Continuity of the North Qilian and North Qinling orogenic belts, Central Orogenic System of China: Evidence from newly discovered Paleozoic adakitic rocks

Chien-Yuan Tseng^a, Huai-Jen Yang^{a,*}, Houng-Yi Yang^a, Dunyi Liu^b, Cailai Wu^c, Chiu-Kuang Cheng^d, Cheng-Hong Chen^d, Choon-Muar Ker^a

^a Department of Earth Sciences, National Cheng-Kung University, Tainan, Taiwan, ROC

^b Beijing SHRIMP Centre, Institute of Geology, Chinese Academy of Geological Sciences, Beijing, China

^c Institute of Geology, Chinese Academy of Geological Sciences, Beijing, China

^d Department of Geosciences, National Taiwan University, Taipei, Taiwan, ROC

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ABSTRACT

Adakitic intrusive rocks of ~430–450 Ma were discovered in the North Qilian orogenic belt, the western section of the Central Orogenic System (COS) in China. These adakitic rocks were lower crust melts rather than slab melts as indicated by their crustal Ce/Pb, Nb/U, Ti/Eu, and Nd/Sm ratios and radiogenically enriched ($^{87}\text{Sr}/^{86}\text{Sr}$)_i of 0.7053–0.7066 and $\varepsilon_{\text{Nd}}(t)$ of –0.9 to –1.7. While they are all characterized by low Yb (<1.1 ppm) and Y (<11.5 ppm) abundances with high Sr/Y (>65) and (La/Yb)_N (>13.7) ratios, these adakitic rocks are classified into the low-MgO–Ni–Cr and high-MgO–Ni–Cr groups. The low-MgO samples were derived from partial melting of thickened lower crust, whereas the high-MgO samples were melts from delaminated lower crust, which subsequently interacted with mantle peridotite upon ascent. Adakitic rocks from the adjacent North Qinling orogenic belt also originated from thickened lower crust at ~430 Ma. In addition, the North Qilian and North Qinling orogenic belts both consist of lithological assemblages varying from subduction-accretionary complexes at south to central arc assemblages, which include adakitic rocks, then to backarc phases at north. Such a sequence reflects northward subduction of the Qilian and Qinling oceans. In these two orogenic belts, the occurrence of adakitic rocks of common origin and ages together with the similarities in tectonic configurations and lithological assemblages are considered to be the evidence for the continuity between eastern Qilian and western Qinling, forming a >1000 km Early Paleozoic orogenic belt. In such a tectonic configuration, the Qilian and Qinling oceans that subducted from south possibly represent parts of the large “Proto-Tethyan Ocean”. This inference is supported by the coexistence of Early Paleozoic coral and trilobite specimens from Asia, America and Australia in the North Qilian orogenic belt. Post-400 Ma volcanic rocks occur in the North Qinling orogenic belt but are absent in the North Qilian orogenic belt, indicating that these two orogenic belts underwent distinct evolution history after the closure of the Proto-Tethyan Ocean (~420 Ma).

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1. Introduction

China is tectonically composed of the South China block and Tibet Plateau on the south and the North China and Tarim blocks on the north (Fig. 1a). Between these northern and southern tectonic units is the WNW trending Central Orogenic System (COS) extending from the eastern Sulu through Dabie, Tongbai, Qinling, and Qilian to the western Kunlun orogenic belts (Fig. 1a; Yang et al., 2003). The relationships between these orogenic belts provide critical constraints for reconstructing the tectonic evolution of eastern Asia and the

Gondwana continent (e.g., Metcalfe, 2006; Han et al., 2009). The tectonic framework of the Sulu and Dabie orogenic belts is well established on the wide occurrence of ultrahigh pressure eclogites and gneisses resulted from the collision between the North and South China blocks at ~220 Ma (e.g., Ayers et al., 2002; Katsube et al., 2009; Zhang et al., 2009). Based on SHRIMP zircon ages and tectono-stratigraphy, Zhai et al. (2007) extended the Sulu–Dabie orogenic belt to the Korean Peninsula. In contrast, tectonic models for the western COS are mostly limited to individual orogenic belts (e.g., Ratschbacher et al., 2003; Wang et al., 2003; Li et al., 2007; Ye et al., 2008). An important result is that the Kunlun and Qilian orogenic belts represent a suture zone from the closure of the “Proto-Tethyan Ocean” (Bian et al., 2004; Tung et al., 2007). However, the eastern extension of this suture zone remains unclear due to insufficient geological, petrological and geochronological constraints on the relationships between

* Corresponding author. Department of Earth Sciences, National Cheng-Kung University, Tainan, Taiwan 70101, ROC. Tel.: +886 6 2757575x65429; fax: +886 6 2740285.

E-mail address: hjyang@mail.ncku.edu.tw (H.-J. Yang).

Fig. 1. (a) Simplified geological map of the north Qilian and Qinling orogenic belts, NW China (modified from Wang et al., 2005a; Chen et al., 2006), and sampling localities. (b) Detailed geological map enlarged from the rectangular area in (a). The contact zone (described in text) that separates the Northern and Southern terranes is roughly in between the Precambrian–Paleozoic contact at the south and the Leigongshan–Laohushan–Shennmutou lineament at north.

these western orogenic belts, hindering the development of an integrated tectonic model.

In this paper we identify adakitic rocks from the North Qilian orogenic belt and propose that they are petrogenetically and geochronologically related to the adakitic rocks from the easterly adjacent North Qinling orogenic belt. This result together with similarities in tectonic configurations and major constituent lithological assemblages leads to an interpretation that the North Qilian and North Qinling orogenic belts were an undivided tectonic unit formed by the northward subduction of the “Proto-Tethyan Ocean” during Early Paleozoic time.

2. Geological setting

The Qilian Mountain System forms a ~300 km wide and ~1200 km long WNW-trending belt at the northern margin of the Tibetan Plateau (Fig. 1a). It is composed of the North Qilian orogenic belt and the South Qilian block (Fig. 1a). The South Qilian block consists of Precambrian basement overlain by Paleozoic sedimentary sequence (Song et al., 2006). The North Qilian orogenic belt has been proposed to be an island arc system formed by northward subduction of an ancient ocean, the Paleo-Qilian ocean (Xu et al., 1994; Yang et al., 2002). Its eastern section, which bounds to the western end of the

Qinling orogenic belt with an ambiguous tectonic relationship (Fig. 1a), is divided into the Northern and Southern terranes by a <10 km wide contact zone (Fig. 1b; Yang et al., 2002; Wang et al., 2005a). The Southern terrane represents the remnant of oceanic lithosphere, whereas the Northern terrane is composed of arc assemblages (Yang et al., 2002). The contact zone consists mainly of Early Paleozoic volcanic and sedimentary assemblages, which underwent epidote–amphibolite facies metamorphism (Yang et al., 2002). The studied samples were collected from three felsic intrusive bodies at the east of the Northern terrane (Fig. 1b). Among them, the Shenmutou outcrop is the largest with an area of ~50 km², whereas the Leigongshan and Daiqiangu outcrops are <10 km². These outcrops are lithologically homogeneous without obvious mineral layering.

3. Sample descriptions

Samples were collected from Daiqiangu, Leigongshan, and Shenmutou along the Laohushan range (Fig. 1b). All the rocks are coarse-grained. The Daiqiangu and Leigongshan samples are tonalites, consisting of 40–50% quartz and 60–50% plagioclase. The plagioclase grains (An_{33–42}) are euhedral and slightly saussuritized.

The Shenmutou samples contain 20–30% quartz, 40–30% alkali feldspar and 40–30% plagioclase; therefore, are classified as quartz monzonite. The plagioclase (An_{13–18}) is euhedral to subhedral and is commonly zoned. The alkali feldspar is perthite. In all samples, rounded quartz grains mostly occur as an interstitial phase. Accessory minerals, including hornblende, biotite, titanomagnetite and ilmenite, are <10%. Epidote, chlorite, titanite and prehnite occur as secondary minerals. The slightly uralitized hornblende grains are euhedral–subhedral, containing small inclusions of quartz and plagioclase. The biotite has Mg# of 60–62 and is commonly replaced by chlorite and prehnite. The rim of titanomagnetite is sometimes altered to titanite and chlorite. Samples were analyzed for bulk major and trace element abundances (Table 1) and Sr and Nd isotope ratios (Table 2). In addition, zircon grains in two samples were separated for U–Pb age determination (Table 3).

4. Analytical methods

4.1. Bulk major and trace element abundances

Major element concentrations were determined by X-ray fluorescence (XRF). Abundances of trace elements were measured by

Table 1
Major and trace element abundances for Qilian adakitic rocks.

Sample	SMT-01	SMT-02	SMT-05	SMT-07	SMT-12	LGS-07	LGS-10	LGS-15	D38
<i>Major element concentrations in wt.%</i>									
SiO ₂	66.20	67.40	66.88	66.39	68.38	62.59	63.51	64.19	65.57
TiO ₂	0.38	0.40	0.40	0.36	0.31	0.35	0.40	0.28	0.32
Al ₂ O ₃	16.25	16.26	16.28	16.12	15.89	17.06	17.23	17.91	17.11
FeO ^t	2.75	2.58	2.99	2.83	2.37	4.23	4.55	4.20	3.90
MnO	0.05	0.05	0.05	0.05	0.04	0.07	0.09	0.07	0.10
MgO	2.15	2.30	2.32	2.01	1.87	1.50	1.72	1.36	1.55
CaO	2.52	3.79	3.56	3.87	3.46	4.90	6.46	4.47	5.92
Na ₂ O	5.65	4.96	4.64	5.09	4.96	3.12	3.16	2.66	3.45
K ₂ O	2.46	1.99	1.95	1.80	1.88	1.37	1.21	1.42	1.06
P ₂ O ₅	0.18	0.18	0.19	0.18	0.16	0.10	0.12	0.11	0.10
Total	98.59	99.92	99.27	98.70	99.32	95.30	98.45	96.66	99.07
L.O.I.			0.67			3.21		2.61	
<i>Trace element concentrations in ppm</i>									
Ni	38	39	43	39	35	11	3	20	8
Cr	51	54	77	52	42	25	9	40	9
V	58	69	68	66	55	91	97	86	82
Ba	974	826	735	809	715	713	776	841	801
Rb	84.7	55.2	58.9	51.7	53.1	47.9	41	62.6	34.9
Sr	696	756	713	766	685	654	735	745	772
Zr	114	92	96	116	81	47	74	61	54
Y	8.78	11.31	7.61	7.73	6.79	5.69	6.95	8.64	6.20
Nb	6.61	9.00	6.04	5.55	5.35	2.00	2.47	2.90	2.43
Cs	1.93	2.97	2.57	3.05	2.31	2.54	2.33	1.91	1.00
Sc	8.5	8.9	8.8	7.3	6.4	8.9	10.0	9.4	9.1
Hf	2.50	2.59	2.65	2.24	2.39	1.37	1.41	1.65	1.69
Ta	1.68	0.76	0.37	0.71	0.35	0.10	1.04	0.16	0.15
Pb	15.0	16.7	16.1	15.2	17.6	15.9	15.7	6.94	14.5
Th	8.64	11.0	9.90	8.41	9.02	2.49	6.30	8.28	12.5
U	1.57	1.69	1.71	1.20	1.26	1.32	2.18	3.18	1.77
La	23.7	28.6	22.4	16.6	17.1	9.83	18.1	25.5	30.6
Ce	41.6	68.3	52.2	43.7	31.3	17.2	30.8	43.8	48.8
Pr	4.38	6.25	5.05	3.71	3.39	2.04	3.23	4.63	4.67
Nd	16.3	23.5	17.6	13.9	12.9	8.1	12.4	15.7	16.0
Sm	2.90	4.36	2.89	2.53	2.51	1.77	2.43	2.61	2.55
Eu	0.84	1.08	0.82	0.75	0.77	0.60	0.72	0.72	0.73
Gd	2.28	3.15	2.18	1.94	1.91	1.44	1.84	1.88	1.76
Tb	0.30	0.43	0.29	0.28	0.25	0.20	0.24	0.26	0.22
Dy	1.66	2.24	1.54	1.40	1.35	1.05	1.37	1.55	1.15
Ho	0.30	0.41	0.28	0.27	0.26	0.21	0.26	0.32	0.22
Er	0.77	1.06	0.75	0.72	0.69	0.60	0.67	0.90	0.57
Tm	0.12	0.15	0.11	0.10	0.09	0.09	0.10	0.14	0.08
Yb	0.68	0.90	0.65	0.60	0.60	0.63	0.63	0.98	0.55
Lu	0.11	0.14	0.11	0.10	0.10	0.11	0.10	0.16	0.09

FeO^t indicates total Fe as ferrous iron. L.O.I. indicates loss on ignition. Shenmutou samples are SMT-01, -02, -05, -07, and -12. Leigongshan samples are LGS-07, -10, and -15. D38 is a Daiqiangu sample.

Table 2
Sr and Nd isotopic data for the Qilian adakitic rocks.

Sample	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm 2\sigma$	$(^{87}\text{Sr}/^{86}\text{Sr})_i$	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\pm 2\sigma$	$\epsilon_{\text{Nd}}(t)$	$\epsilon_{\text{Nd}}(0)$	$T_{\text{DM}}(\text{Ga})$
<i>Quartz monzodiorites from Shenmutou ($t = 430$ Ma)</i>										
SMT-05	0.239	0.706924	11	0.705460	0.099	0.512277	12	−1.7	−7.0	1.16
SMT-07	0.195	0.706695	12	0.705499	0.110	0.512297	11	−1.9	−6.7	1.25
<i>Tonalites from Leigongshan ($t = 453$ Ma)</i>										
LGS-07	0.212	0.707771	11	0.706403	0.132	0.512349	13	−1.9	−5.6	1.49
LGS-10	0.161	0.707607	14	0.706565	0.118	0.512332	12	−1.4	−6.0	1.31
LGS-15	0.243	0.708335	10	0.706766	0.101	0.512257	12	−1.9	−7.4	1.20

$^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios are calculated from Rb, Sr, Sm and Nd concentrations measured by ICP-MS (Table 1). The initial isotopic ratios of SMT and LGS samples were derived using emplacement ages of 430 and 453 Ma, respectively. $\epsilon_{\text{Nd}}(t)$ values were calculated using present-day ($^{147}\text{Sm}/^{144}\text{Nd}$)_{CHUR} = 0.1967 and ($^{143}\text{Nd}/^{144}\text{Nd}$)_{CHUR} = 0.512638. T_{DM} values were calculated using present-day ($^{147}\text{Sm}/^{144}\text{Nd}$)_{DM} = 0.2137 and ($^{143}\text{Nd}/^{144}\text{Nd}$)_{DM} = 0.51315.

inductively-coupled plasma mass spectrometry (ICP-MS), except for Ni, Cr and V concentrations, which were analyzed by XRF. All analyses were carried out at the Washington State University, U.S.A. following the procedures of Johnson et al. (1999). The accuracy and precision are <1% for XRF analysis and ~5% for ICP-MS analysis.

4.2. Sr and Nd isotope analyses

For Sr and Nd isotopic analyses, approximately 70 mg of sample powder was decomposed by HF–HNO₃ mixture. After dryness, samples were re-dissolved in 2 ml HCl (2 N), then, loaded onto the column containing AG50W-X8 resin (100–200 mesh) to separate Sr and REE. The Sr aliquot was purified by eluting through the Sr-Spec resin. The REE fraction was passed through Eichrom Ln resin to extract Nd. Approximately 250 ng of Sr and 50–200 ng Nd were loaded onto double filaments (Re–Ta), and measured by Thermal Ionization Mass Spectrometer (TIMS; Finnigan MAT262 at National Cheng-Kung University). The total procedural blanks for Sr and Nd were both

<200 pg. Mass fractionations were corrected using $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ and $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$. During the course of this study, the mean $^{87}\text{Sr}/^{86}\text{Sr}$ ratio was 0.710245 ± 11 ($n = 10$) for NBS 987 and 0.705011 ± 12 ($n = 12$) for BCR-1, whereas the mean $^{143}\text{Nd}/^{144}\text{Nd}$ ratios for LaJolla and BCR-1 were 0.511855 ± 12 ($n = 12$) and 0.512625 ± 24 ($n = 7$), respectively.

4.3. Zircon SHRIMP U–Pb dating

Two samples, SMT-05 and LGS-07 were crushed and sieved. Zircon grains were separated by heavy liquids and magnetic separator and finally purified by hand-picking. They were mounted with epoxy and ground to half of their original grain sizes then polished. Reflected-light and transmitted-light photomicrographs and cathodoluminescence images show that most zircon grains are euhedral with long axis of 80–200 μm and aspect ratios of 2–3. The separated zircon grains were dated using in-situ U–Pb isotope data obtained from SHRIMP II at Beijing SHRIMP Centre (Williams, 1998). Radiogenic lead

Table 3
U–Th–Pb SHRIMP data for zircon in two Qilian adakitic rocks.

Spot	U (ppm)	Th (ppm)	$^{232}\text{Th}/^{238}\text{U}$	$^{206}\text{Pb}^*$ (ppm)	Common Pb (%)	$^{206}\text{Pb}^*/^{238}\text{U}$	$\pm \%$	$^{207}\text{Pb}^*/^{206}\text{Pb}^*$	$\pm \%$	$^{206}\text{Pb}^*/^{238}\text{U}$ age (Ma)
<i>Quartz monzodiorite from Shenmutou</i>										
SMT-1	287	145	0.52	16.9	0.64	0.0681	2.1	0.0502	2.4	424.4 ± 8.6
SMT-2	345	188	0.56	19.8	0.17	0.0666	2.1	0.0549	1.4	415.4 ± 8.3
SMT-3	269	114	0.44	16.5	0.15	0.0711	2.1	0.0556	2.5	442.6 ± 8.9
SMT-4	436	275	0.65	26.3	0.34	0.0700	2.1	0.0523	1.9	436.3 ± 8.9
SMT-5	288	171	0.61	17.6	0.70	0.0709	2.1	0.0507	3.5	441.6 ± 8.9
SMT-6	346	189	0.56	19.8	0.66	0.0663	2.1	0.0540	2.9	413.8 ± 8.3
SMT-7	387	173	0.46	22.7	0.37	0.0680	2.2	0.0538	2.4	423.9 ± 8.9
SMT-8	371	235	0.65	22.1	0.32	0.0692	2.1	0.0526	2.6	431.4 ± 8.9
SMT-9	473	245	0.53	28.9	0.23	0.0710	2.0	0.0542	1.6	442.3 ± 8.7
SMT-10	342	233	0.70	20.2	0.75	0.0681	2.1	0.0519	3.2	425.0 ± 8.5
SMT-11	1378	231	0.17	683	0.05	0.577	2.0	0.2778	0.37	2935 ± 47
SMT-12	381	235	0.64	23.2	0.34	0.0705	2.1	0.0538	1.8	439.3 ± 8.7
SMT-13	1239	94	0.08	274	0.04	0.2570	2.0	0.0950	0.33	1475 ± 26
SMT-14	319	186	0.60	18.6	0.06	0.0680	2.1	0.0556	1.6	424.0 ± 8.5
SMT-15	296	163	0.57	17.9	1.01	0.0697	2.1	0.0533	3.5	434.3 ± 8.8
SMT-16	281	86	0.32	41.1	0.35	0.1697	2.0	0.1228	1.9	1011 ± 19
SMT-17	1211	296	0.25	451	0.06	0.4333	2.0	0.1634	0.30	2321 ± 39
<i>Tonalite from Leigongshan</i>										
LGS-1	422	83	0.20	26.7	0.26	0.0735	2.0	0.0552	2.0	456.9 ± 9.0
LGS-2	898	330	0.38	55.2	0.12	0.0715	2.0	0.0549	1.1	445.3 ± 8.7
LGS-3	1313	592	0.47	81.8	0.10	0.0724	2.1	0.0549	0.88	450.9 ± 8.9
LGS-4	559	171	0.32	35.5	0.19	0.0737	2.0	0.0557	1.6	458.4 ± 9.0
LGS-5	395	91	0.24	25.3	0.35	0.0743	2.1	0.0548	2.5	462.2 ± 9.2
LGS-6	425	118	0.29	26.6	0.10	0.0729	2.0	0.0549	1.6	453.4 ± 8.9
LGS-7	1229	462	0.39	75.7	0.14	0.0716	2.1	0.0550	0.78	445.8 ± 8.9
LGS-8	588	202	0.35	37.0	0.21	0.0732	2.0	0.0555	1.4	455.2 ± 8.9
LGS-9	292	94	0.33	18.1	0.36	0.0716	2.1	0.0550	2.3	445.8 ± 9.0
LGS-10	1238	592	0.49	78.9	0.09	0.0741	2.0	0.0554	1.0	460.9 ± 9.0

Data were obtained using SHRIMP at Beijing Centre. Radiogenic lead Pb^* corrected for common Pb using ^{204}Pb . The average $^{206}\text{Pb}^*/^{238}\text{U}$ ages for samples LGS-07 and SMT-05 are 453.4 ± 5.6 Ma and 429.9 ± 6.1 Ma, respectively. These Paleozoic ages are interpreted to be the emplacement ages. Some zircon grains from sample SMT-05 have inherited cores with $^{206}\text{Pb}^*/^{238}\text{U}$ ages of 1011–2935 Ma, which were not included in the average age.

Fig. 2. (a) Plot of REE abundances normalized to the values of C1 chondrite (Sun and McDonough, 1989). (b) Abundances of trace elements normalized to the values of primitive mantle (Sun and McDonough, 1989), then arranged in an order of increasing compatibility during melting of mantle peridotites.

was corrected for common Pb using measured ^{204}Pb . Zircon standard SL 13 (572 Ma, 238 ppm U) was used to calibrate U, Th and Pb concentrations of samples and TEMORA standard. Then, the TEMORA standard (417 Ma) was used to calibrate the Pb/U and UO/U values (Black et al., 2003). The ages of the analyzed zircons were calculated using SQUID and ISOPLOT program (Ludwig, 2000). Two sigma uncertainties are reported for the mean weighted ages (Table 3).

Fig. 3. Sr/Y vs. Y and (La/Yb)_N vs. Yb plot showing that the analyzed samples are within the fields for adakites. Subscript N indicates normalization to the chondritic value. The average continental crust compositions (Rudnick and Fountain, 1995) plot within the field of island arc. Therefore, contaminating arc magmas with continental crustal materials cannot generate the low Y and Yb contents and high La/Yb and Sr/Y ratios of the adakitic rocks.

Fig. 4. Left panels are cathodoluminescence (CL) images of selected zircon grains for SHRIMP U–Th–Pb dating with some analyzed spot indicated. Most zircon grains separated from samples SMT-05 and LGS-07 are pyramidal and prismatic crystals with magmatic oscillatory zonings. Right panels are U–Pb concordia diagrams of zircon analyses. Error bars are 2-sigma errors (at 95% confidence level).

these source-contaminated ocean ridges but resembling the crustal values (Fig. 6), further strengthening their continental affinity.

Although all the samples have similar trace element abundances and Sr and Nd isotope ratios, they are discernible in some other compositional parameters. Specifically, the Shenmutou samples have higher MgO, SiO₂, Na₂O, Cr, and Ni contents than the Leigongshan and Daiqiangu samples (Table 1). Were these differences caused by fractional crystallization as postulated for lavas from the Negros arc of the Philippines (Solidum et al., 2003)? If so, the evolved melts should contain low MgO contents for amphibole and biotite removal and low Sr contents for feldspar removal. However, Sr contents of the low MgO and high MgO samples are comparable (Table 1), inconsistent with the control of fractional crystallization.

Compositions of experimental melts provide constraints on the petrogenesis of the Qilian adakitic rocks. In the MgO–SiO₂ plot, the Leigongshan and Daiqiangu samples plot within the field of partial melts from eclogites, typical of adakitic rocks derived from thickened lower crust (Fig. 7). In contrast, the Shenmutou samples deviate from the experimental melts to higher MgO at a given SiO₂ content falling within the field for slab-derived adakites (Fig. 7). The high MgO, Ni and Cr contents in slab-derived adakites relative to that of experimental melts have been attributed to interaction with mantle peridotite upon ascent

(e.g., [Con](#)

Fig. 5. $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs. $\epsilon_{\text{Nd}}(t)$ plot comparing the Qilian and Qinling adakitic rocks with other adakitic rocks derived from subducted oceanic slab (data sources: Defant et al., 1992; Kay et al., 1993; Sajona et al., 2000) and continental lower crust (data sources: Muir et al., 1995; Petford and Atherton, 1996; Topuz et al., 2005; Wang et al., 2005b; Wen et al., 2008). Subscript i indicates initial ratio. “ t ” indicates emplacement ages and values for the Qilian and Qinling samples are described in the text, whereas those for others are in the “data sources”.

oceanic basalts and eclogites in the North Qilian orogenic belt indicates the existence of Paleo-Qilian ocean (Song et al., 2006). These ophiolites and oceanic basalts represent the ~490–550 Ma Qilian oceanic crust (Shi et al., 2004; Tseng et al., 2007), which subducted to the eclogite stability field at ~460 Ma (Song et al., 2006). Subsequently, the adakitic melts were generated during ~430–450 Ma (Table 3; Fig. 4). This evolutionary sequence is consistent with deriving adakitic melts from lower crust thickened by subduction-related compression. If the overriding lower crust extended to the depths where its adjacent subducted slab dehydrated, the slab-derived fluids could migrate to the lower crust to trigger melting (Wang et al., 2005b). Alternatively, heat for lower crust melting could be provided by underplating of earlier (~450 Ma) mantle-derived arc magmas (Wang et al., 2005a).

Being the source of the adakitic magmas, the lower crust must be mafic and deeper than the eclogite stability field, >40 km (Rapp and Watson, 1995; Petford and Atherton, 1996). The present-day lower crust beneath the eastern Qilian area does extend to the depth of ~45 km but is composed of andesitic layers (Liu et al., 2006), apparently inconsistent with the requirement of mafic sources. However, it should be noted that the present-day crustal thickness and constituent lithologies do not reflect that in the Paleozoic time.

Fig. 7. MgO vs. SiO_2 plot comparing adakitic rocks from the Qilian and Qinling orogenic belts and those proposed to derive from delaminated continental lower crust (Xu et al., 2002; Gao et al., 2004; Wang et al., 2004a,b; Guo et al., 2005;

experimental melts with low MgO, Ni, and Cr abundances, precluding interaction with peridotites (Fig. 7). The petrogenetical similarities between the Qilian and Qinling adakitic rocks suggest that these two orogenic belts, even the two larger scale mountain systems, were subjected to similar tectonomagmatic processes during Early Paleozoic time. This inference is supported by the general similarities in their tectonic configurations and lithological assemblages. Specifically, the Qilian and Qinling Mountain Systems were both formed by northward subduction of oceanic lithospheres. They are both divided into northern and southern parts by narrow ophiolitic belts (Fig. 1a), which had active margins on the north and passive margins on the south during Early Paleozoic time (Meng and Zhang, 2000; Zhang et al., 2001; Yang et al., 2002). The southern parts of these two mountain systems consist of Precambrian basement overlain by Paleozoic sedimentary–volcanic sequence (Zhang et al., 2001; Song et al., 2006). The northern parts known as the North Qilian and North Qinling orogenic belts are characterized by the occurrence of strongly deformed and slightly–moderately metamorphosed volcanic–sedimentary sequences together with mafic–ultramafic complexes, intermediate–felsic plutons, Precambrian basements and the 700–800 Ma granitoid fragments (Zhang et al., 2001; Ratschbacher et al., 2003; Wang et al., 2005a; Chen et al., 2006; Tseng et al., 2006). In summary, the overall consistencies in tectonic configurations and constituent lithological assemblages with the occurrence of adakitic rocks of similar ages and petrogenesis are an unlikely coincidence but suggest the continuity between eastern Qilian and western Qinling forming a >1000 km Early Paleozoic orogenic belt. The apparent topographic discontinuity between these two orogenic belts might reflect later (<420 Ma) tectonic activities.

The developments of the Qilian and Qinling orogens were closely related to the subduction of ancient oceanic lithospheres. Although the existence of the Paleo–Qilian and Qinling oceans has long been recognized, the tectonic architectures of these ocean basins have been debated. Xia et al. (1996) suggested that the Qilian ocean was a small oceanic basin, while Zhang et al. (1997) argued that it was a matured large ocean, probably comparable to the modern West Pacific. Following the recognition of the continuity between the Qilian and Qinling orogens, it is inferred that the adjacent ~500 Ma Qilian and Qinling oceans (Shi et al., 2004; Pei et al., 2005; Tseng et al., 2007) were parts of a large ocean basin rather than two separated small ones. This inference is consistent with paleontological evidence. Cambrian–Ordovician fossil biota from the North Qilian orogenic belt contain coral and trilobite specimens from North China, South China, North America, Siberian, Baltic Sea, and Australia (Zhou et al., 1996). The occurrence of southern continent specimens implies that the inferred ~500 Ma large ocean might correspond to the “Proto-Tethyan Ocean” that separated several small continents in the north from the larger Gondwana continent in the south. While we propose that the North Qilian and North Qinling orogenic belts were an un-divided tectonic unit during ~550–430 Ma, it is emphasized that post 400 Ma volcanic rocks occur in the North Qinling orogenic belt but are absent in the North Qilian orogenic belt. Apparently, these two orogenic belts underwent distinct evolution history since the closure of the Proto-Tethyan Ocean (~420 Ma; Stampfli and Borel, 2002).

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