**Remnants of late Permian–Middle Triassic ocean islands in
northern Tibet: implications for the late-stage evolution of
the Paleo-Tethys Ocean**

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ABSTRACT

In this paper we present new data for the Tianquan (TQ) and Dabure (DB) ocean islands in the western segment of the Longmuco–Shuanghu–Lancangjiang suture zone, northern Tibet, including the results of major and trace element analyses, zircon U–Pb dating, and Hf isotope analyses. Our aim was to assess the genesis of these ocean islands and to consider the implications for the tectonic evolution of the region as a whole. Both TQ and DB retain an ocean-island-type double-layered structure comprising a volcanic basement (basalt and andesite) and an oceanic sedimentary cover sequence (conglomerate, limestone, and chert). The basalts and andesites in the TQ and DB are enriched in light rare earth elements and high field strength elements (Nb, Ta, Zr, Hf, and Ti), yielding chondrite-normalized REE patterns and primitive-mantle-normalized trace element patterns that are similar to those of ocean island basalts. Given the small and generally positive $\epsilon_{\text{Hf}}(t)$ values of the TQ andesites (+4.25 to +6.22) and DB andesites (−0.59 to +1.97, mostly >0), we conclude that the basalts were derived from the partial melting of garnet peridotite in the mantle and that the andesites were formed by fractional crystallization of the mafic parent magma derived from the garnet peridotite mantle. The ascending magmas underwent varying degrees of fractional crystallization but were not contaminated by crustal material. These features indicate that both TQ and DB are typical ocean islands that formed in an ocean basin. Geochemical analyses of cherts from TQ and DB show that they contain terrigenous material, indicating the proximity of a continental margin. The andesites of TQ contain zircons that yield two U–Pb ages of 251 Ma. Given that ages

of 246, 247, and 254 Ma had been reported previously, we conclude that TQ formed during the late Permian–Early Triassic. The andesites of DB contain zircons that yield U–Pb ages of 242 and 246 Ma. Taking into account the youngest age of 244 Ma from the DB basalt, we conclude that DB formed during the Middle Triassic. These data, combined with the geological history of the region, indicate that the development of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean continued after the early Permian and that the closure of this ocean was diachronous from east to west. The eastern segment of the ocean closed during the Early Triassic; however, the western segment remained at least partially open until the Middle Triassic, although the ocean was relatively small at this time. The ocean finally closed in the Late Triassic.

Key words: Tibetan Plateau; Longmuco–Shuanghu–Lancangjiang suture zone; ocean island; late Permian–Middle Triassic

1. Introduction

The Longmuco–Shuanghu–Lancangjiang suture zone (LSLSZ) crosses the middle Tibetan Plateau and lies between the Northern Qiangtang block to the north and the Southern Qiangtang block to the south (Fig. 1a). The zone is an important Paleozoic suture in terms of the development of the Tibetan Plateau (Li, 1987; Li *et al.*, 2008; Wu, 2013; Zhang *et al.*, 2013; Zhang *et al.*, 2014a; Xu *et al.*, 2015; Wang *et al.*, 2016; Zhai *et al.*, 2016). Since the 1980s, many researchers have provided

detailed information on the ophiolites, strata, and volcanic rocks of the LSLSZ and surrounding areas, and it was initially suggested that the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean was open at least by the Cambrian, expanded rapidly during the Ordovician to Middle Devonian, and was reduced in size after the Late Devonian due to subduction (Li, 1987, 2008; Li *et al.*, 2007a, 2008; Wang *et al.*, 2008; Zhang *et al.*, 2009, 2013; Yang *et al.*, 2011; Zhai *et al.*, 2011, 2013, 2016; Wu *et al.*, 2013, 2016; Hu *et al.*, 2014a; Jiang *et al.*, 2015; Fan *et al.*, 2016; Wang *et al.*, 2016). In recent years, evidence from ophiolites in the suture zone, and from the basement and sedimentary formations surrounding the suture zone, has indicated that the LSLSZ represents the main oceanic basin of the Paleo-Tethys Ocean on the Tibetan Plateau, separating Gondwana from Eurasia (Li, 2008; Wu *et al.*, 2013; Fan *et al.*, 2015a; Zhai *et al.*, 2016). Therefore, in addition to providing information on the early formation and evolution of the Tibetan Plateau, further investigation of the LSLSZ would provide a better understanding of the evolution of the Tethyan tectonic domain (Li, 1987, 2008; Li *et al.*, 2008; Wu *et al.*, 2013; Zhai *et al.*, 2016). However, the difficulties in undertaking fieldwork in the harsh climate of the region, combined with the complex geological evolution of the LSLSZ, mean that many aspects of the evolution of the suture zone remain poorly known.

The youngest ages obtained from the LSLSZ ophiolites are early Permian (Wu, 2013). Apart from late Permian–Early Triassic flysch deposits (Fan *et al.*, 2016), there appears to be a general absence of oceanic material younger than the early Permian in this suture zone. This has resulted in many key questions, such as: Did the

89 Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean exist after the early Permian?

90 If so, how did this ocean continue to develop? And when did this ocean close?

91 In this paper, we describe for the first time the newly discovered Tianquan (TQ)
92 and Dabure (DB) ocean islands from the western segment of the LSLSZ (Fig. 1a).
93 Our detailed analysis shows that TQ formed during the late Permian–Early Triassic,
94 and DB during the Middle Triassic; together, they provide late Paleozoic–Mesozoic
95 oceanic records of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean. We
96 present the results of detailed geochemical, geochronological, and Lu–Hf isotopic
97 analyses of basalts, andesites, and cherts of TQ and DB, and we use the data to
98 discuss the genesis and evolutionary history of these rocks. Based on the data and
99 regional geology, we also discuss the late-stage evolution of the
100 Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean during the late Paleozoic
101 and Mesozoic, as well as its eventual closure.

103 2. Geological background

104 The Tibetan Plateau is the highest continental plateau in the world. It lies at
105 ~4500 m above sea level and covers an area of ~2.6 million km² between
106 southwestern China and northern India (Wang *et al.*, 2013a). The plateau is located in
107 the eastern section of the Alpine–Himalayan tectonic domain and it has a complex
108 geological history that includes the formation and evolution of the Paleo-, Meso-, and
109 Neo-Tethys oceans (Pan *et al.*, 1997; Yin and Harrison, 2000; Kapp *et al.*, 2005;
110 Torsvik and Cocks, 2013; Zhu *et al.*, 2013). The closure of the Tethys Ocean occurred

111 along four main suture zones on the Tibetan Plateau (from south to north): the
112 Indus–Yarlung Zangbo suture zone (IYZSZ), the Bangong–Nujiang suture zone
113 (BNSZ), the Longmuco–Shuanghu–Lancangjiang suture zone (LSLSZ), and the
114 Jinshajiang suture zone (JSSZ; Fig. 1a). These suture zones divide the plateau from
115 south to north into the Tethys Himalayan, Lhasa, Southern Qiangtang, Northern
116 Qiangtang, and Bayan Har–Garze blocks (Fig. 1a; Pan *et al.*, 1997, 2006, 2012; Yin
117 and Harrison, 2000; Li *et al.*, 2006, 2016; Wang *et al.*, 2008; Zhai *et al.*, 2013; Zhu *et*
118 *al.*, 2013; Chen *et al.*, 2014; Hu *et al.*, 2014b; Zhang *et al.*, 2014b; Ding *et al.*, 2015;
119 Xu *et al.*, 2016; Zhai *et al.*, 2016).

120 The LSLSZ crosses the middle part of the Tibetan Plateau in an east–west
121 direction and separates the Southern Qiangtang block from the Northern Qiangtang
122 block (Fig. 1a). This suture zone starts in Longmuco in the west and terminates at
123 Shuanghu in the east, but the nature of its eastward extension remains debated (Zhang
124 and Tang, 2009; Yang *et al.*, 2011; Hu *et al.*, 2014b; Chen *et al.*, 2016). Two different
125 views have been proposed: (i) the LSLSZ extends southeastwards and links with the
126 Amdo–Basu high-pressure metamorphic belt within the BNSZ (Pan *et al.*, 2012), and
127 (ii) the LSLSZ connects with the North Lancangjiang suture zone (NLSZ) and the
128 Changning Menglian suture zone (CMSZ; Li *et al.*, 2009; Yang *et al.*, 2011; Wang *et*
129 *al.*, 2013b, c; Hu *et al.*, 2014b). Given the similar ages of ophiolites in the LSLSZ, the
130 NLSZ, and the CMSZ (Fang *et al.*, 1998; Feng, 2002; Li *et al.*, 2008, 2010; Wang *et*
131 *al.*, 2008; Wang *et al.*, 2013b, c; Zhai *et al.*, 2013), and the similar basement rocks,
132 sedimentary formations, metamorphic rocks, and volcanic arc rocks in their

surrounding areas (Kapp *et al.*, 2003; Wang *et al.*, 2006; Li *et al.*, 2009; Qi *et al.*, 2009; Zhai *et al.*, 2011; Hu *et al.*, 2014b), we agree with the view that the LSLSZ connects to the NLSZ and the CMSZ. The LSLSZ–CMSZ represents the main oceanic basin of the Paleo-Tethys Ocean in the eastern section of the Alpine–Himalayan tectonic domain, separating Gondwana from Eurasia (Li, 1987, 2008; Fang *et al.*, 1998; Li *et al.*, 2009; Yang *et al.*, 2011; Wang *et al.*, 2013b, c; Hu *et al.*, 2014b; Zhai *et al.*, 2016).

The present study area is located in the Tianquan region, ~240 km north of Gerze County, in the western segment of the LSLSZ (Fig. 1a). The area contains a number of regional structural features and ductile shear zones that are oriented approximately E–W, and the rocks are strongly deformed (Fig. 1b). It also contains numerous fragments of ophiolites and oceanic islands, as well as sediments of the Dawashan, Wangguoshan, Riwanhaka, Zhanjin, and Tianquanshan formations (Fig. 1b). There also occur high-pressure metamorphic rocks such as blueschist and eclogite (Fig. 1b). The Middle to Late Ordovician Dawashan Formation formed in a rift environment and is dominated by graywackes, shales, and bimodal rocks. The early Carboniferous Wangguoshan Formation is dominated by basalts, basaltic andesites, andesites, dacites, rhyolites, and pyroclastics (Jiang *et al.*, 2015). The early Carboniferous Riwanhaka Formation is dominated by marine conglomerates, limestones, and sandstones, and most of the limestones contain abundant warm-water fossils such as corals and fusulinids of Cathaysian affinity (Fan, 1988; Li, 2008; Zhang *et al.*, 2009, 2013). The late Carboniferous to early Permian Zhanjin Formation

consists of bathyal to abyssal flysch deposits, dominated by greywacke, shale, and glacial marine diamictite that formed within the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean (Fan *et al.*, 2015a). The late Permian to Early Triassic Tianquanshan Formation is newly defined here, and is dominated by graywackes, shales, cherts, andesites, basalts, and conglomerates (Fan *et al.*, 2016).

Ophiolitic rocks are exposed in the Taoxinghu, Tuobeiling, and Heijishan areas of the study region (Fig. 1b). These rocks have been overprinted by Cenozoic tectonothermal events and are generally present as fault-controlled blocks or slices, suggesting that no single complete ophiolite sequence has been preserved. The Taoxinghu ophiolite contains cumulate gabbros, pillow basalts, gabbroic dike swarms, and plagiogranites (Wu *et al.*, 2009; Hu *et al.*, 2014a). SHRIMP U–Pb dating of zircons from the cumulate gabbros yields a mean age of 467 ± 4 Ma, indicating that the Taoxinghu ophiolite formed in the Middle Ordovician (Zhai *et al.*, 2010). The Tuobeiling ophiolite contains a relatively complete ocean-crust sequence of metaperidotites, cumulate gabbros, plagiogranites, pillow basalts, and siliceous sediments. The plagiogranites were intruded into the cumulate gabbro units, and this ophiolite has been interpreted as the eastern extension of the Taoxinghu ophiolite (Wu *et al.*, 2009). Plagiogranite samples from the Tuobeiling ophiolite yield Cambrian zircon U–Pb ages of 504.8 ± 4.2 and 491.6 ± 1.5 Ma (Hu *et al.*, 2014a). The Heijishan ophiolite, which was first investigated by Zhai *et al.* (2013), is dominated by gabbros, basalts, and plagiogranites, and U–Pb dating of zircons from the cumulate gabbros yields ages of 443 ± 3 Ma (Silurian), 357 ± 3 , and 355 ± 3 Ma (Carboniferous; Zhai *et*

177 *al.*, 2013; Zhang *et al.*, 2014c).

178 The ocean islands in the study area include TQ and DB (Fig. 1b), which were
179 newly found during this study. TQ is intermittently exposed in the Duquangou,
180 Tianquan, Tianquanshan, and Pingfengling areas in the northern part of the study area
181 (Fig. 1b). TQ extends more than 40 km E–W and is up to 0.5 km wide (Fig. 1b). DB
182 is located in the Dabure area as an isolated block of 700 × 400 m in size (Fig. 1b).

184 3. Field and petrographic observations

185 Geological transects across the Duquangou, Tianquan, Tianquanshan, and
186 Dabure areas provide representative geological cross-sections of the Tianquan and
187 Dabure ocean islands (Figs 1b, 2a–d).

188 The rocks of TQ have been strongly deformed and dismembered, occurring as
189 fragments with different rock assemblages (Figs 2a–c, 3a), although all fragments
190 retain an ocean-island-type double-layered structure comprising a basaltic basement
191 and an oceanic sedimentary cover sequence (Fig. 2a–c). Several geological cycles
192 have been identified among the different fragments of TQ, and all start with basaltic
193 units that are followed by conglomerate, limestone, or chert; some of these cycles are
194 conglomerate-free. Conformable contacts with the flysch deposits of the
195 Tianshanquan Formation were found in the Duquangou and Tianquanshan fragments
196 (Fig. 2a, c; Fan *et al.*, 2016). The dominant lithologies encountered within the TQ are
197 as follows.

198 (1) Basalt. Basalt is present in all the TQ fragments (Fig. 2a–c) and is

199 dominantly either massive basalt (Fig. 3b) or pillow basalt (Fig. 3c). The massive
200 basalts occur as individual units that are generally thicker than 3 m (Fig. 3b). The
201 pillows are generally 40 × 80 cm to 60 × 100 cm in size (Fig. 3c). The cores and rinds
202 of the pillows contain different components, and the rinds contain amygdales of
203 calcite. Both the pillow basalts and the massive basalts are porphyritic, comprising
204 pyroxene and plagioclase phenocrysts in an intergranular matrix (Fig. 3d). These
205 basalts have undergone lower greenschist facies metamorphism, resulting in some
206 pyroxene phenocrysts being altered to chlorite (Fig. 3d).

207 (2) Andesite. Andesite is found only in the Tianquanshan fragments of the TQ
208 (Fig. 2c). The TQ andesite is grayish in the field (Fig. 3e) and porphyritic, consisting
209 of 30%–50% feldspar phenocrysts (Fig. 3f).

210 (3) Limestone. Like the basalt, limestone is found in all the TQ fragments (Fig.
211 2a–c) and is present as thin or thick layers (Fig. 3g) that record deposition in an
212 isolated, clear-water sedimentary environment associated with a carbonate platform.

213 (4) Chert. Chert is widely distributed in the TQ, especially in the Tianquan and
214 Pingfengling fragments. The cherts are mostly thin-bedded (<10 cm bed thickness)
215 and gray to gray–black in color (Fig. 3h); they are laterally extensive, which is typical
216 of cherts that form in an ocean basin.

217 (5) Conglomerates. Conglomerates are found mainly in the Duquangou,
218 Tianquanshan, and Pingfengling fragments of the TQ, and they can be divided into
219 basaltic conglomerates that contain gravel and matrix that are both formed from
220 basaltic debris, and another type of conglomerate that contains minor amounts of

limestone in addition to basaltic debris (Fig. 3i). The significant deformation in this area means that the limestone fragments in the conglomerates have been stretched out (Fig. 3i). These two types of conglomerate provide evidence of a sedimentary environment with a restricted provenance, proximal deposition in a far-shore-type setting, and rapid accumulation of sediments in the ocean immediately surrounding the island.

Some of the areas along the transects contain thin layers (<20 cm) of chert interbedded with limestone layers of similar thickness, as well as limestones interbedded with basalts (Fig. 3g), providing direct petrological evidence that the TQ is a typical ocean island.

The DB has rock assemblages similar to those of the TQ, and is dominated by basalt (Fig. 3j, k), andesite (Fig. 3l, m), limestone (Fig. 3n), chert (Fig. 3n), and conglomerate (Fig. 3o). However, the DB basalts and andesites are generally aphanitic (Fig. 3k–m). The conformable contact between the basalts and andesites can be seen in the field and under the microscope (Fig. 3m). Thin layers (<20 cm) of chert interbedded with limestone layers of similar thickness are commonly observed (Fig. 3n).

In summary, both the TQ and DB retain ocean-island-type double-layered structures that comprises a volcanic basement (basalt and andesite) and an oceanic sedimentary cover sequence (conglomerate, limestone, and chert), typical of ocean islands.

4. Analytical methods

4.1 Whole-rock major and trace element analyses

Nine basalt, nine andesite, and ten chert samples from the TQ, along with seven basalt, nine andesite, and ten chert samples from the DB, were collected for whole-rock major and trace element analyses. Samples were trimmed to remove weathered surfaces, cleaned with deionized water, crushed, and powdered in an agate mill. Major element compositions were analyzed by inductively coupled plasma–optical emission spectroscopy (ICP–OES; Leeman Prodigy) with high-dispersion Echelle optics at the China University of Geosciences, Beijing, China. Loss on ignition (LOI) values were determined by heating 1 g of sample in a furnace at 1000 °C for several hours before being cooled in a desiccator and reweighed. Analytical uncertainties for the major elements are <1 wt.%. Trace element compositions were analyzed by inductively coupled plasma–mass spectrometry (ICP–MS; Agilent–7500a) at the China University of Geosciences. The analytical accuracy and precision during trace element analysis were determined from standards AGV–2 (US Geological Survey standard) and GSR–3 (Chinese national geological reference standard). The analytical accuracy is <5 wt.% for most elements. The values obtained for the standards are in good agreement with recommended values (Govindaraju, 1994). For details on the analytical procedures, see Hu *et al.* (2013) and Zhai *et al.* (2013).

4.2 *In situ* zircon U–Pb LA–ICP–MS analyses

Two andesite samples of the TQ, and one basalt sample and two andesite samples of the DB were selected for *in situ* zircon U–Pb analyses. The zircons were separated by conventional heavy liquid and magnetic techniques at the Special Laboratory of the Geological Team of Hebei Province, Langfang, China. The internal zircon structures were characterized using cathodoluminescence (CL) at the Institute of Physics, Peking University, Beijing, China, to select analytical points for analysis by laser ablation–inductively coupled plasma–mass spectroscopy (LA–ICP–MS). U, Th, and Pb isotopes, and trace element zircon analyses were conducted by LA–ICP–MS at the Geological Laboratory Centre, China University of Geosciences using a spot size of 36 μm for each sample, and using He as a carrier gas. Reference and internal zircon standards 91500 (Wiedenbeck *et al.*, 1995) and NIST610 (29Si) were used for instrument calibration, and common Pb was corrected following Anderson (2002). For further details of the analytical techniques, see Yuan *et al.* (2004). Reported uncertainties for the age analyses are given as 1σ values, with weighted mean ages at the 95% confidence level. Isotopic data were processed using the GLITTER (version 4.4) and Isoplot/Ex (version 3.0) programs (Ludwig *et al.*, 2003).

4.3 *In situ* zircon Hf isotope analyses

Ten zircon grains from the TQ andesite and twelve from the DB andesite, which were used for U–Pb dating, were selected for *in situ* zircon Hf isotope analyses. The data were collected using a NEPTUNE Plus MC–ICP–MS (Thermo Fisher Scientific,

Germany) with a Geolas 2005 excimer ArF laser ablation system (Lambda Physik, Göttingen, Germany) at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences, Wuhan, China. Analyses were conducted using single spot ablation with a spot size of 44 μm , 20 s of background signal acquisition, and 50 s of ablation signal acquisition. Detailed information on the operating conditions and analytical methods can be found in [Hu *et al.* \(2012\)](#). The $^{176}\text{Hf}/^{177}\text{Hf}$ value for the zircon standard (91500) was 0.282299 ± 31 ($2\sigma_n$, $n = 40$), which is similar to the standard values (0.282302 ± 8 and 0.282306 ± 8 , 2σ) determined by the solution method ([Goolaerts *et al.*, 2004](#); [Woodhead *et al.*, 2004](#)). Off-line selection, integration of analyte signals, and mass bias calibrations were performed using the ICP–MS DataCal program ([Liu *et al.*, 2010](#)).

5. Results

5.1. Whole-rock major and trace element geochemistry

Whole-rock major and trace element data for the TQ and DB basalts, andesites, and cherts are given in Appendix A1–A6. The TQ and DB basalts have undergone lower greenschist facies metamorphism and varying degrees of carbonate alteration, resulting in variable values of LOI and changes in the concentrations of mobile elements (e.g., Na, K, Ca, Cs, Rb, Ba, and Sr) compared with protolith values. However, concentrations and ratios of immobile elements (e.g., REE, Nb, Ta, Zr, Hf, Ti, and P) and transition metal elements (e.g., V, Ni, Cr, and Fe) have not been affected by these processes and can therefore be used to investigate the petrogenesis

and tectonic setting of the samples.

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5.1.1. TQ basalts and andesites

Nine samples of TQ basalt have variable contents of SiO₂ (43.8–52.0 wt.%), high contents of TiO₂ (3.00–4.82 wt.%), Fe₂O₃ (10.1–14.2 wt.%), MgO (6.03–8.99 wt.%), CaO (9.82–13.7 wt.%), and Na₂O (2.08–3.24 wt.%), low contents of K₂O (0.79–3.01 wt.%), and low Mg[#] values (48–62). Nine samples of TQ andesite have high contents of SiO₂ (54.5–58.3 wt.%), Al₂O₃ (15.3–17.3 wt.%), Na₂O (6.01–8.42 wt.%), Fe₂O₃ (7.44–8.66 wt.%), and TiO₂ (1.85–2.44 wt.%), low contents of K₂O (0.45–2.66 wt.%) and MgO (2.20–2.78 wt.%), and low Mg[#] values (37–40). The TQ basalts and andesites have much higher TiO₂ concentrations (3.00–4.82 wt.% and 1.85–2.44 wt.%, respectively) than typical island-arc volcanic rocks (0.58–0.85 wt.%) and mid-ocean ridge basalt (MORB; 1.0–1.5 wt.%), but similar concentrations to those found in OIB (2.20 wt.%; Hishashi, 1954; Irvine *et al.*, 1971; Winchester *et al.*, 1977; Pearce, 1983; Li, 1993).

The samples of TQ basalt and andesite have high total REE concentrations (209–330 ppm and 719–859 ppm, respectively). These samples have light rare earth element (LREE)-enriched chondrite-normalized REE patterns (La_N/Yb_N = 14.9–19.2 and 21.1–22.8, respectively; Fig. 4a), and are enriched in high field strength elements (HFSE; Nb, Ta, Zr, Hf, and Ti), yielding primitive-mantle-normalized trace element patterns similar to those of OIB (Fig. 4b).

Most of the TQ basalt samples are classified as alkali basalt on a SiO₂ vs.

Zr/TiO₂ immobile element classification diagram (Fig. 5; Winchester *et al.*, 1977), whereas the TQ andesites are classified as phonolite and trachyandesite on the same diagram. All the samples of TQ basalt plot in the intraplate alkali basalt field on the Ti/100 vs. Zr vs. (Y × 3) (Fig. 6a; Pearce and Cann, 1973) and (2 × Nb) vs. Zr/4 vs. Y (Fig. 6b; Meschede, 1986) diagrams, and in the ocean island basalt field on the Nb/Th vs. Nb and La/Nb vs. La diagrams (Fig. 6c–d; Li, 1993), indicating they formed from OIB-type magmas.

5.1.2. DB basalts and andesites

Seven samples of DB basalt have high contents of TiO₂ (2.29–5.20 wt.%), Fe₂O₃ (10.0–17.8 wt.%), Al₂O₃ (11.1–13.0 wt.%), and CaO (7.33–10.67 wt.%), low contents of MgO (3.63–4.18 wt.%), Na₂O (1.33–3.75 wt.%), and K₂O (0.79–3.01 wt.%), and low Mg[#] values (35–64). Nine DB andesite samples have high contents of SiO₂ (57.9–61.5 wt.%), TiO₂ (3.34–3.76 wt.%), Fe₂O₃ (11.6–12.4 wt.%), Al₂O₃ (8.04–8.84 wt.%), and CaO (5.39–6.33 wt.%), low contents of MgO (2.41–2.98 wt.%), Na₂O (1.94–2.76 wt.%), and K₂O (0.46–1.13 wt.%), and low Mg[#] values (32–37). Similar to the TQ basalts and andesites, the DB basalts and andesites have much higher TiO₂ concentrations (2.29–5.20 wt.% and 3.34–3.76 wt.%, respectively) than typical island-arc volcanic rocks (0.58–0.85 wt.%) and MORB (1.0–1.5 wt.%), but similar concentrations to those of OIB (2.20 wt.%; Hishashi, 1954; Irvine *et al.*, 1971; Winchester *et al.*, 1977; Pearce, 1983; Li, 1993). The samples of DB basalt and andesite have high total REE concentrations (154–367 ppm and 296–340 ppm,

respectively). These samples have LREE-enriched chondrite-normalized REE patterns ($\text{La}_\text{N}/\text{Yb}_\text{N} = 8.35\text{--}8.74$ and $8.43\text{--}9.02$, respectively; Fig. 4c) and are enriched in HFSE (Nb, Ta, Zr, Hf, and Ti), yielding primitive-mantle-normalized trace element patterns that are similar to those of OIB (Fig. 4d).

The samples of DB basalt are classified as alkali basalt or sub-alkaline basalt on a SiO_2 vs. Zr/TiO_2 immobile element classification diagram (Fig. 5; Winchester *et al.*, 1977), whereas the samples of DB andesite are classified as andesite on the same diagram. All the samples of DB basalt plot in the intraplate alkali basalt field on the $\text{Ti}/100$ vs. Zr vs. $(\text{Y} \times 3)$ (Fig. 6a; Pearce and Cann, 1973) and $(2 \times \text{Nb})$ vs. $\text{Zr}/4$ vs. Y diagrams (Fig. 6b; Meschede, 1986), and in the ocean island basalt field on the Nb/Th vs. Nb and La/Nb vs. La diagrams (Fig. 6c–d; Li, 1993), indicating that they formed from OIB-type magmas.

5.1.3. TQ and DB cherts

Ten samples of TQ chert show high contents of SiO_2 (85.0–96.0 wt.%), Al_2O_3 (1.12–5.13 wt.%), and Fe_2O_3 (0.32–3.73 wt.%), and low contents of TiO_2 (0.04–0.27 wt.%), MnO (0.00–0.12 wt.%), MgO (0.15–1.06 wt.%), CaO (0.06–0.21 wt.%), Na_2O (0.11–0.56 wt.%), and K_2O (0.13–0.66 wt.%). These samples have a wide range of total REE concentrations (13–99 ppm) and show North American shale composite (NASC)-normalized REE patterns that are flat, although with positive Ce anomalies ($\text{Ce}/\text{Ce}^* = 1.33\text{--}1.71$; Fig. 7).

Ten samples of DB chert have concentrations of elements that are similar to

those of the TQ cherts; i.e., they have high concentrations of SiO₂ (84.2–95.4 wt.%) and Al₂O₃ (0.82–2.06 wt.%), but are relatively depleted in Fe₂O₃ (0.12–1.32 wt.%), TiO₂ (0.04–0.11 wt.%), MnO (0.00–0.03 wt.%), MgO (0.12–0.22 wt.%), Na₂O (0.01–0.03 wt.%), and K₂O (0.24–0.61 wt.%). These samples have a wide range of total REE concentrations (9.9–32.2 ppm) and show LREE-enriched NASC-normalized REE patterns (La_N/Yb_N = 0.94–3.25; Fig. 7) and slightly negative Eu anomalies (Ce* = 0.78–1.02; Fig. 7).

5.2. Zircon U–Pb ages

5.2.1. TQ andesites

Zircons are abundant within the TQ andesite samples. LA–ICP–MS U–Pb data for these zircons are given in Appendix A7. The zircons are generally hypidiomorphic short prisms (100–200 µm long) with aspect ratios of 1:1 to 2:1. They are typically transparent, colorless to gray, and exhibit magmatic oscillatory zoning in CL images (Fig. 8a). Their Th/U ratios are 0.95–2.41, indicating a magmatic origin (Hoskin and Black, 2000), and the zircons from the andesite yield weighted mean ²⁰⁶Pb/²³⁸U ages of 250.7 ± 1.5 Ma (MSWD = 0.005) and 250.6 ± 1.5 Ma (MSWD = 0.006), respectively (Fig. 8b–c). Considering the ages of 246, 247, and 254 Ma reported previously for the TQ andesite (Fan *et al.*, 2016), we conclude that TQ formed during the late Permian–Early Triassic.

5.2.2. DB basalts and andesites

Zircons are rare within the DB basalt samples. U–Pb data for the zircons are listed in Appendix A8. The zircons are generally euhedral to subhedral (<100 µm long) with aspect ratios of 2:1 to 4:1. They are typically transparent, colorless to gray, and show wide magmatic oscillatory zoning in CL images (Fig. 8a). The zircons have Th/U ratios from 0.17 to 0.81, indicating an origin from mafic magmatism (Hoskin and Black, 2000). Five spot analyses on zircons from the DB basalt samples yielded ages of 930 to 233 Ma (Fig. 8d).

U–Pb data for the abundant zircons in the DB andesite samples are reported in Appendix A8. The zircons are generally transparent to light brown in color and show narrow magmatic oscillatory zoning in CL images (Fig. 8a). These zircons have Th/U ratios ranging from 0.37 to 1.04 with an average of 0.56, which also indicates a magmatic origin, and they yield weighted mean $^{206}\text{Pb}/^{238}\text{U}$ ages of 241.1 ± 1.7 Ma (MSWD = 0.63) and 245.9 ± 1.2 Ma (MSWD = 0.24) (Fig. 8e–f). Taking into account the youngest age of 244 Ma from the DB basalt, we conclude that DB formed during the Middle Triassic.

5.3. Zircon Lu–Hf isotope data

The analytical spots for U–Pb dating of zircons from the TQ and DB andesites were also used for *in situ* Hf isotope analysis (Appendix A9). The initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios for primary zircons of the TQ andesites with ages of ~251 Ma vary from

0.282741 to 0.282794, yielding $\varepsilon_{\text{Hf}}(t)$ values and T_{DM}^{C} model ages (T_{DM1}) of +4.25 to +6.22 and 640 to 724 Ma, respectively. The initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios for primary zircons of the DB andesites with ages of ~242 Ma vary from 0.282612 to 0.282682, yielding $\varepsilon_{\text{Hf}}(t)$ values and T_{DM}^{C} model ages (T_{DM1}) of –0.59 to +1.97 (mostly > 0) and 809 to 923 Ma, respectively. These values indicate that the TQ and DB andesites were all derived from the partial melting of juvenile crust or by fractional crystallization of the mantle-derived mafic parental magma.

6. Discussion

6.1. Petrogenesis of the basalts and andesites

6.1.1. Crustal contamination

Before discussing the magma source of the analyzed basalts and andesite, it is necessary to consider whether they record any crustal contamination. Th and Ta are sensitive indicators of crustal contamination, as it results in a marked increase in Th/Ta ratios (Condie, 1993). All the TQ basalts and andesites, along with the DB basalts and andesites, have relatively low Th/Ta ratios (1.69–2.13, 2.34–2.49, 1.64–1.77, and 1.06–1.10, respectively), similar to those in volcanic rocks derived from primitive mantle (Th/Ta = 2.3; Condie, 1993) and much lower than those of the upper crust (Th/Ta >10), indicating that none of these andesites or basalts record any crustal contamination. Zr/Nb ratios remain nearly constant during fractional crystallization, but change significantly during magma mixing or crustal contamination (Weaver *et al.*, 1996). Although the TQ basalts and andesites, along

with the DB basalts and andesites, have a wide range of SiO₂ contents (43.8–52.0 wt.%, 54.5–58.3 wt.%, 43.2–50.6 wt.%, and 59.3–61.5 wt.%, respectively), there is only a small range in Zr/Nb ratios (6.44–7.34, 5.99–6.20, 5.95–6.90, and 5.36–5.52, respectively), and this is consistent with none of these rocks being affected by crustal contamination. In addition, the TQ basalts and andesites, along with the DB basalts and andesites, have high contents of Nb (40.3–66.3 ppm, 146–227 ppm, 60.2–76.0 ppm, and 73.6–82.4 ppm, respectively), Ta (2.55–4.04 ppm, 8.12–13.9 ppm, 3.52–4.27 ppm, and 4.48–5.03 ppm, respectively), and TiO₂ (3.00–4.82 wt.%, 1.85–2.44 wt.%, 2.29–5.20 wt.%, and 3.34–3.76 wt.%, respectively). Furthermore, these rocks do not show significant Pb enrichment in primitive-mantle-normalized trace element patterns (Fig. 4b, d), further excluding the possibility that they underwent significant crustal contamination, because continental crust (CC) has characteristic depletion in Nb, Ta, and Ti and significant enrichment in Pb (Rudnick and Gao, 2003; Niu *et al.*, 2009). If the TQ and DB basalts and andesites had undergone crustal contamination, a CC signature would have been apparent in the element data, but this is not observed.

6.1.2. Magma source

Various studies have shown that basaltic magmas are commonly derived from the partial melting of mantle peridotite and that their REE patterns are controlled mainly by the contents of garnet and spinel in their magma source rather than by the contents of olivine, clinopyroxene, or orthopyroxene, or by pressure and temperature

(McKenzie and O’Nions, 1991; Beattie, 1994; Horn *et al.*, 1994; Schwandt *et al.*, 1998). In general, basalts derived from spinel lherzolite have flat chondrite-normalized REE patterns with inconspicuous fractionation between the LREE and the heavy REE (HREE). However, basalts derived from garnet lherzolite display obvious fractionation between the LREE and HREE, and high $\text{La}_\text{N}/\text{Yb}_\text{N}$ and $\text{Ce}_\text{N}/\text{Yb}_\text{N}$ ratios (Hart and Dunn, 1993; Hauri *et al.*, 1994).

The TQ and DB basalts have LREE-enriched chondrite-normalized REE patterns ($\text{La}_\text{N}/\text{Yb}_\text{N} = 14.9\text{--}19.2$ and $8.35\text{--}8.74$, respectively; Fig. 4a, c), similar to those of basalts derived from garnet peridotite. The possibility that the magma source of the TQ and DB basalts was garnet peridotite in the mantle is further supported by (i) the high Dy/Yb ($3.18\text{--}3.31$ and $2.43\text{--}2.70$, respectively), Sm/Yb ($5.18\text{--}5.63$ and $2.62\text{--}3.23$, respectively), and La/Sm ($4.00\text{--}4.88$ and $3.49\text{--}3.80$, respectively) ratios of these basalts, and (ii) the fact that the TQ basalts plot in an area on the Sm/Yb vs. La/Sm diagram (Fig. 9a) that corresponds to 1%–5% partial melting of garnet peridotite and the DB basalts plot in an area that corresponds to ~10% partial melting of garnet peridotite (Sun and McDonough, 1989; McKenzie and O’Nions, 1991; Aldanmaz *et al.*, 2000; Zhao and Zhou, 2007).

The small and homogeneously positive $\varepsilon_{\text{Hf}}(\text{t})$ values of the TQ andesites (+4.25 to +6.22) and DB andesites (−0.59 to +1.97, mostly >0) indicate that these andesites were derived from the partial melting of juvenile crust or by the fractional crystallization of a mantle-derived mafic parent magma. Both the TQ andesites and the DB andesites contain high Nb ($146\text{--}227$ and $73.6\text{--}82.4$ ppm, respectively), Ta

(8.12–13.9 and 4.48–5.03 ppm, respectively), and TiO_2 (1.85–2.44 and 3.34–3.76 wt.%, respectively), which are difficult to explain in terms of the partial melting of juvenile crust. We infer, therefore, that both the TQ and DB andesites were formed by fractional crystallization of the mantle-derived mafic parental magma. The low contents of MgO contents (2.20–2.78 and 2.41–2.98 wt.%, respectively) and the low $\text{Mg}^\#$ values (37–40 and 32–37, respectively) of these andesites indicate that they underwent substantial fractional crystallization, which in turn supports the inference that they were formed by fractional crystallization of the mantle-derived mafic parental magma.

Both the TQ andesites and the DB andesites have LREE-enriched chondrite-normalized REE patterns ($\text{La}_\text{N}/\text{Yb}_\text{N} = 21.1\text{--}22.8$ and $8.43\text{--}9.02$, respectively; [Fig. 4a, c](#)) and enrichment in Nb, Ta, and Ti. We infer, therefore, that their mafic parental magma was derived from garnet peridotite in the mantle, similar to the magma source of the TQ and DB basalts. In addition, the TQ andesites and the DB andesites have high Dy/Yb (2.78–3.10 and 2.67–2.72, respectively), Sm/Yb (5.20–5.72 and 3.40–3.48, respectively), and La/Sm (5.18–6.10 and 3.44–3.65, respectively) ratios, and they plot similarly to the basalts in the field of garnet peridotite on the Sm/Yb vs. La/Sm diagram ([Fig. 9a](#)), further indicating that the mafic parental magma source of these andesites was garnet peridotite in the mantle ([Sun and McDonough, 1989; McKenzie and O’Nions, 1991; Aldanmaz *et al.*, 2000; Zhao and Zhou, 2007](#)).

6.1.3. Fractional crystallization

Most of the Cr and Ni concentrations, and $Mg^\#$ values for the TQ basalts (Cr = 107–705 ppm, Ni = 78.8–387 ppm, $Mg^\#$ = 48–62) and the DB basalts (Cr = 32.3–389 ppm, Ni = 41.1–205 ppm, $Mg^\#$ = 35–64) are lower than the values for primary-mantle-derived magmas (Cr = 300–500 ppm, Ni = 300–400 ppm, $Mg^\#$ = 68–76; [Frey *et al.*, 1978](#); [Hess *et al.*, 1992](#)). This indicates that the TQ and DB basalts underwent fractional crystallization of olivine, chromite, and pyroxene ([Wilson *et al.*, 1989](#); [Jung *et al.*, 1998](#)).

The depletion of Ti in the primitive-mantle-normalized trace element patterns of the TQ andesites ([Fig. 4b](#)) suggests the separation of Ti-bearing phases (e.g., ilmenite and/or rutile). In the CaO/Al₂O₃–MgO diagram ([Fig. 9b](#)) the TQ andesites show a significant fractional crystallization of olivine as well as a weak fractional crystallization of pyroxene, whereas the DB andesites show a significant fractional crystallization of olivine and no fractional crystallization of pyroxene. The slightly negative Eu anomalies in the chondrite-normalized REE patterns of the DB andesites ($Eu/Eu^* = 0.88–0.97$; [Fig. 4c](#)) indicate a weak fractional crystallization of plagioclase, whereas the absence of negative Eu anomalies in the chondrite-normalized REE patterns of the TQ andesites ([Fig. 4a](#)) suggests a lack of fractionation of plagioclase.

In summary, the TQ and DB basalts were derived from the partial melting of garnet peridotite in the mantle. The ascending magmas underwent varying degrees of fractional crystallization but were not contaminated by the crust. The TQ and DB andesites formed by fractional crystallization of the mafic parental magma derived

from garnet peridotite mantle. Similar to the TQ and DB basalts, the ascending
magmas of the TQ and DB andesites were not contaminated by the crust.

6.2. Tectonic environment

The TQ and DB both retain an ocean-island-type double-layered structure
comprising a volcanic basement (basalt and andesite) and an oceanic sedimentary
cover sequence (conglomerate, limestone, and chert), indicating that they are both
typical ocean islands. Geochemical and Hf analyses of the TQ and DB basalts and
andesites show that they are all OIB-type volcanic rocks. The fact that their ascending
magmas did not undergo crustal contamination further demonstrates that the TQ and
DB are typical ocean islands that formed in the basin of the
Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean.

Many cherts are chemically precipitated sedimentary rocks, essentially
monomineralic, and composed chiefly of microcrystalline and/or chalcedonic quartz,
with subordinate coarser quartz (Folk, 1980). Cherts may contain up to 95% or more
silica (Hesse, 1988), and they are chemically inert and largely resistant to weathering
and alteration, making them suitable for geochemical studies. Their main elemental
compositions include simple, rare, and trace elements that reflect constituents such as
volcanic detritus, terrigenous clasts, and hydrothermal precipitates (Kato and
Nakamura, 2003; Sugahara *et al.*, 2010; Eker *et al.*, 2012). REEs are generally
immobile in cherts, and REE patterns provide evidence of the tectonic environment
during deposition as well as the site of deposition (Murray, 1994). Cherts are common

1 547 in the TQ and DB, and an analysis of their chemistry could further enhance our

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3 548 understanding of the tectonic environment of the TQ and DB.

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6 549 [Murray *et al.* \(1990\)](#) studied the chert and shale sequences of the Franciscan

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9 550 assemblage, the Claremont Formation, and the Monterey Formation of coastal

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11 551 California, and recognized three depositional regimes: spreading ridge proximal,

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14 552 ocean-basin floor, and continental margin. Back-arc basins, marginal seas,

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17 553 epicontinental seas, and open continental shelves are all continental margin

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20 554 environments ([Murray, 1994](#)). In general, the chemical composition of a bedded chert

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28 557 and by hydrogenous (authigenic) components. In addition, chert that is deposited near

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31 558 a spreading ridge may be influenced by metalliferous activity. Modification of the

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34 559 chemical composition and isotopic systems of chert may also occur during diagenesis.

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36 560 A high SiO₂ content in a bedded radiolarian chert generally indicates a greater

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39 561 contribution of biogenic silica, whereas a high MnO content indicates a greater

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42 562 contribution of hydrogenous components ([Sugisaki *et al.*, 1982](#); [Kunimaru *et al.*,](#)

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45 563 [1998](#)). Given that Al₂O₃ is derived from terrigenous clastic rocks, a high Al₂O₃

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48 564 content in bedded chert indicates a greater contribution of terrigenous material.

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50 565 [Murray \(1994\)](#) proposed a shale-normalized (La/Ce) ratio versus Al₂O₃/(Al₂O₃ +

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53 566 Fe₂O₃) ratio diagram to determine the depositional environment. The advantage of

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56 567 this approach is that it is independent of diagenetic modification. On the

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59 568 (La/Ce)–Al₂O₃/(Al₂O₃ + Fe₂O₃) diagram, cherts that form in ridge-proximal, pelagic,

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and continental margin environments have $(\text{La/Ce})_{\text{SN}}$ (shale-normalized ratios) of 3–4, 1–2.5, and 0.5–1.5, and $\text{Al}_2\text{O}_3/(\text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3)$ ratios of 0.05–0.4, 0.4–0.7, and 0.55–0.9, respectively.

Among the REE features of cherts, a negative Ce anomaly usually indicates a pelagic depositional setting. Strongly negative Ce anomalies are commonly observed for deep-ocean water and not for clastic rocks. Murray *et al.* (1990) and Murray (1994) noted that the Ce anomaly reflects the depositional environment of chert: areas proximal to a spreading ridge yield values of $\text{Ce/Ce}^* = \sim 0.3$ and $(\text{La/Ce})_{\text{SN}} = 3\text{--}4$, whereas ocean-basin floor or pelagic areas yield values of $\text{Ce/Ce}^* = \text{ca. } 0.5$ and $(\text{La/Ce})_{\text{SN}} = 1\text{--}2.5$, and continental margins yield values of $\text{Ce/Ce}^* = \text{ca. } 0.9\text{--}1.3$ and $(\text{La/Ce})_{\text{SN}} = 0.5\text{--}1.4$.

In the cases of the TQ and DB cherts, they are mostly thin-bedded (<20 cm bed thickness) and laterally extensive (Fig. 3h, n), which is typical of cherts formed in an ocean basin. The TQ and DB cherts have high contents of SiO_2 (85.0–96.0 wt.% and 84.2–95.4 wt.%, respectively) and low contents of MnO (0.00–0.12 wt.% and 0.00–0.03 wt.%, respectively), indicating a greater contribution of biogenic silica and a lesser contribution of hydrogenous components. This inference is further supported by the fact that most of the TQ and DB chert samples (except for one DB chert) plot in the non-hydrothermal origin field on the Fe–Al–Mn diagram (Fig. 10a).

The TQ and DB cherts have high contents of Al_2O_3 (1.12–5.13 wt.% and 0.82–2.06 wt.%, respectively), indicating a large contribution of terrigenous material during their deposition. Positive Ce anomalies (Fig. 7; $\text{Ce/Ce}^* = 1.33\text{--}1.71$) in the TQ

cherts and slightly negative Ce anomalies (Fig. 7; $Ce^* = 0.78\text{--}1.02$) in the DB cherts are similar to the Ce anomalies of cherts that form in a continental margin, and they exclude the possibility that the cherts were deposited proximal to a spreading ridge or in a pelagic setting. The high $Al_2O_3/(Al_2O_3 + Fe_2O_3)$ ratios (0.58–0.78 and 0.38–0.93, respectively) and low values of $(La/Ce)_{SN}$ (0.58–0.82 and 1.04–1.49, respectively) of the TQ and DB cherts indicate that they were deposited close to a continental margin. This inference is supported by the fact that most of the TQ and DB chert samples plot in the continental margin field on the $100 \times (Fe_2O_3/SiO_2) - 100 \times (Al_2O_3/SiO_2)$, $Fe_2O_3/(100 - SiO_2) - Al_2O_3/(100 - SiO_2)$, and $Fe_2O_3/TiO_2 - Al_2O_3/(Al_2O_3 + Fe_2O_3)$ diagrams (Fig. 10b–d).

In summary, we conclude that both the TQ and DB are typical ocean islands that formed in the basin of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean, although close to a continental margin.

6.3. Tectonic model: implications for the late-stage evolution of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean

As discussed above, based on the analysis of ophiolites and volcanic rocks in the LSLSZ and surrounding areas, previous studies have proposed that the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean was open at least during the Cambrian and remained open until the early Permian (Li., 1987, 2008; Li *et al.*, 2007a, 2008; Wang *et al.*, 2008; Zhang *et al.*, 2009, 2013; Yang *et al.*, 2011; Zhai *et al.*, 2011, 2013, 2016; Wu *et al.*, 2013; Hu *et al.*, 2014a; Jiang *et al.*, 2015; Fan *et al.*, 2016;

1 613 Wang *et al.*, 2016; Wu *et al.*, 2016). However, because of the general absence from
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3 614 the LSLSZ of oceanic material younger than early Permian, it is debated whether the
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6 615 Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean existed after this time. The
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9 616 following questions remain: How did this ocean develop after the early Permian, if it
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12 617 existed? And when did it close?

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14 618 Based on an analysis of the sedimentary strata, unconformities, and
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17 619 high-pressure metamorphic rocks within and surrounding the LSLSZ, it has been
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20 620 suggested that the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean
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23 621 continued to develop from the early Permian to the Late Triassic and that it finally
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26 622 closed during the Late Triassic (Li *et al.*, 2007b; Li, 2008; Wu, 2013). However, other
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29 623 studies have reported different views on the late-stage evolution of this ocean. Based
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32 624 on an analysis of the early to late Permian volcanic rocks in the Nangqiao and
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35 625 Zuogong areas of the Northern Qiangtang block, some scholars have suggested that
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41 627 subducted from the early Permian to the late Permian, and that the ocean finally
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44 628 closed during the Early Triassic or early Middle Triassic (Yang *et al.*, 2011; Peng *et*
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46 629 *al.*, 2015), not as late as the Late Triassic. Others have proposed that until the Middle
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49 630 Triassic the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys oceanic lithosphere
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52 631 was still being subducted, with final closure in the Late Triassic (Zhai *et al.*, 2011,
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54 632 2013; Wu *et al.*, 2016). These studies reported that slab break-off and exhumation of
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57 633 high-pressure metamorphic rocks occurred during the Late Triassic (Zhai *et al.*, 2011,
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59 634 2013; Wu *et al.*, 2016). Zhang *et al.* (2011) proposed that slab break-off of the
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Longmuco–Shuanghu–Lancangjiang Paleo-Tethys oceanic lithosphere occurred before the Middle Triassic and that closure of the ocean may therefore have occurred in the Early Triassic or late Permian. Based on an analysis of flysch in the Tianquanshan Formation, [Fan *et al.* \(2016\)](#) inferred that the ocean was still a relatively mature and complete ocean basin during the late Permian–Early Triassic. [Fu *et al.* \(2010\)](#) proposed that the LSLSZ was in a continental rift setting during the Late Triassic. It is clear that there are many different views regarding the longevity of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean.

Ocean islands are widespread in both ancient and modern ocean basins, and they are represented by the Hawaiian and Azores archipelagos in modern oceans ([Zielinski and Frey, 1970](#); [Kar *et al.*, 1998](#); [Regelous *et al.*, 2003](#); [Humphreys and Niu, 2009](#); [Timm *et al.*, 2011](#)). Ocean islands, therefore, play an important role in identifying the existence of ancient ocean basins. The presence of the late Permian–Early Triassic TQ and the Middle Triassic DB in the western segment of the LSLSZ indicates that the western segment of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean remained at least partially open until the Middle Triassic. Taking into account the late Permian–Early Triassic flysch deposits in the Tianquanshan Formation ([Fan *et al.*, 2016](#)) and the numerous late Permian–Middle Triassic arc-type volcanic rocks in the Northern Qiangtang block ([Zhai *et al.*, 2011](#); [Hu, 2014](#); [Wu *et al.*, 2016](#)), we infer that the western segment of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean continued to develop normally during the Middle Triassic, when its oceanic lithosphere was being subducted ([Fig. 11a](#)).

1 657 However, the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys oceanic
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3 658 lithosphere had been undergoing subduction and a reduction in size since the Late
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6 659 Devonian (Jiang *et al.*, 2015; Wang *et al.*, 2016), and by the Middle Triassic it had
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9 660 been subjected to these processes for at least 150 Myr. Taking into account the late
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11 661 Permian–Early Triassic opening of the Bangong–Nujiang Meso-Tethys Ocean to the
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14 662 south of the LSLSZ, and its rapid expansion in the Early–Middle Triassic (Fan *et al.*,
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17 663 2014, 2015b; Liu *et al.*, 2016), we infer that the Longmuco–Shuanghu–Lancangjiang
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20 664 Paleo-Tethys ocean basin was not extremely large (Fig. 11a). This inference is
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23 665 supported by the geochemical analyses of the TQ and DB cherts, which show that
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26 666 they contain terrigenous material, indicating in turn that the TQ and DB formed in an
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29 667 ocean basin relatively close to a continental margin.

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31 668 The change in deposition from marine to continental strata in the Southern
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34 669 Qiangtang and Northern Qiangtang blocks occurred during the early–middle Late
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37 670 Triassic (Wang *et al.*, 2013a). High-pressure metamorphic rocks of the LSLSZ were
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40 671 exhumed in the middle Late Triassic (~220 Ma; Li *et al.*, 2006; Dong *et al.*, 2009;
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42 672 Zhai *et al.*, 2013), and an unconformity developed between the continental strata of
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45 673 the middle–late Late Triassic Wanghuling Formation and the ophiolites of the LSLSZ
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48 674 (Li *et al.*, 2007b). We infer, therefore, that the final closure of the
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51 675 Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean occurred during the early
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54 676 Late Triassic and that subsequent orogenic uplift occurred during the middle–late Late
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56 677 Triassic (Fig. 11b).

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58 678 The LSLSZ extends for 2800 km in China. Therefore, it might be expected that
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the timing of closure of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean would vary within the various segments of this zone. Given that Triassic flysch deposits (Fan *et al.*, 2016) and ocean islands (this study) existed in the western part of the LSLSZ, that there is no evidence for the presence of a Triassic ocean basin in the eastern part of the LSLSZ, and that the collision-type volcanic rocks within the LSLSZ and surrounding areas were formed earlier in the east (251–236 Ma) than in the west (225–205 Ma; Hu *et al.*, 2014; Wang *et al.*, 2014), we suggest that the closure of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean was diachronous from east to west (Hu, 2014). The different closure times in different areas can be summarized as Early Triassic in the eastern segment of the LSLSZ and Late Triassic in the western segment.

In summary, we conclude that the development of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean continued after the early Permian and that ocean closure was diachronous from east to west. The eastern segment of the ocean may have closed during the Early Triassic; however, it remained at least partially open in its western segment until the Middle Triassic, although the ocean was relatively small at this time. The Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean finally closed during the Late Triassic.

7. Conclusions

(1) Tianquan and Dabure are typical ocean islands that formed within the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys ocean basin.

(2) The TQ and DB basalts were derived from the partial melting of garnet peridotite in the mantle, and the ascending magmas underwent varying degrees of fractional crystallization but were not contaminated by the crust.

(3) The TQ and DB andesites were formed by fractional crystallization of the mafic parent magma, which was derived from garnet peridotite mantle. Similar to the TQ and DB basalts, the ascending magmas of the TQ and DB andesites were not contaminated by the crust.

(4) The TQ and DB cherts contain terrigenous material, indicating the proximity of the site of deposition to a continental margin.

(5) The development of the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean continued after the early Permian, and it remained at least partially open in its western segment until the Middle Triassic, although the ocean was relatively small at this time.

(6) Closure of the ocean was diachronous from east to west, and the different closure times in different areas can be summarized as Early Triassic in the eastern segment of the LSLSZ and Late Triassic in the western segment.

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Figure captions

Fig. 1. (a) Tectonic framework of the Tibetan Plateau. JSSZ, Jinshajiang suture zone; LSLSZ, Longmuco–Shuanghu–Lancangjiang suture zone; BNSZ, Bangong–Nujiang suture zone; IYZSZ, Indus–Yarlung Zangbo suture zone. (b) Geological map of the study area.

Fig. 2. (a) Field photograph of the Tianquan ocean island (TQ). (b) Field photograph of TQ massive basalt. (c) Field photograph of TQ pillow basalt. (d) Photomicrograph showing porphyritic texture of the TQ basalt. (e) Field photograph of TQ andesite. (f) Photomicrograph showing porphyritic texture of the TQ andesite. (g) Field photograph showing interbedded TQ basalts and limestones. (h) Field photograph of TQ chert. (i) Field photograph of TQ conglomerate. (j) Field photograph of the Dabure ocean island (DB). (k) Photomicrograph showing intergranular texture and carbonate alteration of the DB basalt. (l) Photomicrograph showing aphanitic texture of the DB andesite. (m) Photomicrograph showing the contact between DB basalts and DB andesites. (n) Field photograph of DB limestone and chert. (o) Field photograph of DB conglomerate. Px, pyroxene; Pl, plagioclase. Panels b, e, and f are taken from [Fan *et al.* \(2016\)](#).

Fig. 3. (a)–(c) Geological cross-sections along transects across the Tianquan ocean island in the Duquangou, Tianquan, and Tianquanshan areas of Gerze County, Tibet. (d) Geological cross-section along a transect across the Dabure ocean island in the Dabure area of Gerze County, Tibet.

Fig. 4. (a) Chondrite-normalized REE variation diagram for samples of TQ basalt and andesite. (b) Primitive-mantle-normalized trace element variation diagram for samples of TQ basalt and andesite. (c) Chondrite-normalized REE variation diagram for samples of DB basalt and andesite. (d) Primitive-mantle-normalized trace element variation diagram for samples of DB basalt and andesite.

Fig. 5. Immobile-element-based classification diagram showing the nature of the basalts and andesites analyzed during this study.

Fig. 6. (a) $Ti/100$ vs. Zr vs. $Y \times 3$ diagram for samples analyzed during this study. A, island-arc tholeiite; B, mid-ocean ridge basalt (MORB) + island-arc tholeiite and calc-alkali basalt; C, calc-alkaline basalt; D, intraplate basalt. (b) $2 \times Nb$ vs. $Zr/4$ vs. Y diagram for samples analyzed during this study. AI, intraplate alkali basalt; AII, intraplate alkali basalt and tholeiite; B, E-type MORB; C, intraplate tholeiite and volcanic arc basalt; D, N-type MORB and volcanic arc basalt. (c) Nb/Th vs. Nb diagram for samples analyzed during this study. OIB, ocean island basalt; MORB, mid-ocean ridge basalt; IAB, island-arc basalt. (d) La/Nb vs. La diagram for samples

analyzed during this study.

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Fig. 7. NASC-normalized REE variation diagram for the TQ and DB chert samples.

NASC, North American Shale Composite.

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Fig. 8. (a) CL images of zircons analyzed during this study. (b–c) Concordia diagrams

showing the results of LA–ICP–MS analyses of zircons from the TQ andesites. (d)

Concordia diagram showing the results of LA–ICP–MS analyses of zircons from the

DB basalt. (e–f) Concordia diagrams showing the results of LA–ICP–MS analyses of

zircons from the DB andesites.

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Fig. 9. (a) Sm/Yb vs. La/Sm diagram for the TQ and DB volcanic rocks (modified

after [Zhao and Zhou, 2007](#)). The mantle array (heavy line) is defined by depleted

MORB mantle (DM; [McKenzie and O’Nions, 1991](#)) and primitive mantle (PM; [Sun](#)

[and McDonough, 1989](#)). Melting curves for spinel lherzolite and garnet peridotite

with both DM and PM compositions are after [Aldanmaz et al. \(2000\)](#). Numbers along

lines represent the degree of partial melting. (b) CaO/Al₂O₃–MgO diagram for the TQ

and DB andesites.

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Fig. 10. Geochemical data of TQ and DB cherts plotted on the (a) Al vs. Fe vs. Mn, (b)

100*(Fe₂O₃/SiO₂) vs. 100*(Al₂O₃/SiO₂), (c) Fe₂O₃/(100 – SiO₂) vs. Al₂O₃/(100 –

SiO₂), and (d) Fe₂O₃/TiO₂ vs. Al₂O₃/(Al₂O₃ + Fe₂O₃) diagrams. A, non-hydrothermal

1 1153 chert; B, hydrothermal chert.
2
3 1154
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6 1155 Fig. 11. Schematic model showing the later stages of evolution and final closure of
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9 1156 the Longmuco–Shuanghu–Lancangjiang Paleo-Tethys Ocean. LSLSZ,
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11 1157 Longmuco–Shuanghu–Lancangjiang suture zone.
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Figure 1

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