Post-collisional, potassic monzonite–minette complex (Shahewan) in the Qinling Mountains (central China): $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology, petrogenesis, and implications for the dynamic setting of the Qinling orogen

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Received 30 October 2006; received in revised form 5 June 2007; accepted 22 June 2007

Abstract

The nondeformed, postcollisional Shahewan monzonite–minette complex in the North Qinling orogenic belt, central China, is composed of potassic monzonites and mafic minette dykes. The monzonites and minette dykes were coevally intruded as evidenced by a zircon U–Pb age of 211 ± 2 Ma for the monzonites and a $^{40}\text{Ar}/^{39}\text{Ar}$ age of 209.0 ± 1.4 Ma on the biotite from the minette dykes. These dates also indicate that the Shahewan complex postdated shortly the cessation of convergent deformation in the Qinling orogenic belt. The K-feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ MDD modeling result combined with the zircon U–Pb age reveals a rapid average cooling history ($11.5^\circ\text{C}/\text{Ma}$) for the complex from 211 to 150 Ma, implying a fast exhumation within a short period of time.

The minettes take features of ultrapotassic rocks. Their geochemical characteristics, such as high MgO (Mg# up to 80), low TiO_2 (0.38–0.62 wt.%) and FeO (2.52–3.42 wt.%), enrichment in LILEs and LREEs, and depletion in HFSE, together with primitive initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.70444–0.70562) and slightly negative $\varepsilon_{\text{Nd}}(t)$ (–3.4 to –2.5), suggest that they were derived from an enriched refractory lithospheric mantle. The monzonites have the similar geochemical (incompatible element enrichment, Nb, Ta, Ti depletion) and isotopic (initial $^{87}\text{Sr}/^{86}\text{Sr}$: 0.70513–0.70646; $\varepsilon_{\text{Nd}}(t)$: –0.5 to –3.6) imprints, indicating they were derived from a common source.

The dynamic setting in which the Shahewan complex formed is discussed. The Shahewan complex formed most likely in a postcollisional extensional setting. Its formation requires a high enough temperature to partially melt an enriched, refractory mantle source. Considerable uplift is necessitated for the rapid exhumation. These may be explained by convective mantle thinning or slab-breakoff mechanisms. The enrichment features in the lithospheric mantle source can be interpreted by injection of small percentage (<1%) partial melts from the convective mantle, or by addition of the metamorphic fluids from the subducted South China lithosphere.

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Keywords: The Qinling orogen; Potassic monzonite–minette complex; Petrogenesis; $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology; Dynamic setting

1. Introduction

The Qinling–Dabie orogenic belt, which separates the eastern Asian continent into the North and South China

Blocks, is crucial to understanding the evolution of the Eastern Asian dynamics. Magmatic rocks in the Qinling belt provide a record of the thermal and geochemical evolution of the deep lithospheric root in this developing orogen. The evolving tectonic regimes of orogenic belts are typically marked by changes in the composition of associated magmatism, therefore granitic intrusions in the

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Qinling belt may provide a tool to pry into the dynamic setting at the deep. Postcollisional, potassic magmatism is a common feature in many collisional orogenic belts around the world, although its origin is poorly understood (Turner et al., 1996). Clearly, the potassic rocks that postdated collision offer a potentially important window into the deep compositional structure of the Qinling orogen.

Various earlier studies have been carried out on the granitic rocks in the Qinling orogenic belt (e.g. Zhang et al., 1996; Lu et al., 1999; Sun et al., 2002a; Wang et al., 2005). However these studies remain controversial about the petrogenesis of these rocks and the dynamic context. The mechanisms proposed for the generation of these rocks are diversified and may be classified into three groups. Firstly, these rocks may have originated by partial melting of the crustal rocks but mixed with the primitive magma derived from the mantle (Wang et al., 2005). Secondly, the potassic rocks may be the products of the partial melting of the subducted crust of the South Qinling (Zhang et al., 1996). And finally, these rocks are derived from an enriched lithospheric mantle source beneath the North Qinling, triggered by slab-breakoff of the subducted South China plate (Sun et al.,

2002a) or mantle delamination (Lu et al., 1999). These different petrogenetic models reflect the diverse perspectives into the dynamic setting in which these rocks are generated. For example, some authors refer to these rocks as “syn-collisional” (Zhang et al., 1996), some name them “syn-orogenic” or “post-collisional” rocks (Sun et al., 2002a), and some attribute them to “post-orogenic” (Lu et al., 1999).

In this paper, based on experimental data, the Shahewan complex is first described according to age, location, geochemical and isotopic composition, and cooling history, then the constraints on the sources of monzonites and minettes of the complex are discussed, respectively. Finally, these observations are employed to assess the magmatism in the context of geodynamic models for the Qinling orogen during the Late Triassic time.

2. Geological setting and the Shahewan monzonite–minette complex

The Qinling orogenic belt is regarded as the western part of the Qinling–Dabie ultrahigh-pressure metamorphic belt (Fig. 1) (Zhang et al., 1995; Li and Sun,

Fig. 1. Geological sketch of the Qinling orogenic belt showing the locations of the granitoids and the Shahewan complex (compiled after Zhang, 1994; and Lu et al., 1999; and from new field observations). The insets show the major plates in China and the studied area, and the detail of the Shahewan complex and the sample sites.

1996; Meng and Zhang, 1999), and an important tectonic domain in eastern Asia (Mattauer et al., 1985; Yang et al., 1986; Zhang et al., 1989, 1995; Xu et al., 2000; Meng and Zhang, 2000). The present studies propose that the studied area of this belt could be divided into the South and North Qinling zones, and the South and North China Blocks (Fig. 1). Two sutures have been identified, i.e. the Shangdan suture between the South and North Qinling zones, and the Mianlue suture between the South Qinling zone and the South China Block (Zhang et al., 1989, 1995; Zhang, 1994; Meng and Zhang, 1999) (Fig. 1).

The tectonic evolution history of the Qinling orogenic belt is seriously disputed. Some geologists suggest two stages of development: Caledonian oceanic crust convergence and, Indosinian detachment in the upper crust (Mattauer et al., 1985; Xu et al., 2000; Zhang et al., 1989, 1995; Ratschbacher et al., 2003). Others argue that the Qinling orogenic belt was built up during the Indosian orogeny by the collision between the North and South China blocks (Sengor, 1985; Hsü et al., 1987; Reischmann et al., 1990; Kröner et al., 1993; Li and Sun, 1996; Sun et al., 2002a,b).

In the Qinling orogenic belt, granitoids are widely distributed, especially in the north Qinling zone (Fig. 1). These granitoids were intruded mainly at three separate stages. The granites at the first stage with ages of 444–357 Ma (U–Pb, Lu, 2000; Sun et al., 2002b), whose distribution is constrained to the North Qinling zone, are characterized by intensive deformation. The granites at the second stage, intruded during 220–205 Ma (U–Pb, Lu, 2000; Sun et al., 2002a) and scattered in both the South and North Qinling zones, are nondeformed. The third stage granites are contained in the North Qinling zone and the North China Block, and were formed during the late Mesozoic time.

The Shahewan monzonite–minette complex occurs on the Shangdan suture with an outcrop area of 120 km² (Fig. 1). Its country rocks include the gneiss of the high-grade metamorphosed Qinling Group, low-grade metamorphosed volcanic rocks of the Paleozoic Danfeng Group (487–397 Ma, Reischmann et al., 1990; Xiao et al., 1988), and the sandy slate of the Devonian Liuling Group. The contacts between the complex and country rocks are approximately vertical. Xenoliths of the older country rocks are commonly seen in the marginal phase.

The Shahewan complex comprised petrologically massive nondeformed medium- to coarse-grained hornblende-biotite monzonites with alkali feldspar megacrysts, most of which are mantled by plagioclase (rapakivi texture, Lu et al., 1999). The monzonites should be classified as I-type due to bearing hornblendes; mafic dykes and enclaves of minettes can be seen widely. Usually, the dykes range from a few centimetres to several metres in width and mingled with the monzonites, forming swarms of enclaves (Lu et al., 1999).

3. Analytical techniques

3.1. Major and trace elements, and Sr, Nd isotopes

Major element's abundance was determined by using X-ray fluorescence spectrometry (XRF) method with a Philips PW 1400 spectrometer at IGGCAS (Institute of Geology and Geophysics, Chinese Academy of Sciences). Analytical uncertainties range from 2% to 5%. Trace elements including REE elements were analyzed using ICP-MS method also at IGGCAS, with analytical uncertainties less than 5%.

Sr, Nd isotopic analyses were conducted at Chenggong University, Taiwan. The concentrations of Rb, Sr, Sm and Nd were obtained by using isotope dilution method; a mixed ⁸⁷Rb–⁸⁴Sr–¹⁴⁹Sm–¹⁵⁰Nd spike solution was used. Isotopic ratio measurements were made on a multicollector MAT-262 thermal ionization mass spectrometer. Measured Sr isotope ratios were normalized to ⁸⁶Sr/⁸⁸Sr = 0.1194; the ⁸⁷Sr/⁸⁶Sr of the Sr standard NBS-607 was 1.200393 ± 12 (2σ, n = 6). Measured Nd ratios were normalized to ¹⁴⁴Nd/¹⁴⁶Nd = 0.7219; the average value of the ¹⁴³Nd/¹⁴⁴Nd ratios of La Jolla standard is 0.511863 ± 7 (2σ, n = 6). Analytical uncertainties are estimated to be <0.5–1.0% for both ⁸⁷Rb/⁸⁶Sr and ¹⁴⁷Sm/¹⁴⁴Nd ratios. Blanks during the course of this study averaged 0.4 ng for Rb, 0.2 ng for Sr, 0.06 ng for Sm and 0.05 ng for Nd. For plotting on a diagram, the Sr and Nd isotope ratios were age corrected to 211 Ma.

3.2. Zircon U–Pb analyses

One sample (QM-2) from the monzonites was chosen for zircon U–Pb dating. The analytical work was performed at the Laboratory of Isotope Geology, Tianjin Institute of Geology and Mineral Resources.

The zircon fractions were separated from the crushed and ground rocks using a continuous sieve, a Wifley table, heavy liquids and a Frantz isodynamic magnetic separator. The zircons were removed stepwise and acting non-magnetically at a current of 1.7 A, and 2.5° vertical and 5° horizontal tilt. The concentrates were further split into size-fractions from which the most transparent, prismatic and inclusion-free crystals were hand-picked. The analyzed zircons were usually transparent and pale-rosy colour with a length-width ratio of 2.5:1, and well developed euhedral prisms and simple pyramidal endings. Four fractions of zircons were analyzed according to their differences in color and morphology.

The chemical procedure, including the decomposition of the crystal and the extraction of uranium and lead, was performed essentially according to the method described by Krogh (1973). A mixed ²⁰⁸Pb–²³⁵U spike was added to the solution. Both uranium and lead were loaded on Re filaments using nitric acid and phosphoric acid/silica gel techniques respectively. Both uranium and lead isotopic ratios were measured on a VG354 mass spectrometer with a Daly collector. Mass-fractionation correction of 0.1% amu^{−1} for Pb was made, based on measurements of NBS ARM 981.

Common lead in the sample was corrected for using the growth curve of [Stacey and Kramers \(1975\)](#) with the following ratios: $^{206}\text{Pb}/^{204}\text{Pb} = 16.1$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.4$ and $^{208}\text{Pb}/^{204}\text{Pb} = 35.8$. The calculation of elemental and isotopic compositions is processed using the program of PBDAT ([Ludwig, 1991a](#)) and the average mean $^{206}\text{Pb}/^{238}\text{U}$ age is calculated using the ISOPLOT program ([Ludwig, 1991b](#)). The decay constants recommended by the IUGS Subcommittee on Geochronology ([Steiger and Jäger, 1977](#)) are applied. The total blank of lead and uranium are, on average, 0.03–0.05 ng and 0.002–0.004 ng, respectively.

3.3. $^{40}\text{Ar}/^{39}\text{Ar}$ analyses

$^{40}\text{Ar}/^{39}\text{Ar}$ step-heating experiments were carried out on a biotite from the minette dyke sample MT-1 at IGCCAS (Institute of Geology and Geophysics, Chinese Academy of Sciences), and on a K-feldspar from the monzonite sample QM-2 at National Taiwan University, Taiwan.

The minerals with grain sizes between 200 and 280 μm were obtained from the usual mineral separation techniques by using heavy liquids, Frantz magnetic separator, and finally handpicked under a binocular to remove all visible impurities. Aliquots of mineral separates were wrapped in aluminum foil and stacked in quartz vial (at IGCCAS) or in an aluminum canister (at Taiwan National University) along with the neutron flux monitor GA1550 biotite (at IGCCAS) or LP-6 biotite (at Taiwan National University) which have K–Ar ages of 98.5 ± 0.8 Ma ([Spell and McDougall, 2003](#)) or 127.7 ± 1.4 Ma ([Odin, 1982](#)), respectively. The samples were irradiated at the H8 position in the 49-2 Reactor in Beijing or the VT-C position of the Tsing-Hua Open-Pool Reactor (THOR) at Tsing-Hua University, Taiwan. At both laboratories, samples were heated stepwise with a double vacuum resistance furnace. The released gas was purified with Zr–Al getters. The isotopic composition was measured using a MM5400 mass spectrometer (at IGCCAS) or a Varian-MAT GDI50 mass spectrometer (at Taiwan National University). After corrections for mass discrimination, system blanks, radiometric interference, $^{40}\text{Ar}/^{39}\text{Ar}$ ages were calculated according to $^{40}\text{Ar}^*/^{39}\text{Ar}_\text{K}$ ratios and the J value obtained by analyses of the monitors. Plateau ages were determined from three or more contiguous steps, comprising $>50\%$ of the ^{39}Ar released, revealing concordant ages at the 95% confidence level (2σ).

4. Results

4.1. Major and trace elements, and Sr–Nd isotopes

The chemical compositions of studied samples are shown in [Table 1](#). The monzonite samples are relatively primitive with SiO_2 ranging from 64% to 67%. They have high contents of alkalis, with $\text{K}_2\text{O} = 4.08\text{--}5.00\%$ and $\text{Na}_2\text{O} = 3.84\text{--}4.50\%$, but low abundances in total Fe ($\text{Fe}_2\text{O}_3 + \text{FeO}$) (3.0–4.2%), MnO (0.05–0.20%), TiO_2

(0.53–0.68%) and P_2O_5 (0.25–0.32%). Al_2O_3 ranges from 14.10 to 15.28%. Variations of some selected oxides (FeO , Al_2O_3 , TiO_2 , SiO_2 in wt.%) and trace elements (Ni, V in ppm) with MgO (in wt.%) are shown in [Fig. 2](#). Chondrite-normalized REE shows relatively smooth patterns without Eu anomalies ([Fig. 3](#)), the inclination to the right suggests enrichment of the light REE. In the spidergrams, the monzonites are enriched in incompatible trace elements such as Rb, Th, Ba ([Fig. 3](#)), and depleted in Nb, Ta and Ti.

The minettes show much lower SiO_2 contents (52.18–57.86%), but K_2O is as high as $\sim 4.6\%$. The MgO contents are in a range of 4.96–9.65, and calculated Mg# of $\sim 70\text{--}80$. In the primitive mantle normalized spidergrams ([Fig. 3](#)), the minettes also show enrichment in incompatible elements with negative Sr anomalies. The REE patterns in [Fig. 3](#) indicate that the LREE contents are enriched against HREE with weak negative Eu anomalies ([Fig. 3](#)).

Whole-rock Sr–Nd isotopic data are listed in [Table 1](#) and shown in [Fig. 4](#). It manifests that the monzonites and minettes show quite similar Sr–Nd initial isotopic ratios (calculated at $t = 211$ Ma): monzonites in a range from 0.70513 to 0.70646 for $^{87}\text{Sr}/^{86}\text{Sr}$ and from 0.51218 to 0.51234 for $^{143}\text{Nd}/^{144}\text{Nd}$, minettes in a range from 0.70444 to 0.70562 for $^{87}\text{Sr}/^{86}\text{Sr}$ and 0.51219 to 0.50224 for $^{143}\text{Nd}/^{144}\text{Nd}$, respectively. These ratios exhibit in a cluster close to the depleted mantle quadrangle ([Fig. 4](#)). Other granitoids of the Qinling belt (data from [Zhang et al., 1991](#)) are plotted in [Fig. 4](#) either for comparison.

4.2. Geochronology

The zircon U–Pb analysis data are listed in [Table 2](#) and the concordia diagram is shown in [Fig. 5](#). The data from four zircon fractions of sample QM-2 are located on the concordia curve with a tight range of $^{206}\text{Pb}/^{238}\text{U}$ ratios, yielding a weighted mean age of 211 ± 2 Ma, which is interpreted as the time of emplacement.

$^{40}\text{Ar}/^{39}\text{Ar}$ dating on a biotite from the sample MT-1 was performed in order to constrain the emplacement time of minette dykes. The biotite displays a fairly flat age spectrum with a well-defined plateau, giving an age of 209.4 ± 1.4 Ma (2σ) over 95% of $^{39}\text{Ar}_\text{K}$ released ([Table 3](#), [Fig. 6a](#)). This age is quite consistent with the zircon U–Pb age of the host monzonite. Theoretically the $^{40}\text{Ar}/^{39}\text{Ar}$ age just indicates the closure time of the Ar isotope system within the mineral during cooling and is later than its emplacement time, therefore the emplacement time of the minette dykes in the Shahewan complex should be rather close to the formation time of their host rocks. Therefore, we conclude that the minette dykes and their host monzonites of the Shahewan complex are coeval.

4.3. Thermochronology

In order to estimate the cooling rate after the emplacement of the monzonites, K-feldspar separates from sample QM-2 were analyzed for the Ar isotope distribution. The

Table 1
Chemical compositions and Sr, Nd isotopes of the Shahewan pluton

Sample	Monzonite						Minette			
	QM-2	QM-3	QM-4	QM-5	QM-6	QM-7	MT-1	MT-2	MT-4	MT-5
<i>Major elements (wt.%, by XRF)</i>										
SiO ₂	67.36	67.00	65.38	64.18	64.20	64.00	54.90	55.50	57.86	52.18
TiO ₂	0.53	0.62	0.68	0.68	0.59	0.59	0.44	0.38	0.41	0.62
Al ₂ O ₃	14.10	14.02	14.63	15.28	14.70	14.40	14.56	15.87	14.36	11.33
Fe ₂ O ₃	1.78	1.00	1.33	1.18	1.02	1.05	1.12	1.23	1.04	1.40
FeO	2.17	2.38	2.37	2.52	2.41	2.57	3.42	3.16	2.52	3.30
MnO	0.05	0.11	0.12	0.11	0.19	0.20	0.18	0.14	0.15	0.05
MgO	2.05	2.04	2.15	2.16	2.20	2.20	5.42	4.96	5.55	9.65
CaO	2.63	2.62	2.94	3.08	5.50	3.00	5.52	5.51	4.85	5.87
Na ₂ O	4.08	4.13	4.24	4.18	3.84	4.50	4.40	3.95	4.37	2.63
K ₂ O	4.18	4.08	4.14	4.51	4.18	5.00	4.70	4.02	4.96	4.62
P ₂ O ₅	0.25	0.30	0.29	0.27	0.25	0.32	0.31	0.30	0.24	0.42
LOI	0.58	0.48	0.80	1.10	0.70	1.25	3.67	3.02	2.56	7.20
Total	99.76	98.78	99.07	99.25	99.78	99.08	98.64	98.04	98.87	99.27
Mg [#]	50.7	54.5	53.5	53.6	56.0	54.6	70.3	69.1	75.5	80.3
<i>Trace elements (in ppm, by ICP-MS)</i>										
Nb	8.00	12.30	13.80	11.30	15.21	14.96	16.71	14.80	15.28	17.64
Zr	193	240	190	180	163	196	120	130	155	161
Y	12.10	11.40	14.30	13.80	13.48	11.90	23.20	14.20	21.21	22.24
Sr	749	360	430	460	676	676	284	321	351	410
Rb	113	88	100	101	110	113	84	96	105	80
Co	10.84	11.32	9.29	8.87	9.22	10.35	20.22	22.40	29.54	20.91
V	60	66	55	48	61	53	100	115	127	109
Ta	1.24	0.84	1.14	0.83	0.95	1.14	1.10	0.74	1.22	0.55
Ni	40	42	37	33	49	42	52	48	46	46
Cr	158	56	57	68	100	85	95	112	88	123
Cu	6.80	7.23	6.21	6.34	5.72	7.43	5.81	6.35	6.51	7.24
Hf		7.90	6.40	6.40	4.60	5.90	4.54	6.72	4.77	5.40
Ba	1443	970	1080	1620	1149	1523	1127	1630	1737	1517
Ga	23.13	13.13	12.46	10.72	20.31	17.90	15.61	17.52	17.98	16.87
Th	14.27	11.20	15.85	15.34	16.50	13.22	11.54	15.87	14.18	13.42
U	4.17	3.62	5.25	5.14	4.47	3.75	3.34	4.55	5.12	3.84
<i>Rare earth elements (in ppm, by ICP-MS)</i>										
La	38.60	49.40	50.40	42.30	34.77	35.05	58.30	52.90	87.04	66.27
Ce	70.5	74.6	94.7	77.4	66.7	62.9	115.2	87.6	134.3	113.5
Pr	7.85	10.60	11.70	10.40	7.48	6.75	10.89	9.21	11.73	11.33
Nd	28.1	27.5	32.9	30.2	26.1	22.4	42.3	33.7	49.2	43.5
Sm	5.22	5.68	6.94	6.26	4.42	3.89	8.28	6.75	7.98	7.16
Eu	1.08	1.29	1.39	1.42	1.49	1.55	1.71	1.06	1.14	1.43
Gd	3.85	3.67	4.32	4.20	4.04	3.74	4.48	4.31	3.96	4.27
Tb	0.60	0.57	0.67	0.63	0.52	0.65	0.58	0.54	0.66	0.62
Dy	3.12	2.51	3.28	3.14	2.58	2.40	2.38	1.96	2.31	2.04
Ho	0.58	0.54	0.69	0.66	0.49	0.44	0.48	0.45	0.66	0.57
Er	1.39	1.32	1.66	1.65	1.39	1.26	1.23	0.96	1.16	1.21
Tm	0.19	0.21	0.25	0.24	0.19	0.18	0.21	0.15	0.22	0.18
Yb	1.11	1.29	1.36	1.25	1.26	1.08	1.73	1.50	1.79	1.98
Lu	0.20	0.19	0.21	0.19	0.18	0.15	0.27	0.25	0.29	0.32
<i>Sr, Nd isotope</i>										
⁸⁷ Sr/ ⁸⁶ Sr*	0.70574	0.70513	0.70548	0.70646	0.70612	0.70523	0.70447	0.70562	0.70444	0.70492
¹⁴³ Nd/ ¹⁴⁴ Nd*	0.51218	0.51224	0.51234	0.51223	0.51221	0.51230	0.51224	0.51224	0.51219	0.51221

result displayed an age spectrum characterized by a gradient of apparent ages increasing from ~150 to ~210 Ma (Fig. 6b). The abnormally old ages of the first few steps are possibly due to the excess Ar present in the margin of the mineral (Fig. 6b). Such increasing age pattern of K-feldspar may record its cooling history between 350–150 °C, and can be modeled by the multi-domain diffusion (MDD) theory (Lovera et al., 1989, 1991) although its

validity is questioned because of the complicated microstructures within K-feldspar (Parsons et al., 1999; Reddy et al., 1999, 2001) and the potential influence from other diffusion mechanisms (Lee, 1995). A slab diffusion geometry is thought to be justified because argon loss from K-feldspar during in-vacuo step-heating experiments can be best described by volume diffusion in a slab geometry (Lovera et al., 1997). After appropriate adjustment of the

Fig. 2. Various oxide plots: ((a) FeO; (b) Al₂O₃; (c) TiO₂; and (e) SiO₂ in wt.%) and trace element plots (d, V, and f, Ni, in ppm) vs. MgO (in ppm) for minettes and monzonites from the Shahewan complex. The open circle denotes minettes and the solid circle is for monzonites.

Fig. 3. Chondrite-normalized REE patterns and Primitive-mantle (PM) normalized spidergrams for monzonites (a,b) and minettes (c,d) of the Shahewan complex. Elements are arranged in the order of decreasing incompatibility from left to right for the spidergrams. The chondrite and primitive mantle values (in ppm) are from Sun and McDonough (1989).

various model parameters, a model age spectrum can be obtained that closely matches the experimental results. The best-fit modeling parameters, such as the active energy

(E), size (ρ) and percentage of released gas (ϕ) of K-feldspar are summarized in Table 4. The modeling results are shown graphically in Fig. 7.

Fig. 4. $\epsilon_{\text{Nd}}(T)$ vs. $(^{87}\text{Sr}/^{86}\text{Sr})_i$ diagram for the Shahewan complex. Other granites with the ages ranging between 214 and 209 Ma (Zhang, 1994) are also shown for comparison.

The zircon U–Pb age and K-feldspar $^{40}\text{Ar}/^{39}\text{Ar}$ MDD modeling result of sample QM-2 show a two stage cooling history for the Shahewan complex: 17.7 °C/Ma during the emplacement time (211 Ma to 180 Ma) and 7.7 °C/Ma from 180 to ~150 Ma (Fig. 8). The average cooling rate of the Shahewan complex is ~11.5 °C/Ma, implying a fast exhumation process of it.

5. Discussion

5.1. Genesis of minette

The minettes generally have high MgO (max. 9.65 wt.%; Mg# = 70–80, Table 1), and compatible elements in olivine and clinopyroxene such as Ni (max. 52 ppm, Table 1), and V (max. 127 ppm, Table 1), suggesting a magma source in the upper mantle. The mantle source mark in these minettes is also shown by the isotopic data. The weakly enriched initial $^{87}\text{Sr}/^{86}\text{Sr}$ and slightly negative $\epsilon_{\text{Nd}}(t)$ (Table 1, Fig. 4), close to the domain of the depleted mantle in Fig. 4, favour a lithospheric mantle source. Although these dykes are also characterized by “crustal-like” trace element features, e.g., the enrichment in light rare earth (LREE) and other incompatible elements, depletion in Nb, Ta, P, and Ti (Table 1, Fig. 3), these are widely explained to have originated from an enriched lithospheric mantle source, similar to those from Sulu–Dabie region and Jiaodong Peninsula (Yang et al., 2005; Guo et al., 2004). The marked ‘negative’ anomalies of the high field strength elements, such as Nb, Ta, Ti and P in the primitive mantle-normalized element abundance plot (Fig. 3d), indicate that the magmas were derived from a source that was previously spiked in LILE and LREE by slab derived hydrous fluids and melts (Sun et al., 2002b), similar to those in the Sulu collisional belt (Guo et al., 2004).

The geochemical features such as the increasing of FeO, TiO₂, V and Ni, the decreasing of Al₂O₃, SiO₂ with decreasing MgO (Table 1, Fig. 2), suggest that the mine-

Table 2
Zircon U–Pb analytical data of the Shahewan pluton (QM-2)

Zircon fraction	Weight (μg)	U (ppm)	Pb (ppm)	Isotopic ratios						Age (Ma)	
				$^{206}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{206}\text{Pb}$	$^{206}\text{Pb}/^{238}\text{U}$	2σ	$^{207}\text{Pb}/^{235}\text{U}$	2σ	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$
(1)	40	502	31	101	0.2130	0.03305	41	0.2298	41	210	210
(2)	35	657	26	531	0.2027	0.03331	38	0.2316	38	211	212
(3)	30	759	30	655	0.1992	0.03380	42	0.2320	39	212	212
(4)	30	469	35	74	0.2254	0.03302	48	0.2293	45	209	210

Fig. 5. Concordia diagram showing data of the U–Pb analyses of zircon fractions from the monzonites of the Shahewan complex. The data locate together on the Concordia curve giving a weighted average age of 211 ± 2 Ma.

tttes may be the result of pyroxene-dominated fractionation of a mafic magma. Based on the data, therefore, this primitive magma should have low SiO_2 (<52.18 wt.%), high MgO (>9.65 wt.%), and elevated Mg-number (~80), TiO_2 (>0.62 wt.%), Ni (>52 ppm), V (>127 ppm), similar to the experiment melts from depleted peridotite (Falloon et al., 1988), indicating a dominantly refractory lithospheric mantle contribution to the magma. The high K_2O contents (4.02–4.96 wt.%) of the minettes require a potassic phase in the source region. Melts in equilibrium with phlogopite are widely considered to have significantly higher Rb/Sr, Th/U and lower Ba/Rb values than those formed from amphibole-bearing sources (Furman and Graham, 1999; Mengel and Green, 1989; Turner et al., 1992, 1996; Wang et al., 2006). Low Rb/Sr (<0.3), Th/U (<3.5) and relatively high Ba/Rb (>~4.0) ratios strongly imply that the parental magmas of the minettes formed through partial melting of an amphibole-bearing source. In addition, the high Zr/Hf (20~30) ratios also suggest that amphibole were involved in the formation of the mafic magmas, which had been observed in the nearby Sulu region (Yang et al., 2005). Therefore, it can be concluded that the minettes were produced by crystal fractionation of a magma derived from partial melting of an amphibole-bearing, refractory lithospheric mantle source.

5.2. Genesis of monzonite

The monzonites have the similar initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios to those of the minettes, indicating they were derived from a common source. The weakly enriched initial $^{87}\text{Sr}/^{86}\text{Sr}$ and slightly negative $\varepsilon_{\text{Nd}}(t)$ (Table 1, Fig. 4), close to the domain of mantle array (Fig. 4) but far from the lower crust (Qinling Group) and upper crust, favour a lithospheric mantle source contribution to the magma. In the major element vs. MgO diagrams (Fig. 2), the FeO, SiO_2 , Al_2O_3 , V and Ni of the monzonites plot along an extension of the minettes trend, implying that they

were the products of crystal fractionation of a mafic magma similar to the parental melts of the minettes. However, the absence of negative Eu anomalies (Fig. 3) in samples with lower initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios indicates that the parental magmas have not experienced plagioclase fractionation. Recent studies on granitic rocks in a number of orogenic belts have identified broad positive arrays between Rb/Sr ratios of individual granite samples and the time-integrated Rb/Sr ratios of their source rocks inferred from their model neodymium ages (Kemp and Hawkesworth, 2003). This means that the Rb/Sr ratios of granites can reflect the Rb/Sr ratios of their source rocks. Therefore the source of the monzonites would have the low Rb/Sr ratios (<0.2), also implying an amphibole-bearing source.

The incompatible element contents (Table 1 and Fig. 3) are more typical of crustal-derived melts. From these chemical features the possibility remains that these monzonites were contaminated during transport through the continental crust. However, the low initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios and absence of variation of these values with SiO_2 suggest that the parental melts of monzonites did not assimilate crustal materials apparently, and kept a closed system crystal fractionation during magma ascent. The increase of CaO and the decrease of MgO with increasing SiO_2 contents (Table 1) indicate a pyroxene-dominated fractionation process. Therefore, as discussed above, the minettes were derived from an enriched, refractory lithospheric mantle. Thus, the monzonites could have been resulted by pyroxene-dominated fractionation from a parental mafic magma derived from lithospheric mantle.

5.3. Tectonic implications

The nondeformation of the Shahewan complex indicates that it had emplaced after the cessation of convergent deformation. It is common that orogenic granites widely occur in orogens (Turner et al., 1992), which are referred to as “postorogenic” suite. According to the comprehensive studies, orogenesis along the Shangdan suture zone terminated roughly by Middle Triassic (Sengor, 1985; Hsü et al., 1987; Zhang et al., 1989, 1995; Reischmann et al., 1990; Huang and Wu, 1992; Kröner et al., 1993; Mattauer et al., 1985; Xu et al., 2000; Ratschbacher et al., 2003). The metamorphic ages of Mianlue ophiolite (242–221 Ma, Li and Sun, 1996) and the blueschist (240 Ma, Yin and Nie, 1993; 232–216 Ma, Mattauer et al., 1985) suggest that the convergent deformation ended by ~216–220 Ma in the eastern Qinling, quite similar to the ultrahigh-pressure metamorphic (UHP) rocks in the Dabie–Sulu belt (244–221 Ma) (Li et al., 2000; Chavagnac and Jahn, 1996; Hacker et al., 1998). Recently obtained U–Pb ages for the nondeformed granite suites in the south Qinling indicate that they were formed between 220 and 205 Ma (Sun et al., 2002a), immediately postdating the cessation of convergent deformation. These data verify the inference that the last collision along the Qinling belt happened at the Late Triassic time (Meng and Zhang, 2000;

Table 3

Ar isotopic analyses on the biotite from MT-1 and K-feldspar from OM-2

T(°C)	Cum. ³⁹ Ar	⁴⁰ Ar* (%)	³⁶ Ar/ ³⁹ Ar	³⁷ Ar/ ³⁹ Ar	³⁸ Ar/ ³⁹ Ar	⁴⁰ Ar/ ³⁹ Ar	⁴⁰ Ar/ ³⁶ Ar	Date (Ma)
<i>MT-1 biotite</i>								
650	.001	26.9	.9457E-01	.5617E-01	.3174E-01	.3778E+02	.3994E+03	102.2±1.8
730	.004	62.3	.4034E-01	.6732E-01	.2230E-01	.3199E+02	.7930E+03	203.0±7.7
800	.015	64.1	.3710E-01	.3988E-01	.2103E-01	.3055E+02	.8234E+03	198.3±1.4
850	.081	92.7	.5522E-02	.4403E-02	.1486E-01	.2241E+02	.4059E+04	209.8±1.4
900	.131	93.1	.5224E-02	.3821E-02	.1478E-01	.2228E+02	.4265E+04	209.3±1.4
950	.180	96.3	.2377E-02	.3279E-02	.1423E-01	.2148E+02	.9038E+04	209.7±1.4
1000	.249	98.2	.1306E-02	.2834E-02	.1399E-01	.2118E+02	.1622E+05	209.9±1.4
1050	.298	98.1	.1366E-02	.3248E-02	.1399E-01	.2118E+02	.1551E+05	209.7±1.4
1100	.314	98.6	.9733E-03	.4143E-02	.1380E-01	.2108E+02	.2165E+05	209.8±1.4
1150	.478	98.9	.7867E-03	.3341E-02	.1376E-01	.2099E+02	.2669E+05	209.6±1.4
1180	.639	98.8	.8165E-03	.3193E-02	.1386E-01	.2099E+02	.2570E+05	209.4±1.4
1220	.772	98.8	.8273E-03	.2893E-02	.1390E-01	.2100E+02	.2538E+05	209.5±1.4
1260	.860	98.8	.8704E-03	.3663E-02	.1393E-01	.2099E+02	.2412E+05	209.3±1.5
1300	.925	98.8	.8240E-03	.7955E-02	.1375E-01	.2098E+02	.2546E+05	209.3±1.4
1350	.988	98.8	.8400E-03	.7670E-02	.1372E-01	.2094E+02	.2493E+05	208.9±1.4
1500	1.000	98.6	.9530E-03	.1082E-01	.1378E-01	.2103E+02	.2206E+05	209.4±1.4

Sample mass = 25.3 mg

J-value = 0.00594060 ± 0.00004160

Integrated date = 209.2 ± 1.4 Ma

QM-2 K-feldspar

450	.001	78.7	.3225E-01	.1759E+01	.5076E+00	.4438E+02	.1376E+04	238.6±85.8
450	.003	42.1	.3246E+00	.5989E+00	.2326E+00	.1659E+03	.5110E+03	449.3±15.1
500	.006	72.4	.1197E+00	.3180E+00	.1586E+00	.1279E+03	.1068E+04	573.2±16.6
500	.010	63.6	.5297E-01	.1737E+00	.8700E-01	.4296E+02	.8110E+03	188.8±12.4
550	.015	99.9	.1735E-04	.1169E+00	.7713E-01	.5077E+02	.2927E+07	336.5±4.7
550	.025	99.8	.9745E-03	.1569E+00	.3841E-01	.2484E+02	.2549E+05	170.6±2.4
600	.034	100.0	.9861E-05	.1380E+00	.3405E-01	.3141E+02	.3185E+07	215.4±8.2
600	.047	97.2	.2127E-02	.6715E-05	.2631E-01	.2218E+02	.1043E+05	150.5±4.1
650	.059	98.2	.1128E-01	.1297E+00	.2976E-01	.2799E+02	.2482E+04	171.3±4.5
650	.078	100.0	.4599E-05	.6505E-01	.1933E-01	.2277E+02	.4951E+07	158.6±2.5
700	.095	99.9	.9179E-03	.8016E-01	.2016E-01	.2538E+02	.2765E+05	174.2±2.6
700	.120	98.8	.1670E-02	.8519E-01	.1624E-01	.2310E+02	.1383E+05	157.5±0.4
750	.142	98.6	.1211E-02	.9478E-01	.1695E-01	.2438E+02	.2014E+05	167.0±2.1
750	.175	98.2	.1419E-02	.8442E-01	.1521E-01	.2379E+02	.1677E+05	162.6±0.9
800	.202	100.0	.3258E-05	.8515E-01	.1655E-01	.2469E+02	.7578E+07	171.4±1.6
800	.240	100.0	.2317E-05	.8594E-01	.1701E-01	.2458E+02	.1061E+08	170.7±0.6
850	.268	99.7	.2509E-03	.9486E-01	.1523E-01	.2521E+02	.1005E+06	174.4±1.3
850	.305	100.0	.2375E-05	.9477E-01	.1389E-01	.2538E+02	.1069E+08	176.0±0.4
875	.325	100.0	.4443E-05	.6152E-01	.1346E-01	.2554E+02	.5748E+07	177.0±2.6
900	.346	98.1	.1713E-02	.4222E-01	.1610E-01	.2563E+02	.1496E+05	174.3±0.5
925	.368	100.0	.3805E-05	.4875E-01	.1515E-01	.2578E+02	.6774E+07	178.6±1.0
950	.392	97.1	.2539E-02	.7769E-01	.1733E-01	.2545E+02	.1002E+05	171.5±1.7
975	.416	96.4	.2888E-02	.3149E-01	.1850E-01	.2534E+02	.8776E+04	170.1±0.9
1000	.441	98.8	.1059E-02	.1112E+00	.2039E-01	.2540E+02	.2399E+05	174.1±1.9
1000	.470	99.7	.1453E-03	.8714E-01	.1982E-01	.2501E+02	.1721E+06	173.3±2.3
1025	.489	100.0	.4679E-05	.6965E-01	.1725E-01	.2523E+02	.5392E+07	175.0±0.6
1025	.514	97.2	.2370E-02	.9579E-01	.1828E-01	.2519E+02	.1063E+05	170.1±0.7
1050	.531	99.7	.1903E-03	.2305E-01	.2075E-01	.2565E+02	.1348E+06	177.4±1.2
1050	.557	99.9	.3354E-05	.4237E-02	.1725E-01	.2570E+02	.7663E+07	178.1±1.1
1050	.358	100.0	.4424E-05	.5209E-01	.1514E-01	.2607E+02	.5892E+07	180.5±1.1
1050	.607	100.0	.2908E-05	.2908E-01	.1609E-01	.2628E+02	.9035E+07	181.9±1.0
1075	.630	100.0	.3842E-05	.2703E-01	.1444E-01	.2651E+02	.6902E+07	183.4±1.4
1075	.669	99.2	.8038E-03	.1997E-01	.1337E-01	.2655E+02	.3303E+05	182.1±1.4
1075	.699	100.0	.2947E-05	.1245E-01	.1379E-01	.2679E+02	.9093E+07	185.3±0.6
1075	.745	100.0	.1902E-05	.6907E-02	.1396E-01	.2669E+02	.1403E+08	184.6±1.2
1100	.782	99.9	.2330E-05	.2330E-05	.1400E-01	.2701E+02	.1159E+08	186.7±0.5
1100	.832	100.0	.1751E-05	.3566E-01	.1613E-01	.2708E+02	.1546E+08	187.2±0.6
1100	.904	100.0	.1208E-05	.1799E-01	.1487E-01	.2720E+02	.2251E+08	187.9±0.4
1100	.965	99.2	.8002E-03	.2574E-01	.1686E-01	.2933E+02	.3665E+05	200.4±0.7
1100	.987	99.9	.4018E-05	.4018E-05	.1799E-01	.3012E+02	.7495E+07	207.0±1.8
1150	.996	100.0	.9862E-05	.7524E-01	.1104E-01	.3164E+02	.3208E+07	216.8±3.8

(continued on next page)

Table 3 (continued)

T(°C)	Cum. ³⁹ Ar	⁴⁰ Ar* (%)	³⁶ Ar/ ³⁹ Ar	³⁷ Ar/ ³⁹ Ar	³⁸ Ar/ ³⁹ Ar	⁴⁰ Ar/ ³⁹ Ar	⁴⁰ Ar/ ³⁶ Ar	Date (Ma)
1200	1.00	98.2	.1978E-02	.1521E+00	.2074E-04	.3422E+02	.1730E+05	229.8±8.0
Sample mass = 45.8 mg								
J-value = 0.00404072 ± 0.00000400								
Integrated date = 182.6 ± 0.3 Ma								

Note. Cum.³⁹Ar denotes the cumulative of ³⁹Ar during the stepheating; ⁴⁰Ar*% means the percentage of radiogenic ⁴⁰Ar at every step.

Fig. 6. ⁴⁰Ar/³⁹Ar age spectra for biotite (a) from MT-1 and K-feldspar (b) from QM-2. The width of boxes of the spectra represents the relative fraction of ³⁹Ar_k released in step heating and the height the error (2σ) of the age of the corresponding step.

Table 4				
Parameters (activation energy <i>E</i> , relative domain size, log (<i>D</i> ₀ / <i>ρ</i> ²), percentage of released gas (<i>φ</i>)) used for K-feldspars'Arrhenius modeling				
Domains	<i>E</i> (kcal/mol)	Relative domain size	log (<i>D</i> ₀ / <i>ρ</i> ²) (s ⁻¹)	<i>φ</i> (³⁹ Ar%)
1	48.5 ± 1.6	0.00145	6.84	9.82
2	48.5 ± 1.6	0.00369	5.56	13.376
3	48.5 ± 1.6	0.03810	5.55	10.877
4	48.5 ± 1.6	0.03978	3.22	4.508
5	48.5 ± 1.6	0.03984	3.21	5.518
6	48.5 ± 1.6	0.04012	3.21	13.570
7	48.5 ± 1.6	0.24031	3.13	14.549
8	48.5 ± 1.6	0.80642	2.17	14.858
9	48.5 ± 1.6	1.00000	1.34	12.925

Sun et al., 2002c). Although the time lag between magmatic pulses does not necessarily imply any lag between the end of deformation and the first appearance of nondeformed granites, these data and the zircon U–Pb ages show that the Shahewan monzonites emplaced (211±2 Ma) ~ 10 m.y. later than the collision of the Qinling–Dabie belt. Therefore, the Shahewan complex should be classified as “post-collisional” suite.

Petrological and geochemical data emphasize that the Shahewan complex was derived by partial melting of the enriched refractory lithospheric mantle source. They have geochemical features similar to postcollisional lavas emplaced in the other parts of the Tethyan orogenic belts,

Fig. 7. Experimental and MDD modeling diagrams for QM-2 K-feldspar age spectrum: (a) Experimental and theoretical age spectra, (b) calculated cooling paths with a inset showing the 90% confident interval of the total distribution of cooling histories and the 90% confident interval of the median of the distribution.

Fig. 8. Compilation of the cooling history of the Shahewan complex. The values of cooling rates are shown.

such as the Spanish lamproites (Nelson et al., 1986), Sulu ultrapotassic granitic rocks (Yang et al., 2005) and Tibetan shoshonitic rocks (Turner et al., 1996; Miller et al., 1999). The high-K magmatism is considered as a mark of postcollisional extensional setting (Bonin, 1990; Bonin et al., 1998; Turner et al., 1996).

The rapid cooling history obtained from the $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology (Fig. 8) is inconsistent with a pure conductive cooling concluded by McDougall and Harrison (1999) and thus rather reflects an important denudation event undisturbed from 211 to 150 Ma, suggesting a fast exhumation of the Shahewan complex. Absence of sedimentation in the North Qinling offers support that the denudation kept going through the Triassic time. Although the collision between the South China Block (Yangzi Block) and South Qinling, having taken place during the Late Triassic time (Meng and Zhang, 2000; Sun et al., 2002c), affected the whole Qinling belt, the nondeformed extensive granite intrusions during the Late Triassic suggest that the convergent deformation was not dominant as discussed above. During a convergent deformation, the thickening of the crust results in increased gravitational potential energy and extensional buoyancy forces (England and Houseman, 1988, 1989; Sandiford and Powell, 1990; Turner et al., 1992, 1996). It follows that the extension and uplift will occur if the convergent plate driving forces are reduced or the internal buoyancy forces exceed the external driving forces. But the amount of uplift by this isostatic adjustment is only about 1–3 km (England and Molnar, 1990), not enough to exhumate the Shahewan complex to a shallow depth in the upper crust (convert 150 °C to ~3 to ~7 km below the earth surface based on the geothermal gradient of ~20–50 °C/km supposedly). Therefore, the rapid exhumation of the Shahewan complex should result from different mechanics after the collision (see discussion later). Rapid uplift and extension during Cretaceous and Cenozoic times (Meng and Zhang, 2000) indicate that the uplift and extension occurred over a long time and was a

significant process in the North Qinling after the convergent deformation.

The rapid exhumation could cause the rapid pressure release of the emplaced magma from the overlying strata, this may provide an alternative rationale for formation of the rapakivi texture (K-feldspar megacrysts mantled by plagioclase) within the monzonite of the Shahewan complex, although it can also be explained by the magma mixing (Wang et al., 2005). The calculation indicates that the decompression can produce paragenesis that can lead to all of the features of the rapakivi texture (Nekvasil, 1991).

Several models have been proposed to explain the post-collisional magmatism, such as lithospheric delamination (Bird, 1979), convective removal of a thickened thermal boundary layer (Houseman et al., 1981) or the mantle root of the orogen (Turner et al., 1992, 1996), and the breakoff of an oceanic slab from the buoyant continental lithosphere during subduction (Davies and von Blanckenburg, 1995). The breakoff model was employed to interpret the granitic intrusions in the South Qinling following the subduction of the South China (Sun et al., 2002a), supposing the breakoff of the slab happened at a very shallow depth which disturbed the asthenosphere greatly and led to the formation of the orogenic granitoids, because small fractions of K-rich melt can be produced by fluid-absent phengite dehydration-melting of metabasalts (flush melting) (Schmidt and Poli, 2003). Zheng et al. (2003) argue that this mechanism can account for the Late Triassic potassic-ultrapotassic rocks in the Dabie–Sulu orogenic belt and thus classified them as the product of syn-exhumation magmatism. Geochemical and Sr, Nd isotopic data indicate that the Shahewan complex was derived from an enriched lithospheric mantle. In this regard, part of the subducted South China lithospheric mantle and its overlying oceanic basaltic crust after the breakoff is plausibly responsible for generating the observed geochemical and isotopic features for both minettes and monzonites.

An alternative is thinning of the lithosphere and crustal extension induced by convective stability of a thickened mantle boundary layer (Houseman et al., 1981; Turner et al., 1992, 1996). Partial melting of an enriched refractory lithospheric mantle source from which the Shahewan complex was derived requires a sufficiently high temperature. The nondeformed magmatic suites of orogens seem to take the higher magmatic temperature than those deformed orogenic suites (Turner et al., 1992). This means that a significant perturbation in the thermal regime of the orogen is necessary. The cooling history of the Shahewan complex necessitates a considerable extent of uplift which is unlikely to be produced by isostatic response only after the cessation of convergent deformation. The convection mantle thinning (Houseman et al., 1981; Turner et al., 1992, 1996) may be a plausible means of attaining a temperature high enough for melting the lithospheric mantle and providing tremendous uplift.

Thinning of the lithospheric mantle under the North Qinling orogen may be an automatic response to litho-

spheric thickening as a result from the collision. Prior to convective thinning of mantle lithosphere, the driving forces and buoyancy forces were in balance, the uplift was limited and the lithospheric mantle did not melt due to low heating. With the increase in degree of convective thinning of the mantle lithosphere, increase of the potential energy of orogen would preclude further compression deformation and induce uplift and extension. Convective thinning of the lithospheric mantle brought previously cooler and insulated zones of the lithospheric mantle in direct contact with asthenospheric temperatures such that any metasomatized refractory lithospheric rocks underwent partial melting. Thinning or even removal of the root of the mantle lithosphere would result in very high Moho temperatures, and led heat into upper lithosphere, which triggered melting there beneath the Shahewan region.

Although the composition of the subcontinental mantle lithosphere beneath the Shahewan region cannot yet be quantitatively described, the geochemical data of Shahe-wan complex indicate that the mantle lithosphere was enriched. The degree of partial melting and compositional heterogeneities may account for this enrichment. It is likely that the mantle lithosphere is enriched or veined by small percentage (<1%) partial melts from the convective mantle that are enriched in light REEs and incompatible elements (Menzies and Murthy, 1980; O'Nions and McKenzie, 1988; Hawkesworth et al., 1990; Turner et al., 1992, 1996). Parts of a mantle lithosphere previously enriched or veined by such partial melts from the asthenosphere will have a lower solidus than the subjacent convective mantle. Resultant partial melt compositions will therefore be primitive, but with light-REE and incompatible elements enriched signatures (Turner et al., 1992, 1996).

Another source for the enrichment within the lithospheric mantle beneath the Shahewan region may be from the subducted South China lithosphere during the Triassic time, which may happen near the boundary between continental and oceanic lithospheres (Sun et al., 2002a). Although the process of continental subduction is characterized by the relative lack of fluids based on H and O isotopic evidences (Zheng et al., 2003), the introduction metamorphic fluids derived from the decompression exsolution from amphibole, phengite and clinopyroxene could metasomatize and vein the overlying lithosphere.

6. Conclusions

The Shahewan monzonite–minette complex is a product of postcollisional magmatism. The zircon U–Pb dating on the monzonite and $^{40}\text{Ar}/^{39}\text{Ar}$ timing on the mafic minette dykes indicate that they emplaced coevally. $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronology on K-feldspar from the monzonite gives a two stage, rapid cooling history, suggesting a fast uplift after the cessation of the convergent deformation in the North Qinling during the Late Triassic time. Isotopic and geochemical data show that the complex was derived from

the partial melting of an enriched lithospheric mantle source. Two models, including the subduction model of the South China lithosphere beneath the North Qinling and the convective mantle thinning model, are assessed to construe the geodynamic setting of the Qinling orogenic belt during the Late Triassic time.

Acknowledgement

The authors are indebted to Dr Da-Ren Wen and Prof. Sun-Lin Chung from the Geology Department of National Taiwan University, for carefully carrying out Sr and Nd isotopic analyses. Damian Tootell is so kind for improving the English usage. We are very grateful to W.D. Sun and an anonymous reviewer for their thorough and helpful reviews that have significantly improved the manuscript. This work is supported by the Chinese National Natural Science Foundation (Grants No. 40673048).

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