Neoproterozoic active continental margin of the Cathaysia block: Evidence from geochronology, geochemistry, and Nd–Hf isotopes of igneous complexes

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ABSTRACT

Previous studies have suggested that northwest Cathaysia was a passive continental margin during Neoproterozoic time, given the absence of Neoproterozoic magmatism. However, this view has been challenged by recent discoveries of mafic and felsic magmatic rocks with geochemical affinities to continental arc or back-arc volcanic rocks from the 1.0–0.9 Ga Wuyi–Yunkai belt, indicating the development of an early Neoproterozoic continental margin arc and back-arc system in the Cathaysia block. In this paper we present new age data for the Miaohou and Shanhou arc complexes, demonstrating that the active continental margin of the Cathaysia block lasted until ca. 830 Ma. The Miaohou complex is composed of gabbro, diorite, and granite, the Shanhou complex of diorite and granite, and both complexes are bound by the Jiangshao Fault. Twelve samples from the Miaohou and Shanhou complexes give zircon U–Pb ages of ca. 830 Ma, indicating that the magmatism of the Miaohou and Shanhou complexes were coeval. The Miaohou gabbro has high zircon $\varepsilon_{\text{Hf}}(t)$ (+2.8 to +9.6) and whole-rock $\varepsilon_{\text{Nd}}(t)$ (+3.9) values together with the geochemical features of volcanic arc basalts, while the Miaohou and Shanhou diorites and granites show large variations and high values of zircon $\varepsilon_{\text{Hf}}(t)$ and $\varepsilon_{\text{Nd}}(t)$, indicating assimilation and fractional crystallization by subduction-related fluids, partial melting of the mantle wedge, followed by varying degrees of assimilation and fractional crystallization (AFC) as the magmas ascended. These processes provide a reasonable explanation for the Neoproterozoic arc magmatism on the northeastern margin of the Cathaysia block, represented by the Miaohou and Shanhou complexes.

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1. Introduction

South China is an important geological region in eastern Asia, comprising the Yangtze Block to the northwest and the Cathaysia block to the southeast (Fig. 1a). These two blocks are thought to have amalgamated in the early Neoproterozoic along the Pingxiang–Jiangshan–Shaoxing Fault, and the collision of the two blocks gave rise to the Jiangnan Orogeny (also called the “Sibao” or “Jinning” Orogeny) (Fig. 1a) (Charvet, 2013; Zhang and Zheng, 2013; Zhao, 2015; Zhao and Cawood, 2012; Zheng et al., 2013). Although the timing of the collision remains controversial, most researchers believe that the final collision of the Yangtze and Cathaysia blocks occurred in the early Neoproterozoic, with ages either in the range 900–880 Ma (e.g., Greentree et al., 2006; Li et al., 1995, 2002, 2007, 2008a, 2009; Ye et al., 2007), correlating the Neoproterozoic Jiangnan Orogeny in South China with the global Grenvillian-aged orogenic events (Li et al., 2002, 2008b), or in the range 860–820 Ma (e.g., Chen et al., 2014; Shu, 2012; Wang et al., 2006, 2007, 2012a, 2012b, 2013a, 2014a; Yao et al., 2013, 2014a; Zhao, 2015; Zhao and Cawood, 1999; Zhao and Cawood, 2012; Zhao et al., 2011; Zhou et al., 2002), due to the existence of ca. 860–820 Ma arc rocks in both the western and eastern segments of the Jiangnan Orogen (Chen et al., 2014; Wang et al., 2012a, 2014a; Yao et al., 2013, 2014a).

The main geodynamic scheme of the South China amalgamation, as proposed by previous researchers, can be summarized as follows: the closing of an ocean basin that existed in the
Neoproterozoic through its northwards subduction beneath the Yangtze Block and the subsequent collision of the Yangtze and Cathaysia blocks, giving rise to the active continental margin of southeast Yangtze, the passive continental margin of northwest Cathaysia, and the Jiangnan Orogeny (e.g., Li et al., 2009; Wang et al., 2012a; Yao et al., 2013, 2014a; Ye et al., 2007). However, recent reports of magmatic rocks along the Wuyi–Yunkai belt (Table 1) have challenged such a passive continental margin model for the Cathaysia block. The evidence of magmatism includes: (1) calc-alkaline metamafic rocks at ca. 984–969 Ma along the Wuyi–Yunkai belt, with geochemical affinities to continental arc or back-arc basalts (Wang et al., 2013b; Zhang et al., 2012); (2) felsic rocks in Nanling and Yunkai with the geochemical features of continental arc acidic volcanic rocks, and which are represented by the 970 ± 8 Ma Jingnan volcanic rocks dominated by rhyolites and rhyolitic graywackes (Shu et al., 2008a); (3) ca. 982–909 Ma peraluminous granites in the Wuyi–Yunkai belt which might be linked petrogenetically to a post-collisional setting in an early arc/back-arc regime (Wang et al., 2013b); and (4) the Chencai hornblende gneiss with a protolith that is considered to be the product of subduction magmatism at 879 ± 10 Ma (Yao et al., 2013). All these observations indicate that during Neoproterozoic time, an active continental margin, with arc and back-arc magmatism, existed along the northwestern margin of

Fig. 1. (a) Geological sketch map of the distribution of representative Neoproterozoic rocks in South China (modified after Cawood et al., 2013; Wang et al., 2013b). (b) Geological map of the Miaohou and Shanhou complexes. Sample locations are also indicated.
Table 1  
List of reliable ages of Neoproterozoic igneous rocks in Cathaysia block.

<table>
<thead>
<tr>
<th>Batholith or group</th>
<th>Rock type</th>
<th>Age (Ma)</th>
<th>Method</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chencai</td>
<td>Hornblende gneiss</td>
<td>879 ± 10 Ma</td>
<td>LA-ICP-MS zircon</td>
<td>Yao et al. (2014b)</td>
</tr>
<tr>
<td></td>
<td>Basalt</td>
<td>857 ± 7 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td>Shu et al. (2011)</td>
</tr>
<tr>
<td>Chatian</td>
<td>Amphibolite</td>
<td>969 ± 13 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td>Wang et al. (2013b)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>984 ± 6 Ma</td>
<td>LA-ICP-MS zircon</td>
<td></td>
</tr>
<tr>
<td>Zhuji</td>
<td>Ophiolitic gabbro</td>
<td>858 ± 11 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td>Shu et al. (2006)</td>
</tr>
<tr>
<td>Plagioclase amphibolite</td>
<td></td>
<td>845 ± 10 Ma</td>
<td>*Ar/Ar hornblende</td>
<td>Shui et al. (1986)</td>
</tr>
<tr>
<td>Tehinite</td>
<td>Amphiolite pyroxenite</td>
<td>832 ± 7 Ma</td>
<td>*Ar/Ar hornblende</td>
<td>Kong et al. (1995)</td>
</tr>
<tr>
<td></td>
<td>Huangshan quartz diorite</td>
<td>818 ± 6 Ma</td>
<td>LA-ICP-MS zircon</td>
<td>Wang et al. (2012b)</td>
</tr>
<tr>
<td>Zhangcun</td>
<td>Metaphyolite</td>
<td>838 ± 5 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td>Li et al. (2010)</td>
</tr>
<tr>
<td>Lipu</td>
<td>Gabbro-diorite</td>
<td>841 ± 6 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td></td>
</tr>
<tr>
<td>Miaoou</td>
<td>Gabbro, diorite, granite</td>
<td>828 ± 11 – 834 ± 14 Ma</td>
<td>LA-ICP-MS zircon</td>
<td>This study</td>
</tr>
<tr>
<td>Shanhou</td>
<td>Diorite, granite</td>
<td>830 ± 12 – 832 ± 9 Ma</td>
<td>LA-ICP-MS zircon</td>
<td></td>
</tr>
<tr>
<td>Shibun</td>
<td>Metadiabase</td>
<td>978 ± 11 Ma</td>
<td>Sims U-Pb zircon</td>
<td>Wang et al. (2013b)</td>
</tr>
<tr>
<td></td>
<td>Gabbro</td>
<td>970 ± 10 Ma</td>
<td>LA-ICP-MS zircon</td>
<td>Shu et al. (2011)</td>
</tr>
<tr>
<td>Miamianshan Group</td>
<td>Gabbro</td>
<td>847 ± 8 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Gabbro</td>
<td>836 ± 7 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td></td>
</tr>
<tr>
<td>Wanquan Group</td>
<td>Metavolcanics</td>
<td>788 ± 27 – 800 ± 14 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td>Li et al. (2010)</td>
</tr>
<tr>
<td>Zhenghe</td>
<td>Diabase</td>
<td>795 ± 7 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td>Shu et al. (2008b)</td>
</tr>
<tr>
<td>Masba</td>
<td>Metagranite</td>
<td>909 ± 10 – 916 ± 6 Ma</td>
<td>SIMS and LA-ICP-MS U-Pb zircon</td>
<td>Wang et al. (2014b)</td>
</tr>
<tr>
<td>Xunwu Group</td>
<td>Metagranite</td>
<td>963 ± 11 – 982 ± 27 Ma</td>
<td>LA-ICP-MS zircon</td>
<td></td>
</tr>
<tr>
<td>Jingnan</td>
<td>Rhyolite</td>
<td>972 ± 8 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td>Shu et al. (2008a)</td>
</tr>
<tr>
<td>Yunkai Group</td>
<td>Plagioclase amphibolite, Metagabbro</td>
<td>824 ± 77 Ma</td>
<td>Sm-Nd isochron</td>
<td>Peng et al. (2006)</td>
</tr>
<tr>
<td></td>
<td>Metabasalt</td>
<td>980 ± 8 Ma</td>
<td>LA-ICP-MS zircon</td>
<td>Wang et al. (2013b)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>978 ± 19 Ma</td>
<td>SHRIMP U-Pb zircon</td>
<td>Zhang et al. (2012)</td>
</tr>
<tr>
<td></td>
<td>Metagranite</td>
<td>926 ± 28 – 954 ± 14 Ma</td>
<td>LA-ICP-MS zircon</td>
<td>Wang et al. (2014b)</td>
</tr>
</tbody>
</table>

Cathaysia (Cawood et al., 2013; Wang et al., 2013b, 2014a; Zhao, 2015).

However, most of the previous studies were limited in scope, and more precise ages and geochemical data are required before the tectono-magmatic evolution of this newly recognized Cathaysian active continental margin can be elucidated. We have therefore conducted a detailed geochronological and petrological study on the Miaoou and Shanhou complexes in order to better understand the formation, growth, and evolution of the Neoproterozoic active continental margin of the Cathaysia block.

2. Geologic background and samples

Unlike the Yangtze Block with its Archean–Paleoproterozoic crystalline basement, the Cathaysia block contains almost no exposures of Archean basement, and the Paleoproterozoic migmatic, granitoid gneisses, and amphibolites, which only crop out in the NW Zhejiang and NE Fujian provinces, form most of the basement (e.g., Fujian BGMR, 1985; Xia et al., 2012; Yu et al., 2009, 2012; Zhejiang BGMR, 1989; Zheng et al., 2011). However, the Neoproterozoic volcano-sedimentary sequences (Fig. 1a), which were metamorphosed under the greenschist to lower amphibolite facies, are more widespread in a number of areas where they are known by a variety of names: the Chencai Group in central Zhejiang, the Longquan Group in southwest Zhejiang, the Mayuan and Miamianshan groups in northwest Fujian, the Shenshan Group in southern Jiangxi, the Taoxi Group in northeast Guangdong, and the Yunkai Group in central-west Guangdong (Charvet et al., 2010; Zhao, 2015; Zhao and Cawood, 2012). These metavolcanic rocks represent basaltic and rhyolites, with minor andesites and dacites, and some display arc affinities in terms of their geochemical compositions. In addition, small amounts of amphibolite, metabasalt, and metagabbro, and even ophiolitic mélanges are exposed sporadically, or occur as lenses, pods, and fragments, which might represent the relics of a Neoproterozoic subduction system in the Cathaysia block (e.g., Shu, 2012; Shu et al., 2006, 2008a, 2008b, 2011; Wang et al., 2011, 2013b, 2014b; Yao et al., 2014b; Zhang et al., 2012; Zhao, 2015).

The Miaoou and Shanhou complexes are located to the north of Jinhua city in western Zhejiang Province, and the complexes crop out over areas of ~2.84 km² and ~27.2 km² respectively. Due to the Quaternary sedimentary cover, their relationships with other adjacent geological formations are not clear. The Miaoou complex is composed of granite, mainly found in the east of the complex, diorite in the middle and west (and scattered in the east), and gabbro in the middle of the complex. The Shanhou complex lies to the east of the Miaoou complex, and is made up mainly of diorite, with minor granite scattering around the margins. Samples were collected from the Miaoou and Shanhou complexes for this study (Fig. 1b), and sample locations (latitudes and longitudes) and the mineral assemblages of the samples are listed in Table 2. The Miaoou and Shanhou granites have similar mineral assemblages of K-feldspar + quartz + plagioclase + biotite. Accessory minerals are apatite, magnetite, Fe–Ti oxides, zircons, and sometimes sericite. The Miaoou granite is fine- to medium-grained with micrographic intergrowths of quartz and alkali feldspar, while the Shanhou granite is porphyritic; a high-temperature and low-pressure environment of crystallization is indicated for these two granites. The Miaoou and Shanhou diorites are fine- to medium-grained and contain plagioclase + quartz + K-feldspar + biotite + amphibole; the accessory minerals are epidote, titanite, apatite, zircons, magnetite, and Fe–Ti oxides. The Miaoou gabbro is fine- to medium-grained and dominated by the mineral assemblage amphibole + plagioclase + clinopyroxene + orthopyroxene + quartz with accessory titanite, Fe–Ti oxides, zircons, and apatite. Representative samples were analyzed for whole-rock major and trace elements as well as whole-rock Sr–Nd isotope compositions, and zircons separated from the samples were analyzed for trace elements, U–Pb dating, and Hf–isotope compositions.

3. Analytical methods

3.1. U–Pb dating and trace elements of zircons

Zircons were extracted using standard density and magnetic separation techniques. Random zircon grains were handpicked...
under a binocular stereomicroscope and mounted in a 1.4 cm diameter epoxy disk, and polished to expose the central parts of the grains. In order to characterize the internal structures of the zircons and to choose appropriate target sites for U–Pb and Hf isotope analyses, cathodoluminescence (CL) images were obtained using a JEOL JSM-7000F scanning electron microscope at State Laboratory of Microstructures, Nanjing University. The operating conditions for the CL imaging were at 15 kV and 20 nA.

Zircon U–Pb dating were carried out with or without simultaneous determination of trace elements using an Agilent 7500a ICP-MS equipped with a New Wave 213 nm laser sampler in the State Key Laboratory of Mineral Deposits Research, Nanjing University. Details of instrument settings and analytical procedures follow Jackson et al. (2004). Analyses were carried out with a beam diameter of 25 μm, 5 Hz repetition rate, and energy of 10–20 J/cm². Data acquisition for each analysis took 100 s (40 s on background and 60 s on signal). The raw ICP-MS data were processed using GLITTER (van Ackerbergh et al., 2001). Common Pb was corrected according to the method proposed by Andersen (2002). The age calculations and plotting of concordia diagrams were made using Isoplot (ver. 3.23) (Ludwig, 2003). The well-characterized Mud Tank zircon was analyzed frequently to monitor the reproducibility and the stability of instrument. The analyses of Mud Tank during the U–Pb dating with simultaneous determination of trace elements yielded a weighted mean 206Pb/238U age of 733 ± 9 Ma (2σ; n = 52), while the analyses of Mud Tank during the U–Pb dating without simultaneous determination of trace elements yielded a weighted mean 206Pb/238U age of 727 ± 4 Ma (2σ; n = 20). Both of them are in accordance with the TIMS results (732 ± 5 Ma; Black and Gulson, 1978).

Zircon trace elements analyses were simultaneously obtained during zircon U–Pb dating. NIST612 glass was used as an external standard to calculate the trace element contents of the unknowns, with working values recommended by Pearce et al. (1997). We used 26Si as an internal standard to normalize each analysis. Average analytical uncertainty ranges from 10% for light rare earth elements (LREEs) to 5% for other trace elements. For the calculation of Ti-in-zircon temperature, the presence of quartz in all the samples suggests SiO2 activity = 1 (Anderson et al., 2008). TiO2 activity is estimated at 0.75 for the presence of titanite and Fe–Ti oxide in the Miaohou and Shanhou complexes (Hayden and Watson, 2007).

### 3.2. Hf–isotope analysis of zircon

In situ Lu–Hf isotopic analyses of zircon were conducted using a Neptune Plus MC-ICP-MS, connected with a UP 193 nm laser at the State Key Laboratory for Mineral Deposits Research, Nanjing University. Instrumental conditions and data acquisition were comprehensively described by Hou et al. (2007). A stationary spot was used for the present analyses, with a beam diameter of 35 μm. Helium was used as the carrier gas to transport the ablated sample from the laser-ablation cell to the ICP-MS torch via an Ar gas mixing chamber. A 8 Hz repetition rate, pulse energy density of 10.51/cm² were used. In order to evaluate the reliability of the data, zircon standard 91500 was analyzed during the course of this study and yielded a weighted mean 176Hf/177Hf ratio of 0.282296 ± 8 (2σ; n = 36). The measured 176Lu/177Hf ratios and the 176Lu decay constant of 1.867 × 10–11 year–1 (Söderlund et al., 2004) were used to calculate initial 176Hf/177Hf ratios. The chondritic values of 176Lu/177Hf = 0.0336 ± 1 and 176Hf/177Hf = 0.282785 ± 1 (2σ) (Bouvier et al., 2008) were used for calculating fHf values. The depleted mantle Hf model ages (TDM) were calculated using the measured 176Lu/177Hf ratios based on the assumption that the depleted mantle reservoir has a linear isotopic growth from 176Hf/177Hf = 0.279718 at 4.55 Ga to 0.283250 at present, with 176Lu/177Hf = 0.0384 (Griffin et al., 2000). The new continental crust Hf model ages (TNC) were calculated using the measured 176Lu/177Hf ratios based on the assumption that the new continental crust reservoir (island arcs) has a linear isotopic growth from 176Hf/177Hf = 0.279703 at 4.55 Ga to 0.283145 at present, with 176Lu/177Hf = 0.0375 (Dhuime et al., 2011). The TNC provides a better constraint than TDM on when the continental crust generated (Dhuime et al., 2011). We also present a two-stage model age (T2DM or T2NC) for each zircon, which assumes that its parental magma was produced from an average continental crust (176Lu/177Hf = 0.015; Griffin et al., 2002) that was originally derived from the depleted mantle or island arcs.

#### 3.3. Major and trace element analyses of whole-rocks

Bulk-rock major element analysis was performed using an ARL9800XP+ X-ray fluorescence spectrometer (XRF) in the State Key Laboratory of Mineral Deposits Research, Nanjing University. The glass discs were prepared by fusion of a mixture with an alkali flux consisting of a 66.33:33.17:0.5 mixture of lithium tetraborate, lithium metaborate and lithium bromide at 1050 °C. Analyses were carried out with an accelerating voltage of 50 kV and a beam current of 50 mA. Standards (GSR-3) were prepared using the same procedure to monitor the analytical precision. The analytical precision is generally better than 2% for all elements. Trace element analyses of most samples were measured in the State Key Laboratory of Mineral Deposit Research, Nanjing University. For trace element analyses, ca. 50 mg of powder was dissolved in high-pressure Teflon bombs.
using a HF+HNO₃ mixture. Rb was used as an internal standard to monitor signal drift during ICP-MS analyses. Trace element concentrations were determined using a Finnigan Element II ICP-MS. The precision of ICP-MS analyses is <10% for all trace elements and <5% for the majority of elements. For detailed procedures of the trace element analyses, see Gao et al. (2003). Whole-rock trace element analyses of other few samples were carried out at the State Key Laboratory of Geological Processes and Mineral Resources, China University of Geosciences in Wuhan. For trace element analyses, ca. 50 mg were dissolved in distilled HF+HNO₃ (3:1) in Savillex Teflon screw-cap capsules at 100 °C for 2 days, dried and then digested with 6 M HCl at 150 °C. Three duplicates of three standards (AGV-1, GSR-3 and DNC-1) were prepared by the same procedure to monitor the analytical precision. The solutions were measured for trace elements using an Agilent 7500a ICP mass spectrometer (MS). The discrepancy among triplicates is less than 10% for all elements. Analyses of standards are in agreement with the recommended values. The more detailed analytical procedures were referred to Lin et al. (2000).

3.4. Sr–Nd isotope analysis of whole-rocks

Sr isotopic compositions were measured using a Finnigan Tri- ton TII thermal ionization mass spectrometer (TIMS) in the State Key Laboratory of Mineral Deposits Research, Nanjing University following the methods of Pu et al. (2004, 2005). Nd isotopic compositions were measured using the Neptune (Plus) MC-ICP-MS at the State Key Laboratory of Mineral Deposits Research, Nanjing University. For whole-rock Sr–Nd isotope analyses, ca. 50 mg of powder was dissolved in the same way as for trace element analyses. Rb–Sr and Sm–Nd were separated using AG50W×8 resin and various eluents. The rare earth elements (REEs) were first separated from Rb–Sr by conventional cation exchange chromatography using HCl as an eluent. Rb and Sr were then separated and purified using a mixed eluent of pyridinium and DCTA complex. Sr and Nd were separated and purified using HIBA as an eluent through a small volume of cation exchange resin (0.6 mL). 87Sr/86Sr and 143Nd/144Nd ratios are reported as measured, after normalization to $^{86}$Sr/$^{88}$Sr = 0.1194 and $^{146}$Nd/$^{144}$Nd = 0.7219 (O’Nions et al., 1977), respectively, to correct for instrumental fractionation. The long-term measurements of the Japan JNdI-1 Nd standard gave $^{143}$Nd/$^{144}$Nd = 0.512056 ± 0.000008 (2σ) and for the NIST SRM 987 Sr standard yielded $^{87}$Sr/$^{86}$Sr = 0.710248 ± 0.000004 (2σ). For the calculation of $^{143}$Sr/$^{86}$Sr), $\epsilon_{Nd}(t)$ and Nd model ages, the following parameter were used: $\lambda_{Sm} = 1.42 \times 10^{-11}$ year$^{-1}$ (Minster et al., 1982); $\lambda_{Nd} = 6.54 \times 10^{-12}$ year$^{-1}$ (Lugmair and Marti, 1978); $^{147}$Sm/$^{144}$Nd)$_{CHUR} = 0.1960 \pm 4$, $(^{143}$Nd/$^{144}$Nd)$_{CHUR} = 0.512630 \pm 1$ (2σ) (Bouvier et al., 2008); $(^{143}$Nd/$^{144}$Nd)$_{DM} = 0.513151$, $(^{147}$Sm/$^{144}$Nd)$_{DM} = 0.2136$ (Liew and Hofmann, 1988). The $^{147}$Sm/$^{144}$Nd value of 0.118 for average continental crust (Jahn and Condie, 1995) was used for the mantle extraction model age ($T_{DM}$) for the source rocks of the magmas.

4. Analytical results

4.1. Zircons U–Pb geochronology and trace elements

The locations of selected samples for zircon U–Pb dating are shown in Fig. 1b. CL images of representative zircons are shown in Fig. 2. The trace element and age results for samples of these two complexes are graphically shown in Figs. 3 and 4 (also see Table 3, Appendix Tables 1 and 2). Zircons separated from the selected

![Fig. 2. CL images of representative zircons from the Miaohou and Shanhous complexes. Small solid circles are spots for U–Pb isotope analyses, and big dashed circles are spots for Hf isotope analyses. Age and $\epsilon_{Nd}(t)$ values are also shown for each spot.](image-url)
Fig. 3. Chondrite-normalized REE patterns of zircons from the Miaohou and Shanhou complexes. The chondrite values are from Sun and McDonough (1989).

Table 3
Summaries of the zircon U–Pb ages, trace elements and Hf isotopic results of the samples from Miaohou and Shanhou complexes.

<table>
<thead>
<tr>
<th>Complexes</th>
<th>Rock type</th>
<th>Sample</th>
<th>U–Pb age (Ma)^a</th>
<th>U–Pb age (Ma)^b</th>
<th>εHf(t)</th>
<th>TDM (Ga)</th>
<th>TDM2 (Ga)</th>
<th>Ce(IV)/Ce(III)^c</th>
</tr>
</thead>
<tbody>
<tr>
<td>Miaohou</td>
<td>Granite</td>
<td>MH09</td>
<td>830 ± 9</td>
<td></td>
<td>+3.2 to +8.1</td>
<td>1.06–1.25</td>
<td>1.20–1.51</td>
<td>10–257</td>
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<tr>
<td></td>
<td></td>
<td>MH13</td>
<td>833 ± 12</td>
<td></td>
<td>+1.1 to +6.0</td>
<td>1.17–1.34</td>
<td>1.33–1.65</td>
<td>26–729</td>
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<tr>
<td></td>
<td></td>
<td>MH16</td>
<td>832 ± 9</td>
<td>828 ± 4</td>
<td>+0.6 to +5.3</td>
<td>1.19–1.37</td>
<td>1.38–1.67</td>
<td>19–454</td>
</tr>
<tr>
<td></td>
<td></td>
<td>MH20</td>
<td>830 ± 10</td>
<td></td>
<td>+0.1 to +6.8</td>
<td>1.12–1.44</td>
<td>1.28–1.71</td>
<td>37–442</td>
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<tr>
<td></td>
<td></td>
<td>Av. Zircon</td>
<td></td>
<td></td>
<td>+3.7 ± 0.4</td>
<td></td>
<td></td>
<td>145</td>
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<tr>
<td></td>
<td>Diorite</td>
<td>MH07</td>
<td>832 ± 16</td>
<td></td>
<td>+4.8 to +12.2</td>
<td>0.90–1.19</td>
<td>0.94–1.41</td>
<td>40–2265</td>
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<tr>
<td></td>
<td></td>
<td>MH14</td>
<td>832 ± 13</td>
<td>831 ± 4</td>
<td>+3.5 to +6.9</td>
<td>1.10–1.24</td>
<td>1.28–1.49</td>
<td>16–1527</td>
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<tr>
<td></td>
<td></td>
<td>MH18</td>
<td>834 ± 14</td>
<td></td>
<td>+1.7 to +9.5</td>
<td>1.00–1.32</td>
<td>1.11–1.61</td>
<td>17–242</td>
</tr>
<tr>
<td></td>
<td></td>
<td>MH24</td>
<td>831 ± 8</td>
<td></td>
<td>+2.6 to +10.2</td>
<td>1.00–1.46</td>
<td>1.07–1.55</td>
<td>42–753</td>
</tr>
<tr>
<td></td>
<td>Gabbro</td>
<td>MH15</td>
<td>828 ± 11</td>
<td>831 ± 4</td>
<td>+3.4 to +9.8</td>
<td>0.99–1.27</td>
<td>1.09–1.50</td>
<td>24–457</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Av. Zircon</td>
<td></td>
<td></td>
<td>+5.6 ± 0.4</td>
<td></td>
<td></td>
<td>310</td>
</tr>
<tr>
<td>Shanhou</td>
<td>Granite</td>
<td>SH01</td>
<td>832 ± 9</td>
<td></td>
<td>+0.9 to +11.2</td>
<td>0.94–1.36</td>
<td>1.00–1.66</td>
<td>29–816</td>
</tr>
<tr>
<td></td>
<td></td>
<td>SH02</td>
<td>832 ± 9</td>
<td>829 ± 5</td>
<td>+0.7 to +11.2</td>
<td>0.93–1.36</td>
<td>1.00–1.67</td>
<td>37–707</td>
</tr>
<tr>
<td></td>
<td>Diorite</td>
<td>SH04</td>
<td>830 ± 12</td>
<td>830 ± 4</td>
<td>+4.1 to +10.9</td>
<td>0.94–1.21</td>
<td>1.02–1.45</td>
<td>99–2300</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Av. Zircon</td>
<td></td>
<td></td>
<td>+8.4 ± 0.6</td>
<td></td>
<td></td>
<td>502</td>
</tr>
</tbody>
</table>

^a zircon U–Pb ages measured simultaneously with trace elements.
^b zircon U–Pb ages measured simultaneously without trace elements.
^c zircon Ce(IV)/Ce(III) is calculated after Ballard et al. (2002).
Fig. 4. U–Pb Concordia diagram of representative samples from the Miaohou and Shanhou complexes.
samples are light yellow, prismatic or ellipsoidal in shape with aspect ratios of 1.5–4.0, and approximately 100–250 µm long with well-developed oscillatory zonation. Some euhedral zircons show prisms with {110} > {100} and pyramids with {211} > {101} or {211} < {101}, and others are ellipsoidal shape. The trace elements are analyzed simultaneously with the U–Pb dating by LA-ICP-MS (Fig. 2).

4.1.1. Miaohou granite (MH09, MH13, MH16 and MH20)

A total of 23 analyses on 23 zircons from MH09 give high Th/U values (0.54–2.20) (Appendix Table 2), and plot on or close to the concordia curve, yielding a weighted mean 206Pb/238U age of 830 ± 9 Ma (MSWD = 0.066; Table 3, Fig. 4). The simultaneous determinations of trace elements show that they are enriched in HREEs with positive Ce anomalies (Ce/Ce* = 2.25–18.02) and negative Eu anomalies (Eu/Eu* = 0.31–0.66) except one sample with positive Eu anomalies (Eu/Eu* = 1.15) (Appendix Table 1; Fig. 3).

Zircons from MH13 show high Th/U values (0.99–3.84) and are consistent with their magmatic origin (Appendix Table 2) with a weighted mean 206Pb/238U age of 833 ± 12 Ma (MSWD = 0.052; Table 3, Fig. 4). Except that MH13-02 has high contents of LREEs which may be caused by minor inclusions (e.g., apatite, melt inclusions), the simultaneous determinations of trace elements show that the majority of these zircons are enriched in HREEs and have positive Ce anomalies (Ce/Ce* = 2.55–73.86) and negative Eu anomalies (Eu/Eu* = 0.20–0.76) (Appendix Table 1; Fig. 3).

Twenty-two analyses of 22 zircons from MH16 were obtained with simultaneous determination of U-Pb dating and trace elements, showing high Th/U values (1.13–3.02) (Appendix Table 2) and give a weighted mean age of 832 ± 9 Ma (MSWD = 0.081; Table 3, Fig. 4). Most of the zircons are enriched in HREEs, and have positive Ce anomalies (Ce/Ce* = 5.27–61.81) and negative Eu anomalies (Eu/Eu* = 0.21–0.58) (Appendix Table 1; Fig. 3). Probably influenced by tiny inclusions (e.g., apatite, melt inclusions), three analyses (MH16-03, 17 and 20) show relatively flat REE patterns (Fig. 3). The 18 analyses of U-Pb age data without simultaneous determination of trace elements yield a more precise weighted mean age of 828 ± 4 Ma (MSWD = 0.15; Table 3, Fig. 4).

Nineteen simultaneous analyses of U–Pb dating and trace elements on nineteen grains from MH20 give higher and varying Th/U values (1.06–3.30) (Appendix Table 2) and are grouped and concordant, defining a weighted mean 206Pb/238U age of 830 ± 10 Ma (MSWD = 0.053; Table 3, Fig. 4). As typical magmatic zircons, they exhibit positive Ce anomalies (Ce/Ce* = 6.86–58.59) and negative Eu anomalies (Eu/Eu* = 0.13–0.56) and high contents of HREEs (Appendix Table 1; Fig. 3).

According to the Ti-in-zircon thermometer (Ferry and Watson, 2007), calculated Ti-in-zircon temperatures are 746–934 °C for these four samples, with average of 781 °C, 801 °C, 805 °C and 805 °C. The calculated zircon Ce(IV)/Ce(III) for these four samples are 10–729 with average of 145 (Table 3).

4.1.2. Miaohou diorite (MH07, MH14, MH18 and MH24)

Nineteen U–Pb analyses of zircons from MH07 show variably high Th/U (0.49–2.56), indicating an igneous origin (Appendix Table 2), and all of them are concordant with a weighted mean 206Pb/238U age of 832 ± 16 Ma (MSWD = 0.12; Table 3, Fig. 4). The simultaneous determinations of trace elements show that 17 analyzed zircons are enriched in HREEs with positive Ce anomalies (Ce/Ce* = 1.72–27.56) and negative Eu anomalies (Eu/Eu* = 0.11–0.75) except MH07-18 with positive Eu anomalies (Eu/Eu* = 1.20) (Appendix Table 1; Fig. 3). Two other analyzed zircons (MH07-04, 05) have high contents of LREEs which may be affected by tiny inclusions (e.g., apatite, melt inclusions) (Fig. 3).
Zircons from the MH14 have high and varying Th/U values (0.91–1.69) (Appendix Table 2). The 25 analyses plot on the concordia curve and yield a weighted mean $^{206}$Pb/$^{238}$U age of 832 ± 13 Ma (MSWD = 0.025; Table 3, Fig. 4). Also, the simultaneous determinations of trace elements show that they are enriched in HREEs with positive Ce anomalies (Ce/Ce* = 6.15–27.27) and negative Eu anomalies (Eu/Eu* = 0.25–0.65) (Appendix Table 1; Fig. 3). The U–Pb age data without simultaneous determination of trace elements for this sample yield a more precise weighted mean age of 831 ± 4 Ma (MSWD = 0.105; Table 3, Fig. 4).

A total of nine simultaneous analyses of U–Pb dating and trace elements on nine zircons from MH18 give high Th/U values (1.26–2.55) (Appendix Table 2), and they plot on or close to the concordia curve, yielding a weighted mean $^{206}$Pb/$^{238}$U age of 834 ± 14 Ma (MSWD = 0.31; Table 3, Fig. 4). These zircons also exhibit positive Ce anomalies (Ce/Ce* = 23.5–28.90) and negative Eu anomalies (Eu/Eu* = 0.33–0.90) and high contents of HREEs indicating an igneous origin (Appendix Table 1; Fig. 3).

Zircons separated from MH24 show high Th/U values (0.37–4.52) and are consistent with their magmatic origin (Appendix Table 1). Twenty-two analyses measured with simultaneous determination of trace elements yield a weighted mean $^{206}$Pb/$^{238}$U age of 831 ± 8 Ma (MSWD = 0.111; Table 3, Fig. 4). Except one zircon (MH24-06) with higher La content than Ce, all other zircons from this sample are enriched in HREEs with positive Ce anomalies (Ce/Ce* = 3.85–40.41) and negative Eu anomalies (Eu/Eu* = 0.38–0.86) except MH24-04 and MH24-25 with positive Eu anomalies (Eu/Eu* = 1.07 and 1.60 respectively) (Appendix Table 1; Fig. 3).

Ti-in-zircon temperature calculations show 717–900°C for these four samples with average of 760°C, 792°C, 815°C and 774°C for MH07, MH14, MH18 and MH24, respectively. The calculated zircon CeIV/CeIII for these four samples are 16–2265 with average of 310 (Table 3).

4.1.3. Miaohou gabbro (MH15)

Twenty-one simultaneous analyses of U–Pb dating and trace elements on nineteen grains from MH15 give high and varying Th/U values (0.84–3.04), indicating an igneous origin (Appendix Table 2). The bulk of the analyses are grouped and concordant, defining a weighted mean $^{206}$Pb/$^{238}$U age of 828 ± 11 Ma (MSWD = 0.089; Table 3, Fig. 4). Their chondrite normalized REE patterns are enriched in HREEs with positive Ce anomalies (Ce/Ce* = 2.33–14.00) and negative Eu anomalies (Eu/Eu* = 0.34–0.97) except MH15-20 with high contents of La, weak positive Ce anomalies and positive Eu anomalies which may be affected by thin inclusions (e.g., apatite, melt inclusions) (Appendix Table 1; Fig. 3). The 21 analyses of U–Pb age data without simultaneous determination of trace elements yield a more precise weighted mean age of 830 ± 4 Ma (MSWD = 0.16; Table 3, Fig. 4). Ti-in-zircon temperature calculations show the zircons crystallization temperatures are 713–840°C, with an average of 769°C. The calculated zircon CeIV/CeIII for these samples are 99–2300 with average of 502 (Table 3).

4.2. Zircons Hf–isotopes

The zircon Hf isotopes were measured on the same grains used for U–Pb dating (Fig. 2). Analytical results of the Lu–Hf isotopic compositions are given in Table 3, Appendix Table 3 and illustrated in Fig. 5.

4.2.1. Miaohou granite (MH09, MH13, MH16 and MH20)

Zircons from MH09, MH13, MH16 and MH20 show similar Hf isotope compositions. The $\varepsilon$Hf(t) and initial $^{176}$Hf/$^{177}$Hf ratios for zircons in these four samples are relatively homogeneous. The $\varepsilon$Hf(t) values vary from +0.1 to +8.1, clustering within the range of +2 to +7 (Fig. 5). In addition, the initial $^{176}$Hf/$^{177}$Hf ratios range from 0.282264 to 0.282489, corresponding to the two-stage depleted mantle Hf model ages ($T_{DM}$) of 1.20 Ga to 1.71 Ga or the two-stage new continental crust Hf model ages ($T_{NC}$) of 0.99 Ga to 1.52 Ga (Table 3; Appendix Table 3).

4.2.2. Miaohou diorite (MH07, MH14, MH18 and MH24)

The zircons from MH07, MH14, MH18 and MH24 also exhibit similar $\varepsilon$Hf(t) values and initial $^{176}$Hf/$^{177}$Hf Hf ratios. The $\varepsilon$Hf(t) values vary in a large range from +1.7 to +12.2, but clustering within the range of +3 to +8 (Fig. 5). In addition, the initial $^{176}$Hf/$^{177}$Hf ratios range from 0.282304 to 0.282603, corresponding to the two-stage depleted mantle Hf model ages ($T_{DM}$) of 0.94 Ga to 1.61 Ga or the two-stage new continental crust Hf model ages ($T_{NC}$) of 0.72 Ga to 1.42 Ga (Table 3; Appendix Table 3).
4.2.3. Shanhou gabbro (MH15)

The zircons from MH15 show high $\varepsilon_{HF}(t)$ values of +3.4 to +9.8 with a weighted mean of $+7.0 \pm 0.7$ and initial $^{176}Hf/^{177}Hf$ ratios (from 0.282356 to 0.282536) (Table 3; Appendix Table 3; Fig. 5). Similar to above zircons, these zircons also give the two-stage depleted mantle Hf model ages ($T_{2DM}$) of 1.09–1.50 Ga or the two-stage new continental crust Hf model ages ($T_{2NC}$) of 0.88 Ga to 1.30 Ga (Table 3; Appendix Table 3).

4.2.4. Shanhou granite (SH01 and SH02)

Zircons from SH01 and SH02 show heterogeneous Hf-isotope compositions with varying $\varepsilon_{HF}(t)$ values of +0.7 to +11.2, clustering within the range of +2 to +10 (Fig. 5). The initial $^{176}Hf/^{177}Hf$ ratios of these two samples are varying from 0.282281 to 0.282577, corresponding to the two-stage depleted mantle Hf model ages ($T_{2DM}$) of 1.00 Ga to 1.67 Ga and the two-stage new continental crust Hf model ages ($T_{2NC}$) of 0.78 Ga to 1.48 Ga (Table 3; Appendix Table 3).

4.2.5. Shanhou diorite (SH04)

Zircons from sample SH04 with a $^{206}Pb/^{238}U$ age of ca. 830 Ma show high $\varepsilon_{HF}(t)$ values of +4.1 to +10.9 with a weighted mean of $+8.4 \pm 0.6$ (2 SD) or initial $^{176}Hf/^{177}Hf$ ratios (from 0.282376 to 0.282569) (Table 3; Appendix Table 3; Fig. 5), corresponding to the two-stage depleted mantle Hf model ages ($T_{2DM}$) of 1.02 Ga to 1.45 Ga or the two-stage new continental crust Hf model ages ($T_{2NC}$) of 0.80 Ga to 1.26 Ga (Table 3; Appendix Table 3).

4.3. Whole-rock major and trace elements

Whole-rock major and trace element analyses for representative samples from the Miaohou and Shanhou complexes are listed in Table 4.

All the samples of the Miaohou and Shanhou granites show high SiO$_2$ (between 67.26 and 79.75 wt.%) and relatively high total alkalis (K$_2$O + Na$_2$O) ranging from 4.82 to 8.90 wt.%, with the majorities plotting in the granite field on the total alkali–silica (TAS) diagram (Fig. 6a). Based on the molar ratios of Al$_2$O$_3$/(CaO + Na$_2$O + K$_2$O) (A/CNK) and Al$_2$O$_3$/(Na$_2$O + K$_2$O) (A/NK) (Fig. 8a), the Miaohou granite straddles a boundary between metaluminous and weakly peraluminous (A/CNK = 0.86–1.08), plotting in the field of I-type granites. The Shanhou granite is metaluminous to weakly peraluminous (A/CNK = 0.85–1.06), and also plot in the I-type granites field. All samples of the Miaohou and Shanhou granites have relatively higher Na$_2$O (3.07–5.41 wt.%) than K$_2$O (1.75–4.06 wt.%) and very low K$_2$O/Na$_2$O (0.57–0.86), characteristic of mantle-derived materials. Following Frost et al. (2001)’s classification (Fig. 6c and d), these two granites vary from calcic to alkaline with various MAI (Na$_2$O + K$_2$O–CaO) (0.47–7.90), belonging to magnesian granites with low FeO$^T$/[FeO$^T$ + MgO] (0.60–0.80), similar to those of Cordilleran granitoids. In the K$_2$O–SiO$_2$ diagram (Fig. 7), all samples of the Miaohou and Shanhou granites plot in the calc-alkaline to high-K calc-alkaline fields. Their negative P$_2$O$_5$–SiO$_2$ correlation reflects apatite fractionation, which is often interpreted to be characteristic of I-type granites (Fig. 7). The P$_2$O$_5$ and TiO$_2$ saturation thermometric calculations indicate that these two granites have relatively high initial temperatures (Fig. 7; Watson and Harrison, 1984; Green and Pearson, 1986), which is in accordance with their Zr saturation temperatures (725–812°C, with an average of 778°C) (Table 4; Watson and Harrison, 1983) and Ti-in-zircon temperatures (average of 745–805°C; Appendix Table 1). The Miaohou and Shanhou granites with total REE ranging from 72 to 154 ppm show similar chondrite-normalized REE patterns and relatively enrichment of LREEs over HREEs with significant to weak negative Eu anomalies (Eu/Eu$^*$ = 0.27–0.94), which likely reflect variably fractionation of feldspar (Fig. 9). The (La/Yb)$_{nu}$ values of these two granites are high (3.86–15.59) (Table 4). In the primitive mantle-normalized trace element diagrams (Fig. 9), they are all characterized by positive Th, U, K, Pb, Zr and Hf and negative Nb and Ta anomalies and marked depletion in P and Ti, which indicates that fractionation of apatite and Fe–Ti oxides also occur during magma evolution. However, all samples of the Miaohou and Shanhou granites have similar REE patterns and trace element diagrams to the continental arc ADR (andesite, dacite and rhyolite), suggesting they were formed in active continental margin.

The Miaohou and Shanhou diorites are silica oversaturated, containing quartz, with varying SiO$_2$ (55.55–63.75 wt.%). They have relatively high K$_2$O + Na$_2$O (6.00–6.89 wt.%), and mainly plot in the diorite field on the TAS diagram (Fig. 6a). Most of these samples are calc-alkaline (Fig. 6b) and show relatively low K, plotting in the calc-alkaline field on K$_2$O–SiO$_2$ diagram (Fig. 7). These two diorites straddle from calcic to alkaline with various MAI (Na$_2$O + K$_2$O–CaO) (−1.06 to 4.92) (Fig. 6c; Frost et al., 2001). The P$_2$O$_5$ and TiO$_2$ saturation thermometric calculations indicate that these two diorites have relatively higher initial temperatures than the two granites (Fig. 7; Watson and Harrison, 1984; Green and Pearson, 1986), which is in accordance with their higher Ti-in-zircon temperatures (average of 760–815°C; Appendix Table 1). All samples of the two diorites show variably elevated REE abundances but similar REE patterns with (La/Yb)$_{nu}$ values ranging from 2.85 to 12.55 (Fig. 9). Also, they display weak negative or positive Eu anomalies (Eu/Eu$^*$ = 0.57–1.09), which likely reflect insignificant fractionation of feldspar. On primitive mantle-normalized trace element diagrams, these two diorites show enrichment in K, Pb, Zr and Hf, but depletion in Nb, Ta, and Ti (Fig. 9). They exhibit large variations in trace element abundances, particularly for MH18 with positive Th and Nb anomalies and the highest SiO$_2$ (63.75 wt.%) relative to other samples, and share the similar REE and trace element characteristics with the Miaohou and Shanhou granites while other samples show more closely REE and trace element characteristics resemble the continental arc andesites.

Both samples collected from the Miaohou gabbro contain quartz. Sample MH15 has higher SiO$_2$ (51.85 wt.%) than MH17 (50.03 wt.%) (Table 4). On the TAS diagram (Fig. 6a), these two samples straddle a boundary between the alkaline and subalkaline field. Samples MH15 and MH17 have variable K$_2$O (2.62 and 0.99 wt.% respectively) and K$_2$O/Na$_2$O (0.80 and 0.33 wt.% respectively) (Fig. 7; Table 4). Their variably high MgO (7.60 and 4.84 wt.% respectively), FeO$^T$ (6.80 and 9.44 wt.% respectively), Ni (77.44 and 16.16 ppm respectively) and Cr (286.2 and 54.92 ppm respectively).
### Table 4
Chemical compositions of representative samples from Maohou and Shanhou complexes.

<table>
<thead>
<tr>
<th>Sample</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>Fe₂O₃</th>
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<th>CaO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>Sr</th>
<th>Y</th>
<th>Ga</th>
<th>Zr</th>
<th>TiO₂</th>
<th>Nb</th>
<th>Ba</th>
<th>Yb</th>
<th>Ho</th>
<th>Eu/Eu*</th>
<th>Temperature (°C)</th>
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<tr>
<td>Maohou</td>
<td>69.02</td>
<td>15.06</td>
<td>2.37</td>
<td>0.05</td>
<td>1.38</td>
<td>0.74</td>
<td>1.20</td>
<td>0.05</td>
<td>74.44</td>
<td>106.56</td>
<td>780.57</td>
<td>5.96</td>
<td>2.20</td>
<td>1.38</td>
<td>0.13</td>
<td>4.20</td>
<td>205.78</td>
<td>117.20</td>
<td></td>
</tr>
<tr>
<td>Shanhou</td>
<td>78.39</td>
<td>15.11</td>
<td>2.39</td>
<td>0.06</td>
<td>1.20</td>
<td>0.85</td>
<td>1.20</td>
<td>0.06</td>
<td>72.02</td>
<td>105.02</td>
<td>770.57</td>
<td>6.01</td>
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<td>0.14</td>
<td>4.20</td>
<td>205.78</td>
<td>117.20</td>
<td></td>
</tr>
</tbody>
</table>

**Notes:**
- **A/CNK** = molar Al₂O₃/(CaO + Na₂O + K₂O).
- **Eu/Eu**° = Eu/Eu°(Sr/Y + Gd/Y).
- Temperature (°C) is calculated after Watson and Harrison (1983).

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respectively) are consistent with their mafic nature and with their being derived/evolved from mantle melts (Wilkinson and Le Maitre, 1987). They have relatively low total REE (53 and 35 ppm respectively), and LREE enrichments (i.e., (La/Sm)_n > 1) with small to negligible Eu anomalies (Eu/Eu^* = 1.09 and 1.13 respectively) (Fig. 9). Their primitive mantle-normalized trace element systematics resemble, to a certain extent, continental arc basalts with enrichments in K but depletion in Nb, Ta and small anomalies in Ba, Pb, Sr, P and Ti (Fig. 9) although we are aware that gabbros may not be melt, but cumulate and they can have melt-like (but not melt) compositions because of incomplete element concentration in trapped interstitial melt (Niu et al., 2002). Nevertheless, sample MH17 has higher Rb, Ba, K and lower Nb, Ta than MH15, suggesting the possibility that MH17 may have been contaminated by more crustal material.

4.4. Sr–Nd isotopes of whole-rocks

Sr and Nd isotope data for whole-rock samples of the Miaohou and Shanhou complexes are given in Table 5. The Miaohou and Shanhou granites have very similar Nd isotopic compositions with εNd(t) from +0.3 to +3.1, and two-stage depleted mantle Nd model ages (T_{2DM}) from 1.24 to 1.48 Ga. However, the Miaohou granite exhibits higher 87Rb/86Sr ratios of 0.85–2.31 with lower 87Sr/86Sr, ranging from 0.69426 to 0.69987, while the Shanhou granite gives lower 87Rb/86Sr ratios of 0.35–0.50 with higher 87Sr/86Sr, ranging from 0.70146 to 0.70306. However, those samples show no anticipated inverse (87Sr/86Sr) vs. εNd(t) correlation, which could be erased by post-emplacement mobilization of Sr, but are more likely on account of a larger uncertainties of calculated (87Sr/86Sr)_i ratios due to the higher Rb/Sr ratios samples. The Miaohou diorites have large variations in initial 87Sr/86Sr ratios ranging from 0.69704 to 0.70452 and high εNd(t) values ranging from +1.1 to +3.6. The initial 87Sr/86Sr and εNd(t) values of Shanhou diorite are 0.70282 and +5.5 respectively. Also, sample MH18 gives the lowest (87Sr/86Sr)_i 0.69704 with the highest 87Rb/86Sr ratios (1.24) among these diorites. The Miaohou and Shanhou diorites have relatively younger T_{2DM} ages of 1.05 Ga to 1.41 Ga. The Miaohou gabbro has high εNd(t) value (+3.9), low initial 87Sr/86Sr ratio of 0.70175 and young TDM age of 1.19 Ga. In the εNd(t) vs. age diagram (Fig. 10a), the Nd isotopic compositions of these two complexes plot between depleted mantle and Cathaysia basement, implying that they involved both mantle-derived and crustal materials. The Miaohou and Shanhou complexes are also well distributed along the terrestrial array of εNd(t) vs. εEu(t) diagram (Vervoort and Blichert-Toft, 1999), and most of the samples plot into an overlap between ocean island basalts (OIB) and global lower crust (Fig. 10b).

5. Discussion

5.1. Timing of Neoproterozoic magmatism in the Cathaysia block

The evidence for Neoproterozoic magmatism in the Cathaysia block comes firstly from the ca. 984–969 Ma mafic–ultramafic bodies that are distributed mainly along the Wuyi–Yunkai belt in the areas of Longquan, Qingyuan, Zhenghe, Suichang, Jian’ou, Jianyang,
and Xinyi, where the country rocks are quartz schist, gneiss, and migmatite (Table 1). The igneous rock types include gabbro, diabase, basalt, and andesite, and serpentinite and pyroxenite are also found (Wang et al., 2013b; Zhang et al., 2012). Zhang et al. (1998) also reported 972 Ma rhyolitic rocks in the Nanling domain as well as synchronous dacite porphyry in the Yunkai domain. In the Nanling domain at Jingnan (northeast Guangdong) and Hezi (south Jiangxi), rhyolite and granodiorite yield SHRIMP zircon U–Pb and Pb–Pb ages of 972 ± 8 Ma and 996 ± 29 Ma, respectively (Liu et al., 2001; Shu et al., 2008a, 2011; also see Table 1). Wang et al. (2014b) reported a set of SIMS and LA-ICP-MS zircon U–Pb ages of ca. 982–909 Ma for peraluminous granites in the Wuyi–Yunkai belt of the Cathaysia block (Table 1). These data indicate the existence of very early Neoproterozoic (ca. 1.0–0.9 Ga) magmatism along the Wuyi–Yunkai domain in the interior of the Cathaysia block.

Yao et al. (2014b) obtained a LA-ICP-MS U–Pb zircon age of 879 ± 10 Ma for the protolith of the Chencai hornblende gneiss. Also, previous studies demonstrated the existence of a series of mafic rocks with zircon SHRIMP U–Pb ages of 857–836 Ma in the Zhuij, Chencai, Jian’ou, and Zhenghe areas (Table 1; Shu et al., 2011). These data are in agreement with some of the previous age data, including the SHRIMP zircon U–Pb age of 858 ± 11 Ma for an ophiolitic gabbro (Shu et al., 2006), the 841 ± 6 Ma zircon SHRIMP U–Pb age for the Lipu gabbro–diorite (Li et al., 2010), and the SHRIMP U–Pb zircon ages of 857 ± 0.2 Ma and 853 ± 4 Ma on Jian’ou metabasalt and tuff (Shu et al., 2008b). Our geochronological

![Fig. 7. Harker diagram of major element compositions of the Miaohou and Shanhou complexes: K2O vs. SiO2 diagram after Peccerillo and Taylor (1976), the dotted line represents the division between potassic alkaline and shoshonitic suites after Calanchi et al. (2002); P2O5 vs. SiO2 diagram, the dotted line represents isothermal apatite solubilities for melts after Watson and Harrison (1984), the l-type granites trend follows Chappell (1999); TiO2 vs. SiO2 diagram, the dotted line represents isothermal Ti-rich phase solubilities for melts after Green and Pearson (1986). Symbols are the same as those in Fig. 5.]

![Fig. 8. (a) A/NK vs. A/CNK diagram (after Maniar and Piccoli, 1989) and (b) ACF diagram (after White and Chappell, 1977). Symbols are the same as those in Fig. 5.]

results for 12 samples from the Miaohou and Shanhao complexes give zircon U–Pb ages of ca. 830 Ma (Table 3; Fig. 4). All the analyzed zircons show high Th/U ratios (mostly >0.4) and an enrichment in the HREEs with positive Ce and negative Eu anomalies, indicating their magmatic origin; the only exceptions were a few spot analyses which may have been affected by small inclusions (Appendix Table 1; Fig. 3). These ages can therefore be interpreted as the crystallization ages of the different igneous lithologies in the Miaohou and Shanhao complexes, and clearly their formation ages are the same within error. It is always difficult to distinguish which block the plutons within Jiangshao Fault belong to. Nevertheless, the Miaohou and Shanhao complexes were emplaced significantly later than the ca. 970–890 Ma volcanic rocks of the Shuangxiwu Group and granitoids intruded into the group in Zhejiang Province, which are thought to be part of the southeastern margin of the Yangtze Blocks (Li et al., 2009; Ye et al., 2007), but coevally with the mafic to felsic magmatism in the Cathaysia segment of Zhejiang (e.g., Lipu gabbro–diorite, Zhuji gabbro and Zhangcun rhyolite; also see Table 1). Thus, we believe that the Miaohou and Shanhao complexes developed on the northwestern margin of the Cathaysia block.

However, the metavolcanic rocks, diabase, and bimodal volcanic rocks in the Wuyi domain yield the ages of ca. 820–790 Ma (Li et al., 2005, 2010; Shu et al., 2008b, 2011; Wang et al., 2012b; also see

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**Fig. 9.** Chondrite-normalized REE patterns and primitive mantle-normalized multiple trace element diagrams of the Miaohou and Shanhao complexes. The chondrite values and primitive mantle values are from Sun and McDonough (1989). E-type MORB, N-type MORB and OIB are from Sun and McDonough (1989), continental arc basalts and Continental arc andesites are from Kelemen et al. (2003). Continental arc ADR (andesite, dacite and rhyolite) is from Drummond et al. (1996). Symbols are the same as those in Fig. 5.
Table 5 compositions of samples and Shanhou complexes.

<table>
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<tr>
<th>Sample</th>
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<th>MH14</th>
<th>MH15</th>
<th>MH16</th>
<th>MH18</th>
<th>MI20</th>
<th>MH24</th>
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<th>SH02</th>
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<td>0.70316</td>
<td>16.7</td>
<td>0.512400</td>
<td>3.6</td>
<td>0.711751</td>
<td>0.000004</td>
<td>0.707335</td>
<td>0.000007</td>
<td>0.707209</td>
</tr>
<tr>
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<td>0.000007</td>
<td>0.70418</td>
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<td>1.96</td>
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</tr>
<tr>
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<td>0.70146</td>
<td>2.31</td>
<td>2.31</td>
<td>0.000013</td>
<td>0.707480</td>
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<td>1.38</td>
<td>1.47</td>
<td>1.48</td>
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<td>0.000005</td>
<td>0.707335</td>
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<tr>
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</table>

5.2. Petrogenesis of the Miaohou and Shanhou complexes

The two samples of the Miaohou gabbro are high in Al₂O₃ (16.11 and 16.95 wt.% respectively), but low in MgO (<8 wt.%), and are characterized by an enrichment in LILEs and LREEs but a depletion in high field strength elements (HFSEs) (Fig. 9). Like many other gabbros, the Miaohou gabbro was probably formed from mantle-derived basaltic magma, which is consistent with the mantle isotopic signatures of the gabbroic samples; i.e., high whole-rock εNd(t) (>0, +3.9) (Fig. 10a) and high zircon εHf(t) (>0, +2.8 to +9.6) (Fig. 5). Such high εNd(t) and εHf(t) values indicate that the parental basaltic magma of the gabbro was derived from isotopically depleted mantle (Fig. 10b). However, the gabbros have chondrite-normalized REE and primitive-mantle-normalized trace element patterns that are typical of continental arc basalts, which suggests that they were derived from a mantle wedge that had been metasomatized by subduction-related fluids (Fig. 9). Given that zircon is very refractory, the Ce(IV)/Ce(III) anomaly (Fig. 9), is a sensitive and robust indicator of magmatic oxidation state (Ballard et al., 2002). The relatively high Ce(IV)/Ce(III) ratios (>100; Table 3) of the gabbro indicates high oxygen fugacity, consistent with that the gabbro was derived from a hydrated, oxidized convergent margins mantle source (Kelley and Cottrell, 2009). Furthermore, the enrichment in incompatible elements (e.g., La/SmCN > 1) is consistent with the parental basaltic magma being derived by a low degree of melting of a metasomatized source. It is worth noting that sample MH17 shows higher contents of Rh, Ba, and K than MH15, but lower Nb and Ta, implying that more crustal material was involved in MH17. Moreover, the lower SiO₂ content indicates that MH17 was contaminated by mafic lower crust.

The Miaohou and Shanhou diorites are mainly calc-alkaline (Figs. 6b and 7) with low K₂O/Na₂O ratios (0.28–0.61) and chondrite-normalized REE and primitive-mantle-normalized trace element patterns that are characterized by an enrichment in LREEs and LILEs, a significant positive Pb anomaly, a weak positive Zr–Hf anomaly, and negative Nb–Ta, P, Ti, and Eu anomalies, thus resembling continental arc andesites (Fig. 9). Like the Miaohou gabbro, both the Miaohou and Shanhou diorites have depleted Sr–Nd–Hf isotopic features with positive values of εNd(t) (+1.1 to +5.5) and εHf(t) (+1.7 to +12.3) (Figs. 9 and 10a–b), and relatively high zircon Ce(IV)/Ce(III) ratios (>100; Table 3) suggesting that their parental melts were also calc-alkaline with high oxygen fugacity and derived from a metasomatized mantle. The εHf(t) values of zircons from these two diorites show a large variation from +1.7 to +12.3) and εNd(t) also varies by almost five εNd units, which is consistent with processes in an open magma-chamber system at crustal levels. Therefore, the metasomatized mantle-derived melt parental to these two diorites may have been mixed with crustal components, producing the observed isotopic variability in the resultant diorites. However, the high εHf(t) and εNd(t) values suggest that the crustal contribution, if any, was either very small or due to juvenile lower crust.

The Miaohou and Shanhou granites show relatively low values of A/CNK (<1.1; Fig. 8a) and K₂O/Na₂O (0.57–0.86), and there is a negative correlation between P₂O₅ and SiO₂ (Fig. 7); most of the samples plot in the I-type field on the ACF diagram (Fig. 8b). The P₂O₅ and TiO₂ saturation temperatures, Zr saturation temperatures, and Ti-in-zircon temperatures all indicate relatively high initial temperatures for these two granites. All these data indicate that the two granites belong to I-type granite categories. I-type granites are usually derived either by the partial melting of mafic to intermediate meta-igneous crustal rocks that had not undergone surface processes, or by the mixing of crustal
materials with mantle-derived magmas (Chappell, 1999; Collins, 1996; Gray, 1984; Kemp et al., 2007; Xia et al., 2014). However, pure meta-igneous crustal source cannot explain the linear array of major elements (Fig. 7), the superposition of chemical compositions from I- to S-type granites (Fig. 8b), and the large variation in SiO\textsubscript{2} content from 67.26 to 79.75 wt.%. The two granites also exhibit MALI and FeO\textsuperscript{T}/(FeO\textsuperscript{T} + MgO) ratios similar to those of Cordilleran granitoids (Fig. 6c–d), together with chondrite-normalized REE and primitive-mantle-normalized trace element patterns similar to those of continental arc ADR (andesite, dacite, and rhyolite), with pronounced negative Nb–Ta, P, and Ti anomalies (Fig. 9). The fact that these two granites show large variations in the high values of zircon ε\textsubscript{Hf}(t) (+0.1 to +11.3) and ε\textsubscript{Nd}(t) (+0.3 to +3.1) (Figs. 5 and 10a–b) indicates that they were formed by the mixing of a melt, derived from a metasomatized and depleted mantle, with material derived from a juvenile lower crust.

The different lithologies in the Miaohou and Shanhou complexes (gabbro, diorite, and granite) were emplaced coevally (within error) and there are strong correlations between the major elements plotted in Fig. 7, suggesting close relationships among these lithologies. Compositional variations may have resulted from varying degrees of partial melting, fractional crystallization, magma mixing, or wall-rock contamination. As shown in the La/Sm vs. La diagram (Fig. 11a), the Miaohou and Shanhou complexes plot in areas that define both mixing (or partial melting) and fractional crystallization trends. Moreover, there is no negative correlation in the La/Yb vs. Dy/Yb plot (Fig. 11b), which argues against the possibility that the parental magmas of the different lithologies were produced by varying degrees of partial melting of a uniform source. Their Ba/Th and (La/Sm)\textsubscript{C} ratios also resemble those of normal arc rocks (Fig. 11c), indicating that fluids derived from a dehydrating subducted slab played a major role in transporting elements from the slab to the wedge. Nevertheless, there are still some granitic samples that overstep the field of normal arc rocks (Fig. 11c), implying the presence of sediments or sediment-derived melts during the transport of elements from slab to wedge (Tatsumi, 2006). Pearce (2008) suggested using the Th/Yb vs. Nb/Yb diagram to highlight any crustal contamination. In this diagram, the variation trends of the Miaohou and Shanhou complexes indicate magma–crust interaction (contamination) due to wedge melting together with assimilation and fractional crystallization (AFC) (Fig. 11d). The various La/Nb ratios (0.6–4.5) and negative ε\textsubscript{Nd}(t) values fit the characteristics of melts from a metasomatized mantle, as summarized in previous studies (Peng et al., 2006; Wang et al., 2013b; Zhang et al., 2012), but there are also indications of crustal assimilation in the mantle-derived melts. According to the ε\textsubscript{Nd}(t) vs. ε\textsubscript{Hf}(t) diagram (Fig. 11f), the rocks deviate from the trend of metasomatized mantle or crustal assimilation, suggesting that fractional crystallization also played an important role in generating various compositions among different lithologies, suggesting that the assimilated crust was mainly juvenile lower crust. Consequently, we suggest that a low degree of partial melting of a mantle wedge that had been metasomatized by subduction-derived fluids, combined with assimilation of juvenile lower crust during magma ascent and fractional crystallization, produced the different lithologies of the Miaohou and Shanhou complexes.

5.3. Processes of fractional crystallization during magma differentiation

The different lithologies of the Miaohou and Shanhou complexes have similar Nd–Hf isotopes (Fig. 10b), indicating a relatively homogeneous source and excluding the possibility of significant wall-rock contamination except for the assimilation of juvenile lower crust. Therefore, the compositional variations of the different lithologies were probably caused by fractional crystallization. In the Harker diagrams (Fig. 7), K\textsubscript{2}O and Na\textsubscript{2}O increase, and Al\textsubscript{2}O\textsubscript{3}, CaO, FeO\textsubscript{T}, MgO, and MnO decrease with increasing SiO\textsubscript{2} from the gabbro to granite samples, indicating the fractionation of mafic minerals. P\textsubscript{2}O\textsubscript{5} and TiO\textsubscript{2} decrease with increasing SiO\textsubscript{2} from the diorite to granite samples (Fig. 7), implying the fractionation of Ti-bearing minerals and apatite during later differentiation.

Correlations among Ni, V, and Cr (Fig. 12a–b) suggest that the parental magmas of the Miaohou and Shanhou complexes have undergone varying degrees of pyroxene and hornblende fractionation, whereas olivine is unlikely to have been an important fractionating phase in the dioritic to granitic magmas. The steep decrease in Ni and V from the diorite to granite samples implies, therefore, the fractionation of magnetite or Fe–Ti oxides. Fractionation of low-Mg amphibole would have lowered the Nb/Ta and Dy/Yb ratios in the remaining melt because its D\textsubscript{Nb}/D\textsubscript{Ta} and D\textsubscript{Dy}/D\textsubscript{Yb} ratios are >1 (Sisson, 1994; Tiepolo et al., 2001). A positive correlation between Nb/Ta and Dy/Yb in the Miaohou and Shanhou complexes implies amphibole fractionation (Fig. 12c). Eu anomalies in the gabbro and diorite vary from insignificant to positive (Fig. 9), being consistent with a considerable delay in the crystallization of plagioclase and thus forming high-alumina differentiates under high partial pressures of H\textsubscript{2}O in the early differentiation.
stage (e.g., Schiano et al., 2004; Sisson and Grove, 1993). Nevertheless, the significant negative Eu anomalies, and the correlations between Sr/Y and Eu/Eu* in the diorite and granite samples, are indicative of varying degrees of plagioclase fractionation during later differentiation (Fig. 12d), because plagioclase is characterized by high positive Sr and Eu anomalies together with an excessive depletion in all other REEs and Y (McKay et al., 1994). As a para-
luminous mineral, biotite has extremely high partition coefficients for V, but low partition coefficients for Th (Bea et al., 1994). Thus, the fractionation of biotite will give rise to negative correlations between V/Th and SiO2/Al2O3 ratios among the different lithologies (Fig. 12e). Ti-bearing minerals such as ilmenite and sphene might be other important late fractionated phases, as suggested by the decrease of TiO2 with increasing SiO2 (Fig. 7). The fractional crystallization of titanium-rich minerals commonly generates negative Nb-Ta and Ti anomalies (e.g., Xiong et al., 2005). However,
there is no significant correlation between Nb/La and Ti/Yb, which argues against the significance of fractional crystallization of Fe–Ti oxides (Fig. 12f). In fact, the extensive fractionation of feldspars can increase Nb/La ratios and counterbalance the decreasing Nb/La ratios, especially in the granite samples (Fig. 12f). Therefore, varying degrees of fractional crystallization of Fe–Ti oxides, magnetite, pyroxene, hornblende, biotite, plagioclase, and apatite would be the main mechanism for producing the variations in the chemical compositions of the different lithologies in the Miaohou and Shanhou complexes.

5.4. Implications for the tectono-magmatic evolution of the Neoproterozoic active continental margin of Cathaysia

As mentioned above, recent studies have identified a series of ca. 1.0–0.9 Ga mafic to felsic magmatic rocks along the Wuyi–Yunkai belt (Liu et al., 2001; Shu et al., 2008a, 2011; Wang et al., 2013b, 2014b; Zhang et al., 1998, 2012; also see Table 1), representing an early Neoproterozoic continental margin arc and back-arc system, as recently advocated by Cawood et al. (2013) and Wang et al. (2013b, 2014a). In this paper we have described some
associated gabbros, diorites, and I-type granites that yield zircon U–Pb ages of ca. 830 Ma, and these rocks have the geochemical signatures of active continental margin magmatism with positive $\varepsilon_Hf(t)$ and $\varepsilon_{Nd}(t)$ (Fig. 10b). The Miaohou gabbros fall in the calc-alkaline field on the $Y/15-La/10-Nb/8$ diagram (Fig. 13a) with Th/Ta ratios in the range 6–20, indicating a continental arc setting (Gorton and Schandl, 2000). Although the Miaohou and Shanhou diorites and I-type granites straddle the fields of volcanic arc and within-plate magmatism on the Th/Ta vs. Yb and Rb/30–Hf–Ta*3 diagram (Fig. 13b–c), they have geochemical affinities to Cordilleran batholiths (Fig. 6c–d), with almost all plotting in the volcanic arc field on the $Y+\text{Nb}$ vs. Rb diagram (Fig. 13d). They also exhibit REE and spidergram patterns similar to those of felsic continental arc volcanic rocks with pronounced negative Nb–Ta, Sr, P, and Ti anomalies (Fig. 9). Taken together, the data lead us to conclude that a magmatic arc developed along the northwestern margin of the Cathaysia block during early Neoproterozoic time.

Shu et al. (2011) and Yao et al. (2014b) described a series of metamorphic rocks with ages of 879–836 Ma in the Zhuji, Chencai, Jian’ou, and Zhenghe areas (Table 1), and our geochronological studies of the Miaohou and Shanhou complexes imply that the Cathaysia continental arc and back-arc system lasted until ca. 830 Ma.

The presence of the continental arc-like Miaohou and Shanhou complexes means that the final collision of the Yangtze and Cathaysia blocks did not take place until after ca. 830 Ma. In fact, the widespread Proterozoic metasedimentary sequences in the Jiangnan Orogen can be divided into folded sequences and an overlying cover sequence separated by an angular unconformity which has been considered to mark the Neoproterozoic orogenic event that saw the amalgamation of the Yangtze and Cathaysia blocks (Wang et al., 2014a). The folded metasedimentary sequences may have been deposited in a foreland basin (Wang et al., 2007) and then deformed under conditions of the lower greenschist facies. Considering the geological correlations and new dating results for detrital zircons in the folded basement sequences and for magmatic zircons in the interlayered volcanic rocks and intrusive granitoids, the folded metasedimentary basement sequences in the Jiangnan Orogen might have been deposited during the period of 860–825 Ma, and the maximum depositional age of the underlying sequence constrains the timing of the final amalgamation to later than ca. 825 Ma (Wang et al., 2007, 2013, 2014; Zhao and Cawood, 2012; Zhao, 2015). The existence of continental rifting or within-plate magmatism at ca. 820–790 Ma (e.g., Li et al., 2005, 2010; Shu et al., 2008; Wang et al., 2012; also see Table 1), and especially the formation of the $818 \pm 9$ Ma sequence of the Mamianshan bimodal volcanic rocks (SHRIMP zircon U–Pb age; Li et al., 2005), also suggest that the closure of the Cathaysia arc and the back-arc system took place no later than ca. 820 Ma.

The integration of recently obtained geochronological, geochemical, and isotopic data for the Neoproterozoic igneous rocks in the South China Block provides insights into the tectono-magmatic evolution of the Neoproterozoic active continental margin of Cathaysia. The main conclusions can be summarized as follows (Fig. 14). At ca. 1000 Ma, the southeastward subduction of an oceanic plate under the Cathaysia block resulted in dehydration of
the subducted sediments, metasomatism of the mantle wedge by subduction-related fluids, and the consequent partial melting of the metasomatized mantle wedge to produce the primary basaltic melt of the arc. The mantle-derived magmas formed a juvenile lower crust by the process of underplating, which provided the heat source for crustal melting that produced the felsic magmas. Later, at ca. 830 Ma, a low degree of melting of the metasomatized mantle wedge produced the basaltic parents of the Miaohou and Shanhou complexes. During magma ascent, varying degrees of crustal assimilation and fractionation crystallization (AFC) formed the different lithologies of the Miaohou and Shanhou complexes (gabbro, diorite and granite). At ca. 825 Ma, the ocean between the southeast margin of the Yangtze Block and the northwest margin of Cathaysia was finally closed by subduction, leading to the amalgamation of the Yangtze and Cathaysia blocks to form the Jiangnan Orogen.

6. Conclusions

(1) The Miaohou complex is composed of gabbro, diorite and granite, while the Shanhou complex is composed of diorite and granite; both complexes are bound by the Jiangshao Fault. The emplacement of these plutons from the Miaohou and Shanhou complexes coevally occurred at ca. 830 Ma.

(2) The Miaohou gabbro has the geochemical features of volcanic arc basaltic rocks, while the Miaohou and Shanhou diorites and granites have geochemical affinities to Cordilleran batholiths. Their continental-arc-like geochemical signatures, and their relatively high εNd(t) and εHf(t) values, suggest that the different lithologies of the Miaohou and Shanhou complexes were produced by low degrees of partial melting of a mantle wedge that had been metasomatized by subduction-related fluids, followed by varying degrees of assimilation and fractional crystallization (AFC) during ascent.

(3) The different lithologies of the Miaohou and Shanhou complexes have similar Nd–Hf isotopic characteristics, and the diverse compositional variations in major and trace elements among the different lithologies indicate varying degrees of fractional crystallization (involving Fe–Ti oxides, magnetite, pyroxene, hornblende, biotite, plagioclase, and apatite) as the main mechanism for producing the variations in the chemical compositions of the different lithologies.

(4) At ca. 1000 Ma, the southeastward subduction of an oceanic plate under the Cathaysia block caused the formation of an active continental margin along the northwest margin of Cathaysia together with related magmatic activities. The Cathaysia continental arc and back-arc system lasted until ca. 825 Ma, when the ocean basin was finally closed, leading to the amalgamation of the Yangtze and Cathaysia blocks to form the Jiangnan Orogen.

Acknowledgments

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at http://dx.doi.org/10.1016/j.precamres.2015.08.006.

References


