Provenance of Late Triassic sediments in central Lhasa terrane, Tibet and its implication

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1. Introduction

The geology of Tibetan Plateau has received considerable geologic attention with much of this work being focused on the processes attendant to the ongoing collision of India and Asia (e.g. Molnar and Tapponnier, 1975; England and Searle, 1986; Tapponnier et al., 2001; Ding et al., 2005; Aitchison et al., 2007). In contrast, the pre-collisional and early collisional history of southern Tibet is relatively poorly understood. However, some recent important discoveries have begun to shed new light on this hitherto enigmatic phase. Firstly, the discovery of the Paleozoic eclogite, igneous rocks and metamorphic events at Sumdo to the northeast of Lhasa city suggest the existence of a previously unrecognized Permo-Triassic orogenic zone within the central Lhasa terrane, termed the Sumdo–Cuoqen belt (Chen et al., 2009; Yang et al., 2009; Zhu et al., 2009; C.Y. Dong et al., 2011; Li et al., 2011; X. Dong et al., 2011). Secondly, our recent investigations have suggested that the boundary between India and Asia may extend to the south of the Indus–Yarlung suture zone (IVSZ, Liu et al., 2010, 2012). In addition, new paleomagnetic data have been used to suggest a more complex, multi-phase collisional history than have been traditionally envisaged (Aitchison et al., 2007; Cai et al., 2012; van Hinsbergen et al., 2012).

Finally, Zhu et al. (2011, 2013) have proposed that the Lhasa terrane derives from a continental fragment rifted from the Australian Gondwanan margin.

These discoveries and interpretations call into question some traditional view concerning the pre-and early collisional history of southern Tibet, and raise a raft of new questions. The Paleozoic–Mesozoic sedimentary sequences in Lhasa terrane provide a key to clarify these questions. In this study, we focus on the Late Triassic sedimentary sequences in the central Lhasa terrane. We use petrographic and detrital zircon isotopic data to constrain the provenance of the sediments. We also compare our data with that from the Late Triassic sequences on the southern flank of the Sumdo–Cuoqen belt.

2. Geological setting of Lhasa terrane

Tibet is underlain by four E–W trending geologic terranes. From north to south, these are the Songpan–Ganzi, Qiangtang, Lhasa and Himalaya terranes. Each is separated from the other by inferred sutures that from north to south are (1) Jinshajiang suture (JSS), (2) Bangong–Nujiang suture (BNS), and (3) Indus–Yarlung suture (IVS). These represent the multi-phase Tethyan Ocean relics, respectively (Yin and Harrison, 2000; Yin, 2006; Fig. 1).

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The Lhasa terrane extends more than 2000 km east to west and 200–300 km north–south. In the west it is truncated by the Karakoram strike-slip fault, and in the east it bends southward around the eastern Himalaya syntaxis (Dewey et al., 1988; Yin, 2006). It can be subdivided into the southern, central and northern Lhasa terranes, which are separated by the Shiquan River–Nam Tso mélangé zone (SNMZ) to the north and Luobadui–Milashan Fault (LMF) to the south, respectively (Zhu et al., 2013, Fig. 1). The southern Lhasa terrane is characterized primarily by Mesozoic–Cenozoic intrusive and volcanic rocks of the Gangdese arc. This includes the Early Jurassic Yeba and Tertiary Linzing volcanic–sedimentary rocks (Chu et al., 2006; Mo et al., 2007; Zhu et al., 2008; Ji et al., 2009; Zhu et al., 2010, 2013), as well as minor Mesozoic–Cenozoic sedimentary strata (Leier et al., 2007). Meanwhile, Precambrian crystalline basement occurs locally in the Linzhi and Chayu areas (C.Y. Dong et al., 2011; X. Dong et al., 2011; Zhu et al., 2013). The northern Lhasa terrane comprises mainly Late Jurassic–Early Cretaceous sedimentary sequences (Yin et al., 1988; Leier et al., 2007), with some Mesozoic (mainly Cretaceous) volcanic–sedimentary rocks and plutonic rocks (Zhu et al., 2011, 2013). The northern Lhasa thrust belt was tectonically active from the Cretaceous through to the Eocene (Kapp et al., 2007), prior to and during the early stages of India–Asia collision.

In the central Lhasa terrane, it exposes the Proterozoic to Early Cambrian metamorphic basement in the Nyainqentanglha area, and is covered with widespread Permo-Carboniferous metasedimentary rocks and small amounts of well-exposed Ordovician, Silurian, and Devonian strata, and Triassic limestone as well (Pan et al., 2006; Zhu et al., 2013). Paleozoic coesite-bearing eclogites have been documented in an E–W trending belt at least hundred km long the central Lhasa terrane, also named the Sumdo–Cuoqen belt (Chen et al., 2009; Li et al., 2009; Yang et al., 2009; Li et al., 2011; Fig. 1). The P-T for these eclogite conditions is estimated to be higher than 2.7 GPa and 730 °C, with coesite pseudo-morphs implying high to ultra-high pressure conditions (Chen et al., 2009; L.S. Zeng et al., 2009; Q.G. Zeng et al., 2009; Yang et al., 2009). The geochemistry and Sr–Nd isotopic analysis suggest a typical MORB affinity for the protolith rocks of the eclogites (Chen et al., 2009, 2011; Li et al., 2011). Eclogites in fault contact with the garnet-bearing, muscovite–quartz schist in the Sumdo region have U–Pb zircon ages of 262 ± 5 Ma (Chen et al., 2008), while metamorphic zircon ages of 225–212 Ma are reported from the Nyainqentanglha area (X. Dong et al., 2011). Muscovite and amphibole from the adjacent muscovite–quartz schists and from the eclogites have Mid–Late Triassic 39Ar–39Ar ages in the range of 220–240 Ma (Li et al., 2009, 2011). Magmatism associated with this orogenic event include: (1) the Early and Middle Permian volcanic rocks that are exposed locally from Cuoqen in the west to Ranwu in the east (Geng et al., 2007; Zhu et al., 2010; Fig. 1); (2) the Late-Permian peraluminous granite at Pikang (Zhu et al., 2009); and (3) the Early–Middle Triassic volcanic rocks (Li et al., 2012), and Late Triassic granites found in several areas (Li et al., 2003; Zheng et al., 2003; Kapp et al., 2005; Chu et al., 2006; He et al., 2006; K. Liu et al., 2006; Q.X. Liu et al., 2006; Zhang et al., 2007; Zhu et al., 2011, 2013). The latter may be attributed to the extension after the Late Permian–Triassic compression (Zhu et al., 2013) or the rifting of Lhasa terrane from Gondwana (Fig. 1).

3. Stratigraphy

Our study region is located in the central portion of the Lhasa terrane, in the Damxung-Linzhou area (Fig. 1), in a zone comprising mostly Ordovician to Tertiary strata. The Paleozoic sedimentary strata consist primarily of Carboniferous sandstone, metasandstone, shale and phyllite, with subordinate Ordovician, Silurian and Permian limestone together with interbedded mafic and felsic volcanic rocks. Triassic rocks are defined mainly by the Late Triassic Mailonggang Formation, with rare exposures of Early Triassic strata of the Chaqupu Formation. The Mailonggang Formation includes a large suite of limestone interbedded with sandstone and mudstone, which occur in thrusts contact above Cenozoic clastic sandstone of the K3 Shexing Formation. The Early Jurassic Jialapu Group is composed of conglomerates, mudstone, and partly volcaniclastic sandstones, locally intruded by the Cenozoic granites. The Upper Jurassic Duodigou Group is made up of litoral facies carbonate. Cretaceous and Tertiary strata consist of clastic mudstone, sandstone and local conglomerate units and arkosic fluvial sandstone and mudstone successions (Li, 1990; Leier et al., 2007; Pullen et al., 2008). In central Lhasa, a northward-propagating retroarc thrust belt (Lhasa–Damxung thrust) developed between the Luobadui–Milashan fault and the Shiquan River–Nam Tso mélangé zone (Fig. 1) has emplaced Paleozoic and Mesozoic strata over Cenozoic sequences, and was active between 105 and 53 Ma (Kapp et al., 2007). Tertiary Linzingon volcanic rocks unconformably over the Late Cretaceous clastic rocks (Mo et al., 2007).

In the study region, the Upper Triassic Mailonggang Formation consists of thick-bedded micritic limestones, interbedded with shale and siltstone in the upper parts of the succession, and clastic/lithic sandstone...
in the lower section (Fig. 2). Corals (*Distichophyllia*, *Margarosmilia*, *Margarophyllia*), bivalves (*Trigonia* cf. *Jingguensis*, *Lilangina nobilis*, *Indopecten*, *Pergomidia*) (Fig. 2F), conodonts and occasionally ammonite fragments (*Nevadites*) indicate a time range from Carnian to Norian (*Ji et al., 2003*). Abundant mud cracks indicate near-shore partly tidal depositional environments, while chert debris flows within the limestones and common load structures (see Fig. 2 A, B, C) may be attributed to earthquake related-shaking during deposition. In sum, the assemblage of fossils, lithology and sedimentary structures imply a tectonically active, shallow water mixed carbonate–clastic platform environments.

4. Petrography of Mailonggang Formation sandstones

We sampled seven sandstones (LZ11-2-5-1, LZ11-2-5-2, LZ11-2-6, LZ11-2-7, LZ11-2-8, LZ11-2-9, LZ11-2-11) and several limestone samples from the Mailonggang Formation for petrographic observation, heavy mineral statistics and detrital zircon analysis (see Appendix A). The stratigraphic positions of the sandstone samples are illustrated in Fig. 2 (see Appendix A).

Mailonggang Formation sandstones are typically poorly sorted with mostly angular to subangular shapes, implying low maturity consistent with relatively proximal provenances. Quartz (71–74%) includes both monocrystalline and polycrystalline (chert) grains with Qm/Qt value ranging of 59.6%–75%. Feldspar constitutes 3–11% of the rock and is commonly twinned. Lithic clasts (16–24%) are commonly defined by volcanic fragments, sedimentary clasts, metamorphic lithics, and occasionally serpenitized ultramafic fragments reported by *Li* (1990). For individual samples a total of 300–400 points were counted using the Gazzi–Dickinson method (Dickinson and Suczek, 1979) and plotted on ternary diagrams. The samples studied in this work plot within the collisional orogen field on a Qp-Lv-Ls diagram and within the recycled orogen provenance field on Qt-F-L and Qm-F-Lt diagrams (Fig. 3).

Detrital heavy mineral suites are composed primarily of zircon, apatite, tourmaline, rutile, anatase, leucoxene, magnetite, limonite, and sporadic biotite, chromite, amphibole and barite and epidote. Using the heavy mineral assemblages analysis proposed by *Morton and Hallsworth* (1994), we calculated the index values (ATi, RZi, and CZi, respectively) (see Appendix A). A wide variation of the ATi values in these samples, range from 2 to 100 with especially low values in the samples from the middle part of section, indicate that some samples were probably derived from the recycled sediments (but it can't be completely excluded the greater apatite loss makes the ATi values lower during transport, deposition and diagenesis); and some with high ATi values indicate that igneous rocks were major contributors. The RZi values are relatively low (mostly around 5) in the sequence. CZi values are generally very low in the section, with the exception of one sample (LZ11-2-8) which indicates a locally significant contribution from ultramafic rocks. Generally, the heavy mineral assemblages of sandstones are consistent with a mixed provenance includes volcanic, magmatic, sedimentary and ultramafic rocks, and maybe some metamorphic rocks.

Fig. 2. Sedimentary column of Mailonggang Formation showing the characteristic of lithology, structures and fossils. A. Load structure on the bottom of limestone; B. Chert debris in the limestone; C. Debris flow in the limestone layer; D. Convolute and horizontal bedding in the sandstone; E. Micrograph of the sandstone; F. Coral fossil from the limestone.
5. Detrital zircon analysis

5.1. Analytical methods

Zircon crystals were obtained from crushed rock using heavy liquid and magnetic separation techniques, and mounted in epoxy resin. Cathodoluminescence (CL) images were used to check the internal structures of individual zircon grains and to select potential target domains for in-situ U–Pb dating and Hf analyses. U–Pb and Hf isotope analyses were conducted at the School of Earth Sciences, University of Melbourne. An Agilent 7500a quadruple inductively-coupled-plasma-mass-spectrometer (QICP-MS), coupled to a 193 nm ArF excimer laser was used to collect U–Pb data, while Hf isotopes were collected by using Nu Plasma Multi Collector-ICPMS, also coupled to a 193 nm ArF excimer laser. Analytical methods followed the procedures outlined in Woodhead et al. (2004, 2007), Eggins et al. (2005), and Paton et al. (2010). The Plesovice, TEMORA and 91500 zircon standards were analyzed along with zircon unknowns to correct for U–Pb fractionation. Data reduction and fractionation correction for U–Pb analyses were undertaken using the Iolite Wavemetrics Igor Pro data analysis software (Hellstrom et al., 2008). Concordia plots were processed using ISOPLOT 3.0 (Ludwig, 2003). The εHf and model ages reported here use $^{176}$Lu decay constant $\lambda = 1.867 \times 10^{-11}$ a$^{-1}$, $^{176}$Lu$^{177}$Hf$_{CHUR} = 0.0332$, $^{176}$Hf$^{177}$Hf$_{CHUR} = 0.282772$, $^{176}$Lu$^{177}$Hf$_{DM} = 0.0332$ and $^{176}$Hf$^{177}$Hf$_{DM} = 0.282772$ (Woodhead et al., 2004).

5.2. U–Pb detrital zircon chronology

A total of 517 rounded–subhedral zircon grains with sizes between 50 and 250 μm were analyzed. Cathodoluminescence (CL) images and Th/U ratios show two distinct zircon types. By far the most abundant zircon type (~96%) has strong oscillatory zones and Th/U ratios ~4 (and up to 5.24) that we interpret to be of a magmatic origin. The high Th/U zircons yielded 493 concordant U–Pb ages (by 90–110%) in the range of 3458–212 Ma, with five major age peaks ~304 Ma, ~560 Ma, ~1155 Ma (three biggest), ~1564 Ma, and ~1750 Ma. The youngest age cluster is in the range of 240–212 Ma (Figs. 4, 5, and Appendix A) and corresponds to a period of significant magmatism in the Lhasa terrane (Li et al., 2003; Zheng et al., 2003; Kapp et al., 2005; Chu et al., 2006; He et al., 2006; K. Liu et al., 2006; Q.X. Liu et al., 2006; Zhang et al., 2007; Zhu et al., 2011, 2013). The second type, comprising only about 4% of the population, is unzoned or only weakly zoned has low Th/U (0.009–0.09), and $^{206}$Pb/$^{238}$U ages ranging from 438.7 Ma to 1932.4 Ma. This population probably derives from the metamorphic sources.

5.3. Hf isotopes

A total of 156 grains from the Mailonggang Formation were analyzed for in-situ Hf isotopic abundance. The zircons fall into four broadly defined groups: 1) the youngest group with ages 240–212 Ma has $\varepsilon_{Hf(t)}$ values of −4.2 to +9.4, with $T_{DM}$ ages in the range of ~1008–484 Ma; 2) a group with Paleozoic ages (~512–315 Ma) with $\varepsilon_{Hf(t)}$ values of −16.5 to +8.1 with $T_{DM}$ ages in the range of ~1643–809 Ma; 3) the −1255–530 Ma group typically has $\varepsilon_{Hf(t)}$ values in the broad range of −31.4 to +11.0 and $T_{DM}$ ages in the range of 2.7–1.1 Ga; and 4) the −3471–1431 has $\varepsilon_{Hf(t)}$ values in the range of ~13.4 to 2.5 and $T_{DM}$ ages of 3.7–1.6 Ga, and a few zircon crystals in this group have more positive $\varepsilon_{Hf(t)}$ values (~4.7 to +9.6) (Fig. 7 and Appendix A).

6. Provenance of the Late Triassic sediments

6.1. Overview of zircon U–Pb ages and Hf isotope data from Lhasa terrane and its relevant terranes

In this section we summarize other zircon U–Pb ages and Hf isotopic data collected from elsewhere in Tibet to help constrain the likely provenance of the Mailonggang Formation (Table 1).

The Qiangtang terrane is located north of Bangong–Nujiang suture zone. Recent detailed geochronologic studies indicate magmatism was active during Paleozoic–Early Mesozoic (~530–202 Ma) (Kapp et al., 2003; Zhai et al., 2007, 2010; Pullen et al., 2011, Zhu et al., 2013). The age distributions of the detrital zircons from the Paleozoic metadolerites exhibit two main peaks at ~754 and ~942 Ma, characterized by $\varepsilon_{Hf(t)}$ values of ~37.9–16.2 and $T_{DM}$ from 0.7 to 4.0 Ga (Kapp et al., 2003; C.Y. Dong et al., 2011; Zhu et al., 2011).

Lhasa terrane basement rocks (as shown in Fig. 1) have been dated at ~912–501 Ma (Hu et al., 2005; Guynn et al., 2012; Zhu et al., 2013), Late Permian to Early Jurassic magmatic rocks from the central and southern parts of the Lhasa terrane (~265–190 Ma) (Chu et al., 2006; Wen et al., 2008; Ji et al., 2009; Zhu et al., 2009, 2013; references therein, Fig. 1) contain zircons with $\varepsilon_{Hf(t)}$ values (~17.3 ~ +6.7) and two-stage Hf model ages ($T_{DM}$) from 0.3 to 2.5 Ga (Chu et al., 2006; Zhang et al., 2007; Ji et al., 2009; Zhu et al., 2009, 2011, 2013). Several pre-Permain volcanic central Lhasa terrane sequences have $\varepsilon_{Hf(t)}$ values
Fig. 4. U–Pb concordia plots for detrital zircon analyses from the Upper Triassic Mailonggang Formation.

Fig. 5. U–Pb age probability plots for detrital zircon analyses from the Mailonggang Formation.
in range of $-13.9$ to $+7.5$ with $T_{DM}$ values ranging from $0.9$ to $2.3$ Ga (Dong et al., 2010; Zhu et al., 2013). X. Dong et al. (2011) reported the U–Pb zircon ages of $\sim 225$–$212$ Ma indicated the metamorphic event occurred in Nyainqentanglha during Triassic near this study region. In the southern Lhasa terrane, Early Jurassic volcanic rocks (both felsic and mafic, with minor amounts of andesite) in the Yeba Formation dated at $\sim 190$–$172$ Ma (Zhu et al., 2008). Pre-Mesozoic inherited zircons and detrital zircons are common in Lhasa, showing $\varepsilon_{Hf}(t)$ values in range of $-38.1$ to $+15.6$ with $T_{DM}$ ages ranging from $0.3$ to $3.9$ Ga (Chu et al., 2006; Leier et al., 2007; Ji et al., 2009; Zhu et al., 2009, 2011).

To the south of the Indus–Yarlung suture zone, crystalline basement outcrops in the Greater Himalaya (dominantly Precambrian in age) and in the Tethyan Himalaya as a series of the discontinuous belt of gneiss domes with U–Pb ages of about $500$ Ma (Lee et al., 2000; Decelles et al., 2004; Gehrels et al., 2006; Myrow et al., 2010). The Paleozoic magmatic rocks are reported in Greater Himalaya (Zanskar, Mandi, Yadong, Abor), and Tethyan Himalaya (Gyirong, Selong, Kangmar, Mabja) (Spring et al., 1993; Garzanti et al., 1999; Miller et al., 2001; Cawood et al., 2007; Zhu et al., 2010). Detrital zircons from the Permian–Precambrian elastic strata in these region have four major age peaks at $\sim 540$, $\sim 950$, $\sim 1500$–$1700$ Ma, and $2500$ Ma and have a wide range of $\varepsilon_{Hf}(t)$ values ($-29.5$ to $12.6$) with varying $T_{DM}$.
of the paleogeographic constraints. The existence of a Late Triassic deep Himalayan terranes (Fig. 1; Yin, 2006) would seemingly preclude an ob-
Permanent and pre-Permian volcanic occurrences in the central Lhasa 
et al., 2013). Parts of the Paleozoic zircons are comparable to the 
rocks (Figs. 1, 6, 7), while parts of the negative 
the derivation of the Lhasa terrane and Triassic Gangdese magmatic arc 
values of ~1170 Ma while for the Himalaya, age peaks are at ~480, ~840, and 
Lhasa, and Himalaya). However, there are some important differences 
clude major Paleozoic, Late Neoproterozoic, Late Mesoproterozoic peaks 
6.2. General provenance considerations

Detrital zircon ages of the samples from Mailonggang Formation in-
clude major Paleozoic, Late Neoproterozoic, Late Mesoproterozoic peaks 
and a broad distribution of Early Neoproterozoic—Early Paleoproterozoic. 
This range of zircon ages is common in all Tibet terranes (Qiangtang, 
Hlasa, and Himalaya). However, there are some important differences 
between these terranes. For example, the major Precambrian age peaks 
in the Qiangtang terrane occur at ~ 550 Ma, ~800 Ma, ~950 Ma, and 
~2500 Ma. For the Lhasa terrane the major age peaks are at ~550 and 
~1170 Ma while for the Himalaya, age peaks are at ~480, ~840, and 
~950 Ma (Fig. 6).

The absence of significant detrital zircons in the Mailonggang For-
modation with age peaks of ~800, ~950 Ma and ~2500 Ma precludes a 
Qiangtang terrane provenance as the main source for Mailonggang For-
modation (Fig. 6). Furthermore, the immature nature of the Mailonggang 
metasandstone also suggests a proximal source. Lastly, paleogeographic 
constraints also argue against the possibility that the Mailonggang sed-
iments were derived from the Qiangtang terrane, since there is evidence 
that Qiangtang terrane was not juxtaposed with the Lhasa terrane until 
Late Jurassic or Early Cretaceous (Allègre et al., 1984; Yin and Harrison, 
2000; Zhu et al., 2013; references therein).

Similarly, a Himalayan provenance can be precluded mainly because 
of the paleogeographic constraints. The existence of a Late Triassic deep 
water flysch (Langjixue Group) in IYSZ deposited between Lhasa and 
Himalayan terranes (Fig. 1; Yin, 2006) would seemingly preclude an ob-
vious route for Himalayan sediments to be transported to the shallow 
water carbonate–clastic platforms in the Lhasa terrane from south 
to north. Furthermore, measured paleocurrents in Langjiexue Group 
show a southward paleo 
flow (Li et al., 2004), which also indi-
cates that the sediments of Himalaya didn’t reach into the Lhasa terrane 
to the north.

The youngest cluster of U–Pb zircon ages (~240–212 Ma) with εHf(t) 
values of ~4.2 to ~9.4 from the Mailonggang Fm. are compatible with 
the derivation of the Lhasa terrane and Triassic Gangdese magmatic arc 
rocks (Figs. 1, 6, 7), while parts of the negative εHf(t) zircons could 
derive from the metamorphic rocks (C.Y. Dong et al., 2011; X. Dong 
et al., 2011), and/or from the southern Lhasa terrane as well (Zhu 
et al., 2013). Parts of the Paleozoic zircons are comparable to the 
Permian and pre-Permian volcanic occurrences in the central Lhasa 
terrane with εHf(t) values in range of ~ -13.9 ~ +7.5 (Zhu et al., 2011).

The distribution of Paleozoic zircons in this study with two main 
peaks at ~550 and ~1170 Ma is consistent with the patterns of the de-
trital zircons with the varied εHf(t) values (~ -38.1 ~ 15.6) and εHf(t) 
age (~0.3 ~ -3.9 Ga) from Permo-Carboniferous sandstones in the central 
Lhasa terrane (Leier et al., 2007; Pullen et al., 2008; Zhu et al., 2011; 
Figs. 5, 6). These indicate the igneous, recycled sedimentary and a bit of 
metamorphic rocks in central Lhasa terrane provided the sources, which 
quite agrees with the results of petrography and heavy mineral 
analysis.

Our new zircon data, combined with petrography and heavy mineral 
analysis, indicate the proximal igneous rocks, recycled sediments and 
some ultramafic, metamorphic rocks are the mainly potential sources 
for the Mailonggang Formation. These sources match with the rock 
units in the central Lhasa terrane and the Sumdo–Cuoqen belt (Fig. 1). 
Meanwhile, the Sumdo–Cuoqen belt has been initially defined as a 
new Late Permian–Triassic orogenic belt in the central Lhasa terrane, 
indicated by the Late Paleozoic high or ultrahigh metamorphic belt with 
igneous rocks and Early Mesozoic thrust fault system (Chen et al., 
2009; Li et al., 2009; Yang et al., 2009; Li et al., 2011; X. Dong et al., 
2011). The Mailonggang sequence is in depositional contact with or 
located immediately southeast of the Sumdo area. Generally, according 
to these evidences and our results, we suggest the Upper Triassic 
Mailonggang Formation was derived from the Late Permian–Triassic 
orogenic belt along Sumdo–Cuoqen belt in central Lhasa terrane.

7. Discussion

One issue concerns the relationship between the Upper Triassic 
Mailonggang Formation and the Langjixue Group. Li et al. (2010) pro-
posed that the Triassic Langjixue Group on the south side of Indus–
Yarlung ophiolites (Fig. 1) originated along the southern margin of the 
Lhasa terrane in a forearc or backarc setting (Li et al., 2010). Our analysis 
suggests that both the Mailonggang and Langjixue sequences were 
likely derived from the Late Permian–Triassic Sumdo–Cuoqen belt in 
the Lhasa terrane. This contention is supported by the similarities in de-
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Late Triassic

Gondwana

Passive continental sequences

Langjiexue G.

Mailonggang F.

Late Permian-Triassic orogenic zone

Subduction on the northern edge of Gondwana

Fig. 8. The sketch tectonic model of the Upper Triassic sequences on the south margin of the Lhasa terrane.

Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.gr.2013.06.019.

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Nelson (1993) has the northern margin of Indian block extending no further north than the Main Himalayan Thrust (MHT) along the south margin of the Langjiexue Group, rather than all the way to the IYSZ. The K. Liu's et al. (2006), Q.X. Liu's et al. (2006) interpretation is consistent with the notion that Langjiexue Group is part of the Lhasa terrane. We conclude that the Langjiexue Group and Mailonggang Formation were likely both deposited on the south margin of Lhasa terrane (Fig. 8) probably before the initial drifting of Lhasa terrane from Gondwana. As mentioned earlier, the newly recognized Late Permian–Triassic orogenic belt in the central Lhasa terrane exhibits Mid–Late Triassic cooling ages. We propose that the erosional denudation of this belt provided the proximal source of immature terrigenous sediment observed both in the Mailonggang Formation and the Langjiexue Group.

A key question is whether the Lhasa terrane originally derives from the Indian (e.g., Allègre et al., 1984; Yin and Harrison, 2000) or Australian sector of the Gondwana margin (Zhu et al., 2011, 2013). In the former case, a simple model for the early evolution of Neo-Tethys sees the Lhasa terrane rifted from India along the southern edge of Lhasa terrane, thereby segmenting the formerly contiguous Langjiexue Group and Mailonggang Formation during Late Triassic–Early Jurassic period. In this scenario, the Lhasa terrane subsequently collided with the Qiangtang Terrane in the Early Cretaceous (Yin and Harrison, 2000; Yin, 2006). A key observation supporting this hypothesis is the presence of rift-related Late Triassic–Early Jurassic magmatism in the Lhasa terrane (Zhang et al., 2007; Zhu et al., 2013), which indicates that rifting had continued well after the Permian (Garzanti et al., 1999; Chauvet et al., 2008). Similar ages and styles of magmatism are observed along Gondwana margin fragment in East Indonesia (Java), New Guinea and Northeast Australia as a precursor to eventual Jurassic breakup (Crowhurst et al., 2004; Hill and Hall, 2003; Metcalfe, 2011; Van Guts et al., 2013).

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.gr.2013.06.019.

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