethyan oceanic crust. This proposal was challenged by more recent studies on the 133–123 Ma Indus-Yarlung ophiolites due to key differences from the archetypal ophiolite sequences (Wu et al., 2014). The unconformable contact between the peridotites and the Xigaze group sediments (Huang et al., 2015; Maffione et al., 2015), and the narrow age range of 133–123 Ma (Liu et al., 2016) were used to argue that the ophiolites represent oceanic lithosphere formed during forearc hyperextension, rather than remnants of the Neotethys.

**PETROGRAPHY**

Appinites are unique rocks that are formed mainly in convergent margins around the world. Appinite suites consist of a group of coeval hypabyssal and/or plutonic rocks that range from mafic to felsic in composition, with hornblende being the dominant mafic mineral. Typically, hornblende minerals occur as both large prismatic phenocrysts and in the finer-grained matrix, which suggests that the magma source was wet (Murphy, 2013). The appinitic pluton in this study is ~400 m long. The total width of the pluton is obscured by the overlying sedimentary rocks and is intruded by a younger granitic pluton (Figs. 3A–3C). Within the appinitic pluton, the mineral structures exhibit a vertical variance in composition. Hornblende accumulated in the bottom portion of the pluton (Fig. 3A) and occur as phenocrysts with plagioclase occupying the interstices (Figs. 4B and 4H). Sometimes, the hornblende crystals occur as large aggregates with subrounded shapes (Figs. 4A and 4D), and they may have been metamorphosed during crystallization or later baking by the intrusive granitic pluton. Large phenocrysts and prismatic hornblendes are surrounded by small plagioclase laths or are crosscut by plagioclase laths in the matrix (Figs. 4C and 4G). In the photomicrograph in Figure 4G, the hornblendes show clear euhedral shapes but comb-like cuspate tails, which were formed by crosscutting of later crystallized prismatic plagioclase. This observation suggests that the hornblendes were crystallized before the plagioclase. In short, the earlier-formed hornblende occupied more of the available space, which resulted in plagioclase formation along the margins or in the interstitial space of the hornblende crystals. In addition, no solid-state deformation evidence (such as recrystallized tails, kinking, undulose extinction, etc.) was observed in the hornblende or plagioclase, which suggests that their crystallization was controlled by magmatic flow without strain accumulation, rather than solid-state flow (Vernon, 2000). We also note that there was no significant “loss on ignition” (LOI) for these rocks, and no altered biotite, sericite, or epidote, which imply that any metamorphic effects were quite minor. More distal from the contact with the granitic rocks, the appinitic rocks have primary igneous textures, with hornblende showing subhedral shapes (Fig. 4I).

In the central part of the pluton, alternating cumulate layers are observed. These cumulate layers are composed of dark hornblende-cumulate layers (hornblendite) and light plagioclase-cumulate layers (Fig. 5). Fresh samples were taken from different parts of this pluton, which included the meta-appinitic rocks near the lower contact with the granite and unmetamorphosed appinitic samples from the central portion of the body. Cumulate layers (dark or light) were not sampled because they do not provide a true measure of the magma source composition.

**ANALYTICAL METHODS**

**Whole-Rock Geochemistry**

Bulk-rock major elements were measured by X-ray fluorescence spectrometry (ME-XRF), and trace elements were measured by inductively coupled plasma–mass spectrometry (ICP-MS) and inductively coupled plasma–atomic emission spectrometry (ICP-AES) at the ALS Chemex Co., Ltd., Guangzhou, China. Samples were fused with lithium metaborate-lithium tetraborate flux, which also included an oxidizing agent (lithium nitrate), and then the samples were poured into a platinum mold. The resultant disk was then analyzed by XRF spectrometry. XRF analysis was performed in conjunction with an LOI analysis at 1000 °C. The analytical data from both determinations were combined to produce a “total.” For XRF, a prepared sample was added to the lithium metaborate/lithium tetraborate flux, mixed well, and fused in a furnace at 1025 °C. Then, the resulting melt was cooled and dissolved in an acid mixture containing nitric, hydrochloric, and hydrofluoric acids. This solution was then analyzed by ICP-MS. A prepared sample was digested with perchloric, nitric, hydrofluoric, and hydrochloric acids. The residue was topped up with dilute hydrochloric acid, and the resulting solution was analyzed by ICP-AES. The results were corrected for spectral interelement interferences. The detailed analytical procedures followed the work of Ma et al. (2017). The analytical results are presented in Data Repository Table DR1.

**Whole-Rock Sr-Nd Analysis**

High-precision isotopic (Sr, Nd) measurements were carried out at Nanjing FocuMS Technology Co., Ltd., Nanjing, China. Geological rock powder was digested in high-pressure polytetrafluoroethylene (PTFE) bombs. Strontium and Nd were purified from the same digestion solution by two-step column chemistry. The first exchange column, with Bio-Rad AG50W-X8 and Sr Spec resin, was used to separate the Sr and rare earth elements (REEs) from the sample matrix. Neodymium was separated from the other REEs on the second column using Ln Spec-coated Teflon powder. The Sr- and Nd-bearing elutions were dried and redissolved in 1.0 mL of 2 wt% HNO3. Small aliquots of each were analyzed using an Agilent Technologies 7700x quadrupole ICP-MS (Hachioji, Tokyo, Japan) to determine the exact available contents of Sr and Nd. Diluted solutions

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*GSA Data Repository Item 2018191: Text File DR1 presents the analytical procedures for analysis of bulk-rock major and trace elements, bulk-rock Sr-Nd isotopes, electron microprobe analysis (EMPA) of mineral compositions, LA-ICPMS zircon U-Pb dating, and LA-MC-ICPMS zircon Lu-Hf isotopes. Table DR6 shows the analytical results and relevant calculated parameters. Table DR7 presents collected age data for the Middle Triassic to Jurassic igneous rocks in the Gangdese magmatic belt, southern Tibet. The data repository item is available at http://www.geosociety.org/datarepository/2018, or on request from editing@geosociety.org.*
(50 ppb Sr, 50 ppb Nd, doping with 10 ppb Tl) were introduced into a Nu Instruments Nu Plasma II multicollector (MC) ICP-MS (Wrexham, Wales, UK) using a Teledyne Cetac Technologies Aridus II desolvating nebulizer system (Omaha, Nebraska, USA).

Raw data for isotopic ratios were corrected for mass fractionation by normalizing to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ for Sr and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ for Nd with an exponential law. International isotopic standards (NIST SRM 987 for Sr, JNdi-1 for Nd) were periodically analyzed to correct the instrumental drift (Höppe, 1999; Tanaka and Toda, 2000). Geochemical reference materials U.S. Geological Survey (USGS) BCR-2, BHVO-2, AVG-2, and RGM-2 were treated as quality control (Weis et al., 2006). The analytical results are presented in Table DR2.

For the calculation of the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, $\varepsilon_{\text{Nd}}(t)$ values, and Nd model ages, we used the decay constants of $^{147}\text{Sm} = 6.54 \times 10^{-12}$ yr$^{-1}$ (Lugmair and Marti, 1978) and $^{87}\text{Rb} = 1.42 \times 10^{-11}$ yr$^{-1}$ (Steiger and Jäger, 1977); chondrite uniform reservoir (CHUR) $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of 0.1967 (Jacobsen and Wasserburg, 1980); CHUR $^{143}\text{Nd}/^{147}\text{Nd}$ ratio of 0.512638 (Goldstein et al., 1984); and $^{143}\text{Nd}/^{147}\text{Nd}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ depleted mantle ratios of 0.513151 and 0.2136 (Liew and Hofmann, 1988), respectively.

**Electron Microprobe Analysis (EMPA)**

Based on detailed petrographic observations, representative hornblende, plagioclase, ilmenite, and magnetite from thin sections were...
selected for electron microprobe analyses (EMPA). Mineral chemical compositions were determined using a JEOL JXA-8230 electron microprobe with a 5 μm beam spot (15.0 kV accelerating voltage, 20 nA beam current) at the Institute of Mineral Resources, Chinese Academy of Geological Sciences, Beijing, China. The analytical results are presented in Table DR3.

**Zircon LA-ICP-MS U-Pb Dating**

Zircon U-Pb ages were measured using an Agilent 7500a ICP-MS attached to a NewWave 213 nm laser-ablation system with an in-house sample cell at the State Key Laboratory for Mineral Deposits Research, Nanjing University, Nanjing, China. The laser light beam with a diameter of ~32 μm and a repetition rate of 5 Hz under a 70% energy condition was chosen for the spot analysis. Isotope mass fractionation was normalized through an external standard GEMOC GJ-1 with a 207Pb/206Pb age = 608.5 ± 1.5 Ma (Jackson et al., 2004), and the analytical accuracy was monitored using the Mud Tank zircon standard, which has an intercept age of 732 ± 5 Ma (Black and Gulson, 1978).

Zircon analyses were carried out in runs of 15 analyses, which included 5 zircon standards and up to 10 sample spots. The U-Pb age results were calculated from the raw signal data using the online software package GLITTER (ver. 4.4; www.mq.edu.au/GEMOC). The 204Pb could not be measured due to a low signal and interference from 204Hg in the gas supply; thus, common Pb correction was performed using the EXCEL program ComPbCorr#3–15G (Andersen, 2002). The U-Th-Pb age calculations and concordia plotting were completed using the ISOPLOT/Ex program (ver. 2.49; Ludwig, 2001). Zircon Th and U concentrations were calculated by comparing the relative signal intensity between the zircon samples and standard zircon GJ-1 (Th = 8 ppm, U = 330 ppm) using the EXCEL program Data Templatev2b from the Australian Research Council National Key Centre for Geochemical Evolution and Metallogeny of Continents.
Zircon LA-ICP-MS Hf Isotopic Analysis

Zircon Hf isotopes were analyzed with a 193 nm laser attached to a Neptune MC-ICP-MS (Thermo Finnigan, Bremen, Germany) at the Key Laboratory of Continental Tectonics and Dynamics, Institute of Geology, Chinese Academy of Geological Sciences, Beijing, China. Most LA-MC-ICPMS analyses were carried out with a beam size of ~32 μm. A decay constant for $^{176}\text{Lu}$ of $1.867 \times 10^{-11}$ yr$^{-1}$ (Söderlund et al., 2004) was adopted to analyze the initial $^{176}\text{Lu}/^{177}\text{Hf}$ ratios, while the chondritic values of $^{176}\text{Lu}/^{177}\text{Hf} = 0.0336$ and $^{176}\text{Hf}/^{177}\text{Hf} = 0.282785$ were chosen to obtain the $\varepsilon_{\text{Hf}}(t)$ values (Bouvier et al., 2008). The single-stage model age ($T_{\text{DM1}}$) was calculated relative to the depleted mantle with present-day values of $^{176}\text{Hf}/^{177}\text{Hf} = 0.28325$ and $^{176}\text{Lu}/^{177}\text{Hf} = 0.0384$ (Griffin et al., 2000). A two-stage continental model age ($T_{\text{DMC}}$) was calculated by projecting the initial $^{176}\text{Hf}/^{177}\text{Hf}$ of zircon back to the depleted mantle growth curve using $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$ for the average continental crust (Griffin et al., 2000, 2002). The analytical results are presented in Table DR5.

WHOLE-ROCK GEOCHEMISTRY

Whole-rock major- and trace-elemental geochemistry was carried out on 16 fresh samples. The analytical techniques are provided in Data Repository Text DR1, while the results are listed in Table DR1. In addition, the previously reported results of Meng et al. (2016) are also shown in this table. Whole-rock Sr-Nd isotopes were analyzed for six samples with the smallest LOI (~0.44–0.54 wt%). The results are shown in Table DR2. The contents of SiO$_2$ ranged between 43 and 55 wt%, and most samples were in the 46–50 wt% range. Other oxide contents also showed variable concentrations, such as Al$_2$O$_3$ (17.2–20.5 wt%), MgO (3.5–8.1 wt%), and Na$_2$O (1.0–4.7 wt%). This compositional variation is consistent with our field observations. In a discrimination plot of FeO$^+$-MgO-(Na$_2$O + K$_2$O),
The appinitic samples show a close affinity to typical arc-related mafic cumulate tholeiitic rocks (Fig. 6A), and they fall into the tholeiitic field on the FeO'/MgO versus SiO_2 diagram (Fig. 6B). The samples have very similar trace-element patterns, with only slight variations (Figs. 6C and 6D). The similar trace-element patterns, in conjunction with the wide range of SiO_2 values, imply that the appinite was formed through a continuous fractional crystallization process (Murphy, 2013) and in concert with the variation of rock types in the field and mineral structures. In general, the appinitic suite consists of a group of coeval hypabyssal and/or plutonic rocks, ranging from ultramafic to felsic in composition, in which hornblende is the dominant mafic mineral (Murphy, 2013).

On chondrite-normalized REE diagrams, the light (L) REEs are enriched relative to the heavy (H) REEs and show a slightly right-dipping slope. The REE patterns show a large distribution, but there are similar trends between samples with slightly positive Eu anomalies (Fig. 6C). This phenomenon is most likely related to the accumulation of plagioclase (Fig. 4B). On the primitive mantle–normalized trace-elemental diagrams (Fig. 6D), all the appinitic rocks are characterized by a strong enrichment in large ion lithophile elements (LILEs) such as Cs, Ba, and U, but there are remarkable negative anomalies in the high field strength elements (HFSEs), such as Nb and Ta, relative to the neighboring elements. Furthermore, some samples are depleted in Zr and Hf.

The appinitic rocks have narrow ranges of (_87Sr/86Sr) ratios (0.703262–0.703505), (_143Nd/144Nd) values (0.512716–0.512747), and positive ε Nd(t) values (+6.9 to +7.5; Fig. 7A). In the diagram of ε Nd(t) versus (_87Sr/86Sr), the studied appinitic rocks fall into the field of Neotethyan ophiolites (Fig. 7A), revealing that the magma source of these appinitic rocks was derived from depleted mantle.

MINERAL CHEMICAL COMPOSITIONS, GEOTHERMOMETRY, AND GEOBAROMETRY

Mineral compositions were determined by EMPA. The detailed analytical techniques are discussed in Data Repository Text DR1, and the analyzed results are listed in Table DR3. In this study, representative hornblende, plagioclase, ilmenite, and magnetite were selected for EMPA analysis.

The larger magmatic hornblende and plagioclase pairs in the central portion of the appinite were selected for EMPA analysis (Figs. 4H and 4I). The hornblendes have high Al2O3, FeO, MgO, and CaO concentrations (Table DR3), are classified according to the nomenclature of Leake et al. (1997), and have a mineral formula calculation based on 23 oxygens. The hornblendes fall into the magnesiohornblende and tschermakite fields (Fig. 7B), according to the nomenclature of Leake et al. (1997), typical of magmatic calcic amphiboles. All the hornblendes are characterized by (Na +...
K<sub>A</sub> < 0.50, Ca<sub>A</sub> < 0.50, and Ca > 1.50 on the M<sub>3</sub> sites. The plagioclases exhibit a very narrow range of albite (Ab) and anorthite (An) values. In the following calculations of the hornblende-plagioclase geothermometer and geobarometer, the mean composition of plagioclase is used.

Hornblendes are stable across a wide pressure-temperature (P-T) range between 1 and 23 kbar and between 400 °C and 1150 °C (Blundy and Holland, 1990); however, the hornblende geothermometer and geobarometer are only valid at pressures >2 kbar (Johnson and Rutherford, 1989; Schmidt, 1992). The hornblende and hornblende-plagioclase geothermometers are based on the values of Si and Al cations in the tetrahedral positions in hornblende (Blundy and Holland, 1990; Holland and Blundy, 1994; Ridolfi and Renzulli, 2012), while the hornblende geobarometer is calculated using empirical and experimental equations according to the Al contents of hornblendes (Anderson and Smith, 1995; Blundy and Holland, 1990; Hammarsstrom and Zen, 1986; Holland and Blundy, 1994; Hollister et al., 1987; Johnson and Rutherford, 1989; Schmidt, 1992).

Based on the edenite-tremolite reaction, Blundy and Holland (1990) proposed an empirical hornblende-plagioclase thermometer that could only be applied to quartz-bearing intermediate to felsic igneous rocks. In their seminal study, Holland and Blundy (1994) recalibrated their previous hornblende-plagioclase thermometer to two thermometers, in which thermometer A (edenite-tremolite reaction) is applicable to quartz-bearing metabasites and thermometer B (edenite-rich-richterite reaction) is for quartz-free igneous rocks. In the present study, all results based on these three equations are presented in Figure 8A. In addition, hornblende thermometers from Ridolfi and Renzulli (2012) were also adopted to calculate the crystallization temperature of the appinitic pluton. Thermometer A is only based on the chemical composition of hornblende, whereas thermometer B is calibrated by independent pressure data. There are no independent pressure data for our appinite, and so we used pressure data for the metamorphic rocks of the Sangri Group in Nymo (Sun et al., 2011), which may be coeval with the sequences adjoining our appinite. The temperature ranges span from 750 °C to 900 °C (Fig. 8A).

The EMPA data for ilmenite and magnetite are listed in Table DR3. Ilmenites have relatively high TiO<sub>2</sub> contents, ranging from 49.6 to 52.9 wt%. In contrast, the magnetites have very low TiO<sub>2</sub> contents, which hinders their use in Fe-Ti oxide thermometry (Evans et al., 2016; Hora et al., 2013). The Mn/Mg ratios in ilmenites are a function of temperature and therefore can be used to approximate the crystallization temperature of the magma (Evans et al., 2016). The ilmenite Mn/Mg ratios in the appinite suggest that they crystallized at temperatures no greater than 770 °C (Fig. 8B). In contrast, the magnetite yielded temperature estimates far below those of the ilmenites, which implies a later crystallization. The relationship between the temperature estimates is consistent with the mineral textures and contact relationships (Figs. 8C–8F).

Several Al-in-hornblende barometers have been proposed (Blundy and Holland, 1990; Hammarstrom and Zen, 1986; Hollister et al., 1987; Johnson and Rutherford, 1989; Schmidt, 1992). Temperature also plays an important role in the Al-content of hornblendes (Blundy and Holland, 1990). Therefore, Anderson and Smith (1995) presented a new formula for the Al-in-hornblende barometer that factors in pressure, temperature, and oxygen fugacity. The new formula introduces a temperature correction for the hornblende-plagioclase thermometer proposed by Blundy and Holland (1990). In this study, the temperature correction was employed from the hornblende-plagioclase thermometer of Holland and Blundy (1994). The crystallization pressures obtained from these hornblende-plagioclase barometers are equivalent to the emplacement pressures, and the emplacement depth was obtained by a conversion constant of 1 kbar = 3.7 km of crust (Tulloch and Challis, 2000). All aforementioned formulas were used to calculate the emplacement depth, and the results fall into the range 14.5–19.5 km (Fig. 9).

**ZIRCON U-Pb GEOCHRONOLOGY AND Hf ISOTOPES**

The analytical techniques for zircon geochronology and Hf isotope analysis are provided in Data Repository Text DR1. The zircon U-Pb analyses are presented in Table DR4. Some zircon grains from the appinite showed broad and banded zoning, while others showed oscillatory zoning in cathodoluminescence (CL) images (Fig. 10). Moreover, nearly all of the zircon grains displayed euhedral prismatic shapes (~50–150 μm in length and ~40–80 μm in width; Fig. 10). These observations indicate that these zircons are magmatic in origin.

In total, 15, 15, 15, and 10 analyses obtained for the zircon grains from samples xm74, xm75, xm76, and xm77, respectively, yielded concordia...
Figure 8. Geothermometry results for the appinite. (A) Results of hornblende and hornblende-plagioclase thermometers, where results were calculated using 1—hornblende-plagioclase thermometer of Blundy and Holland (1990), 2—hornblende thermometer without independent pressure data of Ridolfi and Renzulli (2012), 3—hornblende thermometer with independent pressure data of Ridolfi and Renzulli (2012), 4—hornblende-plagioclase thermometer equation A (with quartz) of Holland and Blundy (1994), and 5—hornblende-plagioclase thermometer equation B (with or without quartz) of Holland and Blundy (1994), respectively. (B) Results of ilmenite thermometer of Evans et al. (2016). (C–F) Backscattered electron (BSE) images showing the analysis spot positions on the ilmenites and magnetites. Abbreviations: Ilm—ilmenite; Mag—magnetite.
ages of 214 ± 2.6 Ma (mean square of weighted deviates [MSWD] = 1.1), 214.8 ± 2.6 Ma (MSWD = 0.33), 213.2 ± 8.3 Ma (MSWD = 1.02), and 220 ± 4.1 Ma (MSWD = 0.85), respectively (Fig. 11A). All zircons had high Th/U ratios (0.8–2.4), which is consistent with their magmatic origin (Hoskin and Black, 2000; Corfu et al., 2003; see also Table DR4).

The analytical results of the zircon Lu-Hf isotopic compositions are listed in Table DR5. Fifty-five analyses of the dated U-Pb zircons from the four samples yielded 176Lu/177Hf ratios between 0.000864 and 0.005021 and 176Hf/177Hf ratios between 0.282973 and 0.283156. The εHf(t) values were calculated using the crystallization age of the pluton (215 Ma; weighted mean age for four samples), and they varied from +11.1 to +17.5, with the corresponding depleted mantle Hf model ages (TDM1) mainly falling into the range of 320–220 Ma (Fig. 11B; Table DR5).

DISCUSSION

Late Triassic Intra-Oceanic Subduction in the Neotethys

Source and Subduction Signature of the Appinite

The cumulate appinite suite in this study consists of magmatic hornblende and plagioclase, implying its generation from a wet basaltic magma (Gaetani and Grove, 1998; Murphy, 2013; Sisson and Grove, 1993). A water/fluid-enriched condition is implied by the presence of hornblende, enrichment in Al2O3, Eu, and Sr, and depletion of HFSEs (Table DR1). A water-enriched condition has the effect of suppressing plagioclase crystallization and fractionating olivine and clinopyroxene and is consistent with our findings (Feig et al., 2006; Sisson and Grove, 1993; Smith et al., 2009). The field photos and photomicrographs reveal that the plagioclases occur as the interstitial phase between the hornblende phenocrysts (Figs. 4A, 4B, and 4H). Microscopic textures illustrate that the plagioclases crosscut the hornblendes and were crystallized as small laths, implying that they postdate the hornblendes (Figs. 4C and 4G). These observations are consistent with the appinite having been crystallized from a wet (water-enriched) magma.

The hornblende and hornblende-plagioclase thermometers yielded a crystallization temperature range of 750–900 °C (Fig. 8A) for the appinite, which is far below the crystallization temperatures of most mafic rocks. Magmatic ilmenite coexists with hornblende and plagioclase, and its formation temperature represents the temperature conditions under which the appinite was formed. Based on the experimental thermometer...
formula of Evans et al. (2016), the maximum apparent temperature of the appinite was ~770 °C (Fig. 8B), which is considerably lower than the crystallization temperature of a normal basaltic magma (Lee et al., 2009; Wan et al., 2013). Generally, if the magma is wet, the crystallization temperature will be sharply decreased (Wan et al., 2013). Thus, the geothermometer results support the interpretation that the magma source of the studied appinite was water-enriched.

A high water content (>3 wt%) in magma is required for hornblende to crystallize (Sisson and Grove, 1993). The water content in lithospheric and asthenospheric mantle is less than 0.1 wt%, and it is less than 0.001 wt% in continental crust (Williams and Hemley, 2001). Therefore, the most likely source of additional hydration for the magma would be from the mantle wedge in a subduction zone (Gaetani and Grove, 1998; Hirschmann, 2006; Murphy, 2013; Shaw et al., 2008). In wet magmas, high oxygen fugacity (O2) and oxidized conditions would be characteristic and result in the early-forming phase of vanadium titanomagnetite (Figs. 8C–8F; Garcia and Jacobson, 1979; Sun et al., 2013).

Diagnostic features of subduction-related arc magma for the studied appinite are implicit in the elevated Th, La, and LILEs relative to the HPSEs and negative Nb and Ta anomalies (Figs. 6C and 6D), implying the introduction of slab-derived melts and/or fluids into the mantle wedge (Hawkesworth et al., 1993; Hawkins and Ishizuka, 2009; Pearce, 2008; Prouteau et al., 2001; Reagan et al., 2010). Furthermore, all the investigated hornblendes are characterized by Σ(Ca + Na) ≥ 1.00, Na < 0.50, and Ca ≥ 1.50 on the M1 site (Leake et al., 1997), and they fall into the field of calcic amphiboles that are typical of magmatic origins (Fig. 7B). In general, calcic amphiboles (hornblendes) are typical of I-type intrusions and therefore indicate the existence of a subduction system (Clemens and Wall, 1984).

The abundance and elemental ratios of LREEs, Ba, Th, and Sr are useful proxies for determining the extent of sediment input into the arc magma source (Hole et al., 1984). In the present study, the Ce/Ce* ratios ranged from 0.95 to 1.05, where Ce* is reached by extrapolating between La and Nd (clustering ~1.0; Figs. 12A and 12B). These results are comparable to those of primitive arcs such as Vanuatu, Palau, and New Britain, but they differ markedly from those of the Lesser Antilles, which implies an insigniﬁcant sediment input (Hawkins and Ishizuka, 2009). Similarly, the apinites yielded low Th/La (<0.2) and Th/Yb (<0.6) ratios (Table DR1), indicating a greater role for aqueous ﬂuids relative to sediment input (Woodhead et al., 2001). Last, the samples plot parallel to the Ba/Th and Sr/Nd axes (Figs. 12C and 12D), which, in combination with their low La and Th levels relative to HREEs, argues for aqueous ﬂuids from the dehydration of a subducting mid-ocean-ridge basalt (MORB) crustal slab rather than from melted subducted sediments (Labanlieh et al., 2012; Woodhead et al., 1998).

Tectonic Setting of the Cumulate Appinite

The petrological and geochemical results presented here reveal that the parental magma of the appinite was derived from a water-enriched depleted mantle that was metasomatized by subducted slab-derived ﬂuids/melts. The depleted whole-rock Sr-Nd isotopes (Fig. 7A) and highly positive εNd(t) values (Fig. 11B) suggest a depleted mantle source.

Tectonic discrimination diagrams reveal that the appinite was formed in an island-arc or volcanic-arc ﬁeld (Fig. 13). To distinguish between a continental arc and an intra-oceanic arc, additional analyses are required. Condie (1989) proposed using a plot of La/Yb versus Th/Yb to differentiate these two settings, and our samples fall close to the intra-oceanic arc ﬁeld (Fig. 14A). In the Th/Yb versus Sr/Nd plot, our samples are similar to typical intra-oceanic arc ﬁelds such as the South Sandwich, Vanuatu, New Britain, and Tonga Island arcs, where the introduction of sediments to the magma source is insigniﬁcant (Fig. 14B). This conclusion is conﬁrmed by further discrimination of the diagram Th/Yb versus Nb/Yb (Pearce, 2008), in which all samples show a close afﬁnity to the Mariana arc-basin system (Fig. 14C; Pearce et al., 2005) and island-arc tholeiites (Fig. 14D; Pearce, 2014). The whole-rock Sr isotopes ([87Sr/86Sr] = 0.7033–0.7035; Fig. 7A; Table DR2) resemble Mariana arc lavas (0.7033–0.7040; Hole et al., 1984). In addition, this cumulate appinite was emplaced at depths between 14.5 and 19.5 km (Fig. 9), which is comparable to the general depths of arc-related cumulates in an intra-oceanic arc setting (DeBari and Greene, 2011; Leat and Larter, 2003).
Incompatible elements such as TiO₂, Na₂O, and K₂O are concentrated in the melt as mantle melting or crystal fractionation proceeds (Kelley et al., 2010). In basaltic systems, deeper melts are progressively enriched in iron (Klein and Langmuir, 1987), whereas K₂O contents in convergent margin magma sources are strongly affected by subduction-related metasomatism (Ribeiro et al., 2013). Consequently, the Na₂O, TiO₂, and FeO contents are good proxies for the degree and depth of melting when they are corrected for olivine fractionation to infer their Na₈, Ti₈, and Fe₈ contents (Klein and Langmuir, 1987; Ribeiro et al., 2013). Triggered by high degrees of melting at greater depths in the presence of slab-derived fluids, the K₈O/TiO₂ ratios are negatively correlated with the Fe₈ and Ti₈ contents (Klein and Langmuir, 1987; Ribeiro et al., 2013). Triggered by high degrees of melting at greater depths in the presence of slab-derived fluids, the K₈O/TiO₂ ratios are negatively correlated with the Fe₈ and Ti₈ contents (Figs. 15A and 15B), which confirms their genesis from an arc magma source (Ribeiro et al., 2013). Moreover, all K₈O/TiO₂ values lie above the typical normal (N) MORB fields (Figs. 15A and 15B), further demonstrating a subduction component in their magma genesis (Shen and Forsyth, 1995). In general, back-arc mafic magma sources are characterized by more melting at shallower depths compared to MORBs, where Na₈ positively correlates with Fe₈. However, this was not observed in our study. A negative correlation between Na₈ and Fe₈ is shown in Figure 15C, and this implies that magma production occurred at deeper levels (Ribeiro et al., 2013). In Figure 15D (Ti₈ vs. Fe₈), the samples fall closer to the arc field. Here, Ti₈ and Fe₈ are positively correlated, in contrast to the negative correlation between Na₈ and Fe₈ (Fig. 15C). The positive correlation may have resulted from the crystallization of titanomagnetite. All these observations suggest that the appinite was formed in an intra-oceanic arc with little sedimentary input, which is similar to the modern-day settings of New Britain and Vanuatu in the central south Pacific.

**Intra-Oceanic Arc of the Neotethys**

Based on the analysis of the stratigraphy and tectonic framework of southern Tibet, we believe that remnants of Neotethyan oceanic crust must have been preserved within the mélangé zone of the northern Tethyan
Figure 13. Chemical classification diagrams for tectonic setting. (A) Ti vs. Zr, (B) Cr vs. Y, (C) Cr vs. Ce/Sr (after Pearce, 1982), (D) Hf-Ta-Th (after Wood, 1980), (E) Y-Nb-La (after Cabanis and Lecolle, 1989), and (F) FeO/MgO vs. TiO₂ (after Glassley, 1974). The published whole-rock geochemical data are quoted from Meng et al. (2016). Abbreviations: CAB—calc-alkaline basalt; MORB—mid-ocean-ridge basalt; E-MORB—enriched mid-ocean-ridge basalt; IAT—island-arc tholeiite; N-MORB—normal mid-ocean-ridge basalt; OIB—oceanic-island basalt; RT—ridge tholeiite; VAB—volcanic arc basalt; VAT—volcanic arc tholeiite; WPB—within-plate basalt; WPT—within-plate tholeiite.
Himalaya. Candidates include the Bainang and Zedong terranes (Aitchison et al., 2000) or those within the Gangdese magmatic belt. Previously published data, especially those on Triassic–Jurassic radiolarites and ophiolites within the mélangé zones, are compiled in Table DR6. This compilation leads us to propose a model of the Neotethyan Ocean beginning at least by the Middle–Late Triassic (or earlier).

Here, our detailed investigations in the Gangdese magmatic belt revealed that a ca. 215 Ma cumulate appinite pluton occurred in the Quxu Batholith. The appinite exhibits geochemical similarities to island-arc rocks formed in a primitive arc system that was triggered by intra-oceanic subduction. In addition, some synchronous calc-alkaline magmatic rocks have been identified from the Gangdese magmatic belt (Wang et al., 2016), and their generation was correlated with northward subduction of Neotethyan oceanic lithosphere. How do we incorporate these synchronous but different tectonic settings into the same Neotethyan realm?

In our opinion, a double subduction system inferred for the tectonic evolution of the northern Neotethyan Ocean. Double subduction settings have been proposed in a variety of studies, such as the 275–270 Ma westward subduction of the paleo-Pacific plate beneath the continental margins (including Jiamusi block and Eastern Australia); the Dongfanghong and Gympie intra-oceanic arcs (Buckman et al., 2015; Sun et al., 2015); the Cretaceous-age double north-dipping subduction systems in the Kohistan region in the western Neotethys Ocean (Burg, 2011); and the Zedong and Dazhuqu regions of the central Neotethys Ocean (Aitchison et al., 2000; Ali and Aitchison, 2008; McDermid et al., 2002; Ziabrev et al., 2004). Modern analogs include the double west-dipping subduction of the Philippine plate beneath the Asian continent and the Pacific plate beneath the Izu-Bonin-Mariana intra-oceanic arc chain (Deschamps and Lallemand, 2003).

In southwestern Anatolia, the arc-type Turunç basalt and intercalated early Late Triassic cherts from the Lycian Nappes are thought to have been formed in an intra-oceanic arc (Sayit et al., 2015, 2017). Further east, back-arc basalt and overlying Middle Triassic radiolarian chert within the Mersin mélangé of southeast Turkey are argued to have formed within an intra-oceanic subduction system in the Neotethys (Sayit et al., 2017; Tekin et al., 2016). In addition, Middle–Late Triassic subduction-related
volcanic rocks in the Changguo region of southern Lhasa were proposed to have been formed in an active continental margin setting (Wang et al., 2016). Those studies combined with our results favor the existence of a double subduction system in the eastern Neotethyan area during the Late Triassic (Fig. 16).

If we accept that the northward drift rate of the Lhasa terrane was at least 5 cm/yr (Z.Y. Li et al., 2016), along with the opening of the Neotethys during the Permian (Kroner et al., 2016; Torsvik et al., 2012), the Lhasa terrane would have been >2000 km away from the Gondwanan margin (Indian or Australian) by the Late Triassic (ca. 215 Ma; Fig. 17). Numerous Triassic–Jurassic–aged subduction-related calc-alkaline rocks outcrop along the southern margin of the Gangdese magmatic belt (Table DR7). These calc-alkaline rocks are thought to have resulted from the northward subduction of Neotethyan oceanic lithosphere beneath the southern margin of the Lhasa terrane (Guo et al., 2013; Kang et al., 2014; Wang et al., 2016), rather than southward subduction of Bangong-Nujiang oceanic lithosphere beneath the northern margin of the Lhasa terrane. Most of the magmatic rocks in the central-eastern part of the northern margin of the Lhasa terrane cluster between 120 and 110 Ma in age, and there are very few granitoid plutons of ca. 160 Ma age in the western part (Zhu et al., 2016). Considering the substantial shortening of the Lhasa terrane during the Late Cretaceous (Kapp et al., 2007), the proto–Lhasa terrane would have been ~500 km wide from north to south. The proposal that subduction-related Triassic–Jurassic magmatic rocks along the southern margin of the Lhasa terrane were formed by southward subduction of the Bangong-Nujiang oceanic slab is untenable.

There is a second suture zone within the Lhasa terrane that is demarcated by the Sumdo eclogite and MORB-type peridotite (Yang et al., 2007; Z.M. Zhang et al., 2014). The subduction polarity of Sumdo oceanic lithosphere remains ambiguous. Previous geochronological and structural analyses point to the closure of the Sumdo Ocean at ca. 240–220 Ma (H.Q. Li et al., 2011; Yang et al., 2007). Thus, the subduction of Sumdo oceanic lithosphere preceded the Triassic–Jurassic calc-alkaline magmatism observed in the Gangdese magmatic belt.

**Opening of the Neotethys**

The Pangaea supercontinent was surrounded by the paleo–Pacific Ocean during the Carboniferous period (Kroner et al., 2016; Veivers and Tewari, 1995). Throughout the late Carboniferous to Permian, the rifting of Pangaea...
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Figure 16. Geodynamic model for the double subduction system within the Neotethys in the Late Triassic.

Figure 17. Paleogeographic reconstruction of the Neotethys during 250–215 Ma. (A) Early Triassic (ca. 250 Ma) paleogeographic reconstruction of the Neotethyan realm, showing the position of the Lhasa terrane. (B) Late Triassic (ca. 215 Ma) paleogeographic reconstruction of the Neotethyan realm, showing the intra-oceanic subduction system within the Neotethyan Ocean. Figures are modified according to previously published work (Cawood, 2005; Domeier and Torsvik, 2014; Ettensohn, 1997; Faure et al., 2016; Seton et al., 2012; Stampfli and Borel, 2002; Torsvik et al., 2012; Veevers, 2004; Wilmson et al., 2009; Xiao et al., 2015). The relative position of the Lhasa terrane in the Late Triassic (ca. 215 Ma) is from Zhou et al. (2016). The notion of intra-oceanic subduction in the western part of the Neotethyan Ocean is from Sayit et al. (2015, 2017) and Tekin et al. (2016). Abbreviations: AF—African block; AU—Australian block; IA—Indian block; LA—Lhasa terrane; TR—Turkey terrane.
was accompanied by volcanism and rift-related sedimentary rocks (Veevers and Tewari, 1995). As the supercontinent began to break apart, the rifting of the Cimmerides from Gondwana led to the opening of the Neotethyan Ocean during the late Carboniferous to Permian. The rift-related volcano-sedimentary rocks are preserved along the Neotethyan realm from western Turkey through the Arabian margin, southeast Pamir, Tethyan Himalaya, and to the eastern Sibumasu terrane (Domeier and Torsvik, 2014). These rift-related volcanic rocks include Middle–Late Triassic basaltic erupted within a continent-ocean transition zone in the Adiyaman Province of SE Turkey (Robertson et al., 2015), Middle Permian basaltic from the Oman rifted margin (Lapierre et al., 2004), Permian Panjal Traps basalt in India (Chauvet et al., 2008; Shellnutt et al., 2011), Permian basalt in South Tibet (Ganzanti et al., 1999; Ji et al., 2005), and Middle Permian diabase and basalt in the Tethyan Himalaya (Zeng et al., 2012; Zhu et al., 2010).

In the Lhasa terrane, Upper Carboniferous mudstone and sandstone are unconformably or disconformably overlain by earliest Permian coastal-marine conglomerate and sandstone (Zhang et al., 2015; Zhao et al., 2001). As a western continuation of the Lhasa terrane, the Karakoram terrane features an unconformity between the pre-Permian strata and the overlying lowermost Permian sandstone (Zanchi and Gaetani, 2011). These breakup unconformities are likely related to the opening of the Neotethys (Yang et al., 2017; Zanchi and Gaetani, 2011). A layer of basalt occurs in the lower part of the Lower Permian Pangduo Group in the Linzhou Basin, which indicates a rifting environment (Ji et al., 2005). Furthermore, the Permian marine biostratigraphy of the Lhasa terrane is characterized by an admixture of cold- and warm-water fauna. This was most likely linked to the northward drift of the Cimmerian continent through the warm-water Cathaysian fauna (Zhang et al., 2013).

Numerous models for the reconstruction of the Neotethyan realm have been proposed, and there is a general consensus that the opening of the Neotethyan Ocean began during the Early–Middle Permian, and by ca. 250 Ma, the ocean had reached a considerable scale (Fig. 17; Domeier and Torsvik, 2014; Kroner et al., 2016; Stampfli and Borel, 2002; Sun et al., 2015; Torsvik et al., 2012; Veevers and Tewari, 1995; Xiao et al., 2015; Ziaibrev et al., 2004). Based on newly reported paleomagnetic data from Triassic marine sediments in the Lhasa terrane, Zhou et al. (2016) suggested that the Lhasa terrane was located at 16.5°S ± 3.9°S during the Triassic Period, which is at least 2000 km away from the protocontinent (e.g., Indian or Australian). By the Early Jurassic (ca. 180 Ma), a well-dated paleomagnetic pole from the Sangri Group volcanics (Lhasa terrane) indicates that the Lhasa terrane had moved close to the equator during that time (3.7°S ± 3.4°S; Z.Y. Li et al., 2016).

Combined with the early Late Triassic basalt intercalated with pelagic sediments in southwestern Anatolia (Sayit et al., 2015), Middle Triassic radiolarian chert and back-arc basalt in the Mersin melange (southern Turkey; Sayit et al., 2017; Tekin et al., 2016), Late Triassic radiolarian chert in the Danquiu-Xuiguguba melange (southern Tibet; Mo et al., 2005a), and other Middle Triassic radiolarites in the Yarlung-Tsangpo suture zone or southern melanges (Pan et al., 2006; Table DR6), our study indicates that the Neotethyan Ocean was open to a considerable size during the Middle Triassic.

CONCLUSIONS

In this study, we documented a cumulate apolline suite from the Gangdese belt, southern Tibet. The cumulate apolline was emplaced between 220 and 213 Ma at a depth of ~14.5–19.5 km with crystallization temperatures of 750–900 °C. The apolline is composed of major hornblende and minor plagioclase, which suggests it was generated from a water-enriched magma. Isotopic data imply that the magma source was dominated by depleted mantle, which was derived from a mantle wedge metasomatized by the introduction of subduction-derived fluids and with minor addition of sediments. According to its chemical characteristics, the apolline was formed in an intra-oceanic arc setting. In combination with coeval discoveries in southern Turkey, we suggest that a Middle–Late Triassic intra-oceanic subduction system existed within the Neotethys from west to east. Together with the Middle–Late Triassic calc-alkaline volcanic rocks in the Gangdese magmatic belt, the studied apolline lends support to a double subduction system collage in the northern Neotethys, an intra-oceanic subduction system within the Neotethys, and northward subduction of Neotethyan oceanic lithosphere beneath the southern margin of the Lhasa terrane. Based on these observations, we argue that the opening of the Neotethyan Ocean was under way by the Early Triassic.

Although we argue that intra-oceanic subduction of the Neotethys took place during the Late Triassic and that the opening of the Neotethys occurred in the Early Triassic or Permian, we recognize the scarcity of typical rock assemblages expected in an intra-oceanic arc setting. Further geological, geochronological, and geochemical investigations are required to test this model.

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