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.

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INTRODUCTION

The Indus-Yarlung Tsangpo ophiolitic mélange 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 Neotethyan 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 radiolarites 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 synchronous 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, Zhongba 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.

**GEOLOGICAL 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).
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 Tsangpo 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

Appinite suites 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 appinite 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 appinite pluton, the mineral structures exhibit a vertical variance in composition. Hornblende crystals accumulated in the bottom portion of the pluton (Fig. 3A) and occur as phenocrysts with plagioclase occupying the interstices (Figs. 3B and 3C). Sometimes, the hornblende crystals occur as large prismatic 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 plagioclase hornblendes are surrounded by small plagioclase laths or are crosscut by plagioclase laths in the matrix (Figs. 4C and 4D). In the photomicrograph in Figure 4G, the hornblendes show clear euhedral shapes but comblike 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 appinite 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 (hornblende) and light plagioclase-cumulate layers (Fig. 5). Fresh samples were taken from different parts of this pluton, which included the metabappinitic 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-ICPMS zircon Lu-Hf isotopes. Tables DR1-DR8 show the analytical results and relevant calculated parameters. Table DR6 shows compiled age constraints on ophiolites and/or radiolarians within the Indus-Yarlung Tsangpo suture (mélange) zone in the south of the Gangdese magmatic belt, southern Tibet. 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 $^{86}$Sr/$^{88}$Sr = 0.1194 for Sr and $^{146}$Nd/$^{144}$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 $^{87}$Sr/$^{86}$Sr ratios, $\epsilon_{Nd}(t)$ values, and Nd model ages, we used the decay constants of $^{147}$Sm = 6.54 × 10^{-12} \text{ yr}^{-1}$ (Lugmair and Marti, 1978) and $^8\text{Rb} = 1.42 \times 10^{-11} \text{ yr}^{-1}$ (Steiger and Jäger, 1977); chondrite uniform reservoir (CHUR) $^{147}$Sm/$^{144}$Nd ratio of 0.1967 (Jacobsen and Wasserburg, 1980); CHUR $^{143}$Nd/$^{147}$Nd ratio of 0.512638 (Goldstein et al., 1984); and $^{143}$Nd/$^{147}$Nd and $^{147}$Sm/$^{144}$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 examined...
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 \(= 608.5 \pm 1.5 \) Ma (Jackson et al., 2004), and the analytical accuracy was monitored using the Mud Tank zircon standard, which has an intercept age of \( 732 \pm 5 \) 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 \(^{204}\)Pb could not be measured due to a low signal and interference from \(^{204}\)Hg 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.

**Figure 4.** Representative field photos (A, D, E, and F) and photomicrographs (B, C, G, H, and I). (A) Hornblende phenocrysts and aggregates. (B) Small plagioclase laths occupy the place between the hornblende aggregates. (C) Hornblende phenocryst is crosscut by the plagioclase. (D–F) Equigranular textures of the hornblende and plagioclase. (G) Equigranular texture. However, the hornblende crystals show comb-like cuspate tails, implying intense crosscutting by the later-crystallized plagioclase minerals. (H) Small plagioclase crystals occur as the interstitial phase. (I) Subhedral hornblende and plagioclase showing equigranular texture. Mineral abbreviations: Hbl—hornblende; Pl—plagioclase.
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 $\epsilon_{\text{Hf}}(t)$ 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$^T$-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 SiO₂ diagram (Fig. 6B). The samples have very similar trace-elemental patterns, with only slight variations (Figs. 6C and 6D). The similar trace-element patterns, in conjunction with the wide range of SiO₂ 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 (²⁷Sr/²⁶Sr) ratios (0.703262–0.703505), (¹⁴⁴Nd/¹⁴⁴Nd) values (0.512716–0.512747), and positive εNd(t) values (+6.9 to +7.5; Fig. 7A). In the diagram of εNd(t) versus (²⁷Sr/²⁶Sr), 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 are depleted in Zr and Hf.

**Figure 6.** Chemical classification for the appinite. (A) FeO'/MgO-(Na₂O + K₂O) (after Irvine and Barag, 1971), where fields for noncumulate and cumulate rocks are from Beard (1986) and Eyuboglu et al. (2010). (B) FeO'/MgO vs. SiO₂ (after Miyashiro, 1974). (C) Chondrite-normalized rare earth element (REE) patterns. (D) Primitive mantle–normalized spider diagrams. The published whole-rock geochemical data are from Meng et al. (2016).
K)A < 0.50, CaA < 0.50, and Ca > 1.50 on the M4 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; Hammarstrom 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-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 appinites, and so we used pressure data for the metamorphic rocks of the Sangri Group in Nyomo (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 TiO2 contents, ranging from 49.6 to 52.9 wt%. In contrast, the magnetites have very low TiO2 contents, which hinder 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 appinites 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 analyzed in this study 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 zircon grains are magmatic in origin.

In total, 15, 15, 15, and 10 analyses obtained for the zircon grains from samples zm74, zm75, zm76, and zm77, 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 hornblende 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...
In general, calcic amphiboles (hornblendes) are typical of I-type (Clemens and Wall, 1984). In wet magmas, high oxygen fugacity (fO2) 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). Furthermore, all the HFSEs and negative Nb and Ta anomalies (Figs. 6C and 6D), imply 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 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 ([187Sr/186Sr] = 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 (DeBarth and Greene, 2011; Leat and Larter, 2003).
Incompatible elements such as TiO$_2$, Na$_2$O, and K$_2$O 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 K$_2$O contents in convergent margin magma sources are strongly affected by subduction-related metasomatism (Ribeiro et al., 2013). Consequently, the Na$_2$O, TiO$_2$, and FeO contents are good proxies for the degree and depth of melting when they are corrected for olivine fractionation to infer their Na$_8$, Ti$_8$, and Fe$_8$ 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 K$_2$O/TiO$_2$ ratios are negatively correlated with the Fe$_8$ and Ti$_8$ 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 K$_2$O/TiO$_2$ ratios are negatively correlated with the Fe$_8$ and Ti$_8$ 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 K$_2$O/TiO$_2$ ratios are negatively correlated with the Fe$_8$ and Ti$_8$ 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 K$_2$O/TiO$_2$ ratios are negatively correlated with the Fe$_8$ and Ti$_8$ contents (Klein and Langmuir, 1987; Ribeiro et al., 2013).

Figure 12. Source diagrams distinguishing between sediment melting and slab dehydration. (A) Ce/Ce* vs. Th and (B) Ce/Ce* vs. Ba (after Hawkins and Ishizuka, 2009). (C) Ba/Th vs. La/Sm (after Labanieh et al., 2012). (D) Sr/Nd vs. Th/Yb (after Woodhead et al., 1998). Ce/Ce* = 2 x Ce$_N$ / (La$_N$ + Nd$_N$). The chondrite standard values are from Sun and McDonough (1989). The collected whole-rock geochemical data are cited from Meng et al. (2016). The Ce/Ce* ratio fields of primitive arcs (such as New Britain and Vanuatu) lacking evidence for sediment input and Mariana arc volcanoes with some sediment input are cited from Hawkins and Ishizuka (2009).
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 mélanges 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 two 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 Türunç 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 mélange 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 Pangea supercontinent was surrounded by the paleo–Pacific Ocean during the Carboniferous period (Kroner et al., 2016; Vevers and Tewari, 1995). Throughout the late Carboniferous to Permian, the rifting of Pangea...
Late Triassic intra-oceanic arc system within Neotethys

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; Wilmsen 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.
which indicates a rifting environment (Ji et al., 2005). Furthermore, the eruption was accompanied by volcanism and rift-related sedimentary rocks (Veev-XUXUAN MA ET AL. | Late Triassic intra-oceanic arc system within Neotethys mers et al., 2011), Permian basalt in South Tibet (Garzanti 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 during the Early Triassic. The Neotethyan Ocean was under way by the Early Triassic.

CONCLUSIONS

In this study, we documented a cumulate appinitic suite from the Gangdese belt, southern Tibet. The cumulate appinites were emplaced between 220 and 213 Ma at a depth of ~14.5–19.5 km with crystallization temperatures of 750–900 °C. The appinites are 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 appinites were 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 appinite 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 during 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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