

Petrology and geochemistry of metabasalts from the Taoxinghu ophiolite, central Qiangtang, northern Tibet: Evidence for a continental back-arc basin system

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Abstract

Newly discovered ophiolitic metabasalts in the Taoxinghu area of central Qiangtang on the Qinhai-Tibet Plateau, here described for the first time, have important implications for reconstructions of the tectonic history of the Paleo-Tethys Ocean. Most of the metabasalts belong to the tholeiitic basalt series and most have undergone greenschist-facies metamorphism. The distribution of rare earth elements and trace elements shows that the rocks are typical of sub-continental margin arc-basin lavas, similar to the environment of formation of the present-day Okinawa lava, suggesting that the Taoxinghu metabasalts represent the upper portions of a supra-subduction zone (SSZ)-type ophiolite that formed in a continental back-arc basin tectonic environment. The Taoxinghu metabasalts are mainly sourced from a depleted spinel mantle-source region, with a spinel lherzolite content equivalent to partial melting of 6%-25%. In addition, lava compositions were likely affected by melting of sediments during subduction, while the influence of aqueous fluids was minor. Combining with the existing knowledge on the ophiolites of Longmuco-Shuanghu-Lancang suture zone (LSLSZ), an evolutionary model is proposed. The LSLSZ Paleo-Tethys Ocean basin may have started to form during the Cambrian or earlier, and subducted in the early Carboniferous. As subduction proceeded, a continental back-arc basin was developed, the site of generation of most of the Taoxinghu metabasalts. The LSLSZ Paleo-Tethys Ocean finally closed in the Triassic.

1. Introduction

As relicts of ancient oceans, ophiolites indicate the occurrence of suture-zone and post-collisional-zone processes, and are therefore useful for investigating the evolution of orogenic belts and for reconstructing tectonic histories (Savov et al., 2001; Xu et al., 2003). Volcanic rock associations preserved in ophiolites generally offer a means of assessing the processes and conditions of basaltic magmatism through time (Ghazi et al., 2012). Therefore, the study of volcanic rocks provides a key for evaluating the evolution of the oceanic lithosphere.

The Longmuco-Shuanghu-Lancang suture zone (LSLSZ), located in central Qiangtang, on the Qinhai-Tibet Plateau, extends for at least 1000 km from western Longmuco to the east of Changdu (Fig. 1a). The LSLSZ is characterized by ophiolitic fragments, high-pressure metamorphic rocks, mélanges, and volcanic rocks, and is interpreted to represent the relicts of the main Paleo-Tethys Ocean basin (Li, 2008; Wu, 2013). However, some scholars have considered that the LSLSZ represents a rifting environment (Wang et al., 1987; Yin, 1997). In addition, Kapp et al. (2003) proposed that the metamorphic rocks were underthrust southward beneath the Qiangtang continental margin during Triassic low-angle oceanic subduction along the Jinsha suture, and that they were then exhumed to upper crustal levels by Late Triassic-Early Jurassic normal faulting. The controversy about the origins of the LSLSZ remains acti-

ve on account of the paucity of systematic detailed data on LSLSZ ophiolites.

Previous studies of the LSLSZ have focused mainly on the geochemistry and isotopic dating of high pressure (blueschist and eclogite facies) metamorphic rocks (Zhai et al., 2009, 2010; Li et al., 2006, 2009; Dong et al., 2009). Recently, regional geological surveys in Qiangtang have revealed an increasing number of ophiolitic sequences in the region (Zhai et al., 2007; Li, 2008; Wu et al., 2009); these sequences provide important evidences for understanding the tectonic evolution of the LSLSZ.

In this study, we conducted petrological and whole-rock geochemical investigations of the metabasalts from the Taoxinghu ophiolite in the western part of the LSLSZ. The results elucidate the petrogenetic history of the metabasalts and help to constrain the tectonic setting of the Taoxinghu ophiolite, and the evolution history of LSLSZ.

2. Geological setting

Longmuco-Shuanghu-Lancang suture zone is mainly composed of blueschist, eclogite and ophiolite rocks. The blueschist and eclogite occur in Gangmaco, Gemu, Rongma and Shuanghu areas (Zhai et al., 2013) (Fig. 1a), and show subducted E-MORB and OIB affinities, and yield ages of 214-240 Ma (Zhai et al., 2010, 2011; Li et al., 2006, 2009; Pullen et al., 2008; Dong et al., 2009; Zhai et al., 2009). The P-T conditions of the

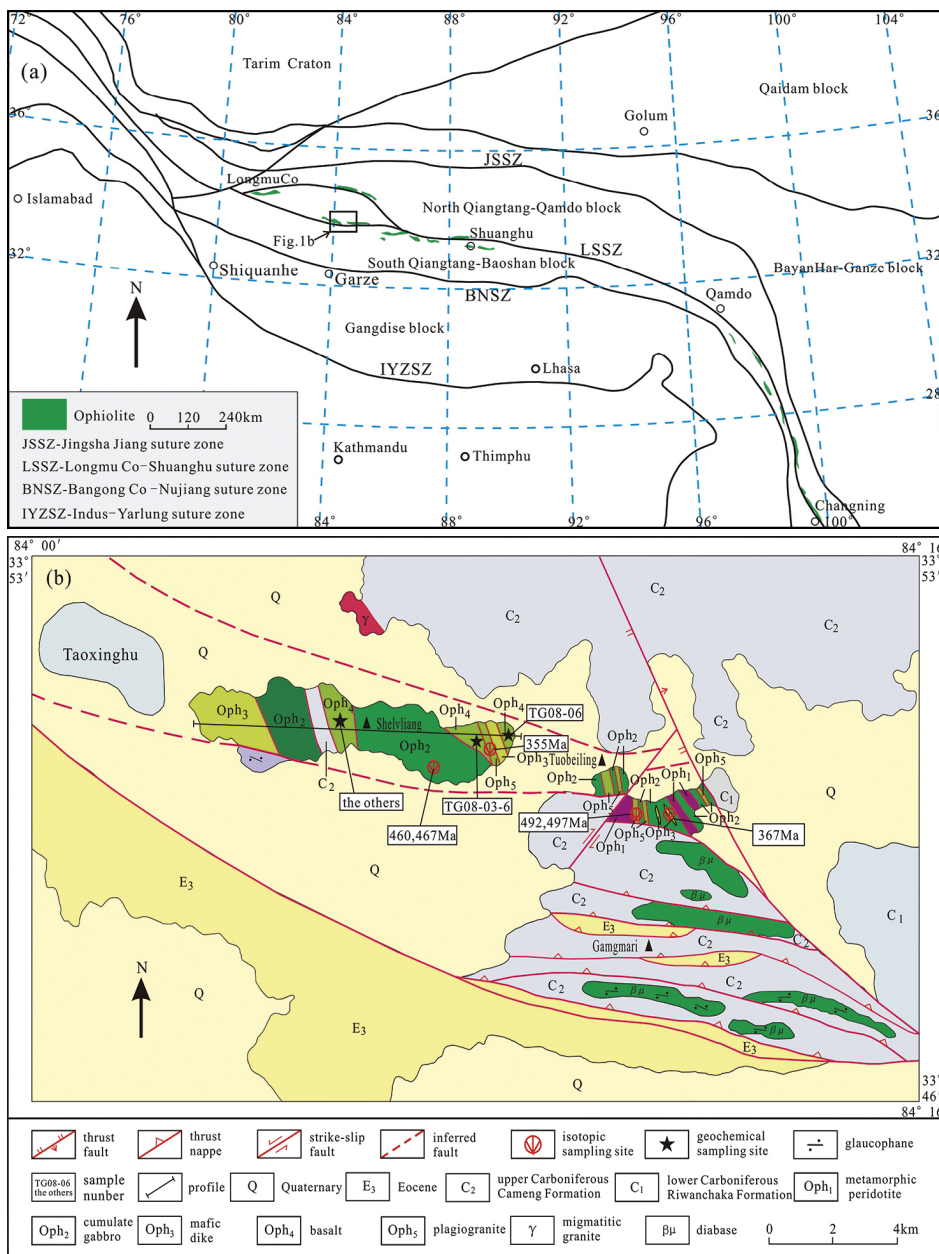


Figure 1b. (a) Map showing tectonic units and the distribution of ophiolite in the Longmuco-Shuanghu-Lancang suture zone, and (b) a schematic geological map of the Taoxinghu ophiolite in central Qiangtang (modified from Wu et al. 2009).

peak eclogite-facies and blueschist-facies metamorphisms are (230-244Ma, Zhai et al., 2011; Pullen et al., 2008) 410-460 °C and 2.0-2.5 GPa (Zhai et al., 2011) and 330-415°C and 9-11.5 kbar (Tang and Zhang 2013), respectively. Recent field mapping shows that the ophiolites from Taoxinghu, Guoganjian-an, Gangmaco and Jiaomuri areas in Qiangtang (Li et al. 2008; Wang et al. 2008; Zhai et al. 2010, 2013). However, no complete sequence of ophiolite has been preserved. Cambrian-Permian zircon U-Pb ages have been obtained from metagabbro, metabasalt and plagiogranite of these ophiolites (e.g. Wang et al., 2008; Zhai et al. 2013), implying a long evolution period for Paleo-Tethys Ocean (Wu, 2013). Geochemical analyses of these ophiolites show a progressive evolution from MORB in Cambrian to SSZ in Permian (Wu, 2013).

The Taoxinghu area in the western part of the LSSSZ is located ~200 km north of Gaize, central Qiangtang (Fig. 1a). The structural style of the Taoxinghu area is dominated by E-W-trending faults, and overthrust structures in the Gangmacuo area (Xie et al., 2010). The upper Carboniferous Cameng Group consists of a low-grade metamorphic sequence, consisting of metamorphosed quartz sandstone, schist, phyllite, crystalline limestone, and mafic volcanic rocks, along with glacial diamictite and some mafic volcanic rocks (diabase and basalt) that have undergone blueschist-facies metamorphism (Li, 2003; Geological Survey Academy of Guizhou Province, 2005). The lower Carboniferous Riwanhaka Group consists of limestone and clastic rock containing a Yangtze-type fauna, including coral and brachiopods. Brecciform cataclastic limestone with calcite veins is observed to the east of the ophiolitic mélangé, and is inferred to be early Carboniferous in age. Paleogene strata, consisting of red sandstone and conglomerate, occur mainly in the southern part of the study area.

The Taoxinghu ophiolite, ~14 km long and 2-3 km wide, is oriented E-W between Shelvliang and Tuobeiling, where it is divided into the Shelvliang and Tuobeiling ophiolites based on outcrop location (Fig. 1b). Taoxinghu ophiolite is dismembered, and the ophiolitic members occur as block and slice and are in fault contact with each other (Fig. 2). The Shelvliang ophiolite consists mainly of cumulate gabbro, cumulate pyroxenite, gabbro dikes, plagiogranite, and metabasalt (Wu et al., 2009; Zhai et al., 2010). The Tuobeiling ophiolite includes metamorphic peridotite, cumulate gabbro, cumulate pyroxenite, gabbroic dikes, and plagiogranite (Wu et al., 2009; Zhai et al., 2010). These ophiolitic rocks have been variably altered; the peridotites are mostly serpentinized, with the modal content of serpentine varying from 60% to 80% (Wu et al., 2013). The cumulate gabbro and gabbro dikes have mostly undergone greenschist- to lower amphibolite-facies

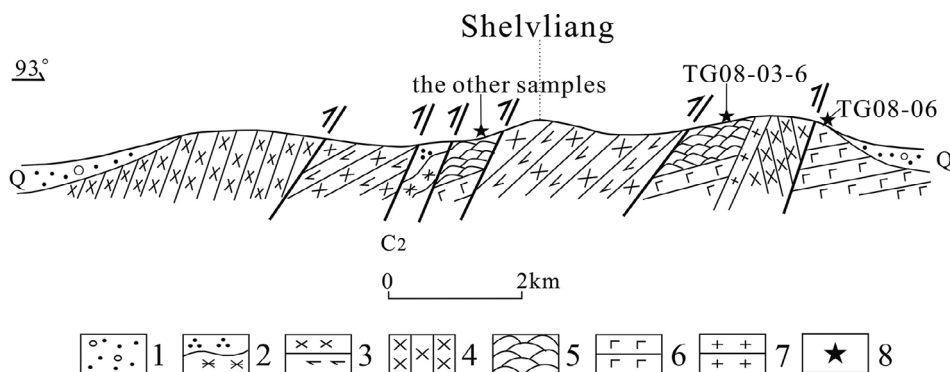


Figure 2. Profile of Shelvliang ophiolite in Chabu town, Gaize country (modified from Wu et al., 2009). 1 – Quaternary sediment; 2 – actinolite quartz schist; 3 – cumulate; 4 – diabase dike; 5 – pillow basalt; 6 – massive basalt; 7 – plagiogranite; 8 – sampling location

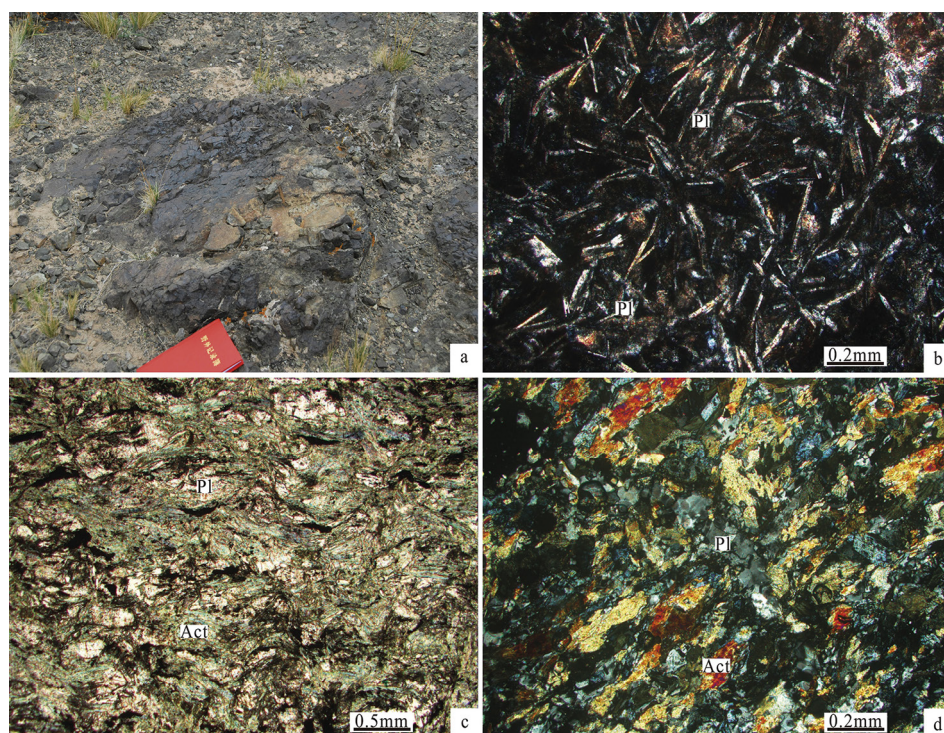


Figure 3. Representative petrographic characteristic of metabasalts from Taoxinghu ophiolite. a-pillow basalt; b-interstitial texture of basalt; c, d-actinolite schist; Pl-plagioclase; Act-actinolite; Zo-zoisite

metamorphism, with pyroxene being replaced by amphibole. Zhai et al. (2010) reported a zircon SHRIMP U-Pb age of 467 ± 4 Ma from these metagabbros, which have similar characteristics to those in the mid-ocean ridge basalts.

3. Petrology

The basalts examined in this study were collected from the Shelvliang ophiolite, and sample locations are shown in Figure 1b and 2. The basalts occur as slices in fault contact with cumulate gabbro, plagiogranite, and Late Carboniferous quartz schist (Fig. 2). The basalts mainly display small to large brecciated pillow structure, with pillows up to 80 cm in diameter (Fig. 3a), and massive basalts are only found in the eastern part of Shelvliang ophiolite (Fig. 2). The fresh basalts mostly show intersertal textures with glass filling the spaces between sub-

hedral crystals of plagioclase (crystal size, 0.2-0.5 mm), and the plagioclase laths show various degree of alteration to clay minerals and sericite (Fig. 3b). Most of the basalts have undergone greenschist-facies metamorphism and contain mineral assemblages of actinolite (50-60%) and plagioclase (40-50%), with minor zoisite and epidote (Fig. 3c, d). The metamorphic minerals are often slightly oriented. These metabasalts have similar mineral assemblages to the metagabbro of Taoxinghu ophiolite and metamorphic mafic rocks from other ophiolites in LLSZ (e.g. Li et al., 2008; Zhai et al., 2010, 2013), suggesting that their metamorphic degree are basically identical.

4. Geochemistry

4.1 Analytical methods

Samples were powdered to less than 200 mesh for whole-rock analyses, performed at the Geological Survey, Hebei Province, Langfang, China. Major element oxide compositions and trace element concentrations were determined on samples TG07 and TG08 at the Chinese Academy of Geological Sciences, Beijing, China. Major element analyses were performed by

X-ray fluorescence spectrometry (XRF) and trace element concentrations were determined by inductively coupled plasma-mass spectrometry (ICP-MS), inductively coupled plasma-optical emission spectrometry (ICP-OES), and X-ray fluorescence spectrometry (XRF). Standards GSR2 and GSR3 were used in analyses of both major and trace elements. The analytical precision (relative standard deviation) was generally better than 2% for major elements and better than 5% for trace elements. The whole-rock elemental composition of sample Tg10 was determined at the China University of Geoscience, Beijing, China. Major and trace element abundances were determined using a PS-950 XRF and an Agilent 7500a ICP-MS, respectively. Analytical details and results of analyses of reference materials are reported in Yu (2011). The analytical results are listed in Table 1.

Sample	Tg10-23-H1	Tg10-23-H2	Tg10-23-H3	Tg10-23-H4	Tg10-23-H5	Tg10-23-H6	Tg10-23-H7	Tg10-23-H8	Tg10-23-H9	Tg10-23-H10	TG08-01-6	TG08-04-6	T07-18	T07-18R	T07-19	TG08-03-6	TG08-06
SiO ₂	54.42	59.40	55.59	58.89	59.16	56.23	53.19	54.54	52.24	51.89	49.16	53.53	48.72	48.70	45.54	49.37	51.59
TiO ₂	1.18	1.05	1.24	1.13	1.09	1.12	1.20	1.30	1.34	1.32	1.28	0.91	1.08	1.09	1.11	1.12	3.80
Al ₂ O ₃	14.41	14.29	13.57	12.82	13.35	13.05	14.57	13.69	14.05	13.98	15.15	16.84	15.04	15.01	15.38	13.79	8.18
Fe ₂ O ₃ T	9.99	7.59	9.26	7.89	8.36	9.73	9.23	9.84	9.42	10.58	10.24	9.15	10.56	10.47	10.15	11.72	12.40
MnO	0.13	0.10	0.14	0.12	0.12	0.14	0.13	0.14	0.13	0.15	0.14	0.16	0.21	0.21	0.16	0.21	0.14
MgO	5.80	4.25	5.35	4.58	5.02	5.69	5.43	5.95	5.52	7.18	6.42	5.46	6.82	6.84	6.32	7.94	8.69
CaO	7.28	4.56	8.27	8.22	4.56	6.99	9.24	6.42	10.55	7.12	8.75	8.60	9.33	9.33	11.60	11.67	6.87
Na ₂ O	3.78	4.08	3.01	2.71	3.78	3.01	2.87	3.26	3.44	3.46	3.36	2.85	3.31	3.31	2.19	2.21	3.81
K ₂ O	0.04	1.35	0.11	0.18	0.96	0.15	0.48	0.33	0.10	0.28	0.23	0.44	0.25	0.25	0.59	0.04	1.43
P ₂ O ₅	0.21	0.23	0.23	0.22	0.21	0.20	0.21	0.24	0.24	0.24	0.22	0.17	0.14	0.14	0.14	0.09	0.44
LOI	2.41	2.42	2.79	2.57	2.62	3.11	3.27	3.81	2.84	3.84	4.40	1.28	4.23	4.26	6.36	1.50	2.08
Total	99.64	99.31	99.55	99.33	99.23	99.42	99.82	99.53	99.86	100.03	99.35	99.39	99.69	99.61	99.54	99.66	99.43
Li	22.48	22.30	23.68	22.10	25.17	28.86	27.24	26.02	20.94	27.62	31.50	5.77	-	-	-	9.28	32.60
Be	-	-	-	-	-	-	-	-	-	-	0.63	0.92	-	-	-	0.24	2.17
Sc	31.22	26.64	32.34	33.64	27.25	32.08	32.96	33.78	34.76	34.02	30.50	22.90	37.50	38.30	38.60	48.50	30.90
V	242	196	236	253	206	236	280	233	281	261	244	223	257	261	270	311	309
Cr	149	109	145	174	127	165	159	173	493	197	172	92	219	216	198	101	558
Co	32.64	24.42	32.22	34.54	26.20	35.60	31.08	35.70	36.08	42.40	40.50	29.90	40.80	41.20	41.60	45.70	51.80
Ni	93.94	68.66	87.36	110	76.31	108	97.56	113	425	131	119	57.80	103	103	89.50	60.30	307
Mn	945	749	981	1044	869	1074	1011	1065	1137	1366	-	-	-	-	-	-	-
Cu	51.01	42.84	57.81	77.62	47.82	72.30	63.25	64.43	72.73	82.37	72.50	49.60	87.20	88.40	105	40.70	153
Zn	52.62	62.00	70.49	83.23	78.83	107	86.92	86.50	68.77	101	91.20	94.60	87.10	87.50	87.90	120	139
Ga	16.91	15.27	16.25	17.35	15.19	17.39	20.26	16.21	19.50	18.13	20.50	20.50	16.20	16.50	17.50	18.40	21.60
Ge	-	-	-	-	-	-	-	-	-	-	0.95	1.08	1.96	1.97	1.69	0.99	0.69
Cs	0.78	3.45	1.62	1.70	2.51	2.46	8.25	4.44	1.18	5.03	2.75	0.43	0.31	0.32	0.49	0.28	1.45
Rb	1.34	40.52	4.17	7.12	28.34	7.91	29.62	15.37	3.70	11.69	10.90	4.11	7.62	7.68	17.90	0.79	42.20
Ba	40.20	538	97.18	77.98	379	145	169	161	30.53	203	72.60	143	62.80	63.00	131	7.04	471
Th	0.72	4.36	0.88	0.88	5.21	0.76	0.84	0.94	0.69	0.68	0.77	0.65	2.77	2.84	2.91	0.11	8.09
U	0.19	1.05	0.21	0.23	1.08	0.19	0.20	0.23	0.20	0.15	0.20	0.10	0.65	0.64	0.62	0.19	2.08
Nb	4.67	7.97	5.52	6.05	7.94	5.14	5.16	5.91	5.93	5.48	4.27	2.79	9.13	9.34	9.73	0.92	49.20
Ta	0.31	0.60	0.35	0.39	0.63	0.33	0.34	0.39	0.38	0.36	0.29	0.19	0.61	0.61	0.62	0.11	3.57
La	8.99	17.81	9.64	10.63	16.82	8.28	8.96	10.40	9.23	8.76	7.04	11.00	11.20	11.60	11.70	1.77	48.90
Ce	20.80	38.52	22.92	24.92	37.56	19.96	21.58	24.48	20.55	19.45	17.10	26.80	23.40	24.40	24.50	6.05	104
Pb	1.08	12.34	1.65	2.83	7.82	1.49	1.86	2.42	1.09	1.61	1.64	4.77	4.10	4.12	6.24	0.72	6.13
Pr	2.87	4.87	3.17	3.43	4.70	2.78	3.01	3.37	2.94	2.78	2.45	3.48	2.90	3.04	3.14	1.09	12.80
Sr	180	381	185	310	398	205	286	365	144	271	312	519	161	164	422	103	306
Nd	13.86	20.74	15.18	16.41	19.88	13.43	14.48	16.05	14.05	13.27	11.90	15.30	12.90	13.50	14.20	6.51	51.80
Zr	86.00	126	94.28	104	134	84.85	87.27	101	95.10	91.44	77.50	37.30	88.60	89.30	92.80	55.40	320
Hf	2.08	3.15	2.36	2.58	3.38	2.11	2.20	2.52	2.25	2.25	2.21	1.33	2.30	2.31	2.40	1.73	8.45
Sm	3.84	4.89	4.21	4.59	4.73	3.78	4.10	4.42	3.83	3.60	3.28	3.27	3.23	3.37	3.52	2.44	10.20
Eu	1.30	1.35	1.39	1.48	1.25	1.23	1.33	1.48	1.32	1.24	1.24	1.02	1.02	1.05	1.16	1.00	3.14
Ti	8157	7288	8829	9348	7626	8516	8799	9008	9454	9366	7672	5455	6474	6533	6653	6713	22777
Gd	4.51	5.06	4.90	5.34	4.89	4.45	4.81	5.19	4.47	4.23	4.31	3.46	3.68	3.73	4.01	3.98	9.14
Tb	0.72	0.77	0.78	0.86	0.75	0.72	0.77	0.84	0.75	0.71	0.69	0.51	0.65	0.65	0.69	0.69	1.24
Dy	4.63	4.90	5.03	5.52	4.77	4.57	4.92	5.36	4.62	4.33	4.57	3.17	4.06	4.13	4.31	5.05	6.71
Y	25.72	27.28	27.70	30.82	26.68	25.32	27.56	29.78	25.75	24.42	23.50	15.70	23.50	23.60	24.80	28.10	26.90
Ho	0.93	0.97	1.01	1.11	0.95	0.91	0.99	1.08	0.95	0.90	0.96	0.63	0.89	0.89	0.94	1.13	1.16
Er	2.69	2.83	2.92	3.21	2.78	2.65	2.84	3.13	2.76	2.61	2.70	1.82	2.41	2.46	2.55	3.20	2.90
Tm	0.42	0.45	0.46	0.51	0.44	0.41	0.44	0.49	0.41	0.39	0.37	0.25	0.37	0.37	0.38	0.47	0.36
Yb	2.51	2.70	2.74	3.02	2.64	2.46	2.62	2.94	2.47	2.35	2.47	1.71	2.37	2.35	2.45	3.17	2.03
Lu	0.37	0.40	0.41	0.44	0.39	0.37	0.38	0.43	0.36	0.35	0.36	0.27	0.35	0.35	0.36	0.45	0.28
Mg [#]	57.49	56.63	57.41	57.51	58.32	57.66	57.83	58.48	57.71	61.26	59.37	58.18	60.08	60.36	59.20	61.22	62.03
Eu/Eu*	0.95	0.83	0.94	0.91	0.79	0.92	0.92	0.95	0.98	0.97	1.01	0.93	0.90	0.91	0.94	0.98	0.99
ΣREE	68.44	106.26	74.77	81.47	102.56	66.00	71.22	79.67	68.72	64.97	59.44	72.69	69.43	71.89	73.91	37.00	254.66
(La/Yb) _N	2.57	4.73	2.52	2.52	4.57	2.41	2.45	2.54	2.68	2.68	2.04	4.61	3.39	3.54	3.43	0.40	17.28
(La/Sm) _N	1.51	2.35	1.48	1.50	2.30	1.41	1.41	1.52	1.56	1.57	1.39	2.17	2.24	2.22	2.15	0.47	3.09
Ce/Zr	0.24	0.31	0.24	0.24	0.28	0.24	0.25	0.24	0.22	0.21	0.22	0.72	0.26	0.27	0.26	0.11	0.33
Ti/Y	317	267	319	303	286	336	319	302	367	384	326	347	275	277	268	239	847
Th/Yb	0.29	1.62	0.32	0.29	1.97	0.31	0.32	0.32	0.28	0.29	0.31	0.38	1.17	1.21	1.19	0.03	3.99
Th/Nb	0.15	0.55	0.16	0.15	0.66	0.15	0.16	0.16	0.12	0.12	0.18	0.23	0.30	0.30	0.30	0.12	0.16
Zr/Nb	18.4	15.8	17.1	17.1	16.8	16.5	16.9	17.1	16.0	16.7	18.1	13.4	9.70	9.56	9.54	60.2	6.50
Nb/Yb	1.86	2.95	2.02	2.00	3.01	2.09	1.97	2.01	2.40	2.33	1.73	1.63	3.85	3.97	3.97	0.29	24.2
Dy/Yb	1.84	1.81	1.83	1.83	1.81	1.85	1.88	1.82	1.87	1.84	1.85	1.85	1.71	1.76	1.76	1.59	3.31
La/Yb	3.59	6.60	3.52	3.52	6.38	3.36	3.42	3.54	3.74	3.73	2.85	6.43	4.73	4.94	4.78	0.56	24.1
Sr/Nd	13.0	18.4	12.2	18.9	20.0	15.2	19.8	22.8	10.3	20.4	26.2	33.9	12.5	12.1	29.7	15.8	5.91

Mg[#]=100×Mg²⁺/(Mg²⁺+Fe²⁺); Eu/Eu*=Eu_N/SQRT(Sm_N×Gd)**Table 1** The concentration of major (wt. %), trace elements (ppm) and their parameters for metabasalts from Taoxinghu ophiolite.

4.2 Metamorphism and elemental mobility

Generally, major elements abundances of metamorphic rocks are inherited from the original protoliths. However, different elements exhibit different behaviors in relation to the degree of alteration and metamorphism. Some elements are relatively mobile, whereas others are relatively immobile (Guilmette et al., 2009). For example, K, Na, Si, and Ca are mobile under conditions of greenschist- and amphibolite-facies metamorphism, while Ti, Al, Mn, and P are relatively stable and Zr, Sc, and Y are very immobile under such conditions (Ghazi et al., 2012; Coish, 1977; Winchester and Floyd, 1977; Pearce and Cann, 1973). In addition, some researchers have suggested that elements with high ionic potential (e.g., K, Ba, Sr, Cs, and Rb) are mobile, while Ta, Nb, trace elements (Cr, Co, Ni, V), and REEs are stable during greenschist- and amphibolite-facies metamorphism (Mullen, 1983; Seewald et al., 1990). Petrological characteristics show that the analyzed samples of the present study experienced greenschist-facies metamorphism. Therefore, when using whole-rock geochemical data of metamorphic rocks to infer the nature and origin of rock units samples, we selected relatively immobile element concentrations and related discrimination diagrams for analysis.

4.3 Geochemistry of the Taoxinghu metabasalts

4.3.1 Major elements

Results of 17 whole-rock geochemical analyses of metamorphic basalt samples are shown in Table 1. Analyses of 16 of the samples show SiO_2 concentrations of 45.54–59.40 wt.%, Al_2O_3 concentrations of 8.18–16.84 wt.%, CaO concentrations of 4.56–11.67 wt.%, MgO concentrations of 4.25–8.69 wt.% (Table 1), and a total alkali content ($\text{Na}_2\text{O} + \text{K}_2\text{O}$) of 2.23–5.51 wt.%. Concentrations of TiO_2 are 0.91–1.34 wt.% (except for sample TG08-06), which are similar to those of basalts from back-arc basins (0.7–2 wt.%, Pearce and Peate, 1995; Gribble et al., 1996, 1998; Leat et al., 2000; Fretzdorff et al., 2002; Sinton et al., 2003). Concentrations of MnO and P_2O_5 are less than 0.5 wt.%, and loss on ignition (LOI) values are 1.28–6.36 wt.%. The $\text{Mg}^\#$ values (56–62) are less than those of primary magma (68–75, Wilson, 1989), suggesting a low degree of crystal differentiation (Table

1). We classified the rocks based on concentrations of P_2O_5 , MgO, Zr, Ti, Nb, and Y. According to the Zr/Ti vs. Nb/Y rock classification diagram of Winchester and Floyd (1977), the analyzed samples range from basaltic to andesitic, and lie close to the subalkaline rock series (except for sample TG08-06, which is an alkaline basalt) (Fig. 4a). The Zr- P_2O_5 diagram (Fig. 4b) shows that one sample plot as alkaline basalts and the others fall into the tholeiitic field.

4.3.2 REEs and trace elements

The 17 samples exhibit different rare earth and trace elements distribution patterns (Fig. 5a, b). Fifteen samples show REE distributions similar to cause of formation. Values of ΣREE are from 59.44 ppm to 106.26 ppm, with mean of 70.06 ppm; the ratio of light REE to heavy REE (LREE/HREE) is 2.62–5.25, with a mean of 3.65; HREE contents are very low (e.g., the Yb content is from 1.71 ppm to 3.02 ppm), suggesting that garnet is present in magma source residues. Chondrite-normalized REE distributions show LREE enrichment, with a negative (right-leaning) slope (Fig. 5a); differentiation of LREEs and HREEs is relatively weak; La_N/Yb_N values are 1.93–4.46, with a mean of 3.08, and there are no obvious Eu or Ce anomalies; Eu/Eu^* values are 0.79–1.01, with a mean of 0.92, indicating that plagioclase has not separated or crystallized from the melt. The REE distribution curves are generally parallel, and means for each sample show that the REEs are derived from a cognate magma. Primitive-mantle-normalized trace element

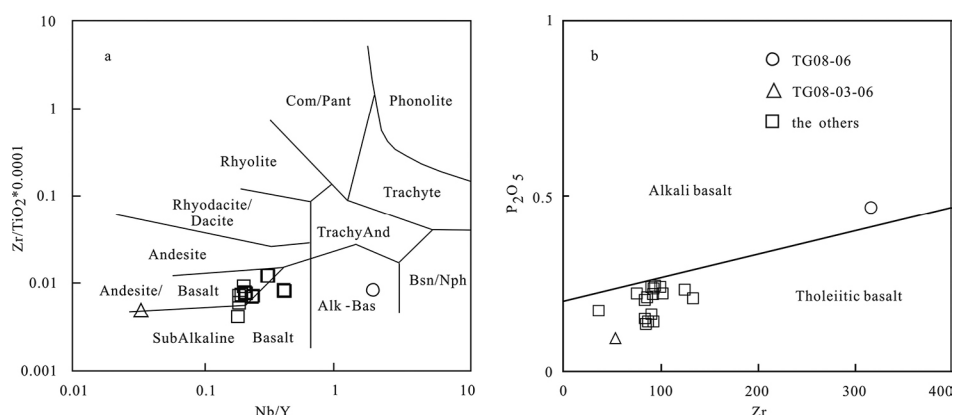


Figure 4. Geochemical classifications for the metabasalts. (a) Diagram of $\text{Zr}/(\text{TiO}_2 \times 0.0001)$ versus Nb/Y (after Winchester and Floyd, 1977). (b) Plot of Zr versus P_2O_5 . The boundary line between tholeiitic and calc-alkaline rock types is from Miyashiro (1974).

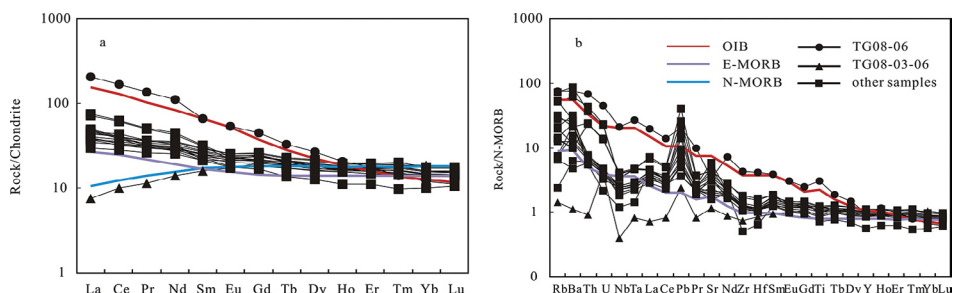


Figure 5. Chondrite-normalized REE diagrams (a) and N-MORB-normalized spider diagrams (b) for metabasalts (normalizing values are from Sun and McDonough, 1989).

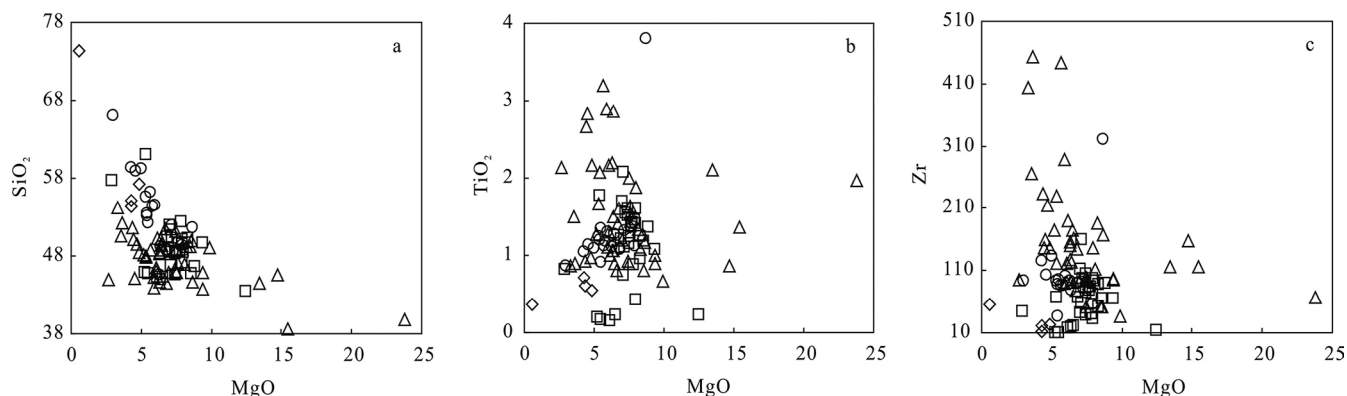


Figure 6. Mafic rocks from LLSZ ophiolites plotted in the Bowen diagrams for SiO₂, TiO₂ and Zr. The legend is shown in Figure 7.

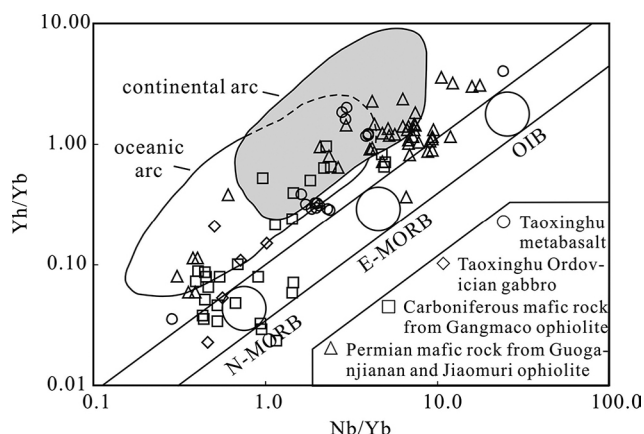


Figure 7. Th/Yb versus Nb/Yb diagram for mafic rocks from LLSZ ophiolites (after Pearce and Peate, 1995). The geochemical data are from the following sources: Taoxinghu Ordovician gabbro (Zhai et al., 2010); Carboniferous mafic rock from Ganmaco ophiolite (Zhai et al., 2013); Permian mafic rock from Guoganjianan and Jiaomuri ophiolite (Zhai et al., 2006)

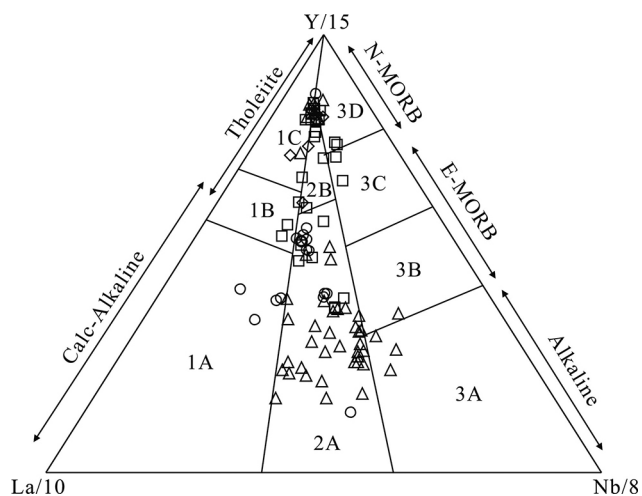


Figure 8. Ternary Y/15-La/10-Nb/8 diagram of Cabanis and Lecolle (1989). 1A, calc-alkaline basalts; 1B, overlap between 1A and 1C; 1C, volcanic arc tholeiites; 2A, continental basalts; 2B, back arc basin basalts; 3A, alkaline basalts; 3B and C, E-MORB; 3D, N-MORB. The legend is shown in Fig.7.

distributions are enriched in Rb, Th, K, Pb, and other elements, and show negative Nb, Ta, and Ti anomalies; normalized REE and trace element distribution patterns are similar to those of island arc basalts (Fig. 5b).

In sample TG08-03-06, the Σ REE content is 37 ppm, the LREE/HREE ratio is 1.04, the La_N/Yb_N ratio is 0.38, the Eu/Eu* value is 0.98, and the Ce/Ce* value is 0.99, indicating no clear fractionation between LREEs and HREEs. Chondrite-normalized REE distributions show depletion in LREEs and almost no Eu or Ce anomalies. Figure 5a shows a depletion in LREEs and flat distributions of HREEs, which are similar to normal mid-ocean ridge basalt (N-MORB)-type REE distributions (Sun and McDonough, 1989). Figure 5b shows that trace elements distributions in the samples are similar to those of N-MORBs. However, compared with N-MORBs, the U, K, and Pb contents of the samples show different degrees of enrichment, wide ranges of variation, and higher concentrations relative to other elements; these differences may be related to the strong activities of these elements or the influence of different degrees of seawater metasomatism. The contents of Nb and Ta show no depletion relative to that of Th, suggesting that the rocks did not form in an island arc setting. Values of Ce/Zr, Zr/Nb, Th/Yb, Zr/Y, Ti/Y,

and $(La/Sm)_N$ (0.11, 60.22, 0.03, 1.97, 238.91, and 0.46, respectively) are similar to those of N-MORBs (0.1, 31.76, 0.04, 2.64, 254, and 0.61, respectively) (Sun and McDonough, 1989).

In sample TG08-06 the Σ REE value is 254.66 ppm, the LREE/HREE ratio is 9.69, the La_N/Yb_N ratio is 16.28, the Eu/Eu* values is 0.98, the Ce/Ce* value is 0.96, and total REE, LREE, and HREE contents are relatively high and show clear differentiation; virtually no Eu or Ce anomalies are observed. Chondrite-normalized REE diagrams show LREE enrichment and a flat HREE distribution curve for heavier elements (Fig 5a). A primitive-mantle-normalized incompatible element distribution diagram shows strong high field strength element (HFSE) differentiation of the "big bump" type (Fig. 5b), similar to that of typical OIB distributions.

5. Discussion

5.1 Comparison with mafic rocks from other LLSZ ophiolites

This section compares the Taoxinghu metabasalt with Taoxinghu metagabbro and then with mafic rocks from other LLSZ ophiolites.

In which section it was mentioned, Taoxinghu metabasalt show a MORB-IAT geochemical characteristic, similar to suprasubduction-zone (SSZ) ophiolite defined by Dilek and Furnes (2011). This characteristic is rather different from that of Taoxinghu metagabbro, which have N-MORB affinities with LREE depleted REE patterns and negative anomalies in Nb-Ta and Ti (Zhai et al., 2010) (Fig. 5a and b). Bowen diagrams also show no relationship between Taoxinghu metabasalt and metagabbro (Fig. 6). However, in the MgO-TiO₂ and MgO-Zr diagrams, the metabasalts closely match Carboniferous and Permian mafic rocks from Guoganjianan, Gangmaco and Gemu ophiolites, which were formed on subduction zone (Zhai et al., 2013; Wu et al., 2014) (Fig. 6b and c).

The geochemical similarity of Taoxinghu metabasalt and Carboniferous and Permian mafic rocks is also well demonstrated in the Nb/Yb-Th/Yb diagram (Fig. 7). The MORB and OIB domains form a diagonal mantle array on this discriminant diagram, whereas magmas that have interacted with continental crust or have a subduction component are displaced to higher Th/Yb values (Pearce, 2008). Taoxinghu metabasalts closely occupy the space between Carboniferous and Permian mafic rocks, and mostly plot between the MORB array and compositional field for typical arc basalts and in

continental arc field. Taoxinghu metabasalts have higher Th/Yb ratios than Carboniferous mafic rocks, showing a stronger subduction-derived component influence. Compared to Permian mafic rocks, Taoxinghu metabasalts are closer to N-MORB, implying their mantle source is more depleted. On the triangular Y-La-Nb diagram, Carboniferous mafic rocks mainly have tholeiite composition and plot in the MORB and back-arc basin basalt fields, whereas Taoxinghu metabasalts and Permian mafic rocks display calc-alkaline character and are rather confined to the area represented by continental arc (Fig. 8). However, most of Taoxinghu metabasalts are closer to back-arc basin basalt field (Fig. 8). These suggest a progressive evolution of LLSZ ophiolites from MORB to IAT.

Dilek and Furnes (2011) showed that suprasubduction-zone ophiolites may evolve in extending, embryonic backarc to forearc environments, forearc settings, and both oceanic and continental backarc basins. So far, no boninites has been discovered in LLSZ. Moreover, the hybrid mixture between MORB-like and arc-like element signatures in Taoxinghu metabasalts is generally acknowledged to be unique to back-arc basin basalt (Volpe et al., 1987; Pearce and Peate, 1995; Pearce and Stern, 2006). Taoxinghu metabasalts mostly show E-MORB typed REE patterns (Fig. 5a), and mainly plot closer to the field

of for continental arc basalts than to oceanic arc basalts on tectonic discrimination diagrams (Fig. 7 and 8). These characteristics are similar to that of basalt from Okinawa back arc basin (Shinjo et al., 1999). Therefore, the metabasalts from the Taoxinghu ophiolites were probably formed in a continental back-arc and continental arc tectonic setting.

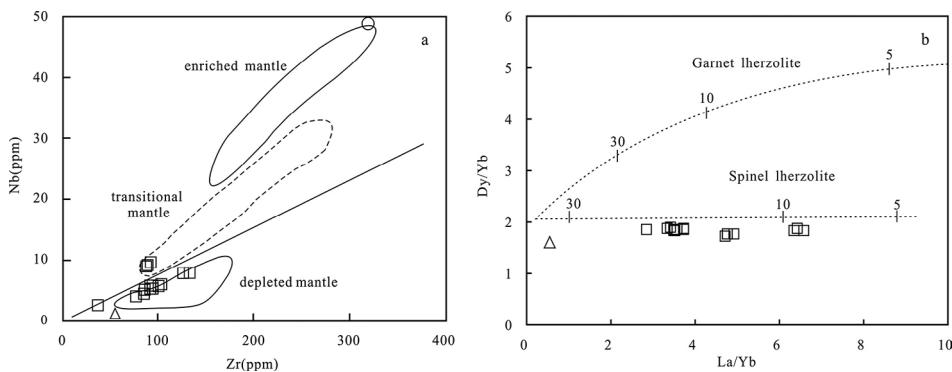


Figure 9. Zr-Nb diagram (a) (Geng et al. 2011) and Dy/Yb-La/Yb diagram (b) (Xu et al. 2001) for Taoxinghu metabasalt. The melting model, mode and partition coefficients are after Xu et al. (2001). Numbers along lines represent the degree of the partial melting.

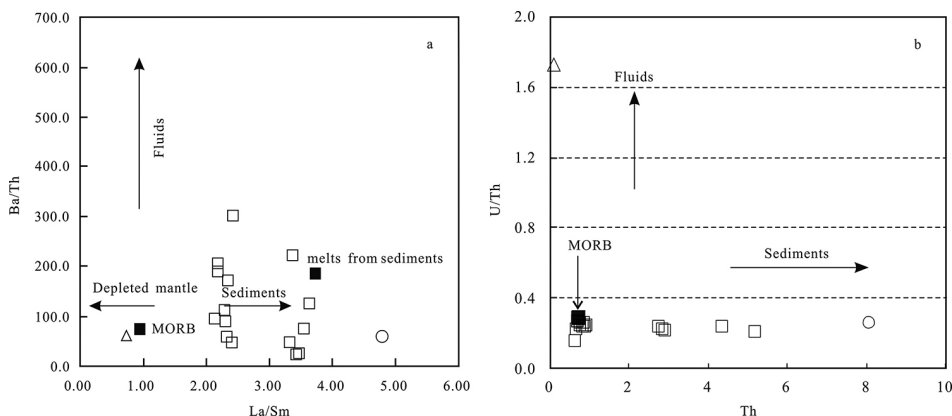


Figure 10. Incompatible element discrimination diagrams for the Taoxinghu metabasalt. (a) Ba/Th-La/Sm diagram (modified after Elliott, 2003); and (b) U/Th-Th diagram (Dilek et al., 2008). The compositions of mid-ocean ridge basalts (MORBs) and magmas resulting from melting of IBMs are given for reference (Hofmann, 1988; Hochstaedter et al., 2001).

5.2 Petrogenesis of the metabasalt

Of the 17 samples, 16 are selectively enriched in large ion lithophile elements (LILEs) and LREEs, and depleted in Nb, Ta, Ti, and HFSEs (Fig. 5b), and 15 samples (all except TG08-03-6 and TG08-06) show relatively high La/Nb ratios (2.9-7.8), Ba/Nb ratios (28.9-134.4), Ba/La (5.2-35.4) ratios, Ba/Th ratios (38-329.3), and Zr/Nb ratios (31.4-79.2), and relatively low Th/Yb ratios (0.1-1). The analyzed samples show that the contents and ratios of elements are similar to those of subduction zone-related rocks (Stern,

2002; Zhou et al., 2004). Generally, such geochemical characteristics are attributed to an enriched mantle with a low degree of partial melting, or the reaction of depleted upper mantle with a mantle plume source, or are influenced by dehydration and release of molten materials during subduction processes (Shilling et al., 1983; Donnelly et al., 2004). However, the 16 samples (all except TG08-06) are depleted in Nb and Ta, showing that the above characteristics are unrelated to the addition of material from a mantle plume source; their TiO₂ concentrations of 0.87%-1.32%, which are lower than the concentrations of a mantle plume source magma, also confirm this. In addition, analyzed samples with high Ba/La and Th/Zr ratios provide evidence for the influence of the fluid component of a subducting plate. Therefore, the enrichment of elements in the Taoxinghu metabasalt and their corresponding ratios may represent contamination by fluids or melting of subducted sediments, or replacement losses caused by the mantle source region.

Elements that are usually inactive (e.g., Zr, Nb, and Yb) were used to identify the characteristics of the magma source (Pearce and Cann, 1973). Except for sample TG08-06, which shows high Nb and Zr contents, the remaining 16 samples show low Nb contents as compared with those of Zr, and show the characteristics of a depleted mantle source (Fig. 9a). Sample TG08-06 has a very high La/Yb ratio (24.1). The La/Yb ratio of TG08-03-6, in contrast, is less than 0.56, and for the remaining 15 samples the La/Yb ratio shows relatively small variations, and, along with high La/Yb ratios (2.85-6.60), show particularly low Dy/Yb ratios (1.52-2.01) (Table 1). On a Dy/Yb-La/Yb diagram (Fig. 9b), the La/Yb value of sample TG08-06 falls outside the scope of the graph, which prevents further analysis of the extent of partial melting of primitive mantle. The diagram shows the remaining 16 samples in the depleted spinel mantle source region, although sample TG08-03-6 may be a spinel lherzolite which is a product of over 30% partial melting of a primitive mantle source. The remaining 15 samples are concentrated near the melting curve representing a spinel mantle source region, or the equivalent of a spinel lherzolite product representing ~6%-25% partial melting of a primitive mantle source.

Previous research has shown that high Ba/Th and low La/Sm ratios are indicative of a mantle source altered by oceanic crustal fluids, and that low Ba/Th ratios are attributed to the melting of sediments (Elliott, 2003). High U/Th ratios are characteristic of a mantle source region that has been influenced by aqueous fluids (Dilek et al., 2008). Therefore, the Ba/Th-La/Sm and U/Th-Th diagrams can be used to consider the influence of aqueous fluids and the contribution of subduction zone materials on the magma source (Fig. 10a and b). With the exception of sample TG08-03-6, the remaining 16 samples reflect the influence of melted sediments, while the influence of aqueous fluids appears to be minimal.

5.3 Implications for the evolution of the Paleo-Tethys in central Qiangtang

Based on an integration of geochemical data and paleontological and radiometric ages from ophiolites along the whole

LSSLZ, as well as related paleomagnetic data and geochemical data of volcanic rocks, we propose a geodynamic model for the evolution of the Paleo-Tethys Ocean in central Qiangtang.

The Taoxinghu ophiolite plagioclase granite yields zircon U-Pb ages of 497 and 492 Ma (Hu et al., 2014a). The ophiolite represents the products of partial melting of young oceanic crust. Cumulate gabbros yield a zircon U-Pb age of 467 Ma, and they have MORB-type geochemical characteristics (Zhai et al., 2010). The Guoganjianan ophiolite (240 km east of Taoxinghu) yields zircon U-Pb ages of 438, 431, and 451 Ma (Li et al., 2008; Wang et al., 2008), and it has geochemical MORB-type signatures (Zhai et al., 2007). In eastern Tibet along the eastern edge of the Longmuco-Shuanghu-Lancang suture zone, gabbro and cumulate gabbro ophiolites yield zircon U-Pb ages of 470, 439, and 454 Ma (Wang et al., 2013). These suggest that the Paleo-Tethys Ocean basin may have started to form during the Cambrian or earlier.

Paleomagnetic data from eastern Tibet-western Yunnan show that drift of the Baoshan terrane to the south and of the Yangtze terrane to the north began in the Late Ordovician (ca. 450 Ma), and that separation of the Baoshan and Yangtze terranes may have started in the early Silurian in the western Yunnan area of eastern Tibet, during early rifting of the Paleo-Tethys Ocean basin (Li, 2003).

Initial rifting of the Paleo-Tethys basin occurred at least as early as the Early Ordovician, and the record of the Paleo-Tethys in the Longmuco-Shuanghu-Lancang suture zone may extend to before the early Cambrian in middle to western Qiangtang. The Taoxinghu and Guoganjianan ophiolites yield zircon U-Pb ages of 355, 348, and 355 Ma, and geochemical signatures show that the rocks are the products of partial melting of subducted oceanic crustal material (Shi et al., 2009; stay published data). The above studies show that the Paleo-Tethys Ocean may have subducted in the early Carboniferous. Taoxinghu metabasalts mainly show geochemical characteristics of back-arc basin lavas, implying that a back-arc basin formed during subduction of Paleo-Tethys Ocean.

Results of paleomagnetic analyses show that accretion of the Baoshan and Yangtze blocks started in the early Carboniferous in the area of eastern Tibet to western Yunnan (Li, 2003). Studies on a high-pressure metamorphic belt, exposed in the southern part of the Longmuco-Shuanghu ophiolite belt and extending over 500 km, have shown that the belt formed during 240-214 Ma (Li et al., 2006, 2009; Dong et al., 2009; Zhai et al., 2009). In addition, angular unconformities at the upper contact of the Guoganjianan ophiolite show that the Longmuco-Shuanghu-Lancang suture zone closed by the time of the earliest deposition of sedimentary cover; zircon U-Pb ages of the basal rhyolite interlayer of the Wanghuling Group of 214 Ma (Late Triassic) constrain the timing of closure of the basin and the formation of the suture (Li et al., 2007). Recently, paleogeographic research on Triassic lithofacies on the Tibetan Plateau in the northern Qinghai region has confirmed that the Paleo-Tethys Ocean had closed by the begin Triassic in central Qiangtang (Zhu et al., 2013).

In conclusion, the Paleo-Tethys Ocean basin, represented by the Longmuco-Shuanghu-Lancang suture zone, may have started to form during the Cambrian or earlier, and until the early Carboniferous the basin was in a stage of expansion. Starting in the early Carboniferous, however, expansion of the basin ceased and the basin had closed by the begin Triassic in midwest Qiangtang.

6. Conclusions

(1) The Taoxinghu ophiolite metabasalts experienced greenschist-facies metamorphism, and were formed in a continental back-arc basin. The ophiolite represents the first discovery of the upper-level lavas of typical SSZ-type ophiolites in the Longmuco-Shuanghu-Lancang suture zone.

(2) The metabasalts were likely formed from a depleted spinel mantle source region, with a spinel lherzolite content equivalent to 6%-25% partial melting; one samples show spinel lherzolite contents equivalent to >30% partial melting, and one sample shows signatures of enriched mantle. The metabasalts reflect the contributions of melted sediment, while the influence of aqueous fluids may have been minimal.

(3) The LSLSZ Paleo-Tethys Ocean opened during the Cambrian or earlier, and subduction began in the early Carboniferous. As subduction proceeding, a continental back-arc basin developed. The LSLSZ Paleo-Tethys Ocean was closed in the Triassic.

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